Limitations in the MJO Initiation Over the Indian Ocean in Limited Area Models

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LIMITATIONS IN THE MJO INITIATION OVER THE INDIAN OCEAN IN LIMITED AREA MODELS

By

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LIMITATIONS IN THE MJO INITIATION OVER THE INDIAN OCEAN IN LIMITED AREA MODELS

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This work encompasses the lessons learned when studying the Madden Julian Oscillation (MJO) under a regional modeling framework. Regional modeling is widely used by the meteorological community; however, few studies outline the limitations of regional models and the challenges one must face when modeling particular tropical phenomena. This is the focus of this research.

This study is divided into four parts, the first part is the development of a technique that can be used in order to discern between successful and unsuccessful simulations of MJO cases of study. This technique, referred here as The Tracking Method, consists of tracking the various eastward-moving features during the simulations and obtaining an optimal track amplitude, speed, and start date, which will represent the most prominent eastward-moving feature on different variable fields. This method was applied to long records of data such that climatological values of amplitude and speed were obtained. The tracking method is applied to each simulation and these are rated against the observation’s tracking values within the climatological deviations. The application of this procedure to the long-term precipitation and wind data show a level of decoupling between the convection and the circulation patterns within in the MJO.

The next three parts of this work are dedicated to the investigation of different aspects that limit the regional models in reproducing the MJO: model physics,
environmental moisture patterns, and initial, bottom, and lateral boundary conditions. The study of the influence of the choice of physics in MJO initiation under a regional model setting was focused on the cumulus and planetary boundary layer schemes. Both play an important role in the moisture re-distribution and convection. The effect that the choice of cumulus and planetary boundary layer scheme has in a regional model affects the numerical simulations so greatly that it may change the water cycle balance during the simulation, in as little as 5 days. We found that the non-linearity of the processes within the model physics play a key role in the entire forecast. Without their mutual improvement, we will not be able to accurately represent the MJO, with broader implications to tropical meteorology in general.

The large-scale moisture patterns and their role during the MJO initiation were studied by using zonal spectral nudging of water vapor mixing ratio. Grid nudging of other variables such as temperature and wind was also performed. Our results show that the correction of moisture in the low and middle vertical levels is sufficient for a realistic MJO simulation. Moreover, the correction of fields such as temperature, wind, and their combination does not result in a successful simulation. Specifically, the error correction of the moisture mean and planetary zonal wavenumbers 1 to 3 will improve an MJO simulation greatly. Beyond these three wavenumbers together, the additional correction of higher wavenumbers, a specific wavenumber alone, temporal, or spatial means are not sufficient for a realistic MJO simulation. The moisture spatial patterns are so important that it is found to be possible to reproduce an MJO precipitation pattern during a period without MJO only by nudging the water vapor mixing ratio.

The study of the influence of initial, bottom and lateral boundary conditions
during MJO simulation in a regional model shows that the precipitation amplitude is dominated by the model physics rather than the initial and lateral boundary conditions configuration. However, the timing of precipitation-triggering and the start of the eastward-moving features are associated with the sea surface temperature updates. Moreover, our results show that updating the SSTs in a non-coupled simulation will lead to the premature triggering of the precipitation on top of the warm SST anomalies associated with the MJO. This differs from the real precipitation and SST anomalies observed during MJO cases, where the warm SSTs are ahead of the convection and cooler SSTs related to cloudiness and cooling by fresh water are below the rain.

Lastly, regardless of the lateral boundary conditions setup (time depended vs time independent) in a regional model setting, there is dominance of the physics errors over the information from the lateral boundary conditions, which are overshadowed. Furthermore, the differences between simulations with different lateral boundary conditions will reach the center of the domain faster through the parameterized processes within the model than one might otherwise expect through advection, and the differences originally closer to the boundaries in the regional model will be amplified under convective situations.

Regardless of the fact that regional modeling is a widely used framework due to its major advantage of increased resolution over a particular area of interest, it is necessary to take into consideration its many shortcomings, especially the strong dependence on the parameterized physics rather than the information in the boundaries in some cases. Going forward, it is also critical to take into account that improved forecasts in the tropics will require further development and improvement of the physics processes
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**FIGURE B.2.** Same as Figure B.1 but for cases 6 to 10.

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Chapter 1

Introduction

1.1. The Madden Julian Oscillation

The Madden Julian Oscillation (MJO) is one of the most predominant modes of intraseasonal variability within the tropics. This oscillation is named after the two scientists whom first documented it in 1971 and 1972: Roland A. Madden and Paul R. Julian (1971, 1972). In their first study, they used 10 years of sounding data from Canton Island and observed a maximum in the spectral power between 40-50 days in the surface pressure field and the zonal winds at 850 and 150 hPa. Later, in their 1972 paper, Madden and Julian used data from 12 sounding stations in the tropics and proposed the horizontal pattern for the MJO (Figure 1.1). In this model, the MJO initiates with a negative pressure anomaly over East Africa and the Indian Ocean. These pressure anomalies increase, and by stage 2 the convection in the Indian Ocean has become deep and reaches the top of the troposphere. There are two anomalous circulation cells to the west and east of the MJO convective center: at low levels with westerly (easterly) wind anomalies and at upper levels with easterly (westerly) anomalous wind for the west (east) circulation cell. Those easterly and westerly anomalies translate to convergence at lower levels and divergence aloft.
**Figure 1.1.** Figure 16 from Madden and Julian 1972: “Schematic of the time and space (zonal plane) variations of the disturbance associated with the 40-50 day oscillation. Dates are indicated symbolically by the letters at the left of each chart and correspond to dates associated with the oscillation in Canton’s station pressures indicated in Fig 11. The mean pressure disturbance taken from Fig 12 is plotted at the bottom of each chart with negative anomalies shaded. The circulation cells are based on the mean zonal wind disturbance presented in Figure 13. Regions of enhanced large-scale convection are indicated schematically by the cumulus and cumulonimbus clouds. The relative tropopause height is indicated at the top of each chart.”

The horizontal extent of the eastern circulation is comparatively small during the initial MJO stages but grows in size as the MJO develops and moves eastward at approximately 5ms⁻¹. Simultaneously, the western cell decreases in size with time.
During stage 4, both zonal circulations have similar horizontal extent, and the convection is vigorous between them. In stage 5, the western circulation has reached a minimum while the eastern cell begins to weaken. During the next stage, 6, the convection, convergence at low levels, and divergence at upper levels have all decreased. In the following 2 stages there is no convection associated with the zonal circulations, and the westerly/easterly signature gets weaker while the remnant signal travels around the rest of the globe.

The MJO conceptual model proposed by Madden and Julian (1972) is still fundamentally valid today. However, despite numerous advances in the study of MJO, the state of the science has not evolved to the level of maturity that has been reached for other tropical meteorological phenomena. The MJO initiation mechanism, the eastward propagation mechanism, and its interaction with the environment and convective populations are just a few of the ongoing debates within the scientific community regarding this phenomenon (Li, 2014; Zhang, 2005).

In comparison with other meteorological phenomena such as hurricanes, the advancement of MJO studies has been hindered by the lack of observations in the Indian Ocean region¹. However, one cannot realistically quantify energy budgets, surface fluxes, and the full dynamics of the MJO from observations alone. Therefore, we must model the MJO in order to study in detail the initiation processes. Yet, the MJO is very difficult to obtain from numerical simulations, especially because of its characteristics: its horizontal extent reaches the planetary scales (zonal wavenumber 1,2 and 3) with temporal periods between 30-90 days, and it involves vigorous deep convection moving

to the east at a slower pace than other tropical waves. However, it presents such types of waves (Kelvin waves) during its mature stage within the MJO convective envelope, as well as in suppressed phase (Roundy, 2008).

1.1.1. MJO Teleconnections

Additionally, regardless of whether the MJO is a large scale phenomenon, it also affects certain regions locally with precipitation and strong winds while it moves through the tropics (Zhang, 2013). Also, the MJO has teleconnections with both weather (continental weather in the extra tropics, hurricanes) and climate systems (El Niño Southern Oscillation, monsoons). Therefore, if we are punctilious, such interactions should be captured by numerical models. In the following paragraphs we mention some of these teleconnections.

The effect of the MJO on the formation and development of tropical cyclones has been subject of investigation in the last 15 years. In the East Pacific, Maloney and Hartmann (2000) found that the westerly wind anomalies associated with the active MJO phase bring less vertical wind shear and greater cyclonic anomalies in the East Pacific. This implies that more tropical systems are likely to occur when there is an active MJO event in the Indian Ocean and West Pacific. The mechanism proposed is based on the development of coupled Kelvin waves (10-17 ms\(^{-1}\), (Roundy, 2008)) which are generated by convection and travel east, modifying the conditions in the East Pacific (Maloney and Hartmann, 2000). Schreck and Molinari (2011) studied two cases in the North Pacific and proposed that the effect of the MJO on tropical cyclones is due to an increase of cyclonic potential vorticity due to the convection and heating within the MJO. In the Atlantic, Klotzbach (2010) found that the MJO imposed its finger print by modifying the large
scale patterns of wind shear and humidity in the Atlantic, Gulf of Mexico, and Caribbean Sea. Specifically, phases 1 to 2 of the MJO are associated with a greater number of named storms, hurricanes, and major hurricanes. Alternatively, phases 6 and 7 are associated with stronger wind shear and lower relative humidity in the region.

Other phenomena that are affected by the MJO, and which can also have an effect on the MJO, are the monsoons. The onset of the South Asian and Australian Monsoon are affected by the phase and location of the MJO, (Lau and Chan, 1986; Wheeler et al., 2009). Lau and Chan (1986) used outgoing longwave radiation (OLR) data and found that the onset of the Indian monsoon is correlated with the propagation of the convection associated to the MJO to the north. Other studies have also suggested that these interactions are affected by El Niño Southern Oscillation (ENSO) and by the inter-annual variability in the monsoons (Lawrence and Webster, 2001).

The MJO effects are not only present in the tropics. The extra-tropics are also influenced by the MJO. In the specific case of the US, weather patterns in North America are greatly affected by the MJO, especially because of the source of moisture to higher latitudes (Jones et al., 2004; Ralph et al., 2010; Riddle et al., 2013). After phases 3 and 7 of the MJO, there are also positive precipitation anomalies in the West Coast of the U.S. (Higgins et al., 2000). Other studies indicate that the MJO has a sub-seasonal effect in the US. Baxter et al. (2014) found that 5 to 20 days following MJO convection in the West Pacific, there is enhancement (decrease) of precipitation over the Southern Plains and Great Lakes regions (Southeast of the US). The mechanism proposed by Baxter et al. (2014) is related to the weakening and southerly movement of the jet stream by trains of Rossby waves after the MJO convection.
Moreover, the MJO and interannual phenomenon such as El Niño Southern Oscillation (ENSO) interact. The exact mechanism for their interaction is still not well defined, however several studies argued that the MJO modulates the ENSO periodicity and timing, and as a consequence its predictability (Kapur et al., 2012; Seiki et al., 2011; Zhang and Gottschalck, 2002).

Finally, there have also been studies on the subject of the MJO and how this would be affected under climate change conditions (Subramanian et al., 2011). However, this is a long road for the scientific community; there are many hindrances that need to be resolved before we are confident in such results. Nevertheless, by improving our insights of the MJO numerical simulations, our knowledge and understanding the MJO and its weather interactions will increase. This will result in a better forecast for both the MJO and related phenomena (Zhang, 2013). With this study, we want to take one more step forward in advancing the state of the science for the MJO.

Because of the spatial and temporal scales of the MJO, it is usually called the crossover between weather and climate in the tropics (Chen and Houze, 1996; Moncrieff, 2010; Zhang, 2005; Zhang, 2013). In this study, we called it the puzzle piece that tie those two different time scales together. As a puzzle piece it needs to complete our understanding of the contiguous pieces together. The MJO is therefore the ultimate test for General Circulation models (GCMs) and Regional Models (RM). Some GCMs manage to reproduce the MJO after the model’s climate has already adjusted for simulations on the order of years. Nevertheless, MJO events in the model are usually too weak and differ from those in the real world. Therefore, the usual approach is an “MJO composite” approach, where the model mean MJO is calculated and characterized (Inness
and Slingo, 2003; Inness et al., 2003; Kim et al., 2009; Maloney, 2002; Maloney and Kiehl, 2002; Waliser et al., 2009; Zhang and Mu, 2005).

1.1.2. MJO Modeling Challenges

The modeling challenges to overcome differ between models and are occasionally dependent upon the specific nature of a particular MJO event. Often, when a global model its able to reproduced the MJO it is consider “good” and it has crossed a difficult benchmark in the modeling world (Bechtold et al., 2008; Hirons et al., 2013; Jung et al., 2011; Miura et al., 2007). Numerical models have become a useful tool in MJO research. However, there is still a lot of work to be done in modeling this phenomenon well enough that we can fully trust the models’ representation of the MJO.

There are not many meteorological phenomena that have middle time scales (weather – climate) that directly affect the entire 360 degrees in the tropical belt. Hurricanes, Typhoons, and easterly waves affect the basin where they form, and their surroundings as they move or make landfall. However, they do not cover such longitudinal extent as the MJO. The Monsoon is another example, its time scales are seasonal by nature of its formation mechanisms. Monsoon systems are generally localized in different areas, and usually we refer to them as the Indian monsoon, Australian monsoon, East Asia monsoon, and others, making reference to the region or continent where they take place. If we compare with the MJO, we do not refer to the MJO as the Indian Ocean MJO, or the Maritime Contient or west pacific MJO. The MJO is one system which propagates around all these areas. Regardless of the different nature of these two phenomena, the MJO and the monsoon often interact. The onset period of the monsoon is altered by the MJO activity (South Asian and Australian monsoon) and
the latitudinal movement of the MJO convection is affected by the monsoon (Australian monsoon) (Lau and Chan, 1986; Wheeler et al., 2009). Besides differences between MJO and the monsoon, the Boreal Summer Intra Seasonal Oscillation (BSIOS) occurs during the break of the Asian monsoon and its period is as the MJO, intraseasonal. The BSIOS consists of an intraseasonal northward movement of the convergence over the equator, which may occur in conjunction with or independently from the MJO during the start of the Asia Summer Monsoon. The BSIOS is, however, different from the MJO.

It is well known in the scientific community that obtaining the precipitation associated with the MJO is an impediment that has not yet been overcome. Both GCMs and RMs struggle with the representation of the precipitation. This deficiency is attributed to the cumulus parameterization schemes (Gustafson and Weare, 2004; Maloney and Hartmann, 2001; Slingo, 1996; Zhang, 2005; Zhang et al., 2006). Each cumulus scheme has its own static and dynamic controls, and assumptions in their closures to feedback from the sub-grid scale processes to the grid scale (Arakawa, 2004). The sub-grid scale precipitation ranges between 70-90% of the total precipitation for a specific simulation, which makes the choice of parameterization a critical step when configuring the numerical model. Most of the studies acknowledge this deficiency to the poor representation of entrainment rate, convection-evaporation processes, and large scale moisture dependency, amongst others (Maloney and Hartmann, 2001).

The fact that the cumulus parameterizations available in GCMs and RMs are not able to accurately reproduce MJO cases of study is concerning. After all, these parameterizations have been widely used to understand various meteorological phenomena in the tropics and mid latitudes: monsoons (Crétat et al., 2011; Flaounas et
al., 2010), marine boundary layer clouds (Zhang et al., 2011), Cyclones (Raju et al., 2011) and many others (Del Genio and Wu, 2010; Derbyshire et al., 2004; Fritsch et al., 1986; Kotroni and Lagouvardos, 2001; Li et al., 2007; Liang et al., 2004).

Numerous studies have approached the problem by tuning a specific cumulus parameterization in order to improve MJO forecasts. The latter includes increasing convective sensitivity to environmental moisture (Bechtold et al., 2008; Hannah and Maloney, 2011; Zhang and Mu, 2005), modifying the assumption of closure (Maloney and Sobel, 2004; Wu et al., 2007), and including convective momentum transport (Khoury et al., 2012; Ling and Li, 2014; Miyakawa et al., 2012). However, when this procedure is used, the mean state of the model drifts away from reality (Kim et al., 2011; Maloney and Hartmann, 2001).

A more artificial approach is to change a certain field in the model that is known to be poorly simulated but critical to the MJO. Examples of such brute-force modification of model simulations include redistributing diabatic heating vertically (Li et al., 2009), nudging water vapor from global analysis product into model simulations (Hagos et al., 2011; Subramanian and Zhang, 2014), and using daily updated sea surface temperature (SST) with intraseasonal signals as the lower boundary conditions (Hagos and Leung, 2011). They all serve one purpose: identify key model biases that should be main targets of improvement. They all misrepresent the causal relationship between intraseasonal signals of MJO convection and the variable modified. If carefully done, this approach may reveal insights to MJO physics. Otherwise, their results are misleading.

Another procedure used to improve MJO simulations is by explicitly including air-sea interaction by coupling an atmosphere model to an ocean. This has been done
using a simple 1-dimensional ocean mixed-layer model (Hendon, 2000; Shinoda and Hendon, 1997; Waliser and Lau, 1999; Woolnough et al., 2007) or a full ocean dynamic model (Fu and Wang, 2004; Inness and Slingo, 2003; Kemball-Cook et al., 2002; Zhang et al., 2006; Zheng et al., 2004). But coupling is not necessary for successful MJO simulations as demonstrated by many atmosphere-only MJO simulations (Kim et al., 2008) and prediction (Ling et al., 2014) studies. This however, this is an ongoing debate of the MJO scientific community.

At last, the ultimate approach is the use of cloud-permitting grid spacing (Grabowski, 2003). However, this shifts the burden of improving the cumulus to improving the microphysics schemes. Even though there have been some success stories (Liu et al., 2009; Satoh et al., 2008), cloud-permitting grid spacing is still very computationally expensive. This is especially true when we want to use GCMs for future predictions. Consequently, the use of cumulus parameterizations is not yet obsolete.

We want to simulate the MJO so we can have answers to the questions that we do not know, as well as to improve both its forecast skill and understanding. Here, we focus most on the initiation processes. Ideally, we want to perfectly reproduce an MJO, just as beautiful as it is in nature. However, up to this day, this is impossible.

In this study we focus on the Regional Models (RM). Regional Modeling is a powerful scientific tool in MJO simulations because of its capability to focus on the area of interest, sometimes follow the event, and its flexibility to increase resolution from an analysis or forecast. However, it is important to study the different limitations, advantages, and disadvantages when the RMs are used to simulate the MJO.

In the following chapters of this work we will study:
i. The physics parameterizations limitations suffered by the regional models when trying to simulate the MJO and their divergence from the mean climate.

ii. The role of a realistic reproduction of the moisture space and time patterns and its relevance to MJO modeling.

iii. The impacts of the selection of the initial, lateral boundary, and bottom conditions in the error propagation and growth in MJO simulations.

In chapter 2, we describe the data and method used for this study. Chapter 3 will introduce a new quantitative technique to quantify the success or failure of MJO simulations. Chapter 4 will discuss the physics parameterization role in simulating the MJO. This will be done by focusing on the water cycle biases and the forecast divergence from the analysis and reanalysis when trying to reproduce a specific MJO case. In Chapter 5, we study the relevance of the moisture in the regional models and its impact on MJO simulations. Specifically, we will focus on the spatial and temporal structure of moisture and its dependence on a successful MJO simulation. Chapter 6 will address the problem of the initial, bottom, and lateral boundary conditions’ role in the error propagation during the MJO convection. Lastly, in chapter 7 we will summarize what we have learned during this investigation.
Chapter 2

Data and Methodology

2.1. Observations and Reanalysis

In this section we describe the data used in this study. The periods of time used for each data set changes depending on time in the numerical simulations we are verifying against. We describe in detail the specific periods of time in each chapter.

Precipitation estimates and sea surface temperatures (SST) from the Tropical Rainfall Measuring Mission (TRMM, Kummerow et al. (1998)) are used in this study. The TRMM satellite mission was proposed at the end of the 1980s (Simpson, 1988), and was planned to start at the end of the 1990s.

The precipitation estimates are obtained through a combination of three instruments onboard the satellite: the TRMM Microwave Imager (TMI), the precipitation radar, and the visible and infrared radiometer system. The TRMM Multisatellite Precipitation analysis provides precipitation estimates at a horizontal resolution of 0.25 x 0.25 degrees in a time interval of 3 hours (Huffman et al., 2007). This is called the 3B42 product; in this study we use 6 hours and daily means of version 7 of the TRMM 3B42 precipitation products. For some of the analysis, the TRMM data was re-gridded to lower resolutions for better comparison with other products and numerical simulations. We also used daily SST data from the TMI in TRMM. The TMI SST retrievals have a spatial
resolution of 45Km (Gentemann et al., 2010; Wentz et al., 2000). Regardless of TRMMs many shortcomings such as temporal and spatial gaps, estimation errors, and resolution, we trust this data set and consider its data to be the benchmark for comparison on which the model simulations should achieve.

Outgoing Longwave Radiation (OLR) from the National Oceanic and Atmospheric Administration (NOAA) is used in this study as a proxy of precipitation for some of the analysis. The NOAA-OLR data set is obtained from polar orbiting satellites, and it is usually referred to as OLR Interpolated data. The interpolation is done to the nearest neighbor when there is missing data from the satellite due to malfunctions or incomplete global coverage. In the end, the product is global, with a 2.5 x 2.5 degrees resolution, and daily values are provided (Liebmann, 1996).

The European Centre for Medium-Range Weather Forecast (ECMWF) Interim Reanalysis (ERAI) is the reanalysis data set most used in this study. The ERAI data are reanalyzed fields from the ECMWF forecast with a horizontal resolution of T255 (~75Km) and 37 vertical levels (Simmons et al., 2007). Table 2.1 enumerates the different parameterizations of convection, microphysics, planetary boundary layer and radiation for ERAI and other reanalyses. ERAI is used to compare numerical simulations of fields such as wind, temperature, water vapor mixing ratio, precipitable water, and others. ERAI is also used as to provide the initial conditions (IC), lateral boundary conditions (LBC), and bottom conditions (BC) to most of the numerical simulations presented here. The numerical simulations which have either IC, LBC, or BC from other reanalysis other than ERAI are named accordingly in each chapter.
The ECMWF deterministic analysis and 10-day forecasts produced twice a day (00Z and 12Z) during the Dynamics of the Madden Julian Oscillation (DYNAMO, Yoneyama et al. (2013)) are also used. The resolution (see Table 2.1) of the ECMWF during DYNAMO allows us to initialize WRF simulations at a higher resolution without the use of nesting techniques. Furthermore, fields can be compared with high-resolution numerical simulations. Moreover, we also use the ECMWF Ensemble Prediction System 15-day forecasts during the DYNAMO field campaign. The ensemble forecasts include one control run and 50 perturbed members with varying model configurations (T639L61 through day 10, T319L61 afterwards), initialized daily (00 and 12Z). We will refer to this analysis as ECDY.

**Table 2.1. Reanalyses and Analysis Physics Schemes**

<table>
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<tr>
<th>Acronym</th>
<th>Resolution</th>
<th>Vertical Levels</th>
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<td>ECMWF Interim Reanalysis</td>
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- Lock et al. (2000)
- Louis et al. (1982)

Another data set used in this study is the special global reanalysis for WCRP and WWRP-THORPEX Year of Tropical Convection Project (YOTC) (Waliser and Moncrieff, 2008; Waliser et al., 2012). The YOTC is an initiative that began in May 2008.
and ended in April 2010. Its major goal was to increase our current knowledge of the tropical convection. This was proposed to be done virtually and with the help of different University and Meteorological Centers around the world (Waliser et al., 2012). During this period, different numerical models were run at high resolutions and special derived satellite products were produced for the campaign (Moncrieff et al., 2007). We chose to use YOTC data because of its inclusion of hydrometeor information (e.g., cloud water and ice contents), heating and moistening or temperature and moisture tendencies by parameterized processes, and high resolution grid spacing that are not all available from other global reanalysis data (Table 2.1). The YOTC data used in this study are the ECMWF high-resolution reanalysis and forecasts; this data has 25Km horizontal resolution and 60 vertical levels, and we will be referred to as YOTC.

The National Centers for Environmental Prediction (NCEP) Climate Forecast System Reanalysis (CFSR, Saha et al. (2010)) is another reanalysis used in this study. CFSR products have a T255 (~75Km) horizontal resolution and 64 vertical levels, with the highest level at 0.266 hPa. This reanalysis was primarily used for the comparison of different temperature and moistening tendencies against model simulations and other reanalysis (Table 2.1).

Lastly, we use the Modern-Era Retrospective Analysis for Research and Applications (MERRA, Rienecker et al. (2011)) reanalysis. MERRA has a vast list of variables which are usually not included in other reanalysis. Similarly to the YOTC data, MERRA variables include moist physics temperature and moisture tendencies, deep and cloud scheme temperature tendencies, and others (Table 2.1). The resolution of this data set is T382 (~38Km) with 68 vertical levels. MERRA is available starting from 1979 to
the present. This reanalysis is made by reanalyzed and assimilated data using the Goddard Earth Observing System (GEOS) atmospheric model (GEOS-AM) and data assimilation system (GEOS-DAS).

2.2. MJO Cases of Study

This research will focus on MJO cases of study. Our motivation to use MJO cases of study is that the GCMs are usually able to obtain statistical measurement and composites of the “Model’s MJO”, (Subramanian et al., 2011; Zhou et al., 2012) and experience much more difficulty in the reproduction of specific MJO events (Subramanian and Zhang, 2014). This leaves us with an unclear knowledge of the MJO, since all the MJO cases are different, and the composite of these only represents the mean MJO behavior.

Two MJO cases of study were selected for this investigation: a) October – December 2009 (MJO-09), and b) November – December 2011 (MJO-DY). MJO-OD09 and MJO-DY are similar in several aspects: 1) both cases occur during boreal winter, 2) there is a successive MJO event after each case (approximately during December 20 2009 and 2011), 3) there are two precipitation branches (one after another), during the events (Figure 2.1), and 4) the daily precipitation values are comparable. Specific characteristics and a description of each case are presented in the following sub sections.

2.2.1. The October – December 2009 MJO Case

MJO-09 was selected because of its intensity and the availability of new high-resolution reanalysis and forecast products during the YOTC period. The approximate

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2 Statistical or “model’s MJO” is referred as a climatology of the MJO obtained usually by band-pass filtering, which is compared to a current MJO climatology, rather than a simulation of a specific event.
duration of this event is from October 25\textsuperscript{th} to December 10\textsuperscript{th}, 2009 (Figure 2.1a). During this MJO event, ENSO is in a weak positive phase (El Niño, warmer than normal temperatures in the East Pacific) (Waliser et al., 2012). Positive ENSO phase is related to

\textbf{Figure 2.1.} Longitude – Time daily precipitation means (mm/day\textsuperscript{-1}) from TRMM for (a) MJO-09 and (b) MJO-11 cases of study. The precipitation is meridionally averaged between 10°N-10°S.
a movement of the MJO convection closer to the Central Pacific, while the upper level circulation also moves to the Central North Pacific. Figure 2.2 shows the values of the standardized index for ENSO (Southern Oscillation Index, SOI, yellow), the North Atlantic Oscillation (NAO, blue), and the Artic Oscillation (AO, pink). From this figure, we can see the negative values of both NAO and AO. The value for all the indexes decreases from October to November, and increase again in December. The negative values in NAO indicate lower pressure in the Icelandic low and the Azores high, as well as a weakening of the jet stream in the extratropics. On the other hand, negative values of the AO are associated with positive geopotential height anomalies in the polar vortex. This indicates a weakening of the vortex, which results in weaker upper level westerlies as well.

Despite the fact that MJO-09 occurred during the virtual YOTC effort campaign, this case has not been studied as in depth as other cases during 2008 – 2010 YOTC. Shelly et al. (2014) studied this event using the Met Office Unified Model Atmosphere. They defined the MJO case to start on October 15th and end on December 6th 2009. Their work presents numerical simulations using coupled and uncoupled 15 day hindcasts of the Met Office Model. The atmospheric coupled version of the model is driven by persistent SST daily anomalies to the seasonal cycle. Different from our study, Shelly et al. (2014) focus on hindcasts simulations, whereas we focus on regular forecast simulations. It is noteworthy to mention that both studies, however, are retrospective. Satoh et al. (2011) studied MJO-09 and its consecutive MJO event using the Nonhydrostatic Icosahedral Atmospheric Model (NICAM, Satoh et al. (2008))

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3 Hindcasts simulations are those in which the simulation is re-initialized after a determined number of days and the forecast is the composite of the hindcasts.
4 Forecast simulations are those in which the simulation is initialized only once for its complete length.
Cloud Resolving Model. They obtained MJO signals and compared their OLR results with satellite data. The resolution of their numerical simulations is 14 and 7Km. With these resolutions they found the microphysics scheme needs to be improved.

2.2.2. The November – December 2011 MJO Case

**FIGURE 2.** Time series of standardized Niño 3.4 (ENSO, yellow), North Atlantic Oscillation (NAO, blue), and Arctic Oscillation (AO, pink) indexes. The periods of the MJO cases (MJO-09 and MJO-11) are highlighted by the vertical solid black lines. The monthly values of the indexes were obtained from the National Weather Service Climate Prediction Center: http://www.cpc.ncep.noaa.gov/.

MJO-DY takes place during the DYNAMO field experiment, which yielded a variety of observations available at particular locations (soundings) and area-covered radar measurements. This is, up to 2014, the most observed MJO event. MJO-DY is also interesting because this case in particular has been a struggle to simulate in forecast mode.
for some GCMs (Kerns and Chen, 2014a; Subramanian and Zhang, 2014) and RMs (Seo et al., 2014). The approximate duration for MJO-DY is around 30 days: November 10th to December 20th. These periods are determined subjectively via the precipitation amounts shown in Figure 2.1b (it is important to note that the proposed investigation would not debate nor propose an initiation time for the events to study). Opposite to MJO-09, both NAO and AO are in their positive phases, whereas ENSO is on its negative phase (La Niña) (Figure 2.2). During La Niña the MJO convection in the Indian Ocean is stronger than during El Niño. The convection in this case moves to the north (Moon et al., 2010).

As previously mentioned, this MJO case has the most observations. Therefore, since the MJO campaign ended in early 2012, there have been numerous studies focused on different aspects of this MJO event (Chi et al., 2014; Fu et al., 2013; Hagos et al., 2014; Johnson and Ciesielski, 2013; Kerns and Chen, 2014a; Kerns and Chen, 2014b). As an example, Kerns and Chen (2014b) found that the initiation of the MJO convection is related to air intrusion across the equator. In their studied they attributed the failure of GFS to accurate reproduced this event because it fails to reproduce this large scale feature, particular to this MJO case.

### 2.3. The WRF Model

Three versions of the WRF numerical model (3.2, 3.3, and 3.4) were used in this investigation. WRF is a fully compressible, nonhydrostatic model with a third order Runge-Kutta time stepping scheme on an Arakawa C-grid (Skamarock, 2008). Different
model domain configurations and grid sizes were used during the development of this study.

The specific lists of the simulations made with each domain will be described in detail in each chapter. Seven domains were included for this study (Figure 2.3). The channel domain (CH) follows Ray et al. (2009) configuration. It covers the entire tropics and the latitudinal band 25°S – 25°N. The horizontal resolution of this domain is 1° and has 28 vertical levels. However there are 2 cases where the number of vertical levels was changed to 35 in order to test its sensitivity (see Chapter 4). The lateral boundary conditions (LBC) in this case are periodic in the east-west direction due to its configuration. Two domains of 50Km over the Indian Ocean are also used. IO_50 masks the Indian Ocean region, and IO_50S is a smaller version of IO_50.

One 36 Km resolution domain (IO_36N) was used as a parent nest domain for the IO_12N. Lastly, a 13Km resolution domain (IO_13) was used over the Indian Ocean region. The spatial configuration of the domains was selected so the Tropics (CH), the Indian Ocean (IO_50, IO_50S, IO_13), and the DYNAMO (IO_12 m IO_13) regions can be covered with different resolutions and LBC arrangements (see Figure 2.3).

**Figure 2.3.** Simulations domains used in this study.
The integration time for the simulations was different in some cases, depending on the length of the MJO event or the day chosen for initialization. The output frequency for all the simulations is 6 hours. The initial, boundary, and surface conditions were obtained from ERAI in the majority of the simulations. Simulations with the domain IO_12 and IO_13 and higher resolutions used the ECDY analysis as BC.

WRF has many options for its physics. As an example, version 3.3 of WRF has 15 options for microphysics, 7 for radiation, 4 surface layer, 5 land-surface models, 12 planetary boundary layer, and 10 cumulus schemes options. This will lead to a number of combinations \( \sim 10^5 \). Since precipitation in the MJO is very important, mostly the choice of cumulus (CU), microphysics (MP), and planetary boundary layer (PBL) will play an important role in the success or failure of the simulation. Together, these schemes define where, when, and how the convection is initiated and its interactions with the environment.

The goal of a numerical simulation is to determine the state of the atmosphere at time \( t_1 \), starting from \( t_0 \). The evolution in time (d/dt) of the meteorological conditions is based on the information in the initial conditions (IC) (state at \( t_0 \)). The evolution in time for a given prognostic variable \( \alpha \) is determined by its tendency Equation:

\[
\frac{\partial a}{\partial t} + \nabla \cdot (a\vec{V}) = \tilde{F}(a) \tag{2.1}
\]

where the second term in the left hand side is the advection term (sometimes called flux convergence). The term in the right hand side (RHS) of Equation 2.1 is the forcing term \( \tilde{F}(a) \). This includes the physics tendencies and other forcing terms from the spatial projection and earth rotation (Skamarock, 2008). \( \tilde{F}(a) \) then, can be divided into two terms:
\( \bar{F}(a) = \bar{H}(a) + \left( \frac{\partial a}{\partial t} \right)_{PHYSICS} \)  

(2.2)

The first term in the RHS corresponds to the forcing of the spatial projections and the earth rotation. The second term in the RHS in Equation 2.2 is the tendency due to the parameterized processes in the PBL, CU, MP, and radiation schemes.

\[
\left( \frac{\partial a}{\partial t} \right)_{PHYSICS} = \left( \frac{\partial a}{\partial t} \right)_{CU} + \left( \frac{\partial a}{\partial t} \right)_{MP} + \left( \frac{\partial a}{\partial t} \right)_{PBL} + \left( \frac{\partial a}{\partial t} \right)_{RA} \]

(2.3)

Consequently, the role of each parameterization in the numerical model is to account for the tendency terms in Equation 2.3.

**2.3.1. Cumulus Schemes**

The idea of the cumulus scheme is to parameterize the effects of the sub-grid scale moist processes and relate it to the large scale before the saturation of the grid scale. A closure is the main assumption that is proposed in order to link the large scale with the cloud processes, which is associated with scales that are larger than the numerical model spatial and temporal scale. Based on the closure, one can obtain the temperature, moisture, and (in some cases) momentum changes produced by the convection.

The closure is important because our results will be affected by how we represent the changes produced by the cumulus clouds which modify the environment by cumulus-induced subsidence. Said in better words by Arakawa and Schubert (1974):

“The latent heat release within the clouds does not directly warm the environment, but it maintains the buoyancy of the clouds against the adiabatic cooling due to the upward motion and cooling produced by the entrainment of drier air and colder air from the environment. Thus the latent heat release within the clouds maintains the vertical mass flux of the clouds and thereby, the cumulus-induced subsidence in the environment. The drying and warming of the environment, by the cumulus induced-subsidence, are the indirect effects of condensation and release of latent heat, but their vertical distributions
can be very different from the vertical distribution of the condensation within the clouds."

Generally, the cumulus schemes are named after the scientist who first proposed the parameterization. In the following subsection, we are going to characterize the most important assumptions of each cumulus scheme used in this work (present in WRF).

**Betts-Miller**

This cumulus scheme is an adjustment scheme (Betts, 1986). This means that the sounding at each column will be adjusted to a reference profile so it will be in quasi-equilibrium with the large scale. Betts (1986) defined quasi-equilibrium as to when it is assumed that the differences between the cloud and the environmental buoyancy are small when deep convection is occurring. The adjustment scheme requires an adjustment time to nudge towards the reference profile. In Betts-Miller, the adjustment time is set to be different in the planetary boundary layer and in the troposphere, with the latter having smaller magnitude than the former. This scheme has two different reference profiles for shallow and deep convection. The proposed reference profiles for Betts and Miller (1986) is a virtual moist adiabat, i.e a parcel follows constant lines of virtual equivalent potential temperature (Betts and Miller, 1986). In other words, this is the theoretical equivalent potential temperature that a dry parcel will have with the same density and pressure of a moist parcel. The choosing of this reference profile came from different observational campaigns in the tropics, as well as from hurricane sounding data (Betts, 1986).

The Betts-Miller scheme does not allow for a shallow and deep convection to occur within the same column. First, the cloud base is found, by testing the buoyancy at different levels starting from the surface. The cloud base is the lowest level tested to be
buoyant with respect to the adjacent upper levels. Secondly, the cloud top is localized by the level above the inversion layer or when the buoyancy of the parcel becomes negative. The third step consists of the distinction between shallow and deep convection. This is done based on the cloud top height. If the cloud top is greater than 700 hPa the convection is considered to be shallow convection, and if the opposite occurs it is considered as deep convection. During shallow convection there is no precipitation; however, it allows for moisture and temperature changes in the vertical. The reference profile during shallow convection depends on a defined mixing coefficient within the shallow layer, which represents how strong the mixing is.

**Betts-Miller-Janjić**

This adjustment scheme is based on the Betts-Miller scheme, however, Janjić (1994) added some changes to the work proposed in 1986 by Betts and Miller. The main change is the addition of a cloud efficiency parameter \( E \). This is a dimensionless parameter which depends on the cloud temperature and on the change in entropy within time steps:

\[
E = c_1 \frac{\bar{T} \Delta S}{c_p \sum \Delta T \Delta p}
\]  

(2.4)

where \( c_1 \) is a non-dimensional constant (5.0 in version 3 of WRF), \( \bar{T} \) is the mean temperature of the cloud, \( \Delta S \), and \( \Delta T \) are the entropy and temperature change during time steps, \( c_p \) the isobaric heat capacity, and \( \Delta p \) the depth of the cloud within model layers (see Equations 3.1 to 3.3 in Janjić (1994)). The denominator in Equation 2.4 (addition of the change in temperature and pressure within the cloud times the heat capacity) is proportional to the precipitation change over a time step (proposed by Betts (1986))
Differently from the Betts-Miller scheme, the moisture and temperature reference profiles are now proportional (inversely proportional) to the cloud efficiency parameter (relaxation time). Therefore, there is not only one reference state but an array of them (Janjić, 1994). The shallow scheme is set up to have the extra constraint of non-negative change in the entropy. In WRF, the closure of this scheme is changed with respect to Betts-Miller. The closure in this case is the removal of the Convective Available Potential Energy (CAPE).

**Kain-Fritsch**
This is a mass-flux scheme (Kain and Fritsch, 1990). This means that it is based on the calculation of the mass that is coming inside and outside of the cloud. This scheme tests for convection enhancement by what is usually called as a trigger function. In this case, the trigger function consists of the addition of a temperature perturbation to a parcel of air located in the closest model level to the ground, and test for parcel buoyancy. The temperature perturbation depends on the large scale vertical velocity and it is usually ~1 or 2 K. If with the perturbation the parcel is buoyant enough, then convection initiates.

The original scheme was proposed by Kain and Fritsch (1990), however, some modifications were later made based on Kain (2004). This scheme parameterized the entrainment and the detrainment rates. The two are inversely related such that a dry environment will propitiate low parcel buoyancy and the entrainment rate will increase with high parcel buoyancy and moist environments (Kain and Fritsch, 1990). The closure of the scheme is CAPE removal. In other words, the moisture and temperature profiles will be modified by changing the updrafts and downdrafts in the clouds until 90% of the CAPE is removed (Kain, 2004). CAPE calculations are based on a diluted parcel ascent.
Another characteristic of this scheme is the fact that the downdraft mass flux depends on the relative humidity. There is a minimum entrainment rate in the downdraft which corresponds to 50% of the updraft entrainment rate’s maximum (Kain, 2004). The maximum entrainment rate will depend on the large scale forcing (large scale vertical velocity), and the mixing rate depends on the cloud radius. Shallow convection is also parameterized, in this case, when the cloud depth is less than 2000m. The changes in the shallow cloud are based on the Turbulent Kinetic Energy (TKE) below. Then, the updraft mass flux at the cloud base is a function of TKE.

This scheme allows for the variation of the cloud radius. The dependence of entrainment rate on the cloud radius and environmental humidity in the downdrafts and vertical velocity in the updrafts make KF a scheme suitable for the representation of multi-scale meso-scale convective systems (MCS).

**Grell**

This scheme has gone under many modifications. However, in this case we are going to focus in the Grell ensemble scheme proposed by Grell et al. (1994). This scheme is based on an ensemble of closures. Each closure is different and the resulting moisture and temperature tendencies are based on the ensemble mean. Grell scheme includes CAPE removal, moisture convergence, and others closures. In this scheme, the subsidence may spread to the neighbor column, which is useful for simulations of less than 10Km. It is also important to mention that there are not quasi-equilibrium closures in the ensemble of this scheme.

**Zhang-McFarlane**

This is a relatively new scheme based on the work of Zhang and McFarlane (1995). This is a mass flux scheme. In this case, besides the moisture and temperature the
moist convection will modify the momentum (zonal and meridional) tendency Equation (Equations 2.1 and 2.2). This is a deep scheme, usually used with the Park and Bretherton shallow scheme (see next section). In the Zhang-McFarlane scheme the closure is based on the exponential consumption of CAPE (CAPE removal) under an undiluted parcel ascent. In this scheme the downdrafts are considered to be saturated. The updraft allows for entrainment through their vertical extent, while the detrainment is parameterized at the top of the updraft. The entrained air is assumed to be saturated and to have the same temperature and pressure as the environment. Condensation is allowed only in the updrafts. The air detrained in the updrafts is evaporated into the environment (Zhang and McFarlane, 1995). The precipitation amount is proportional to the vertical flux of cloud water in the updraft (Zhang and Anthes, 1982). The downdraft mass flux is limited to 65% of the updraft mass flux. As previously mentioned, a very important component of this scheme is the inclusion of momentum transport during convection. This is done by parameterizing the “convection – induced pressure gradient” force (Zhang and McFarlane, 1995).

This scheme was first implemented in the Canadian Climate Centre general circulation model (CCC GCM). However, in 2002, new changes to the Zhang-McFarlane scheme were done (Zhang, 2002) and tested in the National Center for Atmospheric Research (NCAR) Community Climate System Model version 3 (CCSM3). In this case, the parcel ascent is considered to be diluted, and there is also a change in the scheme time scale based on the environmental moisture. Other changes are the inclusion of an environmental humidity threshold of 80% before convection starts. The closure was changed to a quasi – equilibrium closure, where the net change of “convective” CAPE is
in balance with the net CAPE change, so that the convective and the large-scale environmental modifications to CAPE are minor (Zhang, 2002). Zhang and Mu (2005) present results where the new scheme is used to reproduce an MJO composite. In this study we used both modalities of the scheme.

**Park-Bretherton**
This scheme is based on Park and Bretherton (2009). This shallow scheme allows for the production of precipitation, contrary to the shallow schemes that we have discussed so far. It is recommended for this scheme to be used with the University of Washington planetary boundary layer scheme (see next section), because both were develop simultaneously to complement each other (Bretherton et al., 2004). This shallow scheme is a mass flux scheme, in which the closure is based on the Convective Inhibition (CIN) change, based on Mapes (2000).

This parameterization takes into account lateral entrainment and detrainment in its updrafts (it does not include the parameterization of downdrafts), as well as penetrative mixing (overshootting into the inversion layer) (Bretherton and Park, 2009). The entrainment and detrainment rates depend on the cloud height and radius, the value is obtained by numerical experiments using Large Eddy Simulations (LES). There is also momentum entrainment and detrainment. The coupling with the PBL comes mainly from the idea that besides the fact that the updrafts originate in the PBL, the function of bringing air to the top of the PBL is the role of the PBL scheme and not of the cumulus scheme. (see Figure 1 of Bretherton and Park (2009)).

**Tiedtke**
This is a mass-flux scheme formally proposed by Tiedtke (1989). With some modifications, this scheme is the one used in the ECMWF and reanalysis products. This
scheme includes updrafts, downdrafts, turbulent and organized entrainment, shallow convection, and mid level convection. In the updrafts the air is assumed to be saturated and entrainment and detrainment are allowed. The air detrained from the updrafts is assumed evaporate into the environment.

The distinction between shallow and deep convection is based on the height of the cloud base (1500m is the reference value). The entrainment is divided into turbulent and organized. The latter depends on the large scale moisture convergence, and the former is based on imposed entrainment rates. The detrainment occurs in the zero buoyancy level only. The deep convection starts when there is sufficient environmental low level convergence or when the environmental moisture is greater than 90%. The scheme assumes that the shallow convection occurs under large-sale suppressed conditions and it is now proportional to surface evaporation.

The detrainment in the updrafts is design to occur at the edges of the cloud and in the cloud top as outflow. In the updrafts, the entrainment and detrainment rates per unit of height are the same for deep and shallow clouds, $1.0 \times 10^{-4} \text{m}^{-1}$ and $3.4 \times 10^{-4} \text{m}^{-1}$ are the values proposed by Tiedtke (1989). In the downdrafts the value is $2 \times 10^{-4} \text{ m}^{-1}$. The organized entrainment is setup to occur at the cloud base and edges. This is assumed to be directly proportional to the large scale moisture convergence. The default values of entrainment/detrainment rates in WRF are:

Deep, mid, and shallow updrafts= $1 \times 10^{-4} \text{ m}^{-1}$, $1 \times 10^{-4} \text{ m}^{-1}$, and $1.2 \times 10^{-3} \text{ m}^{-1}$ and $2.0 \times 10^{-4} \text{ m}^{-1}$ for downdrafts.

In our study there are some simulations in which the entrainment rates were changed in order to quantify the changes associated with the precipitation and the MJO.
development. We used entrainment/detrainment rates of $1 \times 10^{-4} \text{m}^{-1}$, $1.5 \times 10^{-4} \text{m}^{-1}$, and $210^{-4} \text{m}^{-1}$ for deep clouds, $1 \times 10^{-4} \text{m}^{-1}$ and $2 \times 10^{-4} \text{m}^{-1}$ for midlevel clouds, and $1.2 \times 10^{-3} \text{m}^{-1}$ and $6 \times 10^{-4} \text{m}^{-1}$ in shallow clouds. We will explain in detail these simulations in Chapter 6.

**Arakawa – Schubert Scheme**

The Arakawa-Schubert scheme is not an option in WRF, however, two other parameterizations are based directly on the theory proposed by Arakawa and Schubert (1974). Therefore we decided to dedicate a couple of paragraphs to explain the main characteristics and assumptions of this scheme.

The work of Arakawa and Shubert in 1974 is one of the main forward steps towards the development and improvement of the cumulus parameterizations. Schemes previously cited such as Grell, Tiedtke, and Zhang-McFarlane use some of the principles exposed in Arakawa and Schubert (1974).

This parameterization’s goal is the representation of the changes in the environment that an ensemble of different types of clouds would produce to it. Conceptually, it is assumed quasi-equilibrium between the large scale and the cloud ensemble, i.e in an instant, the cloud ensemble adjustment time is considerably less than the large-scale time scale (Arakawa and Schubert, 1974). The latter can be assumed because the adjustment time in the ensemble of cloud is between $10 \times 10^3 \text{s}$ to $10 \times 10^4 \text{s}$, and the large scale time scale is greater than $10^5 \text{s}$ (Arakawa and Schubert, 1974).

Mathematically, they used a cloud work function, which is an integrated measure of the buoyancy on each ensemble of clouds. This is the work done by the buoyancy force within the cloud. Therefore, the environment is conditionally unstable when the buoyancy or the cloud work function is positive.
In this parameterization the air is assumed to be saturated in the cloud. The air that detrains is assumed to evaporate into the environment. It is also assumed that the entrainment may occur in the complete vertical extension of the cloud and that the detrainment occurs only at the top of the cloud. Detrainment on a thin layer of the cloud top is also parameterized. The cloud base is located at the top of the sub-cloud mixed layer. The radius of each cloud is considered to be constant and the entrainment rate is proposed to be inversely proportional to this radius. Therefore, smaller clouds will have higher detrainment rates than larger clouds. Arakawa and Shubert also proposed that the “...The vertical mass flux below the cloud base should be interpreted as the mass flux of the updrafts associated with the clouds but not in the clouds” (Arakawa and Schubert, 1974).

**Simplified Arakawa-Schubert**

The idea of the Simplified Arakawa-Schubert scheme is the application of the original Arakawa-Schubert scheme under simpler conditions. These conditions were proposed by Pan and Wu (1995), as part of the implementation of this convection scheme into the National Meteorological Center (NMC) Medium-Range Forecast Model (MRF). The parameterization is assumed to be for only one type of clouds rather than an ensemble of different cloud types, as proposed in great detail by Grell et al. (1994). They added parameterization of saturated downdrafts, which is very similar to the parameterization of the updrafts (Pan and Wu, 1995). The updraft (downdraft) entrainment level corresponds to a local minimum of the moist static energy, starting from below 700hPa (above 400hPa). Convection starts if an undiluted parcel ascent would reach the level of free convection within 150hPa from the starting level. A maximum of cloud-base mass flux of $0.1 \text{Kg}(\text{m}^2\text{s}^{-1})^{-1}$ is imposed.
New Simplified Arakawa-Schubert
This scheme is based on changes to the Simplified Arakawa Schubert scheme proposed by Han and Pan (2011). The scheme was updated in order to improve the NCEP Global Forecast System (GFS). The new modifications include the parameterization of shallow convection without downdrafts. The mass flux in the shallow cumulus cloud base depends on the surface buoyancy. The starting level for the shallow convection is set up to be the maximum level of moist static energy within the planetary boundary layer. The entrainment rate during shallow convection is inversely proportional to the cloud height by a factor of 0.3 (0.1 for deep convection). These values are based on LES simulations. The detrainment rate is constant with height (Han and Pan, 2011).

Also, overshooting is now allowed during deep and shallow convection. This makes it possible to represent stronger precipitation events (Han and Pan, 2011). The maximum cloud base mass-flux is parameterized to depend on the model vertical layer thickness and time step. During deep convection, the entrainment rate in the cloud base depends on the relative humidity of the environment. The momentum transport between the clouds and the environment is limited to 55% in both shallow and deep convection.

The undiluted parcel ascent pressure change within the level of free convection is changed from 150hPa (see previous sub-section) to a value between 120 and 180hPa depending on the large scale velocity.

2.3.2. Planetary Boundary Layer Schemes
The parameterization of the planetary boundary layer is particularly important because this layer undergoes the most changes during the day, due to incoming radiation, heating, and cooling during the night. Since the planetary boundary layer gets warmer as the day evolves, the amount or turbulence and mixing will increase.
In numerical models, the PBL scheme is a 1-D scheme, which tries to obtain values for the vertical turbulent (') mixing of temperature ($\theta$), moisture (q), and momentum(u, v, w): $\frac{\partial}{\partial z} (w'\theta')$, $\frac{\partial}{\partial z} (w'q')$, $\frac{\partial}{\partial z} (w'u')$, $\frac{\partial}{\partial z} (w'v')$, respectively. Generally the turbulent mixing transport of heating and moisture is upwards (from the surface layer, to the PBL, to the free atmosphere), whereas the momentum transport is downward (as the greatest winds are in the upper levels). However, under unstable conditions $\left(\frac{\partial \theta_v}{\partial z} < 0\right)$, or mechanically forced conditions (from orography), these may change signs.

There are two main kinds of planetary boundary layer schemes, local and non-local. The local schemes refer to those in which the parameterization takes into account only the mixing produced by contiguous vertical layers. In a nonlocal scheme, the mixing that occurs between layers that are not contiguous is accounted for. This turbulence will have a larger scale than the local turbulence and it is usually called the counter gradient terms. In the following paragraphs we present the most used planetary boundary layer schemes used during this study.

**Yonsei University**

This planetary boundary layer scheme is based on Hong and Pan (1996) and Hong et al. (2006). This scheme has a nonlocal-K closure in the mixed-layer region, while in the rest of the atmosphere the approach is a local-K scheme. Local-K scheme is a parameterization based on local gradients of wind and potential temperature (Hong and Pan, 1996). The Yonsei University scheme (YSU) attempts to parameterize the counter gradient transport in the PBL by having a nonlocal-K closure. The counter gradient transport represent the vertical mixing due to larger scale eddies (larger than the separation of two consecutive vertical levels). The nonlocal mixing processes are only
parameterized for the horizontal winds and temperature processes excluding the moist-processes.

The YSU scheme also accounts for explicit entrainment mixing between the top of the PBL and the free atmosphere. Another important characteristic is that the PBL height is determined by the Bulk Richardson number.

In this scheme, the calculations of the eddy viscosity for the thermodynamic and for the momentum variables are different. In the calculation of the momentum eddy viscosity, the Prandtl number is included. This number accounts for the differences according to the dominant regime (inertial vs thermal). However, the momentum eddy viscosity is not susceptible to those changes, and has the same profile under different regimes. The numerical method used to calculate the diffusion coefficient is based on finite differences centered on a vertical layer.

**Mellor-Yamada-Janjíc**

This scheme is far more complicated than the YSU scheme. This scheme was develop over a 20-year period (Janjić, 1990; Janjić, 1994; Janjić, 2001; Mellor and Yamada, 1982). This is a 2.5 level scheme with an order of 1.5. This means that some variance terms are included in the Equations, as well as covariance between variables. The closure in this case is based on the turbulent kinetic energy (TKE) evolution. TKE only has local mixing in the PBL, which neglects the counter gradient terms available in YSU. MYJ has a master length scale ($\ell$), which is the scale of the eddies to be represented in the parameterization. (Mayor and Yamada 1982, Janjic 1990). The eddy diffusivity coefficients are proportional to the master length scale and TKE, with the effects of rotation and shear included. The effects of shear are included in the Richardson number (CAPE/Shear). Another important characteristic of this scheme is the amount of
hydrometeors for which the turbulent transport is calculated: water vapor mixing ratio, cloud water mixing ratio, ice, and others. This will lead to a more realistic representation of the vertical distribution of the different hydrometeors in the free atmosphere. However, an over-transport of the latter would cause changes in the cloud and radiation fields during the convective events.

**University of Washington (Park-Bretherton)**

The UW scheme constitutes the newest of the schemes used in this study. This scheme is based on the work of Bretherton and Park (2009) as they implemented the scheme in the CAM3 numerical model. Regardless that this scheme was mostly developed in 2009, it includes old ideas such as the down gradient diffusion of momentum and thermodynamic variables, and explicit entrainment rate. This is a local scheme in which the TKE is diagnosed. The new idea in this scheme is the coupling of the PBL processes and the shallow cumulus processes. This scheme was designed so it would capture the shallow cumulus over the tropical ocean. This makes the transition from the PBL to the free atmosphere smoother than in other cases. This is relevant in our study because a good representation of the transition between shallow to deep convection will help in the early stages of development of the MJO. In WRF, ideally, this scheme would be coupled with the Park-Bretherton shallow scheme.

The UW planetary boundary layer scheme represents all the turbulence by using downgradient diffusion fluxes. These fluxes are calculated between vertical layers (Bretherton and Park, 2009). The eddy diffusivity is proportional to the dissipation scale of TKE and the stability corrected length scale. The value of the dissipation scale is calculated using LES simulation results and Mellor and Yamada (1982) formulation.
2.3.3. Microphysics Schemes

In a numerical model, the microphysics are referred to as the parameterization which accounts for grid scale condensation, as well as for ice processes and phase changes. There are 2 main kinds of microphysics schemes, single and double moment. The single moment schemes will only predict the value of atmospheric hydrometeors, whereas the double moment will also predict the rain drop number concentration as an independent variable. Similarly to the cumulus scheme, the microphysics scheme will add tendency terms to the humidity and temperature tendency Equations.

WRF Single moment 3-class

This is a bulk parameterization scheme based on the work of Hong et al. (2004) and Dudhia (1989). This scheme is called a 3-class scheme because it has 3 different categories of moisture: water vapor, rain water, and cloud water mixing ratio, all associated with warm precipitation processes. The latter makes this scheme computationally efficient. Ice (from cloud water) and snow (from rain) amounts are calculated based on ice saturation mixing ratio when the temperature is less or equal to 0 °C. This implies that this scheme does not include supercooled water or mixed processes (water vapor ↔ ice ↔ snow).

This parameterization includes the sedimentation of cloud crystals in which the mean terminal velocity of the ice depends on the air density and the ice mixing ratio. WSM3 also differentiates between the ice nuclei and the ice crystals number concentration. The former is set to depend on the ice mixing ratio with the cloud, whereas the ice nuclei number concentration depends on the exponential function of the difference between the cloud temperature and freezing.
Lastly, the rates for cloud ice initiation, ice deposition, ice accretion by snow, ice aggregation to snow, cloud water autoconversion to rain, and rain (snow) evaporation (sublimation), are taken from Dudhia (1989) and Rutledge and Hobbs (1984).

**WRF Single moment 6-class**
Different from WSM3, this scheme (WSM6) has 6 classes of hydrometeors: water vapor, cloud water, rain water, snow, ice, and graupel mixing ratio. In this case, ice, snow and graupel are separate classes, unlike WSM3. By the inclusion of these 3 hydrometer classes, this parameterization includes mixed processes and supercool water.

The most significant changes between WSM6 from WSM3 were made by Dudhia et al. (2008). The main assumption in this study is that riming will equally affect all snow and graupel particles in the same grid box. Therefore, the fall speed must be that of a partially rimed particle, i.e. weighted by the amount of graupel and snow within the grid box (Dudhia et al., 2008). As a result, the accretion between graupel and snow becomes nonexistent in the scheme. The weighed fall speed Equation (see Equation 1 in Dudhia et al. (2008)) is also applied to accretion processes with cloud and rain water.

**2.4. Grid and Spectral Nudging in WRF**

Formally called four dimensional data assimilation (FDDA), usually called nudging, this is the method by which numerical simulations are approach towards observations or reanalysis fields. The nudging technique is use to correct model errors by forcing specific model fields towards observations (Bengtsson et al., 2004). This approach helps examine the effect of model errors in a given variable. By applying nudging, the model simulation remains under a controlled environment, because the nudging strength and location are easily determined by constant numerical parameters.
(see below). The nudging tendencies are particularly telling about model behaviors (Mapes and Bacmeister, 2012). Examining the effects of fields corrected by nudging provides insights into the processes that need to be better represented for improved simulations. In nudging, for a given prognostic variable $\alpha$, an extra term $\left(\frac{\partial \alpha}{\partial t}\right)_{\text{Nudging}}$ is added to its tendency Equation. Following Equation 2.1:

$$\frac{\partial \alpha}{\partial t} = -\nabla \cdot (a \vec{V}) + \tilde{F}(a) + \left(\frac{\partial \alpha}{\partial t}\right)_{\text{Nudging}} \tag{2.5}$$

The third term in the RHS is the nudging tendency. This is defined by Stauffer and Seaman (1990) as

$$\left(\frac{\partial \alpha}{\partial t}\right)_{\text{Nudging}} = G_{\alpha} W_{\alpha} (\alpha_0 - \alpha) \tag{2.6}$$

In (2.5) $G_{\alpha}$ is a nudging factor (an inverse time scale), which determines the strength of the nudging relative to $\tilde{F}(a)$. $W_{\alpha}$ is a four-dimensional weight function, and $\alpha_0$ is a reference value. This nudging procedure forces simulated $\alpha$ to approach $\alpha_0$ at a certain rate ($G_{\alpha}$) and structure ($W_{\alpha}$). There are two types of nudging techniques: grid and spectral nudging. In grid nudging, (1) is applied equally at all model grids and $\alpha_0$ is the “true” value at a specific time and location, which is often substituted by data from a reanalysis product. In spectral nudging (Miguez-Macho et al., 2004; Miguez-Macho et al., 2005), a Fourier Transform is applied to $\alpha$ and $\alpha_0$ to decompose them into their sets of zonal and meridional modes. The resulting nudging tendency $\left(\frac{\partial \alpha}{\partial t}\right)_{\text{Nudging}}$ is the correction of the selected modes. This is an effective way to examine the sensitivity of a model to the correction of a zonal scale of a given variable. All the spectral nudging experiments were designed to test the sensitivity of MJO simulations to the zonal scale of
corrected water vapor. The simulations using nudging were performed in the CH domain, in order to study the effect of the modes of the planetary scale moisture (see Chapter 5).

In WRF, the reference variable \( a_0 \) can be winds, temperature, water vapor mixing ratio, and geopotential height in the grid nudging module, and geopotential height, temperature, and winds in the spectral nudging. Since we are interested in the role of moisture in the MJO, we consider it necessary to add the option to use spectral nudging of water vapor mixing ratio (not included in the default WRF version). In this study, when nudging is used, the simulations are approaching to linearly-interpolated 6-hourly ERAI data.

A detailed description of the numerical simulations where nudging was performed is given in Chapter 5.
Chapter 3

Tracking the Eastward Propagation of the Madden-Julian Oscillation

3.1. Motivation

In this chapter we present a new technique in order to quantify MJO cases of study. Usually, the measure of the MJO is made by the use of the all-season real-time multivariate MJO (RMM) index proposed by Wheeler and Hendon (2004). This index is based on the two leading empirical orthogonal functions (EOFs) of anomalies in global tropical OLR and upper- and lower-tropospheric zonal winds. The principal components (PCs) of the two leading EOFs defined the MJO phases and amplitude (see Figure 8 of Wheeler and Hendon (2004)). Canonical locations of MJO convection centers can be constructed in a composite based on the RMM index. The phase diagram of the RMM index provides a general sense of eastward propagation of a given MJO event, however, it does not measure quantitatively the phase speed and it cannot tell the precise longitudinal location of the convection center of an specific MJO event. The RMM index provides a straightforward and objective way to extract MJO signals without bandpass filtering and it was developed for real time monitoring of the MJO and has received much
broader applications (Gottschalck et al., 2010; Kang and Kim, 2010; Wheeler and Hendon, 2004). This definition of MJO phases has also been used widely in diagnostics of the MJO and related fields (see a summary in Zhang (2013)). Its suitability for real time applications makes it a primary tool of measuring MJO real time forecast skill by dynamical and statistical models (Gottschalck et al., 2010). However, the inability of the RMM index to accurately describe the location of active convection for individual MJO events is particularly evident at the convective initiation stages of the MJO over the Indian Ocean (Gottschalck, 2013) mainly due to its global approach. Recent studies have shown the large sensitivity of the RMM index to the global patterns of zonal winds rather than precipitation. Straub (2013) studied in depth the limitations of the RMM index and reached the conclusion that this index is not a good measure for MJO convection initiation. Because of its large dependence on the global circulation anomalies rather than OLR (Straub, 2013).

Therefore, MJO convection cannot be accurately monitored by the RMM index immediately after its initiation. Accurately forecasting the local behavior of MJO convection and precipitation is equally important as accurately forecasting global patterns and evolution of MJO winds and OLR. Many MJO impacts in the tropics, such as those on the distribution of fires and floods (Reid et al., 2011; Zhang, 2013), are directly related to the location and intensity of its precipitation. MJO influences on remote regions at higher latitudes depend on the relative locations of MJO convection and the extratropical large-scale circulation, especially the westerly jet (Higgins and Mo, 1997; Higgins and Schubert, 1996; Matthews et al., 2004). By no means the objective of this method is to try to diminish the importance and utility that the RMM has had and will continue to
have. However, this method is a simple and direct measure of MJO convection (and other variables).

The new method that we are introducing provides objective measures of the strength, propagation speed, and timing of MJO precipitation over the Indian and Pacific Oceans. This method can be used to 1) quantify these basic MJO characteristics for a given individual event, 2) compare two or more MJO events in terms of the parameters cited above, 3) summarize statistics of the MJO when applied to climatological data, and 4) measure skill of model simulations and forecasts.

3.2. The Tracking Method

The MJO tracking method is designed to be applied to a specific MJO event, however this may be identified. For every variable relevant to the MJO (precipitation, humidity, wind, and others), two quantities (speed and strength) can be calculated to describe the event in terms of each variable. The quantification of the MJO behavior in different variables is very valuable because it allows an for objective comparison (evaluation) of MJO events (MJO simulations).

Let’s consider a three dimensional variable $\alpha = \alpha(x, y, t)$. We define the meridional mean (time mean) of $\alpha$ as $\langle \alpha \rangle (\bar{\alpha})$. The meridional mean operator is defined to be from $10^\circ S$ to $10^\circ N$. The tracking method is applied to the smoothed anomalies (5-day running mean) of $\alpha$ ($\alpha'$) with respect to the time and meridional mean at each point. Therefore, $\alpha' = \langle \alpha \rangle - \langle \alpha \rangle^\circ$; where $\alpha' = \alpha'(x, t), \langle \alpha \rangle = \langle \alpha \rangle(x, t)$, and $\langle \alpha \rangle = \langle \alpha \rangle(x)$

---

5 Note that if $\alpha' = \alpha'(x, t)$, this indicates that we will have a time-longitude diagram (also call hovmoler diagram).
The tracking method can be applied to different thermodynamic and dynamic variables, however, we will describe this method using precipitation \((P)\): \(P' = \langle P \rangle - \langle \tilde{P} \rangle\) for now on. Where \(P'(x,t)\) are the smoothed anomalies.

The MJO can be identified in a contour map of \(P'\) (Hörmoller or time-longitude diagram) as the eastward moving positive anomalies. The latter has been used in numerous studies. The angle of different diagonal lines following the positive anomalies of \(P'\) will represent the speed of the MJO. In the literature, several studies used an “eyeballing” drawn track to attribute propagation speed to MJO case(s) (Hagos et al., 2011; Lappen and Schumacher, 2014; Powell, 2014; Ray et al., 2009; Seo et al., 2014). However, an explanation as to why such a track was chosen is usually vague. The objective of this method is to look for the track that best describes each particular MJO event among all tracks.

The first step is to draw a series of straight lines with different slopes (from now on called tracks) in the time-longitude diagram. Each track starts at a day \(d\), and it reaches up to a chosen longitude \(L\). Each track represents a possible MJO path on the time-longitude diagram. The number of tracks drawn for a specific event depends the time frequency of the data \(M\) and the number of slopes that want to be tested \(S\). In total there are \(S\) number of tracks drawn for unit of time (one per each slope that needs to be tested) . Therefore the total number of tracks will be \(M \cdot S\). In order to take advantage of the spatial resolution of the data, linear interpolation may be used to obtain precipitation values on the line from nearest off-line grid points. The longitudinal extent on which the tracks are drawn \(L\) will depend on the availability of the data in the case of numerical
model simulations; ideally L should cover the totality of the Indian Ocean. Figure 3.1 shows an idealized tracking diagram for $M=60$, $S=34$, and $L=110$.

Each track $T$ in Figure 3.1 may be identified for three quantities 1) slope, 2) strength, and 3) start date.

1. The strength of the track is defined as the addition of the positive precipitation anomalies along the track, normalized by the number of grid points within L ($N_T$), such that the bias towards the longer tracks gets eliminated:

$$
\|T\| = \frac{1}{N_T} \sum_{i=1}^{N_T} P'_i (> 0) \quad (3.1)
$$

2. The slope ($S$) is the corresponding speed associated with each track. In Figure 3.1 $S$ can be any speed from 4 to 20 ms$^{-1}$.

**Figure 3.1.** Tracking method schematic. The x-axis is longitude, which has a range L. The y-axis is the time (days). The number of days is equal to M. Each line represents a different track with different slopes. The speed associated with each slope is represented by the colors. For this example $L=110$, $M=60$, and $S=34$. The tracks are drawn for

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$^6$ The strength of the track may also be defined as the addition of the negative anomalies. This, if negative values of the variable are associated with MJO. For example OLR.
day=10 in order to make the diagram easy to understand (each day should have the same number of tracks). The speeds to be tested are from 3 to 20 ms\(^{-1}\), with increments of 0.5ms\(^{-1}\) (34 slopes per day).

3. The start day of each track is the day on which each track starts. In Figure 3.1 d=10. Each day has S number of tracks for a total of \(M \cdot S\) tracks, and every track has a magnitude \(|T|\) (Equation 3.1). The track that is best associated to the MJO event will be the one with the largest magnitude (\(|T|\)). For a given MJO event there will be a track T with an amplitude \(|T|\), slope S, and start day d that characterize the MJO event the best.

**Figure 3.2.** Time – Longitude diagram of TRMM smoothed precipitation with respect to its time and meridional (10°S to 10°N) mean (a). Time - Speed diagram of \(|T|\) (b). The propagation speed of the MJO event calculated using the tracking method is 6 m s\(^{-1}\) (black line in 2a, and vertical line in 2b), the strength is 6.724 mm day\(^{-1}\) (black circle in 2b), and the starting day is 10 (black horizontal line in 2b). Day 1 is November 10\(^{th}\), 2011. The gray line (dotted gray line) is the track from applying the tracking method to
the ERAI zonal winds at 850 hPa (NOAA OLR), the resulting track magnitude and MJO speed are $2.4098 \text{ m s}^{-1}$ ($-27.5532 \text{ W m}^{-2}$) and $9.5 \ (7.0) \text{ m s}^{-1}$ respectively.

In the example of Figure 3.2a, the propagation speed of the MJO event measured by the tracking method is $6 \text{ m s}^{-1}$, its strength is $6.724 \text{ mm day}^{-1}$, and the starting day is 11 (November 21st, 2011).

Precise values of the strength, speed and starting date determined by the tracking method all depend upon the tracking range $L$. But once the tracking range $L$ is defined, they are all objectively, quantitatively, and uniquely determined for a given MJO event. These three parameters can then be compared between different MJO events, and between observations and forecasts to measure MJO forecast skill.

This method was originally designed to track MJO precipitation. However, there is no reason why it cannot be applied to other MJO fields. Figure 3.3 shows the hovmoller diagram and the slope-time phase diagram for ERAI U850 anomalies for the same MJO event. In this Figure we applied the tracking method to U850. Two tracks of interest have been highlighted in Figure 3.3a:

(i) The track with the largest amplitude (following the westerly anomalies)

(ii) The track with the closest magnitude to zero (following the convergence between the easterly and westerly winds).

The track with the largest amplitude has a speed of $9.5 \text{ m s}^{-1}$, and the track with the magnitude closer to zero have a speed of $8.0 \text{ m s}^{-1}$ (Figures 3.3a and b). It is expected that the convergence track will be located before the positive wind anomalies. For this case also, the positive anomaly track will be located after the maximum MJO precipitation has passed (gray solid line Figure 3.2a). Because of this timing of the westerly anomalies,
the tracking method was applied to the first 30 days of the simulations (instead of the 25
days used for precipitation).

**Figure 3.** Time – Longitude diagram of the U850 smoothed anomalies with respect to
its time and meridional (10 °S to 10 °N) mean (a). Time - Speed diagram of \( \|T\| \) (b). The
propagation speed of the MJO event calculated using the tracking method is 9.5 ms\(^{-1}\)
(black line in a, and vertical black line in b), the strength is 2.410 ms\(^{-1}\) (black circle in 3),
and the start day is 18 (black horizontal line in 3). The gray line in 3a is the track with the
closest magnitude to zero (0.635 ms\(^{-1}\), gray square in 3). This track has a speed of 8.0 ms\(^{-1}\)
(vertical gray line in b) and the start day is 12 (horizontal gray line in b). Day 1 is
November 10\(^{th}\), 2011.

3.2.1. Distinction Between MJO and No MJO Events

If we apply the tracking method to a simulation of the MJO event shown in
Figures 3.2 and 3.3 we can numerically compare the differences between their amplitudes
and speeds. This allows to quantify the errors. However, because of the nature of the
tracking method, there is always going to be a track with maximum or minimum amplitude. Therefore, we need a quantitative measure to discern between a simulation with or without MJO. In order to do this we use a normalized error distance (D). This error takes into account the MJO climatological variations in speed and amplitude. The value of D can be used to measure how close a simulated MJO event is to the observed in terms of its amplitude and propagation speed in combination.

In order to obtain D, we need the climatological values of the MJO speed and amplitude, and most importantly, their deviations (standard deviation). To obtain such values the tracking method should be apply to a long record (~10 years or more) of MJO events for a variable (In this case, precipitation and U850). Let’s also assume that $\sigma_A$ and $\sigma_S$ are the standard deviations in amplitude and speed (respectively) of all the MJO cases in a determined record for a determine variable $\alpha$. Let $A_{SIM}$ and $S_{SIM}$ are the amplitude and speed obtained from the application of the tracking method to a simulation SIM. Then, $A_{OBS}$ and $S_{OBS}$ corresponds to the results of applying the tracking method to either observations or reanalysis variables (in our case TRMM and ERAI). The normalized error (D) will be given as:

$$D = \sqrt{\left(\frac{A_{SIM} - A_{OBS}}{\sigma_A}\right)^2 + \left(\frac{S_{SIM} - S_{OBS}}{\sigma_S}\right)^2} \quad (3.2)$$

If D > 1, a simulation is considered to have failed to reproduce the observed signals of the MJO event because the difference between the simulated and observed MJO events is greater than expected normal MJO variability in nature, which is measured by the observed one standard deviation here (D=1). The separation of success (D $\leq$ 1) from failure (D > 1) might be too rigid and artificial, even though one standard deviation is commonly used to separate noise from signals. Once this criterion is set, however, it
provides an objective, uniform, and quantitative measure of MJO simulations, which has been lacking in the study of MJO simulations of individual cases. A series of values of $D$ from different simulations are more meaningful than a single one when they are compared to see how they approach 1 (a realistic simulation) or zero (a perfect simulation) when model configurations are modified.

Subsequently, for precipitation, if the observed (TRMM) amplitude for a given MJO event is $A_{\text{TRMM}}$ and the speed $S_{\text{TRMM}}$. The normalized error distance $D$ for the precipitation will be $D^{PCP}$, which is defined as follows,

$$D^{PCP} = \sqrt{\left(\frac{A_{\text{SIM}} - A_{\text{TRMM}}}{\sigma_A}\right)^2 + \left(\frac{S_{\text{SIM}} - S_{\text{TRMM}}}{\sigma_S}\right)^2}$$  \hspace{1cm} (3.3)

A similar $D$ can also be defined for U850. A realistic MJO simulation would have to satisfy a criterion that $D < 1$ for both precipitation and U850. This measure of tracked speed and amplitude avoids subjective assessment of the success or failure of a simulation of an individual MJO event by visual inspection of time-longitude diagrams, which has been the only practice in modeling case studies of the MJO up to recently.

The tracking method emphasizes the local skill of MJO simulations. It also complements the measure of global skill of MJO simulations and forecast based on the RMM1 and RMM2 indexes (Gottschalck et al., 2010; Wheeler and Hendon, 2004). A comparison of local vs. global MJO forecast skills of an operation model for three MJO events during the DYNAMO field campaign (Yoneyama et al., 2013) was given by (Ling et al., 2014).

In the following sections of this chapter, we apply the tracking method to 10 years of climatological data of precipitation and winds. We also apply the tracking to ECMWF deterministic and ensemble forecasts during DYNAMO (see Chapter 2, section 2.1) in
order to evaluate the different lead forecast times of ECMWF and their MJO performance in wind and precipitation.

3.3. Tracking Method Applied to Climatological Data

The tracking method was applied to a long record of daily data from TRMM (precipitation) and ERAI (U850). As mentioned in the previous section, first the MJO events dates should be given. In this study we took the MJO list of events from Ling et al., (2013). The Ling et al. (2013) definition of MJO must satisfy two criteria:

(i) Anomalies (from the seasonal cycle) in precipitation averaged over a large domain over the equatorial Indian Ocean (10˚S–10˚N, 60–90˚E) are above one standard deviation (1.45 mmmday⁻¹) for at least three consecutive days.

(ii) The precipitation anomalies should move eastward up to a certain longitude (100˚E).

According to Ling et al. (2013), 23 MJO events were observed in the 12 spring and winters in 1998 - 2009 (Table 3.1). The MJO cases during the northern hemisphere spring were used in order to avoid the complication of both eastward and northward propagation of MJO signals in the boreal summer monsoon season (Lau and Chan, 1986; Wang and Rui, 1990). Because we think that precipitation is be the most important field when quantify when MJO, we applied the tracking method to precipitation and U850 data during the specified MJO events in Ling et al. (2013). Table 3.1 lists the start day of the 23 MJO events according to Ling et al. (2013)
Table 3.1: MJO tracking results for TRMM precipitation.

<table>
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<th>Day = 0 (Ling et al. (2013))</th>
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<th>L=40°E - 160°E Start</th>
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<td>21</td>
<td>10/9/08</td>
<td>5.36</td>
<td>6.5</td>
<td>1</td>
</tr>
<tr>
<td>22</td>
<td>4/5/09</td>
<td>3.80</td>
<td>6.5</td>
<td>-3</td>
</tr>
<tr>
<td>23</td>
<td>10/30/09</td>
<td>6.53</td>
<td>3.0</td>
<td>-8</td>
</tr>
</tbody>
</table>

| MEAN   | 5.35 | 6.7  | -3 | 5.99 | 6.9  | -2 | 6.39 | 6.9  | 0   |
| MEDIAN | 5.17 | 6.0  | -3 | 5.92 | 6.5  | -2 | 6.40 | 6.5  | -1 |
| STD    | 1.55 | 3.4  | 6  | 1.73 | 2.8  | 5  | 1.79 | 2.7  | 5  |

Since the tracking method is formulated for specific MJO events, some modifications had to be made in order to use the 12 year records.

(i) The seasonal cycle was removed from the data.

(ii) The mean of the previous 120 days for each day were also removed.

(iii) The tracking method was applied to the twenty days before (-20) and thirty days after (+30) of day = 0 from Ling et al. (2013) to 5-day smoothed anomalies (Tables 3.1 and 3.2).

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* This in order to remove inter annual variations. Similar to Wheeler and Hendon (2004).
(iv) The largest track should be found between days -15 and +10 for precipitation and -15 to +20 for U850.

(v) The tracking method was applied for three different longitudinal extents (L): 40°E – 160°W, 40°E – 160°E, and 50°E to 160°E.

The results of applying the method to the precipitation and U850 are listed in Tables 3.1 and 3.2 respectively. The mean amplitude and speed as well as their standard deviation of the 23 selected tracks* listed at the bottom of Tables 3.1 and 3.2. These values will be needed in the following chapters when the normalized distance D is calculated.

* Appendix A and B of this thesis shows the individual results of the application of this method to each case listed in Tables 3.1 and 3.2.
### Table 3.2: MJO tracking results for U850.

<table>
<thead>
<tr>
<th>CASE #</th>
<th>Day = 0 (Ling et al. (2013))</th>
<th>L=40°E - 160°W</th>
<th>L=40°E - 160°E</th>
<th>L=50°E-160°E</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>1/17/99</td>
<td>Strength (ms⁻¹)</td>
<td>Speed (ms⁻¹)</td>
<td>Strength (ms⁻¹)</td>
</tr>
<tr>
<td>1</td>
<td>1.88</td>
<td>5.0</td>
<td>-2</td>
<td>2.03</td>
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<td>2</td>
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<td>0</td>
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</tr>
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<td>12</td>
<td>2.29</td>
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<td>6</td>
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<td>11.5</td>
<td>12</td>
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<td>8</td>
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<td>-12</td>
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</tr>
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<td>10.0</td>
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<td>3.96</td>
</tr>
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<td>11.0</td>
<td>7</td>
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<tr>
<td>23</td>
<td>2.90</td>
<td>7.0</td>
<td>12</td>
<td>2.90</td>
</tr>
</tbody>
</table>

**Mean**

| 2.64 | 8.5 | 6 |
| 2.67 | 8.3 | 8 |
| 2.81 | 7.0 | 9 |

**Median**

| 2.45 | 8.0 | 7 |
| 2.41 | 8.0 | 9 |
| 2.57 | 6.5 | 9 |

**Standard Deviation**

| 0.79 | 4.3 | 7 |
| 0.80 | 3.7 | 4 |
| 0.87 | 3.1 | 4 |
Figure 3.4. Scatter diagram amplitudes (x-axes) and speeds (y-axes) obtained from the tracking of (a) precipitation and (b) U850. The stars (squares) represent the mean (median) of the amplitudes and speeds for each L (Tables 3.1 and 3.2). The lines are the best fit lines.

For any of the three selected L, the speed in both precipitation and U850 seems to be more dispersed than the amplitude, especially in U850 (Figure 3.4). The values of the mean and the standard deviations are close between the three Ls. For neither wind nor precipitation, the addition of any of the means with any of the standard deviation for each case is greater any of the mean speeds, amplitudes, or start day for any of the three regions selected.

As it is expected, the magnitude of the mean and the deviations from the mean are greater as the longitudinal extent of L decreases. There is also agreement between the
different sections on the skewness of the histogram. This is specially noticeable in the Precipitation speeds (Figure 3.5c). Both magnitudes of the precipitation and winds are mostly in agreement, especially in the wind amplitude, in which the maximum corresponds to strengths centered around 2.5 ms\(^{-1}\) (Figure 3.5b).

**Figure 3.5.** Histograms of precipitation (left) and U850 (right) of amplitude (top) and speed (bottom) obtained from the tracking method. Each color is a different longitudinal range used in the tracking.

As mentioned in the previous section, the wind propagation speeds are usually greater than the precipitation (Figures 3.5c and d). Additionally, the differences between the start day in precipitation and U850 are very prominent, and consistent in all three longitudinal extents. According to this, the westerly wind anomalies (maximum in +12 days) may have arrived up to 14 days after the precipitation (maximum between -2 and +3 days) (Figures 3.6a and b).
Figure 3.7 shows the composite of precipitation and U850 anomalies for the 23 MJO events. From this we can observe that the tracking method is able to capture the main MJO behavior in both wind and precipitation out of a long record of MJO simulations. The main differences between the longitudinal extent is noticeable in the amplitude of the MJO (Figures 3.7 b and d).

Here on, the MJO composites shown in Figure 3.7 are going to be referred as the Climatological MJO. Their mean values in magnitude, strength, and start day, as well as their respective mean deviations are going to be useful when trying to categorize an MJO event as successful or not.
Figure 3. Time Longitude diagrams (left) and phase diagrams (right) of the composites of precipitation (top) and U850 (bottom) for all the MJO events listed in Tables 3.1 and 3.2. Each line in the time-longitude plots is the mean track speed for each L. Vertical (horizontal) line in the phase diagrams is the mean speed (magnitude) of the mean track. Day 0 is start day according to Ling et al. (2013).
Regardless of the fact that when we apply this method to the climatological data we can obtain statistics and mean or climatological MJO events. The tracking method shows deficiencies in some of the cases. The main deficiency of this method is the dependence on local precipitation or wind maximums, which will act as attractors for the tracks and ultimately lead to different results when the different longitudinal extents are applied. In the case of the precipitation, cases #1, 6, 14, 15, 16, 17, and 23 lead to different results in precipitation speed mainly because of the presence of localized precipitation maximums which may or may not be crossed depending on L (see appendix A). A similar deficiency is found when the tracking is applied to U850. In this case the scale of the local maximum is larger, however, the results are also affected primarily in terms of speed. Specifically in cases #3, 5, 6, 7, 10, 11, 14, 21, and 23 (see appendix B).

One of the possible solutions to this problem is to add an extra step, which would involved the overall scale of the precipitation pattern as a decision maker step, after finding the track with the maximum magnitude. The magnitude of the track may be a combination of both the addition of all the positive precipitation anomalies, as well as of the scale of those anomalies.

3.4. Evaluation of ECMWF Wind and Precipitation Using the Tracking Method

The quantification of the characteristics of the MJO case during DYNAMO is relevant because this information can be use to validate the vast amount of model data available during the experiment (Yoneyama et al., 2013). Therefore, the tracking method provides a valid tool in order to compare different regional and global model results.
quantitative. The tracking method has been successfully used to evaluate the performance of the global climate models precipitation forecast during DYNAMO (Ling et al., 2014).

In this section we go in depth in the characterization of this event in terms of its convective and dynamical features. Therefore, the tracking method is applied to zonal winds and precipitation. Additionally, we demonstrate the utility of the tracking method by the quantitatively validation of the ECMWF forecasts during DYNAMO.

Figures 3.2 and 3.3 show the resulting tracks from applying the tracking method to the precipitation anomalies from TRMM (Figure 3.2) and from ERAI (Figure 3.3) during the MJO-DY case of study. The speed of the MJO zonal winds associated with the MJO are the fastest with a speed of 9.5 m$^{-1}$ while the precipitation associated to the MJO has the slowest speed 6 m$^{-1}$, which is the closest to the usually proposed speed (5m$^{-1}$) for the MJO to have (Zhang, 2005).

The zonal wind in the other hand shows a different track, this is because the tracking method looks for the maximum track magnitude, which is usually located after the convective center. However, if we take a close look at the track in a time-longitude diagram of the wind anomalies we can see the convergence zone located before the westerly anomalies. If instead of the tracking the positive anomalies of the wind, we track the track with the closest magnitude to zero, we find that the track is now shifted to start closest to the precipitation start (from day ~19 to day ~11, Figure 3.3).

However, the quantification of the MJO strength associated to the winds is usually better visualized by the track following the positive value of the winds rather than the minimum value, because the former will give additional information about the strength of the event. We will track the maximum speed of the tracks in order to obtain a
quantifiable value for the MJO strength in terms of its winds, however, depending in the type of study and analysis, the convergence line may also be tracked.

The MJO tracking method is applied to forecast fields of the ECMWF as a measure of MJO forecast skill. The purpose is to demonstrate the utility of the tracking method in quantifying MJO forecast skill for a given individual event. The ECDY include:

(i) A deterministic forecast of 10 days with 16 km grid spacing and 91 vertical levels (T1279 L91), initialized four times a day (00Z, 06Z, 12Z, and 18Z).

(ii) Ensemble forecast of 15 days using the Ensemble Prediction System (EPS) that includes one control run and 50 perturbed members with varying model configurations (T639L61 through day 10, T319L61 afterwards), initialized daily (00 and 12Z).

(iii) A control run with the fixed high horizontal resolution (T639L61) through out 15 days.

(iv) A control simulation with the fixed low horizontal resolution (T319L61) through out 15 days.

ECDY used persistent SST anomalies as the lower boundary condition through the first 10 days integration and an ocean model is coupled to the atmospheric models afterwards. The results and comparison with TRMM\textsuperscript{9} and ERAI\textsuperscript{10} data are shown in Figure 3.8.

\textsuperscript{9} The method was applied in the same longitudinal range as the deterministic and ensemble forecast (L=40°E-150°E).
\textsuperscript{10} The method was applied in the same longitudinal range as the deterministic and ensemble forecast (L=40°E-150°E).
All the ECMWF ensembles have an MJO strength slightly less than the observed (red line in Figure 3.8a), followed immediately by a sharp decrease for the lead time from 1 day to 2 days. This sharp decrease in predicted MJO strength is related to model adjustment to the initial conditions that are too moist in comparison to the model climatology. After that, the predicted MJO strength continues to decrease but only slightly. The spread of the ensemble forecast in MJO strength remains relatively small (≈15% of the observed strength) and does not increase obviously with the lead time up to 10 days. There is no clear winner among the different model configurations in forecasting the strength of this MJO event. However it is clear that there is an issue with the strength of the MJO even during the first 2 days in the forecast, even at 24h lead time. If the Hovmoller diagrams of the deterministic forecast (see appendix C) are examined carefully one by one, it can be seen that ECMWF tends to produce less precipitation than
TRMM (Figure 3.2) over 2 areas: the western portion of the Indian Ocean (50°-80° E) and over Australia.

The magnitude of the wind in the other hand, is greater than the value obtained for ERAI. This demonstrate the struggle of the capture of the convection by ECMWF. It is important to notice that ECMWF has a superior forecasting skill in both precipitation and wind than other GCMs (GFS) for this particular MJO event (see (Kerns and Chen, 2014a)).

**Figure 3.9.** Histograms of MJO speed (a) and start day (b) for precipitation for each lead time forecast. The colors are the lead time of the forecast. The squares are the ensemble mean values for each lead time. The circles are the values for the deterministic forecast at each lead time. The start is the value from TRMM.

Predicted propagation speeds of MJO precipitation are shown in Figure 3.9a. They start slightly less than the observed (7.5 ms⁻¹) and gradually but continuously increase with the lead time. Most of the ensembles show a speed of 6ms⁻¹ up to lead time
day 2. The deterministic forecast\textsuperscript{11}, in the other hand starts with speeds closer to the observed (Figure 3.9a). As the lead time increases, the majority of the ensembles shows greater precipitation speeds. This can be seen clearly in Figure 3.9 when we look at the ensemble and the deterministic forecast means (filled circles and squares). The progression of lead times (colors) will have greater speeds. The ensemble mean increases 0.5 ms\textsuperscript{-1} per every 2 days of lead time (days 1 to 4). Differently from the ensemble forecast, lead times 7, 8, and 9 of the deterministic forecast show slower propagation speeds.

In the case of the precipitation start day, there is a clear decrease (earlier start day) as lead time increase (Figure 3.9b). Both the deterministic and the ensemble mean very close to the start day in TRMM (red star in Figure 3.9.b). The effectiveness and importance of a ensemble forecasts system rather than the deterministic forecast can be seen in Figure 3.9b. Here we can see that lead times day 7 to 10 of the deterministic forecast show later start days, however, the ensemble means (squares) have a closer start day to TRMM. The weakening of the MJO signal is evident by the start of the capture of another MJO event in the deterministic and ensemble forecast after lead time = day 6 (Figure 3.9b).

When using the MJO tracking method, it seem that the U850 field has a greater spread within the ensembles. This can be seen in Figure 3.10. The propagation speed of U850 increases as the lead time increase in the ensemble mean. After day 7 of lead time, the MJO tracking method is not able the converge on a eastward moving feature in the wind, and the speed is then set to 20 ms\textsuperscript{-1} (see Figure 3.10a). The increase in the ensemble

\textsuperscript{11} Appendix C shows the time-longitude (precipitation and wind) and time – speed (tracks magnitude) diagrams for the ECDY deterministic forecasts at all lead times (day 1 to 10).
The spread is also noticeable, especially after lead time day 5. The increase in the spread can also be seen in the U850 amplitude (Figure 3.8b). Regardless of the decrease in the forecast skill of the ensemble system after lead time day 5, the first 4 lead days of the ensemble forecasts show a U850 propagation speed close to the ERAI values. Figure 3.10b also shows the agreement of the ensembles during the first 5 lead time days. This agreement is also evident on the start day of the eastward westerly wind anomalies (Figure 3.10b). In this case both the ensemble and the deterministic forecast show a later start day of the wind anomalies. This is differently from the precipitation, where the forecasts start day is earlier than the observed (TRMM). This means a progressive decoupled of the wind and the convection, as the lead time increases in the forecast.

**Figure 3.10.** Same as Figure 3.9 but for U850. The start day is the value obtained from ERAI.
3.5. Chapter Summary

An MJO tracking method was introduced. It was designed to quantify the strength, eastward propagation speed, and timing of precipitation for individual MJO events. It can be applied to a number of MJO events to derive MJO statistics. When equally applied to observations and forecast, hindcast, and simulations, it measures model skills of forecasting and simulating the MJO in terms of the three MJO parameters.

The MJO tracking method is complementary to the commonly used RMM index (Wheeler and Hendon 2004). The two have their own pros and cons. The RMM index was designed for real time monitoring of the MJO. The MJO tracking method must be further improved before it can be applied in real time. Perhaps the main weakness of this event is the dependence of the precipitation propagation speed to local maximums of precipitation. Depending on which longitudinal extent is chosen to track a specific MJO case, such event speed may be influenced by the amount and location of these localized events. This can be solved by formulating a new track magnitude quantity which will depend on the scale of the main features as well as on the amount of precipitation.

However, regardless of this deficiency, the use of the tracking method can be beneficial to quantify the differences between MJO simulations and observations or to evaluate MJO simulations. In this case the choice of tracking length (L) should be limited by the numerical model domain (such as in section 3.4).

Another use for this method is to measure to some degree the level of decoupling between the precipitation and the wind anomalies within MJO cases. Figure 3.11 shows
the tracked\textsuperscript{12} propagation speeds of U850 and precipitation for each of the 23 MJO cases, along with their mean and median values.

The tracked mean (median) amplitude and propagation speed are 5.35 (5.17) m\,m\,day\textsuperscript{-1} and 6.65 (6.0) m\,s\textsuperscript{-1} for precipitation, and 2.64 (2.45) m\,s\textsuperscript{-1} and 8.5 (8.0) m\,s\textsuperscript{-1} for U850 for $L=\text{40}^\circ\text{E-160}^\circ\text{W}$. The standard deviation of the amplitude and speed are 1.5 m\,m\,day\textsuperscript{-1} and 3.4 m\,s\textsuperscript{-1} for precipitation, and 0.79 m\,s\textsuperscript{-1} and 4.3 m\,s\textsuperscript{-1} for U850. Figure 3.11 shows certain degree of decoupling between the precipitation and the wind signal within the MJO events. While amplitudes of MJO precipitation and U850 are highly correlated in both observations and simulations, their propagation speeds are not (Figures 3). While the decoupling might be counterintuitive based on the perception that MJO is a product of convection-circulation interaction (Rui and Wang, 1990), it may not be so surprising if one accepts the possibility that MJO dynamics are determined by its intrinsic structure that is independent of convection although the two actively interact (Figure 3.11). Such signals of MJO intrinsic structure not directly associated with convection have begun to be revealed (Ling et al., 2013; Matthews, 2008; Straub, 2013). The MJO circulation appears to be determined by such an MJO dynamic structure, not solely by MJO precipitation. This results is shown in Ulate et al. (2015).

Using the MJO tracking method to objectively identify MJO events has so far led to mixed results (see appendix D). The RMM index treats the MJO as a convection-circulation coupled phenomenon. MJO tracking applies to only one field at a time, hence no coupling implied. For individual MJO events, the RMM index is unable to tell precisely the location or timing of their convection centers, and to measure their eastward

\textsuperscript{12} The tracked propagation speeds for both precipitation and wind were made using TRMM and ERAI data re-gridded to 1 degree horizontal resolution for 2 different longitudinal extents (see ).
propagation speeds. MJO tracking can. Both measure the strength of the MJO. The most
distinct and complementary difference between the two is that the RMM index is a global
measure of the MJO, while MJO tracking is a local one. When both are applied to
measure forecast skills of the same MJO events, interesting and intriguing results emerge.
This method will be used in the following chapters.

**Figure 3.11.** Scatter diagram of propagation speeds of U850 vs precipitation, obtained
by applying the MJO tracking method to ERAI and TRMM data for 23 MJO events
during 1998 - 2009. The longitudinal range (L) used for the tracking method was 40°E -
160°W (blue circles) and 40°E - 160°E (orange circles). The mean (median)
climatological value for each L is marked as a star (square) of each color. The mean
(median) climatological values for precipitation and zonal wind propagation speed are
6.65 and 8.5 (6.0 and 8.0) ms\(^{-1}\) for L=40°E-160°W and 6.9 and 8.3 (6.5 and 8.0) ms\(^{-1}\) for
L=40°E-160°E. The propagation speed (magnitude, not shown) standard deviation
between the 23 cases for L=40°E-160°W is 4.3 m s⁻¹ (0.79 m s⁻¹) for wind and 3.4 m s⁻¹ (1.55 mm day⁻¹) for precipitation; and 3.65 m s⁻¹ (0.80 m s⁻¹) for wind and 2.80 m s⁻¹ (1.73 mm day⁻¹) for precipitation for L=40°E-160°E. The solid orange line is the linear fit between the circles (MJO propagation speed for L=40°E-160°W) using a robust linear least-squares method with bi-square weights (to minimize the effect of outliers points). The solid blue line has the same fit but for MJO speeds obtained using a longitudinal range of L=40°E-160°E (instead of 40°E-160°W). The Equation of the fitted line for L=40°E-160°W (L=40°E-160E), solid orange (blue) line, is y=0.39+5.2 (y=0.44x+5.2). Both coefficients are within the 95% confidence bounds for both fitted lines. The dashed black line is the one-to-one (y=x) line. The black “x” marks the values of wind and precipitation propagation speed for the October- November 2009 MJO case.
Chapter 4

The Water Cycle Over the Indian Ocean and Maritime Continent

4.1. Motivation and Background

The atmospheric water cycle is a fundamental process in the earth system, and as a fundamental process, ideally, it should be captured in a numerical model. In the tropics, the water cycle plays an important role in transporting water vapor and its different phase changes: atmosphere to the ground (precipitation), surface to atmosphere (evaporation), and horizontal and vertical redistribution of water (large scale transport). In the MJO context the water cycle will also play an important role, since changes in precipitation, evaporation and moisture transport in general are key during the initiation and development of the MJO.

The different components of the water cycle can be estimated from observations and simulations of numerical models. The most important quantities related to the water cycle within the numerical model are precipitation, surface evaporation, large-scale transport, and moistening and drying by unresolved (subgrid) processes such as cumulus convection, turbulence, and microphysics. Changes any of these components will affect
global and regional climate (Genio et al., 1991; Held and Soden, 2006; O’Gorman and Schneider, 2008; Trenberth and Guillemont, 1998). Specifically, several studies have shown the principal changes of the water cycle in a global warming context (Genio et al., 1991; Held and Soden, 2000; Trenberth, 1998). Most of these studies agreed that an increase in temperature would lead to a strengthening of the water cycle by increasing the moisture residence time in the atmosphere. This would lead to intense precipitation events, more storms, and therefore more floods (Cubasch et al., 2001; Karl and Trenberth, 2003; Knutson and Tuleya, 2004).

In modeling, any physical process is parameterized, errors arise due to their different closures and assumptions and the water cycle is no exception. Every parameterized process has errors associated with it, and the combined influence and interaction between these errors can be very difficult to piece apart. In the context of the water cycle and more importantly the MJO, the physics errors will affect convection, vertical moisture distribution, large-scale circulation, and others.

The global water cycle has been studied using global reanalysis data and output from numerical simulations of global models. These data and model simulations all have their own respective biases (Andersson et al., 2005; Bengtsson et al., 2004; Hack et al., 2006; Hagemann et al., 2006; Hagos and Leung, 2012; Ling and Zhang, 2013; Trenberth and Guillemont, 1998). These biases are mostly due to misrepresentations by parameterization of physical processes that comprise the water cycle, in addition to model resolutions (Hagemann et al., 2006). The degree to which the water cycle depends upon physical parameterization remains largely unexplored, despite its importance (Johnson, 1984). Previous studies have focused on climate models both coupled and
uncoupled to the ocean (Dai, 2006; Li et al., 2007; Liepert and Previdi, 2012). Common model biases include too frequent precipitation in the tropics (Dai, 2006; Stephens et al., 2010), and too strong precipitation relative to evaporation (Liepert and Previdi, 2012) that indicates a too active water cycle. Connections between biases of tropospheric moisture and precipitation depend on regions (Ma et al., 2013).

In this chapter we focus on the changes in the main water cycle characteristics due its sensitivities in planetary boundary layer and cumulus parameterization schemes. We chose this two parameterizations because the processes parameterized in them are crucial for the evaporation, convection, and moisture redistribution in the atmosphere, which are related to the Water Cycle and the MJO. Cumulus and planetary boundary layer schemes are the largest contributors to the diabatic heating during convection. In WRF these contributions range between 60% (cumulus), 23% (planetary boundary layer), 7% (microphysis), and 10% (radiation), for heating processes and 10% (cumulus), 50% (planetary boundary layer), 20% (microphysics), and 20% (radiation) for cooling processes (Figure 4.1).

The cumulus scheme is responsible for much of the heating in the lower and middle troposphere, whereas the cooling processes are dominated by radiative and microphysics processes. However, in the upper levels, as the moist convective processes decrease, so does the heating produced by the cumulus scheme, and the microphysics and radiations take over; the same occurs for the cooling processes.
**Figure 4.1.** Mean diabatic heating (K day$^{-1}$) vertical profile (a) over the Indian Ocean during an MJO case of study. Percentage (%) of heating (a) and cooling (b) for each parametrized process in WRF. The colors represent each one of the parametrized process.

Sensitivities to their planetary boundary layers (PBL) and cumulus parameterization schemes have been widely studied for different atmospheric phenomena such as hurricanes (Braun and Tao, 2000; Hill and Lackmann, 2009; Nolan et al., 2009), monsoons (Evans et al., 2011; Flaounas et al., 2010), mesoscale convective systems (Crétat et al., 2011), amongst others. Regional models have been used to explore sensitivities of their precipitation to parameterization schemes (Gianotti et al., 2012; Pohl et al., 2011). To the best of our knowledge, there has been no systematic documentation of sensitivities of the large-scale water cycle over the tropical ocean to cumulus and PBL parameterizations.
The purpose of this chapter is to document the sensitivities of the water cycle in the tropics to cumulus and PBL parameterization schemes in a regional model. The focus is the Indian Ocean (IO) and Maritime Continent (MC), which play important roles in the global water cycle system (Bosilovich and Schubert, 2002; Manabe and Holloway, 1975; Roads, 2003). The west IO is the largest source of moisture for the Indian summer monsoon, and evaporation over the southern IO is an important source of water for the Sahel region (Bosilovich and Schubert, 2002). The MC region hosts a tropical precipitation maximum rivaled only by those in the South Pacific convergence zones, Africa and South America (Roads, 2003). Precipitation produced by global climate models often suffers from large biases over the IO and MC (Hack et al., 2006; Hagemann et al., 2006; Neale and Slingo, 2003; Schiemann et al., 2013). This chapter is an extension of a paper published in the Journal of Advances in Modeling Earth Systems (Ulate et al., 2014).

4.2. Model Simulations

**Figure 4.2.** Domains of the simulations and diagnostics. The larger inner domain is referred to as the IOMC domain. The smaller inner domain over the Indian Ocean is the IO domain, and the one over the Maritime Continent is the MC domain. Colors are for SST and gray shades for topography.
The domain configuration of the simulations used in this chapter is 10_50 (red rectangle, Figure 2.3). We choose three diagnostic domains which are outlined in Figure 4.2. The large domain covering both the IO and MC (hereafter referred to as IOMC) is essentially the same as the model domain, except for the exclusion of the 10 grid points nearest to the boundaries. One smaller domain is over the equatorial IO (10°S – 5°N, 65 – 95°E) and another over the MC (10°S – 5°N, 95 – 125°E). Hereafter, we will use IO and MC to refer to both these regions and their model diagnostic domains.

**Table 4.1.** Cumulus and planetary boundary layer schemes used.

<table>
<thead>
<tr>
<th>Schemes</th>
<th>Acronym</th>
<th>Reference</th>
</tr>
</thead>
<tbody>
<tr>
<td>Betts-Miller</td>
<td>BM</td>
<td>Betts (1986); Betts and Miller (1986)</td>
</tr>
<tr>
<td>Kain-Fritsch</td>
<td>KF</td>
<td>Kain and Fritsch (1990)</td>
</tr>
<tr>
<td>New simplified Arakawa-Schubert</td>
<td>NSAS</td>
<td>Han and Pan (2011)</td>
</tr>
<tr>
<td>Park and Bretherton</td>
<td>PB</td>
<td>Park and Bretherton (2009)</td>
</tr>
<tr>
<td>Simplified Arakawa-Schubert</td>
<td>SAS</td>
<td>Pan and Wu (1995)</td>
</tr>
<tr>
<td>Tiedtke</td>
<td>TK</td>
<td>Tiedtke (1989)</td>
</tr>
<tr>
<td>Zhang-McFarlane</td>
<td>ZM</td>
<td>Zhang and McFarlane (1995)</td>
</tr>
<tr>
<td>University of Washington</td>
<td>UW</td>
<td>Bretherton and Park (2009)</td>
</tr>
<tr>
<td>Yonsei University</td>
<td>YSU</td>
<td>Hong et al. (2006)</td>
</tr>
</tbody>
</table>

Eight cumulus schemes and three PBL schemes were used in this study (Table 4.1). These schemes are commonly used in simulations of regional and global weather
and climate. The cumulus schemes are based either on principles of mass flux (KF, SAS, NSAS, TK, ZM) or adjustment (BM, BMJ). Some include treatment of shallow convection (NSAS, SAS, TK, and PB), and others convective momentum transport (TK, ZM, NSAS). The PBL schemes are based on approaches of local turbulent kinetic energy (MYJ, UW) or nonlocal mixing (YSU). There are different versions of the same types of schemes (e.g., BM vs. BMJ, SAS and NSAS) and slight modifications of the same scheme, such as turning off the shallow cumulus scheme in NSAS (NSASnoSH).

A total of sixteen numerical simulations were performed, each with its own combination of cumulus and PBL schemes (Table 4.2). In two runs cumulus schemes were disabled (NoCU-MYJ, NoCU-YSU). These simulations provide direct comparisons of the water cycle simulated using the same cumulus scheme paired to different PBL schemes, and vice versa. Other parameterization schemes remain unchanged in all simulations: the WRF single-moment 6 class (WSM6) the new Rapid Radiative Transfer for General Circulation Models for both longwave and shortwave radiation (Mlawer et al., 1997); the Monin-Obukhov and Monin-Obukhov-Janjic surface layer scheme (Monin and Obukhov, 1954); and the Noah land-surface model (Chen and Dudhia, 2001).
The integration time of all simulations was 92 days, beginning on October 1, 2009, with output every 6 hours. The initial and boundary conditions were from ERAI. The lateral boundary conditions were updated every 6 hours. Over the ocean, time independent SST from the same reanalysis was used as the lower boundary condition (Figure 4.2). When the SSTs are time dependent in a model without coupling to an ocean model, there would be an unrealistic shift in the relative phases between SST, convection center, and its related surface wind. They in reality, interact with each other instead of convection being forced by varying SST. This shift would introduce additional errors in surface fluxes. Therefore we decided not to include a simulation with prescribed SST with time independent fluctuations. All the simulations were initialized and forced at the lateral boundaries by a global reanalysis dataset.

<table>
<thead>
<tr>
<th>SIMULATION NAME</th>
<th>CUMULUS SCHEME</th>
<th>PLANETARY BOUNDARY LAYER SCHEME</th>
</tr>
</thead>
<tbody>
<tr>
<td>BM-MYJ</td>
<td>Betts-Miller</td>
<td></td>
</tr>
<tr>
<td>BMJ-MYJ</td>
<td>Betts-Miller-Janic</td>
<td></td>
</tr>
<tr>
<td>ZM-MYJ</td>
<td>Zhang-McFarlane</td>
<td></td>
</tr>
<tr>
<td>ZMPB-MYJ</td>
<td></td>
<td></td>
</tr>
<tr>
<td>ZMPB-UW</td>
<td></td>
<td></td>
</tr>
<tr>
<td>TK-MYJ</td>
<td>Tiedtke</td>
<td></td>
</tr>
<tr>
<td>TK-YSU</td>
<td></td>
<td></td>
</tr>
<tr>
<td>SAS-MYJ</td>
<td></td>
<td></td>
</tr>
<tr>
<td>SAS-UW</td>
<td>Simplified Arakawa-Schubert</td>
<td></td>
</tr>
<tr>
<td>SAS-YSU</td>
<td></td>
<td></td>
</tr>
<tr>
<td>NSAS-MYJ</td>
<td>New simplified Arakawa-Schubert</td>
<td></td>
</tr>
<tr>
<td>NSAS-YSU</td>
<td></td>
<td></td>
</tr>
<tr>
<td>NSASnoSH-YSU</td>
<td>New simplified Arakawa-Schubert-No Shallow scheme</td>
<td></td>
</tr>
<tr>
<td>KF-MYJ</td>
<td>Kain-Fritsch</td>
<td></td>
</tr>
<tr>
<td>NoCU-MYJ</td>
<td>No cumulus scheme</td>
<td></td>
</tr>
<tr>
<td>NoCU-YSU</td>
<td></td>
<td></td>
</tr>
</tbody>
</table>

The integration time of all simulations was 92 days, beginning on October 1, 2009, with output every 6 hours. The initial and boundary conditions were from ERAI. The lateral boundary conditions were updated every 6 hours. Over the ocean, time independent SST from the same reanalysis was used as the lower boundary condition (Figure 4.2). When the SSTs are time dependent in a model without coupling to an ocean model, there would be an unrealistic shift in the relative phases between SST, convection center, and its related surface wind. They in reality, interact with each other instead of convection being forced by varying SST. This shift would introduce additional errors in surface fluxes. Therefore we decided not to include a simulation with prescribed SST with time independent fluctuations. All the simulations were initialized and forced at the lateral boundaries by a global reanalysis dataset.
4.3. Water Cycle Calculations

There are different techniques in order to characterize and quantify the Water Cycle in different locations. Bosilovich and Schubert (2002) used passive tracers on an atmospheric global climate model in order to quantify regional sources in the water cycle, as it was proposed by Koster et al. (1986) and Joussaume et al. (1986). For this study we approach the problem by explicitly calculating the water budget terms, in the regions of interest: IO and MC. The Indian Ocean is a particular area of interest since numerous large-scale precipitation phenomena occur on it: Madden-Julian oscillation, Kelvin waves, Intra-seasonal oscillations, and the continuous ITCZ passage to the north and south, among others. We diagnosed the water cycle in terms of its bulk (time and domain mean) properties, vertical profiles, and time evolution of its major components through the simulations.

The Equation for atmospheric water in the model is

\[
\frac{\partial q}{\partial t} = -\nabla \cdot q \mathbf{V} - \frac{\partial (q \omega)}{\partial p} + \left( \frac{\partial q}{\partial t} \right)_{\text{PHYSICS}}
\]

(4.1)

where \( q = q(x,y,p,t) \) is mixing ratio (g kg\(^{-1}\)) of water in all phases and hydrometeor types: vapor \((qv)\), cloud water \((qc)\), rain water \((qr)\), ice \((qi)\), snow \((qs)\), and graupel \((qg)\):

\[
q = qv + qc + qr + qs + qg
\]

(4.1a)

The first term on the right-hand side of Equation 4.1 is \( q \) flux convergence by horizontal winds \((u,v)\), and the second term the \( q \) flux convergence by vertical wind \((\omega)\). The third term represents \( q \) tendencies due to subgrid, parameterized physics: cumulus (or deep) convection (CU), shallow convection (SH), microphysics (MP), and boundary-layer processes (PBL).
In order to calculate the terms in Equation 4.1 we assumed a simulation of R number of days on a domain of horizontal resolutions dx and dy with N and M number of grid cells and ktop number of vertical levels. WRF has a terrain-following vertical coordinate denoted by \( \eta \) and defined as \( \eta = \frac{(p_{hs}-p_{ht})}{(p_{hs}-p_{ht})} \), where \( p_{hs} \), \( p_{ht} \), and \( p_{hs} \) are the hydrostatic, top, and surface pressure respectively. The denominator \( p_{hs}-p_{ht} \) is \( \mu \), which accounts for the column mass (Skamarock, 2008). Therefore the change in pressure \( \Delta p \) is defined as

\[
\Delta p = \mu \Delta \eta \quad (4.2a)
\]

For a variable \( \alpha = \alpha(x,y,p,t) \) the time mean is represented as: \( \bar{\alpha} = \bar{\alpha}(x,y,p,t) \), horizontal (domain) mean as: \( \langle \alpha \rangle = \langle \alpha \rangle(p,t) \), and vertical integration as: \( [\alpha] = [\alpha](x,y,t) \). The vertical integration is weighted by the mass of each vertical layer as:

\[
[\alpha](x,y,t) = -\frac{1}{g} \int_{p=\text{surface}}^{p=\text{top}} \alpha dp, \quad \text{where } dp \text{ is the thickness of the vertical layer. Using Equation 4.2a, the vertical integration in model levels is:}
\]

\[
[\alpha] = -\frac{1}{g} \int_{\eta=1}^{\eta=0} \alpha \mu d\eta \quad (4.2b)
\]

The vertical layer thickness in the numerical model is discrete, therefore \( d\eta = \Delta \eta \). This applies to the horizontal plane as well: \( dx = \Delta x \) and \( dy = \Delta y \). In the horizontal plane the grid sizes are characterized by a non-dimensional factor (mapfactor). These factors change according to the projection used. In WRF, \( \Delta x \) and \( \Delta y \) are kept constant in the computational space, and the map factors \( (m_x, m_y) \) are defined as the ratio between \( (\Delta x, \Delta y) \) and the real distance on the earth (Skamarock, 2008). In order to simplify the Equations in this paper, we refer as \( \Delta x \) and \( \Delta y \) as the value already multiplied by this mapfactor.
Equation 4.1 can be averaged over a specific domain of an area A. Then, the first term on the right-hand side of Equation 4.1 can be calculated as a closed line integral over its boundary S,

\[
\frac{1}{A} \int \nabla q \cdot \vec{V} dA = \frac{1}{A} \oint q \vec{V} \cdot \hat{n} dS \quad (4.3)
\]

where the unit vector \( \hat{n} \) is perpendicular to the boundary \( S \). The numerical accuracy of the line integration is much higher than that of the areal integration. In this study, we used the line integration of in Equation 4.3 to estimate the net q flux convergence \( (C) \). We define \( C \) positive if there is net q import into the domain. The line integration in Equation 4.3 consists of 4 terms, each being a line integral along one of the domain’s boundaries: east (\( F_E \)), west (\( F_W \)), south (\( F_S \)), and north (\( F_N \)), respectively:

\[
C(p, t) = \frac{1}{A} \oint q \vec{V} \cdot \hat{n} dS = \frac{1}{A} \left( \int_S q_S v_S dx + \int_W q_W u_W dy - \int_N q_N v_N dx - \int_E q_E u_E dy \right) \quad (4.4)
\]

By averaging over the domain of interest, Equation 4.1 then becomes:

\[
\frac{\partial (q)}{\partial t} = \langle F_S \rangle + \langle F_W \rangle - \langle F_N \rangle - \langle F_E \rangle - \frac{\partial (q \omega)}{\partial p} + \left( \frac{\partial q}{\partial t} \right)_{CU+SH+MP+PBL} \quad (4.5)
\]

where the first four terms on the right hand side correspond to the domain mean of those in the right hand side of Equation 4.4. These domain means (\( F_E \), \( F_W \), \( F_N \), and \( F_S \) ) are calculated from the following Equations:

\[
\langle F_E \rangle_{kt} = -\frac{1}{g A} \sum_{j=1}^{M} (q_{Njkt} u_{Njkt} \mu_{Njkt} \Delta \eta_k) \Delta y \quad (4.5a)
\]

\[
\langle F_W \rangle_{kt} = -\frac{1}{g A} \sum_{j=1}^{M} (q_{1jkt} u_{1jkt} \mu_{1jkt} \Delta \eta_k) \Delta y \quad (4.5b)
\]

\[
\langle F_N \rangle_{kt} = -\frac{1}{g A} \sum_{i=1}^{N} (q_{iMkt} u_{iMkt} \mu_{iMkt} \Delta \eta_k) \Delta x \quad (4.5c)
\]

\[
\langle F_S \rangle_{kt} = -\frac{1}{g A} \sum_{i=1}^{N} (q_{i1kt} u_{i1kt} \mu_{i1kt} \Delta \eta_k) \Delta x \quad (4.5d)
\]
The sub-indexes $ijkt$ represent longitude, latitude, pressure level, and time respectively. The total area of the domain is \( A = NM\Delta x\Delta y \). It should be pointed out that the Equations 4.5 a to b were multiplied by the factor \(-\frac{1}{g}\mu\Delta\eta \) (Equations 4.2a and 4.2b) in order to obtain the transport in units of mm.

The vertical q flux convergence (second term on the RHS of Equation 4.1) was calculated using the following Equation:

\[
(F_V)_{ijkt} = -\frac{\Delta(\omega q)_{ijkt}}{\Delta p_{ijkt}} \quad (4.5e)
\]

Where \( \omega \equiv \frac{dp}{dt} \) is the vertical component of the wind in units of Pas\(^{-1}\) and \( \Delta(\omega q)_{ijkt} = (\omega_{ijkt}q_{ijkt}) - (\omega_{ij(k-1)t}q_{ij(k-1)t}) \). Using Equation 4.2a the denominator of 4.5e can be written as \( \Delta\mu_{ijkt}\Delta\eta_k \) and the units of 4.5e are now KgKg\(^{-1}\)s\(^{-1}\). The domain mean of the vertical q flux \( \left( \frac{\partial(\omega q)}{\partial p} = (F_V)_{kt} \right) \) is calculated by follow Equation 4.5f:

\[
\langle F_V \rangle_{kt} = -\frac{1}{NM} \sum_{i=1}^{N} \sum_{j=1}^{M} \left( \frac{\Delta(\omega q)_{ijkt}}{\Delta\mu_{ijkt}\Delta\eta_k} \right) \quad (4.5f)
\]

The physic tendencies \( \left( \frac{\partial q}{\partial t} \right)_{PHYSICS} \) (third tem in the RHS of Equation 4.1) are obtained directly form the numerical simulations accumulate values every 6 hours. Their domain means does not need any special calculations.

Equation 4.5 can be further integrated vertically for the entire atmospheric column,

\[
\frac{\partial \langle PW \rangle}{\partial t} = [\langle C \rangle] + \langle E \rangle - \langle P \rangle \quad (4.6)
\]

where \( PW = [q] \) is precipitable water, \( C \) the water flux convergence across domain boundaries, which is dominated by water vapor flux convergence (Equation 4.4), \( E \) surface evaporation, and \( P \) surface precipitation. In this study, the units of PW is mm and those of \( E, P, \) and \( C \) are mm day\(^{-1}\). \( P \) includes precipitation produced by deep and
shallow cumulus schemes and microphysics without parameterized convection at each grid point. Which is sometimes referred to as large-scale or explicit rainfall. Convective schemes are sources for cloud water and cloud ice on the resolved scale. Then the microphysics acts on these produced clouds. All but one cumulus parameterization scheme used in this study include shallow cumulus by itself (PB). The ZM scheme was coupled with the PB shallow cumulus scheme in two simulations (ZMPB-MYJ and ZMPB-YSU). Lastly, <E> and <P> are direct output from the model, and would correspond to the vertical integration of the physics tendencies.

The first term in the RHS of Equation 4.6 ([<C>]) was calculated by the vertical integration of each one of its components:

\[
\langle F_E \rangle_t = -\frac{1}{\rho A} \sum_{k=1}^{k_{top}} \sum_{j=1}^{M} (q_{N_{jkt}} u_{N_{jkt}} \mu_{N_{jkt}} \Delta \eta_k) \Delta y
\]  

(4.6a)

\[
\langle F_W \rangle_t = -\frac{1}{\rho A} \sum_{k=1}^{k_{top}} \sum_{j=1}^{M} (q_{1_{jkt}} u_{1_{jkt}} \mu_{1_{jkt}} \Delta \eta_k) \Delta y
\]  

(4.6b)

\[
\langle F_N \rangle_t = -\frac{1}{\rho A} \sum_{k=1}^{k_{top}} \sum_{l=1}^{N} (q_{l_{Mkt}} u_{l_{Mkt}} \mu_{l_{Mkt}} \Delta \eta_k) \Delta x
\]  

(4.6c)

\[
\langle F_S \rangle = -\frac{1}{\rho A} \sum_{k=1}^{k_{top}} \sum_{l=1}^{N} (q_{1_{1kt}} u_{1_{1kt}} \mu_{1_{1kt}} \Delta \eta_k) \Delta x
\]  

(4.6d)

The calculation of the time mean of Equation 4.5 requires to calculate the time mean of E, P, and C, as follows:

\[
\overline{< E >} = \frac{1}{R} \sum_{t=1}^{R} < E >_t
\]  

(4.7a)

\[
\overline{< P >} = \frac{1}{R} \sum_{t=1}^{R} < P >_t
\]  

(4.7b)

\[
\overline{< C >} = \left[ \langle F_W \rangle \right] - \left[ \langle F_E \rangle \right] + \left[ \langle F_S \rangle \right] - \left[ \langle F_N \rangle \right]
\]  

(4.7c)

where the four terms in the RHS of Equation 4.7c were calculated by following Equations 4.8a to 4.8d.
\[
\langle F_E \rangle = - \frac{1}{g_{AR}} \sum_{t=1}^{R} \sum_{k=1}^{k_{top}} \sum_{j=1}^{M} (q_{Njkt} u_{Njkt} \mu_{Njkt} \Delta \eta_k) \Delta y \\
\langle F_W \rangle = - \frac{1}{g_{AR}} \sum_{t=1}^{R} \sum_{k=1}^{k_{top}} \sum_{j=1}^{M} (q_{1jkt} u_{1jkt} \mu_{1jkt} \Delta \eta_k) \Delta y \\
\langle F_N \rangle = - \frac{1}{g_{AR}} \sum_{t=1}^{R} \sum_{k=1}^{k_{top}} \sum_{i=1}^{N} (q_{iMkt} u_{iMkt} \mu_{iMkt} \Delta \eta_k) \Delta x \\
\langle F_S \rangle = - \frac{1}{g_{AR}} \sum_{t=1}^{R} \sum_{k=1}^{k_{top}} \sum_{i=1}^{N} (q_{i1kt} u_{i1kt} \mu_{i1kt} \Delta \eta_k) \Delta x
\]

(4.8a)

(4.8b)

(4.8c)

(4.8d)

It should be noted that before the integration of the moisture flux convergence, we can study the vertical profile of the total mean horizontal water flux convergence in mm day\(^{-1}\) or Kgm\(^{-2}\)day\(^{-1}\). This is defined as:

\[
\langle C \rangle_k = \langle F_W \rangle_k - \langle F_E \rangle_k + \langle F_S \rangle_k - \langle F_N \rangle_k
\]

(4.9)

The moisture flux convergence is positive when there is moisture coming into the domain. The addition of the values of \( C \) in \( p \) will correspond to the vertical integration over the column, since the values are weighted by pressure.

The zonal \((F_E, F_W)\) and the meridional \((F_N, F_S)\) components of \( C \) may be studied separately in order to identify the sources and sinks of moisture. This would be useful to investigate where in the boundary the moisture is entering or leaving the domain. The time mean of the vertical \( q \) flux \((\langle F_V \rangle_k)\) and the physics tendencies \((\frac{\partial q}{\partial t})_{PHYSICS}\) are calculated following Equations 4.7 and 4.8.

Vertical profiles of diabatic heating and moistening, or temperature and water tendencies, from parameterization schemes were saved from the simulations and are available to different degrees from the reanalysis data (Table 2.1). Diabatic heating comes from parameterization schemes of long- and short-wave radiation \([\langle \partial T/\partial t \rangle_{RL}, \langle \partial T/\partial t \rangle_{RS}]\), deep and shallow cumulus processes \([\langle \partial T/\partial t \rangle_{CU}, \langle \partial T/\partial t \rangle_{SH}]\), microphysics \([\langle \partial T/\partial t \rangle_{MP}]\), and boundary layer processes \([\langle \partial T/\partial t \rangle_{PBL}]\).
The moisture and heating tendencies were obtained directly as model output from WRF. Both tendencies were accumulated every six hours during the complete simulations. The change in q due to the physics, will depend on the role of the cumulus, microphysics, and planetary boundary layer scheme (see Equation 2.3), the addition of these three terms is what is usually call as moisture sink\(^{13}\) or Q2,

\[
Q^2 = \left(\frac{\partial q}{\partial t}\right)_{CU} + \left(\frac{\partial q}{\partial t}\right)_{MP} + \left(\frac{\partial q}{\partial t}\right)_{PBL}
\]

(4.10)

Similarly, we can call Q1 to the addition of the temperature tendencies due to the physics processes parameterized in the cumulus, microphysics, planetary boundary layer and radiation schemes. This quantity is usually referred as the heating source\(^ {14}\).

\[
Q^1 = \left(\frac{\partial T}{\partial t}\right)_{CU} + \left(\frac{\partial T}{\partial t}\right)_{MP} + \left(\frac{\partial T}{\partial t}\right)_{PBL} + \left(\frac{\partial T}{\partial t}\right)_{RA}
\]

(4.11)

4.4. Column Integrated Properties of the Water Cycle

We first diagnose the vertically integrated water cycle based on Equation 4.6 for both IO and MC domains. One of the most striking results of the simulations is their large spread in domain mean precipitable water (\(<q>\) or \(<PW>\)) and its large biases in comparison to YOTC in both IO and MC through the 92-day integration period (Figures 4.3a and b). In some simulations, there is quick drift toward dry biases in the MC during the first 20 – 30 days of integration. In the same set of simulations, this drift is subtler in the IO but their deviations from YOTC are as obvious as in the MC. In these simulations with large dry biases (BM-MYJ, BMJ-MYJ, ZM-MYJ, ZMPB-MYJ), \(<q>\) is lower than

\(^{13}\) Moisture sink or Q2 is the terminology usually used after Yanai et al. (1973) work.

\(^{14}\) Note that Q1 in Yanai et al. (1973) does not include the radiation term (see Equations 10 and 11 from Yanai et al. (1973)).
that in YOTC by up to 40%. Dry biases, however, barely exists in a few other simulations (NSAS-YSU, SAS-YSU, TK-YSU).

**Figure 4.3.** Time series of domain mean (a and b) precipitable water $<[q]>$, (c and d) precipitation $<P>$, (e and f) surface evaporation $<E>$, and (g and h) precipitable water
flux convergence <\[C\]\> in the IO (left column) and MC (right column). Day 1 is October 01, 2009.

Most, if not all, simulations that suffer from large dry biases in precipitable water (BMJ-MYJ, BM-MYJ, ZM-MYJ, ZMPB-MYJ, NSAS-MYJ) also produced insufficient domain mean precipitation (<\(P\)>) compared to TRMM and YOTC, while simulations that do not suffer much from dry biases in precipitable water (SAS-YSU, TK-YSU) produced spuriously strong <\(P\)> especially in the IO (Figures 4.3c and d). <\(P\)>\textsuperscript{15} in YOTC and TRMM are very similar to each other, whereas CFSR and MERRA have smaller values, consistent with Trenberth (2001). Spread in precipitation among the simulations appears to be much larger in the IO than MC. This difference comes from large fluctuations in precipitation in the IO in two simulations (TK-YSU, NSAS-YSU). With this exception in mind, we hereafter will use “dry” to describe negative biases in the simulations in both precipitable water (against YOTC) and precipitation (against TRMM), and “moist” or “wet” to describe their positive biases.

Figure 4.3 also shows the time series of domain-averaged surface evaporation (<\(E\)>) and vertically integrated net water flux convergence across the domain boundaries (<\[C\]\>). As for precipitable water and precipitation, fluctuations in <\(E\)> and <\[C\]\> are much larger in the IO than in the MC in some simulations (e.g., SAS-YSU, TK-YSU). All of these unrealistic large fluctuations, including those in <\(P\)> and <\(PW\)>), are related to spurious tropical cyclones in the IO, which are mostly absent in the MC.

During the integration period MJO-09 (see Figure a) initiated in the IO (large precipitation on days 29-48, Figure 4.3c) and propagated eastward through the MC (days

\textsuperscript{15} Appendix E shows the time – longitude diagrams of precipitation for TRMM, YOTC, MERRA and CFSR.
This MJO event induced large fluctuations in precipitable water, water flux convergence (YOTC) and precipitation (TRMM) in both IO and MC (Figures 4.3a, b, c, d, g, and h), as well as in evaporation (YOTC, CFSR, and MERRA) but only in the IO (Figure 4.3e). This suggests that, for this particular MJO event, its enhancement in precipitation was potentially contributed by both evaporation and water flux convergence $<[C]> in the IO (Figures 4.3e and g), but only $<[C]> in the MC (Figures 4.3f and h). This is because evaporation is more constrained over land than over the ocean, and moisture supporting the convective pulse in the MC could only come from the large-scale supply (positive $<[C]>$). Note that mean evaporation in both regions is similar, and the difference is in the response to short-term pulses (Figures 4.3e and f).

None of the simulations reproduced this MJO event (Figures 4.3c and d). The reason for this significant model failure cannot be fully understood here without any successful simulation. We will focus our discussion only on the large-scale biases and errors in the water cycle among all simulations over the 92-day period.

Most simulations with dry biases in precipitable water $<[q]> and precipitation $<P>$ produced insufficient surface evaporation $<E>$ and large-scale water flux convergence $<[C]>$.

Interestingly, time mean precipitation ($\bar{P}$) varies almost linearly with $<\bar{E}>$ and (Figures 4.4c and d) $<[\bar{C}]>$ (Figures 4.4e and f). The burden of supplying moisture for rainfall appears to fall upon evaporation (Figures 4.4c and d) more than flux convergence (Figures 4.4e and f) in some simulations (TK-MYJ, NSAS-MYJ, BMJ-MYJ, KF-MYJ, SAS-MYJ) while in the analyses it is evenly shared between the two. Regardless of their similarities, there are important differences among the (re)analyses worthy of mention, even though smaller than those between simulations. For example,
\( \langle P \rangle_{\text{MERRA}} < \langle P \rangle_{\text{CFSR}} < \langle P \rangle_{\text{YOTC}} \) in the MC. Similar results were documented by Lorenz and Kunstmann (2012) when they calculated the global long-term mean precipitation over land. Nonetheless the differences in time scales (long-term vs intraseasonal) and locations (global vs. MC), it is important to notice that long-term trends over land are also present in shorter time scales. Lorenz and Kunstmann (2012) also found the same result over the oceans \((\langle P \rangle_{\text{MERRA}} < \langle P \rangle_{\text{CFSR}})\). In our study, however, \(\langle P \rangle_{\text{MERRA}} > \langle P \rangle_{\text{CFSR}}\) in the IO. This indicates regional variations in the water budget discrepancies among the reanalysis products.

**Figure 4.4.** Scatter diagrams of the time and domain mean water cycle components in the IO (left column) and MC (right column) domains for (a and b) precipitation \(\langle P \rangle\) vs. precipitable water \(\langle [\bar{q}] \rangle\), (c and d) \(\langle P \rangle\) vs. evaporation \(\langle E \rangle\), (e and f) \(\langle P \rangle\) vs. precipitable water flux convergence \(\langle [\bar{C}] \rangle\), and (g and h) \(\langle E \rangle\) vs \(\langle [\bar{C}] \rangle\). Red lines are
<\tilde{P}> from TRMM. Dotted gray lines are best fits to the scatters based on a least square method.

In simulations with large dry biases, \( < [\tilde{C}] > \) can be almost zero (Figures 4.4g and 3h), but not evaporation (BMJ-MYJ, BM-MYJ, ZM-MYJ, KF-MYJ, Figure 4.4c and 4.4d). Possible differential roles of evaporation and flux convergence in the dry biases is better illustrated in Figure 4.5, which shows scatter diagrams of \( < [\tilde{C}] > / < \tilde{P} > \) vs. \( < \tilde{E} > / < \tilde{P} > \) in the IO (Figure 4.5) and MC (Figure 4.5b), respectively. A point on the diagonal line means the total water (q) tendency is zero and \( < \tilde{P} > = < \tilde{E} > + < [\tilde{C}] > \).

**Figure 4.5.** Scatter diagrams of ratios between time and domain mean precipitable water flux and precipitation \((< [\tilde{C}] > / < \tilde{P} >)\) vs. ratios between mean surface evaporation and precipitation \((< \tilde{E} > / < \tilde{P} >)\) for the (a) IO and (b) MC domains.

A point above (below) it means positive (negative) total water tendency, or domain mean moistening (drying). In YOTC, both \( < [\tilde{C}] > / < \tilde{P} > \) and \( < \tilde{E} > / < \tilde{P} > \) are close to 0.5, with a very slight drying tendency. This implies that in YOTC, flux convergence and evaporation contribute equally to precipitation in both IO and MC.
Similar results are observed in MERRA. CFSR on the other hand has slightly more contribution from $< \bar{E} >$ (~0.57) than from $< \bar{C} >$ (~0.4). In the MC, the reanalysis are in better agreement (Figure 4.5b). We suggest that this could be due to the difficulties in initiating precipitation over the ocean in comparison with over land.

In contrast, in most simulations, $< \bar{E} >/< \bar{P} >$ is greater than $< \bar{C} >/< \bar{P} >$, indicating their precipitation depends more upon surface evaporation than flux convergence. In the simulations that suffer from large dry biases, most moisture for precipitation comes from $< \bar{E} >$ within the domains because $< \bar{C} >$ is near zero (Figure 4.5). This might suggest that the lack of precipitation and moisture is caused by the large-scale flow. However, this flaw of the model would not necessarily lead to dry biases. Had excessive evaporation been produced by the model to compensate the deficit in flux convergence, there would not necessarily be a dry bias. The dry biases, and the wide spread of water cycles among the simulations, must be caused by a combination of the three problems: the flawed precipitation total, insufficient surface evaporation, and insufficient large-scale water flux convergence. The three are related to each other: surface evaporation and large-scale flux convergence both depend on the strength of the circulation, which in turn is closely related to the latent heating released by precipitation.

As mentioned above, the linear variation of precipitation with surface evaporation and flux convergence seen in Figure 4.5 indicates a fundamental relationship between the three variables. When the physics are changed to a combination that provides more precipitation, the relative increases in evaporation and flux convergence to supply the extra precipitation are almost given regardless of the exact changes in the physics schemes. This has been observed when simulation resolution increases (Hagemann et al.,
2006) and when model land-ocean configuration is modified (Neale and Slingo, 2003). However, such changes do not necessarily alleviate the dry biases.

The linearity between precipitation, evaporation, and flux convergence also holds for different averaging domain sizes, including the whole domain IOMC (not shown). An interpretation is that an increase in precipitation corresponds to more diabatic heating; area-averaged vertical mass flux and large-scale horizontal convergence increase in proportion, while surface evaporation makes up the balance by increasing in proportion to both. The implication is that the large-scale convergence in this region is completely determined by the physics leading to convective activity, and the strength of the water cycle is also determined by the physics as a consequence. This control of the large-scale mass, flow, and water budget by physical processes associated with convection is probably a feature of large areas of the tropics.

**4.5. Main Sensitivities to Parameterizations**

The simulated water cycle is more sensitive to PBL schemes than to cumulus parameterization. Figure 4.4 clearly shows that most simulations with large dry biases used the MYJ PBL scheme (triangles), while the YSU scheme was used in most simulations without large dry biases (circles). Simulations with the UW PBL scheme lay in between the two extremes. Three groups of simulations (three clusters in Figure 4.4) can be identified.

Group 1: These simulations suffer from large dry biases in mean precipitable water $\langle \bar{q} \rangle$ and precipitation $\langle \bar{P} \rangle$, insufficient surface evaporation $\langle \bar{E} \rangle$ and very small
flux convergence $\langle [\tilde{C}] \rangle$ in both IO and MC. Most of this group used the MYJ PBL scheme (triangles).

Group 2: They include two simulations that produced relatively small biases in $\langle [\tilde{q}] \rangle$, $\langle \tilde{P} \rangle$, $\langle \tilde{E} \rangle$, and $\langle [\tilde{C}] \rangle$ in both the IO and MC. They are two versions of the NSAS cumulus scheme paired to the YSU PBL scheme (NSAS-YSU, NSASnoSH-YSU). Performance of the simulations in groups 1 and 2 is consistently poor (group 1) or good (group 2) in both IO and MC.

Group 3: They feature large discrepancies between the IO and MC: their precipitation is much lower in the MC than IO, with either wet biases in the IO but not in the MC (TK-YSU, SAS-YSU) or large dry biases in the MC but not in the IO (TK-MYJ, SAS-UW, NoCU-YSU).

Four simulations (NSAS-MYJ, NSAS-YSU, TK-MYJ, TK-YSU) will be examined in more detail. They represent simulations with and without large dry biases and large fluctuations in the water cycle. They include simulations using the same PBL scheme paired with different cumulus schemes and the same cumulus scheme paired with different PBL schemes. They are from each of the three identified groups. NSAS-MYJ (dark blue triangle) belongs to group 1 with consistent large biases in the IO and MC. NSAS-YSU (dark blue circle) is from group 2 with consistent small biases in IO and MC. TK-MYJ (light green triangle) and TK-YSU (light green circle) are from group 3 with large discrepancies or inconsistencies in their water cycles between the IO and MC.

Sensitivities of precipitation distributions to both cumulus and PBL schemes are evident (Figures 4.3c and d). Simulations using the TK scheme in general tend to produce far more precipitation than those using NSAS, with differences more noticeable.
in the IO than in the MC (Figure 4.4). The TK scheme also shows much larger rainfall related to topography (Figure 4.2) than the NSAS scheme. The combination of any of the cited schemes with YSU produces an increase in precipitation (this is also true for the SAS scheme and for the simulations without cumulus scheme NoCU) in both the entire IOMC domain and two local regions.

The differences between the schemes are also noted in the contributions to the precipitation by evaporation ($\langle \bar{E} \rangle / \langle \bar{P} \rangle$) and flux convergence ($\langle [\bar{C}] \rangle / \langle \bar{P} \rangle$). In the IO the fraction $\langle \bar{E} \rangle / \langle \bar{P} \rangle$ ranges between 0.41 (TK-YSU) and 0.80 (NSAS-MYJ), TK-MYJ and NSAS-YSU lay in between with values of 0.55 and 0.59 respectively (Figure 4.5). The values of $\langle \bar{E} \rangle / \langle \bar{P} \rangle$ and $\langle [\bar{C}] \rangle / \langle \bar{P} \rangle$ in the tropics produced by a cumulus ensemble model are 0.24 and 0.76, respectively (Sui, 1994), which are within the range found here.

When the PBL scheme MYJ is switched to YSU, the flux convergence fraction increases between 14% (TK) and 21% (NSAS) in the IO, and 11% (TK) and 24% (NSAS) in the MC. On the other hand, when the cumulus scheme is switch from NSAS to TK using the same PBL scheme, the flux convergence in the IO (MC) increases between 14% (7 %) when YSU is used and 20 % (20%) when MYJ is used.

Spatial distributions of time (92 days) mean precipitation from TRMM and the four selected simulations are shown in the left column of Figure 4.6. In TRMM (Figure 4.6a), an ITCZ south of the equator in the IO is visible. There is a hint of a double ITCZ over the western IO. Precipitation is widely spread over most of the IO. There is a local precipitation minimum in the center of the MC, between Borneo and New Guinea. In the four selected simulations, both strong and weak precipitation in TRMM tends to be
exaggerated (Figures 4.6c, e, g, and i). In other words, precipitation is more concentrated in the simulations than in TRMM. The southern ITCZ is the dominant feature in the IO because of the dry biases to both its south and north. The contrast of wet land and dry seas in the MC, especially south of the equator (except the coastal region western Sumatra) is stunning.

Figure 4.6. Mean precipitation (left column), rain frequency (right column) from TRMM (b), and rain frequency difference (simulation – TRMM) for the four selected simulations (d, f, h, and j). Rain frequency was calculated for rain rates greater than 0.5 mm day$^{-1}$. 
These wet and dry biases are consequences of both unrealistic spatial distributions of rainfall and rain intensity probability produced by the parameterization schemes. Compared to TRMM, all four simulations produced insufficient rain except very light and heavy rain (Figure 4.7). This indicates that wet biases exist because of infrequent (< 1%), very heavy rain (> 40 mm day\(^{-1}\)) produced in certain simulations (e.g., TK-YSU).

**Figure 4.7.** Probability distribution of rain rates in the IO (left) and MC (right) for TRMM and the four selected simulations. The rain bin width is 1 mm day\(^{-1}\). The percentage of rain data included in the range of 0 - 40 mm day\(^{-1}\) is listed in the legends.

Spatial distributions of total rain frequencies are very similar to those of mean rain rates (Figures 4.6a and b). But their biases in the four simulations show distinct spatial patterns (right column of Figure 4.6). It rains too often in both the southern ITCZ of the IO where heavy rain is, and in the western IO and near the lateral boundaries of the domain (except for TK-MYJ) where rain is very light. The dry biases can be produced by reduced rain frequency at all rain rates, for example, at the flanks of the ITCZ where erroneously strong ITCZ rain induces erroneously strong subsidence. They can also be produced by too little infrequent heavy rain while light rain is produced too frequently,
for example, further away from the ITCZ. Apparently, rain frequencies are mainly
determined by cumulus parameterization. Their spatial distributions vary a little with
PBL schemes when TK was used (Figures 4.6d and f), and are only slightly sensitive to
the PBL schemes when NSAS was used (Figures 4.6h and j).

4.6. Vertical Structure of the Water Cycle

The evolution of vertical profiles of atmospheric water is presented as domain
mean perturbations in water vapor mixing ratio ($q_v$). The perturbations ($<dq_v>$) are
deviations from a reference profile. The reference profile was chosen to be a profile of
YOTC $q_v$ averaged over the IOMC domain and the 92-day integration period. The
resulting time series of $<dq_v>$ profiles over the IO and MC are shown in Figure 4.8. In
YOTC, $<dq_v>$ are deviations of the respective IO and MC mean of $q_v$ from its IOMC
mean. Positive and negative $<dq_v>$ from the surface to about 400 hPa are related to the
passage of the MJO event (Figures 4.8a and b). They were not reproduced by any of the
simulations (Figures 4.8c to j).

Time series of $<dq_v>$ from the four selected simulations in Figure 4.8
demonstrate that their contrasts developed almost instantaneously after the initial time in
the IO but took about 15 days in the MC. Both MYJ and YSU produce a lower and drier
PBL in the MC than IO. It is apparent that the boundary layer tends to be drier in
simulations using the MYJ PBL scheme than in those using YSU, and the difference is
particularly noticeable with the NSAS cumulus scheme (Figures 4.8g-j).
**Figure 4.** Time series of perturbations in water vapor mixing ratio ($<dq_v>$) averaged over the IO (left column) and MC (right column) from (a and b) YOTC and the
simulations: (c and d) TK-MYJ, (e and f) TK-YSU, (g and h) NSAS-MYJ, and (i and j) NSAS-YSU. At bottom of each panel are domain-mean precipitation ($<P>$) (black line) and planetary boundary-layer height ($<h>$) (red line), with the straight green line being time mean of YOTC $<\bar{h}>$. Units are g kg$^{-1}$ for $<\bar{q_v}>$, mmday$^{-1}$ for $<P>$, and m for $<h>$. See text for details of the calculation of the perturbations.

The NSAS-YSU simulation tends to create a persistent thin moist layer next to the surface in the IO (Figure 4.8i), and the NSAS-MYJ run has a deeper dry layer in the MC (Figure 4.8h). The dry biases tend to extend upward from the boundary layer more in the MC than IO in all simulations.

The YOTC time series of domain-mean boundary-layer height ($<h>$) shows an increase after the rainfall peak of the MJO in the IO when the lowest troposphere become anomalously dry (Figure 4.8a, bottom panel). This variability is not observed in the MC (Figure 4.8b, bottom panel) or in any simulation (Figures 4.8c-j). The simulations using the MYJ PBL scheme produced $<h>$ higher than that from YOTC in both IO and MC regardless of the cumulus scheme used, which corresponds to a drier PBL. For a given cumulus scheme paired with MYJ, $<h>$ is higher in the IO than in the MC but their PBL is dryer in the MC than in the IO. So there is no consistent relationship between h and PBL moisture biases. The variability in $<h>$ is small in all simulations except at the beginning of the simulations using MYJ (Figures 4.8c and g).

Time mean profiles of $<\bar{d}q_v> (\overline{<dq_v>})$ from the four simulations, YOTC, and the two other reanalysis products are shown Figures 4.9a and b. In both IO and MC, $\overline{<dq_v>}$ in YOTC are positive in most of the troposphere because water vapor is generally higher near the equator, where the IO and MC domains are, than away from the equator
toward the southern and northern boundaries of the IOMC domain, in which the reference profile was calculated.

**Figure 4.9.** Vertical Profiles of time and domain mean (a and b) perturbations of water vapor mixing ratio ($\langle \overline{d q_v} \rangle$), (c and d) horizontal water flux convergence ($\langle \overline{C} \rangle$), and (e and f) vertical water flux convergence ($\langle \overline{\partial (q \omega) / \partial p} \rangle$) in the IO (left column) and MC (right column). Note the different y-axes limits in c and d.
In contrast, there are large negative $\overline{dq_v}$ in the lower troposphere and/or in the PBL in all simulations except TK-YSU. These negative $\overline{dq_v}$ in the simulations, reflected by the dry biases seen in Figure 4.8, are in contrast with the positive $\overline{dq_v}$ in YOTC. Discrepancies among the three reanalysis datasets are also apparent, but the dry and moist biases in the simulations would be evident regardless of which reanalysis dataset they are compared to. The examples shown in Figures 4.9a and b represent the general results from most simulations. Their $\overline{dq_v}$ are dry (negative) in both IO and MC with negative peaks near 900 hPa. Exceptions include moist (positive) biases in the boundary layer and mid-upper troposphere (700 – 300 hPa) only in some simulations using the YSU PBL scheme. It is noticed that the profile of $\overline{dq_v}$ from MERRA is in a similar shape as those from simulations using NSAS cumulus scheme. This is not a coincidence. Cumulus parameterization in the model that produced MERRA is based on a simplified Arakawa-Schubert scheme (Moorthi and Suarez, 1992). This illustrates that biases produced by cumulus schemes cannot be removed even by observational constraints during the data assimilation procedure.

To help understand the reason for the vertical structures of the biases in water vapor mixing ratio, profiles of time and domain mean horizontal water flux convergence $\overline{C}$ are shown in Figures 4.9c and d, with net imports (convergence) into a domain at a given level defined as positive. Similarly, profiles of time and domain mean vertical water flux convergence $\overline{\left(\frac{\partial (q_\omega)}{\partial p}\right)}$ are shown in Figures 4.9e and f. In YOTC, net horizontal water imports occur from the surface to 500 hPa in both IO and MC, with large peaks near the surface. None of the simulations reproduced the near-surface import maxima (Figures 4.9c and d). In the IO, TK-YSU, which produced the largest wet bias
(Figure 4.4) and no dry bias in lower-tropospheric water vapor (Figure 4.8 e), is the only one of the four simulations that suffers from too much horizontal flux convergence at low levels (Figure 4.9c). There is erroneous horizontal flux divergence in the layer of 700-500 hPa in the two simulations using NSAS, and even in CSFR (Figure 4.9c). This implies a spurious shallow overturning circulation, as will be further discussed later.

Vertical water flux convergence is negative within the boundary layer and positive above, consistent to the perception that water is taken out of the boundary layer and transported into the troposphere. There is no model bias there. Model biases come from the amplitudes and level of peaks of vertical flux convergence. Simulation TK-YSU, which suffers the largest wet biases (Figures 4.2 and 4.4), consistently produces too much vertical water transport into the troposphere from the boundary layer (Figure 4.9e), where the spurious water supply comes from erroneous horizontal flux convergence (Figure 4.9c). Its vertical flux convergence peaks are at higher level (350 hPa) than YOTC (450 hPa). Interestingly, NSAS-YSU appears to produce double peaks of vertical flux convergence, one at about 700 hPa, the other at 300 hPa (Figure 4.9f). The vertical flux convergence profiles in the MC (Figure 4.9e) are closer together than in the IO (Figure 4.9f). The latter highlights, once more, the limitations in agreement between (re)analysis and simulations over the Ocean.

In the MC, the low-level horizontal convergence layer is too shallow in all simulations in comparison to that in YOTC (Figure 4.9d). There, the erroneous horizontal flux divergence in the layer of 700-500 hPa occurs in the two simulations using MYJ. Model biases in vertical flux convergence appear to be smaller than in the IO (Figure 4.9e and f). It is intriguing that the two simulations that produced vertical flux convergence
profiles closest to those of YOTC in both the IO and MC do not share any common scheme (TK-MYJ, NSAS-YSU). This underlines the role of combined, not individual, parameterization schemes in detailed features of the water cycle.

4.7. Water and Temperature Tendencies

When examining moistening and heating or water and temperature tendencies due to physical processes (cumulus, PBL, microphysics) in YOTC and the WRF simulations, we must bear in mind that they are all products of parameterization schemes. The difference is that results from YOTC are strongly constrained by observations through data assimilation, while those from the simulations are weakly constrained by ERAI through initial and boundary conditions. For example, diabatic heating profiles differ among the global reanalyses (Ling and Zhang, 2013), most likely because of differences in their parameterization schemes. It is more appropriate to interpret results from YOTC as references instead of reality in their comparisons to simulations.

In YOTC, the total precipitable water tendency due to all parameterized processes (cumulus, microphysics, and turbulence), or Q2, is close to zero in the boundary layer and negative (drying) above the boundary layer in both IO and MC (Figures 4.10g and h). The near zero tendency in the boundary layer is a result of balance between large and opposing effects of moistening by PBL processes (mixing evaporated water vapor up from the surface) and drying by cumulus processes (transporting moisture upward out of the boundary layer by updrafts and transporting drier air into the boundary layer by downdrafts).
FIGURE 4.10. Profiles of time and domain mean precipitable water tendency due to (a and b) cumulus convection $<\text{CU+SH}>$, (c and d) microphysics $<\text{MP}>$, (e and f) planetary boundary layer $<\text{PBL}>$, (g and h) their combination $<\overline{Q_2}>$, and (i and j) contributions from cloud water $<\overline{q_c}>$ (solid lines) and cloud ice $<\overline{q_i}>$ (dotted lines) by the cumulus
schemes (CWI) over the IO (left column) and MC (right). Note the different y-axes and x-axes limits in (e and f) and (i and j), respectively.

In all simulations, cumulus drying in the boundary layer is not strong enough to balance moistening by PBL schemes and give rise to net moistening in the boundary layer (Figures 4.10a, b, e, and f). Cumulus drying also exists above the boundary layer in the troposphere up to 200 hPa, although with smaller amplitudes mostly resulting from compensating subsidence (Figures 4.10a and b). Within or outside the boundary layer, cumulus drying is much weaker in the MC than IO (Figures 4.10a and b). This is consistent with weaker precipitation in the MC than IO, as shown in Figures 4.3c and 4.3d. In some simulations, there are two peaks in tropospheric drying, one near 800 hPa and the other between 600 and 400 hPa. These correspond to two heating peaks related to warm and cold clouds, respectively, which will be discussed below. The NSAS simulations have shallow drying by the microphysics around the PBL top, indicating resolved shallow convection that is absent in TK simulations and YOTC (Figures 4.10c and d).

Tendencies of cloud water and ice (CWI) in the cumulus schemes are shown in Figures 4.10i and j. They are one order of magnitude less than those of the water vapor mixing ratio. However, they are important because they demonstrate effects of warm (liquid water mixing ratio tendencies) and cold (ice water mixing ratio tendencies) clouds. Both the cloud liquid water and ice water tendencies are positive due to the conversion of water vapor (drying) to liquid water and ice during convection. It appears that the TK scheme detrains more low/mid-level cloud as a result of more active shallow convection within the scheme. This is supported by Figures 4.11a and b, where the
microphysics is more active in producing a thin layer of shallow clouds itself in NSAS. Meanwhile, simulations using TK produced warm clouds (solid lines) that are much lower than in the simulations using NSAS in both IO and MC (Figures 4.10i and j).

**Figure 4.11.** Profiles of time and domain mean mixing ratios of (a and b) cloud water, (c and d) ice, snow and graupel; and (e and f) rainwater for the IO (left column) and MC (right column).
Figure 4.12. Profiles of time and domain mean temperature tendency (heating) due to (a and b) cumulus convection ($<\text{CU+SH}>$), (c and d) microphysics ($<\text{MP}>$), (e and f)
boundary-layer processes ($\langle \overline{PBL} \rangle$), (g and h) radiation ($\langle \overline{RA} \rangle$), (i and j) total latent heating ($\langle \overline{CU+SH+MP} \rangle$), and (k and l) the combination of all ($\langle \overline{Q1} \rangle$) over the IO (left column) and MC (right column). Note the different y-axes limits in (e and f).

Diabatic heating profiles are relevant to the water cycle because of their effects on the circulation, and hence surface evaporation and water flux convergence, both in the horizontal and vertical. The sign of simulated flux convergence (Figures 4.9c and 4.9d) must depend on the diabatically driven circulation. Profiles of domain and time mean diabatic heating due to various physical processes in YOTC and the four selected simulations are shown in Figure 4.12. In YOTC, total heating ($Q1$) maintains a similar mean strength in the IO and MC (Figures 4.12k and l), consistent with its precipitation (Figures 4.3 and 4.4). Its peak is at 500 – 400 hPa, suggesting the dominance of deep convection involving ice processes. As for precipitable water and precipitation, total heating $Q1$ is very different among the simulations. The amplitude of $Q1$ in simulations of Group 3 (represented by TK-YSU and TK-MYJ) is considerably smaller in the MC than the IO. $Q1$ profiles from Group 2 simulations (represented by NSAS-YSU) stay close to that of YOTC in both IO and MC. $Q1$ in Group 1 simulations (represented by NSAS-MYJ) is substantially underestimated in the IO and becomes erroneously small in the entire troposphere in the MC because latent heating is just enough to offset radiative cooling there (Figures 4.12k and l). There is large net surface heating in all simulations in the MC, not seen in YOTC (Figures 4.12l), perhaps because PBL heating is not canceled by cumulus and microphysics cooling near the surface (Figures 4.12f and j). This occurs to a much lesser extent in the IO (Figures 4.12e and i).
The mid-tropospheric heating profiles are consistent with their drying profiles, being dependent on how the subsidence or convective mass flux varies between simulations. The microphysics heating (Figures 4.12c and d) and moistening (Figures 4.10c and d) tendency profiles show drying and warming in the mid-upper troposphere in the TK runs, and cooling and moistening in the NSAS runs, indicating that the TK runs have active stratiform clouds producing diabatic heating while NSAS produced stratiform clouds that are more passive with net evaporation. This represents a major distinction between how each of the cumulus schemes produces and interacts with resolved stratiform clouds. By contrast, the YOTC microphysics profile exhibits a stronger cooling below the melting level than the TK runs, but also represent active stratiform clouds above that level. The activity of the mid-level stratiform clouds depends on the cumulus scheme, but whether or not these produce net heating is not alone a determining factor on the overall net heating at those levels, which depends more on the PBL scheme.

The offsetting of the strong PBL and cumulus tendencies in the boundary layer indicate that their interaction is important. The YSU PBL tends to be associated with higher surface evaporation, possibly because it mixes through the boundary layer more efficiently, enabling a drier surface layer. When combined with this PBL scheme, cumulus schemes can respond to this extra moisture flux and become more vigorous as is seen by their greater drying effect near the surface. Overall, these offsets are larger with the YSU PBL, indicating a stronger water cycle controlled by the boundary layer and convection.
4.8. Cloud Distribution and Radiative Impacts

The NSAS runs clearly over-estimate boundary-layer clouds compared to the TK runs and YOTC, which is more severe in the IO than MC (Figures 4.11a and b). The YSU scheme does not allow for liquid water at the top of the PBL, leaving it only to dry mixing, which would not be as effective in removing these clouds. In the case of the MYJ, even though the mixing is parameterized for liquid water and potential temperature, the problem seems to be as large as for YSU, perhaps because of the lack of explicit entrainment of drier air into the PBL. In the case of the TK runs, this explicit shallow convection is perhaps avoided by a more effective shallow scheme within TK. Both are mass-flux based shallow schemes.

There is an apparent underestimation of warm (liquid) cloud above the boundary layer in the lower troposphere in all simulations compared to YOTC in both IO and MC (Figures 4.11a and b); similar results were reported by Dai et al. (2007) for the globe. However, it is not clear to what extent the YOTC analysis is true either. No such consistent biases across all simulations are found for cold (ice) cloud. Cold clouds are excessive in some simulations and insufficient in others (Figures 4.11c and d). Most simulations did not produce the double peaks in cloud ice content seen in YOTC in both IO and MC (Figures 4.11c and d). The peak at 500 hPa is perhaps primarily associated with graupel, while the other at 300 hPa associated with ice and snow. This structure likely depends on the microphysical parameterization used by YOTC, which differs from WSM6 used in our simulations. The larger amount of both warm and cold clouds in the IO than MC is consistent with stronger precipitation in the IO than MC (Figures 4.3, 4.4, and 4.5). Similarly, more clouds are found in the runs using the YSU PBL schemes than
those using MYJ, consistent with the precipitation differences. These precipitation differences are also evident in the vertical distribution of rain water or rain drops (Figures 4.11e and f). TK simulations have greater amounts of mean falling water drops overall. However, NSAS simulations have a double peak of rain water close to 900hPa and 600 hPa, which is more noticeable in the IO (Figure 4.11e) than in the MC (Figure 4.11f). The lower maximum may be link to the resolved-shallow clouds in NSAS. These are also evident from the moistening (Figures 4.11c, d) and heating (Figures 4.12c, and d) profiles of the microphysics scheme.

According to Dai and Trenberth (2004) the differences in the clouds distribution among simulations suggest that the simulations will also have diurnal cycles with different characteristics. The discussions on diurnal cycle variations, however, are not in the scope of this study.

The excessive boundary layer clouds lead to exaggerated cloud-top radiative cooling in the lowest 100 hPa, which is larger than 4 K day$^{-1}$ in the IO and 3 K day$^{-1}$ in the MC in the NSAS simulations (Figures 4.12g and h). It is only up to 1 K day$^{-1}$ in YOTC in both IO and MC, indicating no shallow cloud tops there. Above the boundary layer (900 – 200 hPa), radiative cooling is about 1 – 2 K day$^{-1}$ with small inter-simulation spread, without much vertical structure, and only slightly larger than that in YOTC (~ 1 K day$^{-1}$). There is an indication of more cloud-top cooling around 200 hPa consistent with the microphysical profiles while YOTC clouds appear to extend vertically beyond 100 hPa.
4.9. Chapter Summary

The main results presented in this study can be summarized as:

(i) The strength of the water cycle in the IO and MC, measured by its domain and time-mean precipitable water and precipitation, differs substantially among the 16 simulations with different combinations of cumulus and PBL parameterization schemes, and the large spread of water-cycle strength is mainly toward dry biases in comparison to YOTC.

Simulations with large wet/moist biases (in the IO) produced erroneous spikes in surface evaporation (related to spurious tropical cyclones), too much water vapor import, insufficient mid and lower tropospheric cooling and sometimes erroneous shallow heating by the microphysics (as opposed to cooling in YOTC).

The large differences among the simulations demonstrate the sensitivity of the simulated water cycle to cumulus and PBL parameterization schemes. The MYJ PBL scheme was used in most simulations producing large dry biases in both IO and MC, whereas YSU was used in most simulations producing wet biases in the IO or without large dry biases. Crétat et al. (2011) found similar results over southern Africa.

The biases discussed in this study are not unique to the regional model used. Dry biases over the MC and IO and in the tropics in general have also been found in simulations of global models (Andersson et al., 2005; Hack et al., 2006; Neale and Slingo, 2003; Schiemann et al., 2013). Here, several features distinguish simulations with large dry biases from those without. Simulations with large dry biases suffer from insufficient surface evaporation and large-scale horizontal precipitable water flux convergence or transport, and from erroneous mid and lower tropospheric diabatic
heating. Neale and Slingo (2003) found a similar dry bias over the MC in a GCM, which is due to the misrepresentation of the interaction between land and ocean on diurnal scales.

(ii) Biases produced by a PBL scheme can penetrate into the middle troposphere, and biases produced by cumulus schemes into the boundary layer.

The biases in the boundary layer can be induced by cumulus parameterization schemes as well as PBL schemes. For example, dry biases in the boundary layer exist in simulations using the MYJ PBL scheme regardless of cumulus schemes. When the YSU PBL scheme is used, the boundary-layer dry biases still exist if it is paired to the NSAS cumulus scheme but disappear when the TK scheme is used (Figures 4.8 and 4.9). Tropospheric dry biases are much stronger when the NSAS cumulus scheme is paired to MYJ than to YSU (Figure 4.9).

(iii) The simulated water cycle and its spread and biases differ between the IO and MC.

Biases and errors in the simulated water cycle in the IO and MC are different in several ways. Precipitation, surface evaporation, and water flux convergence are weaker in the MC than IO only slightly in YOTC, but substantially in some simulations. In these simulations large wet biases in the IO disappear in the MC, while dry biases in the IO worsen in the MC. The erroneous drying tendency in simulations with large dry biases is much greater during the first 30 days of integration in the MC than IO. Some simulations overestimate horizontal water import into the IO in boundary layer and lower troposphere, while others underestimate it; some produce erroneous exports in mid troposphere (700 – 400 hPa) in the MC. Moist biases in the boundary layer exist only in
the IO with the YSU scheme (Figure 4.8). Excessive boundary-layer clouds in the NSAS simulations are more abundant in the IO than MC (Figures 4.11a and b). The contrasts in the water cycle between the IO and MC suggest strong influences of the surface conditions on the performance of the parameterization packages.

(iv) Model biases in surface evaporation and water flux convergence are linked to biases in vertical structures of diabatic heating and vertical motion that are central to the large-scale circulations.

**Figure 4.13.** Profiles of time and domain mean (a and b) diabatic heating normalized by precipitation \((\langle Q_f \rangle)/(\langle P \rangle)\) and (c and d) vertical velocities \((\langle \omega \rangle)\) for the IO (left column) and MC (right column).
The erroneous horizontal mid-level moisture export and low-level and boundary-layer import suggest a spurious shallow overturning circulation (Figures 4.9c and d). This spurious shallow overturning circulation can be inferred from profiles of diabatic heating with peaks at an unrealistically low level. Such unrealistic bottom-heavy heating profiles are obvious in simulations using the MYJ PBL schemes (Figures 4.12k and l) but common in many other simulations (Figures 4.13a and b). The bottom heavy heating profiles are mirrored by bottom heavy profiles of the vertical motion that are also shared by many simulations (Figures 4.13c and d). The low-level peak of the vertical motion is part of the shallow overturning circulation.

(v) None of these errors and biases is the reason for the model failure to reproduce the MJO event that occurred in the simulation period. The reason may be in common errors in all or most simulations in comparison to YOTC in both IO and MC regardless of the degree of their dry or moist biases. These common errors in the boundary layers include insufficient net water flux convergence, excessive drying by cumulus schemes dominated by even more excessive moistening by PBL schemes, which leads to their unrealistic low-level moistening in contrast to a weak tendency in YOTC. This suggest that a more realistic representation of local and large scale convection together with improvements in the boundary layer process may help the reproduction of the MJO, as proposed by Neale and Slingo (2003).

Based on these results, we summarize the water cycles in the IO and MC as portrayed by YOTC and their biases in the WRF simulations. In YOTC, precipitation depends roughly equally on surface evaporation and horizontal flux convergence controlled by the circulation in both IO and MC (Figure 4.5). In the MC, surface
evaporation is weaker than in the IO, hence weaker precipitation (Figure 4.4). The corresponding weaker latent heating release (Figs. 4.12i and j) in the troposphere leads to a weaker circulation, which results in a weaker flux convergence. Contributions to the weaker precipitation from surface evaporation and flux convergence remain roughly equal. Amplitudes aside, there is no fundamental difference between the water cycle in the IO and MC. This, however, applies only to the western part of the MC. A complete description of the water cycle for the entire MC must include the Celebes, Banda, and Arafura Seas, Moluccas, and New Guinea.

The WRF simulations with prescribed lateral boundary conditions produce dry biases in the water cycle because their precipitation in the IOMC region relies too much on surface evaporation rather than horizontal flux convergence (Figure 4.6) and their diabatic heating peaks are too low (Figures 4.13a and b). These lead to peaks of upward motions at an unrealistically low level (Figures 4.13c and d). The consequence of these errors is a moisture convergence layer that is too shallow compared to that in YOTC, which results in dry biases in the lower troposphere (Figures 4.9 and 4.10). These dry biases weaken precipitation and reduce its associated surface wind. The corresponding reduction in surface evaporation aggravates the already weakened precipitation. This degrading feedback is responsible for the large dry biases in some simulations (Figure 4.3). For the same reason, dry biases in the simulations are worsened over the MC because of the lack of surface evaporation. Liepert and Previdi (2012) found a similar drying in CMIP3 models. They refer to this drying as atmospheric leaking and propose this may be due to kinetic energy variations due to errors in the different clouds and diffusion parameterizations. On the other hand, cumulus schemes with stronger deeper
diabatic heating profiles maintain their mass flux convergence, which leads to stronger moisture flux convergence. There is a positive feedback where stronger, deeper convection helps itself by generating deep moisture convergence.

The processes that lead to the large spread in the water cycle at the beginning of the simulations can be compared to those when the spread become quasi steady later in the simulations (Figure 4.3). Such a comparison is given in Figure 4.14. This Figure lists domain mean surface evaporation $\langle E \rangle$, precipitation $\langle P \rangle$, and vertical integrated water flux convergence $\langle [\tilde{C}] \rangle$ from all simulations and the three global analysis products in the order of their domain mean precipitable water tendency or vertically integrated water tendency $\langle \partial [\tilde{q}] / \partial t \rangle$ (Equation 4) for the first 20 days and the rest of the integration period in the IO and MC. During the first 20 days (Figures 4.14a and b), whether there is a drying or moistening tendency in a simulation largely relies on flux convergence $\langle [\tilde{C}] \rangle$ while evaporation is similar in different simulations. This is particularly so when the IO is compared to MC. This underscores the central role of diabatic heating profiles in inducing the large-scale circulation and its water flux convergence as discussed earlier. While in general it rains less when the drying tendency is large, there is no one to one correspondence between precipitation and the water tendency. This reinforces the argument that the dry biases are not caused by parameterized precipitation alone.

After the first 20 days of integration (Figures 4.14c and d), the spread of the simulated water cycle is no longer increasing, surface evaporation does not vary much among the simulations, the large variability in precipitation is mainly determined by that of water supply through flux convergence (Figures 4.14c and d). In most simulations, however, water supply through flux convergence is much less than surface evaporation,
while in the reanalyses they are similar, as previously seen (Figure 4.5). This deficit of flux convergence is the main mechanism for maintaining the dry biases through the integrations in some simulations.

![Graphs showing domain and time mean evaporation, water flux convergence, precipitation, and precipitable water tendency](image)

**Figure 4.14.** Domain and time mean evaporation ($<E>$), water flux convergence $<[C]>$, precipitation $<P>$, and precipitable water tendency $<d[q]/dt>$ during the first 20 days (top panels) and the last 72 days of the simulations (bottom panels) for the IO (left panels) and MC (right panels). The values of $<d[q]/dt>$ are multiplied by 10 for better visibility. Units are mm day$^{-1}$. 
Chapter 5

The Role of Water Vapor in MJO Simulations

5.1. Motivation and Background

Realistic MJO simulations have been an unmet challenge for most global models (Gustafson and Weare, 2004; Hung et al., 2013; Lin et al., 2006; Monier et al., 2009; Slingo, 1996; Zhang, 2005; Zhang et al., 2006). It is perplexing that regional models fail to produce MJO signals in precipitation when they are able to produce MJO signals in wind under external forcing of lateral boundary conditions (Ray and Zhang, 2010). This suggests that the lateral boundary conditions may influence the circulation in the model interior but are not sufficient to affect simulated precipitation. One possibility for this to happen is that the water cycle in the models is severely biased and such biases cannot be corrected by the lateral boundary conditions (Ulate et al., 2014). When model cumulus parameterization schemes are sensitive to environmental humidity, biases in the water cycle would ultimately impair MJO signals in precipitation even when signals in the wind might be produced.

This raises an issue of circulation-convection decoupling of the MJO. This decoupling also occurs in observations when eastward propagating signals resembling
those of the MJO exist in precipitation but not in wind (Gottschalck, 2013) and in MJO prediction when skill remains high for the wind but low for precipitation (Ling et al., 2014). This decoupling of the MJO is contradictory to the common perception that the MJO is rooted in coupling of the circulation and convection (Khouider et al., 2012; Kim et al., 2009; Zhang, 2005). For this reason, the possible MJO decoupling of the circulation and convection should not be left without investigation.

This study further demonstrates the decoupling in numerical simulations. It first presents a failed attempt of simulating an observed MJO event using a regional model configured into a tropical channel domain (Ray et al., 2009). Then it introduces two approaches novel to the MJO study. One is a measure of errors in the amplitude and eastward propagation speed of a simulated MJO event. Using this measure in place of visual impression, commonly practiced in numerical case studies, the realism of a simulated MJO event and improvement of its simulations can be objectively and quantitatively assessed. The second novelty is the use of spectral nudging techniques to correct model errors in water vapor to various degrees and thus identify the features in water vapor that are the most important to realistic MJO simulations. Not surprisingly, low and mid tropospheric water vapor of planetary scales turns out to be the only important factor. What is surprising is that the effect of water vapor is so dominant that a phantom MJO event was simulated in a period without any observed MJO signal simply by nudging water vapor toward those from another period with an observed MJO event. This and other results suggest a certain degree of convection-circulation decoupling. This chapter is an extension of a paper accepted for publication pending minor revisions in the Journal of Advances in Modeling Earth Systems (Ulate et al., 2015).
5.2. Numerical Simulations

**Table 5.1.** List of amplitude, propagation speeds, and normalized error distance for TRMM, ERAI and WRF simulations. The simulations with values on italic are those that are plotted in Figure 5.4. The overbar indicates time or zonal mean.

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<thead>
<tr>
<th>Simulation Name</th>
<th>Nudging</th>
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The numerical configuration used in this chapter is CH (see Figure 2.3). The time step is 8 min. The main model physical parameterization schemes are the WSM3, the Rapid Radiative Transfer Model for longwave scheme (Mlawer et al., 1997), the Dudhia (1989) shortwave scheme, the Monin-Obukhov similarity theory (Monin and Obukhov,
1954) surface layer scheme, the unified Noah land-surface model (Chen and Dudhia, 2001), the MYJ planetary boundary layer and the BM cumulus scheme.

Thirty-nine simulations were made in this study, which are listed in Table 5.1. All except two simulations covered 72 days starting on October 20, 2009. MJO-09 event was observed during the period of time (see section 2.2.1). Different nudging approaches (section 2.3) were applied to examine factors that determine the realism of simulations of this MJO event. The two simulations that do not cover the MJO-09 period do not cover any observed MJO signal (see section 5.6). The purpose of these two simulations is to explore the extent to which the observed MJO event during October 20 – December 31, 2009 can be reproduced solely by MJO signals in water vapor.

5.2.1. Grid Nudging

The nudging technique has been used to correct model errors by forcing specific model fields towards observations (see Chapter 2). The prognostic variables to be corrected (\( \alpha \), see Equation 2.4) are wind (u,v), temperature (T), and water vapor mixing ratio (qv) (see Table 5.1). The reference values \( \alpha_0 \) (see Equation 2.4) are taken from the linearly-interpolated 6-hourly ERAI data. The nudging is factor \( G_a = 3.0 \times 10^{-6} \text{s}^{-1} \), which satisfies the numerical stability criteria \( G_a \leq \frac{1}{\Delta t} \) (Hoke, 1976; Stauffer and Seaman, 1990). This is a very weak nudging factor compared to those used in other studies (Hagos et al., 2011; Subramanian and Zhang, 2014). With this nudging factor, \( a \) approaches \( a_0 \) with an \( e \)-folding time of \( T_G = \frac{1}{G_a} = 3.85 \text{days} \). The value of the nudging factor was chosen somewhat arbitrarily. What the authors want to accomplish is a simulation with results closer to the observation than the control simulation. However, we also want a
simulation in which the numerical model is the least\textsuperscript{16} constrained. Nudging was applied only above the planetary boundary layer (PBL, approximately 900m over the Indian Ocean) to avoid diurnal cycle delays and biases (Stauffer and Seaman, 1991). The default weight function \( W_a \) is constant in time, uniform horizontally, 0 from the surface to the top of the PBL, 0.1 at the first vertical level above the PBL top, and 1 for all vertical levels above. This default weight function was modified in three simulations, in which \( W_a \) peaks in the lower (GL), mid (GM), and upper troposphere (GU), respectively (see Figure 5.1). Temperature and wind were corrected using grid nudging only in two simulations (GT and GUV respectively in Table 5.1). Water vapor was nudged in both grid and spectral nudging. Water vapor was also nudged to its time mean (Gtm), zonal mean (Gzm), and time-zonal mean (Gztm). GNall is the only simulation with all three variables (temperature, wind, and water vapor) nudged.

\begin{figure}[h]
\centering
\includegraphics[width=0.5\textwidth]{nudging_weight_function.png}
\caption{Vertical profiles of the nudging weight function \( (W_a) \) for GL, GM, and GU, and GQ simulations.}
\end{figure}

\textsuperscript{16} However “least” constrained may be interpreted under nudging conditions. When nudging is used, there is always going to be a constraint imposed during the numerical simulation. This constraint is to force the result close to the nudging field(s) (in this case the analysis).
Four sets of simulations were made using spectral nudging of water vapor mixing ratio. The first used only a single zonal wavenumber \( k \) in \( S_k \), \( k = 0, 1, 2, \) and \( 3 \). The second set used combinations of the zonal mean \( (k = 0) \) and a given wavenumber \( k \) in \( S_0k \), where \( k = 1, 2, \ldots, 6 \). In the third set, nudging was applied to all except one zonal wavenumber \( k \) in \( S_nok \), where \( k = 0, 1, 2, \ldots, 6 \). In the last set, nudging was applied using a group of zonal wavenumbers from 0 to \( k \) in \( S_0tok \), where \( k = 1, 2, \ldots, 18 \). It is important to notice that in the first group of simulations \( (S_k) \) nudging towards \( k = 0 \) \( (S_0) \) is different from grid nudging towards the zonal mean \( (Gxm) \). In \( S_0 \), only the zonal mean is corrected while other zonal components are intact. In \( Gxm \), each grid point is forced towards the same value, the zonal mean.

All of these spectral nudging experiments were designed to test the sensitivity of MJO simulations to the zonal scale of corrected water vapor.

Our main simulation targets are the strength and speed of eastward propagating signals of the MJO. We used an MJO tracking method to quantitatively measure these two variables (see Chapter 3). The tracked propagation speed and amplitude in a simulation can be compared to those from TRMM for precipitation and ERAI for U850. Differences between simulated and observed speeds and amplitudes measure errors in the simulation. Statistics of observed propagation speeds and amplitudes were derived from TRMM precipitation and ERAI U850 from 23 MJO events observed between 1998 – 2009, which were identified according to Ling et al. (2013). The tracked mean amplitude and speed are 5.35 mm day\(^{-1}\) and 6.65 ms\(^{-1}\) for precipitation, and 2.64 ms\(^{-1}\) and

\[17 \text{ See chapter 3 for details on the MJO cases}\]
8.5 m\text{s}^{-1} for U850. The standard deviation of the amplitude and speed are 1.5 \text{mmday}^{-1} and 3.4 m\text{s}^{-1} for precipitation, and 0.8 m\text{s}^{-1} and 4.3 m\text{s}^{-1} for U850.

The standard deviations of the tracked amplitudes and speeds of the observed MJO events were used as a measure of their climatological variability. This is measured by a “normalized error distance” D.

As mention previously, the nudging factor for the majority of the simulations is $G_a = 3.0 \times 10^{-6}\text{s}^{-1}$. However, we performed some numerical simulations in which the nudging factor was increased to $3.0 \times 10^{-3}\text{s}^{-1}$. In this case $\alpha$ approaches $\alpha_0$ with an $e$-folding time of $T_G = \frac{1}{G_a} = 5.56$ minutes. In most of our simulations, we nudged only the moisture, however, there are some simulations in which the temperature, the wind and both of them were also slightly corrected using grid nudging (see Table 5.1). Water vapor was used in both grid and spectral nudging.

**5.3. Control Simulations Comparison**

All simulations presented in this study aim to reproduce an observed MJO event in October - November 2009. Figures 5.2a and b show respectively time-longitude diagrams of anomalies in TRMM precipitation and ERAI U850 averaged over 10\textdegree N-10\textdegree S. The MJO signals are unmistakable. Based on the tracking method (Chapter 3) the eastward propagation speed of this MJO event is 4.5 m\text{s}^{-1} in precipitation (marked by the black line in Figures 5.2a) and 8 m\text{s}^{-1} in the zonal wind (black line in Figures 5.2b), with an amplitude of 5.3 mm\text{day}^{-1} in precipitation and 2.8 m\text{s}^{-1} in the zonal wind. The apparent difference in the tracked propagating speeds in precipitation and zonal wind will be discussed later.
Figure 5.2. Time–Longitude diagrams of the 5-day running mean anomalies in (left panels) precipitation and (right) U850 averaged over 10°N-10°S from (a) TRMM and (b) ERAI, (c and d) CTRL, and (e and f) GQ. Black lines are tracks determined by the tracking method (see Chapter 3). Table 5.1 lists their speeds and magnitudes. Day 1 is October 20, 2009.
Based on our MJO metrics, the Control simulation (CTRL) failed to reproduce the observed MJO signals in precipitation (Figures 5.2c) and U850 (Figures 5.2d). The tracked eastward propagation speed is 4.5 ms\(^{-1}\) for precipitation and 5 ms\(^{-1}\) for U850. Both are within one standard deviation from those of TRMM and ERAI, respectively. But the tracked amplitudes are 2.3 mm\(\text{day}^{-1}\) for precipitation and 1.6 ms\(^{-1}\) for U850, both smaller than the observed by more than one standard deviation. The normalized error distance D is greater than 1 for both precipitation and U850 (Figures 5.3a, b, open diamonds). Therefore CTRL failed to reproduce the observed MJO event.

Interaction between precipitation/convection and humidity has been conceived as central to the MJO and its simulations (Grabowski and Moncrieff, 2004; Hannah and Maloney, 2011). The possible role of humidity in MJO simulations is explored next through a set of nudging experiments. The general strategy is to correct model errors in humidity, wind and temperature through grid nudging to force the model to produce an MJO event as close to the observed as possible, and then to alter the level of error correction and study how the simulated MJO may deteriorate accordingly. The relative importance of error corrections in the different fields can be assessed through comparisons of the resulting D values.
**Figure 5.3.** Scatter diagrams of tracked speed and magnitude for anomalies in (a) precipitation and (b) U850, normalized by their respective climatological standard deviation (Chapter 3). Vertical and horizontal dotted lines mark the values of TRMM precipitation (a) and ERAI U850 (b). Circles mark one standard deviations from the TRMM and ERAI values (stars). D < 1 for all points within the circles, D > 1 for all points outside the circles.
5.4. Grid Nudging Results

Judged by the value of the normalized error distance $D$, the best simulation of the October-November 2009 MJO event by the WRF model was made when grid nudging of water vapor mixing ration ($q_v$) was applied (GQ for “grid nudging of $q_v$”). The eastward propagating signals in precipitation and U850 produced in GQ are evident in this simulation (Figures 5.2e and f). The tracked speeds and amplitudes are 4.5 m s$^{-1}$ and 4.0 mm day$^{-1}$ for precipitation (Figures 5.2e) and 9.5 m s$^{-1}$ and 3 m s$^{-1}$ for U850 (Figures 5.2f). For both precipitation and U850, $D < 1$ (solid diamonds in Figure 5.3). GQ is then considered a successful simulation. Out of our 39 numerical simulations, GQ produced the best signal of the MJO in both precipitation and wind. GQ is taken as a benchmark to compare to the other simulations.

The success of GQ in reproducing the MJO events is not surprising, based on the previous MJO simulation with humidity nudging (Hagos et al., 2011). What is surprising is that correcting humidity alone is enough for a realistic MJO simulation. If the improved MJO signal in precipitation in GQ can be attributed to parameterized convection reacting to corrected humidity, then the improved MJO signal in U850 must come from the circulation response to the improved MJO signals in convection. When grid nudging was applied to horizontal wind alone (GUV), no improvement was made in either precipitation or U850, and the simulation failed (Figure 5.3, black downward pointing triangles). This is not hard to understand if parameterized cumulus convection is the main source of the failure of CTRL and it is very sensitive to environmental humidity (Arakawa and Schubert, 1974; Derbyshire et al., 2004). Corrected U850 through nudging alone is insufficient to overcome model errors in parameterized convection. Erroneous
precipitation in the model remains and overwhelms the circulation that is corrected by nudging. In this sense, a deficient parameterization scheme not only fails to provide needed effects to produce MJO signals, it actually may work to destroy existing MJO signals (from nudging in our case). A similar result was obtained when temperature alone was corrected through nudging (GT). In this case, $D > 1$ for both speeds and amplitudes in precipitation and U850 (Figure 5.3, black upward pointing triangles), indicating no MJO was produced. When humidity, temperature, and winds are all corrected through nudging (GA for “grid nudging of all fields”), the MJO signal in U850 became close to GQ but the amplitude in precipitation is too weak to be considered a successful MJO simulation (Figure 5.3, black squares). The weak MJO signal in precipitation suggests that the cumulus parameterization scheme used in the WRF simulation responds very differently to environmental humidity and temperature combined from that used in the ERAI data assimilation system. Results from these four nudging experiments (GQ, GA, GUV, GT) demonstrate that simulated MJO precipitation is sensitive only to humidity.

Using a global model, Subramanian and Zhang (2014) found that nudging temperature, humidity, and winds led to a successful MJO simulation. Nudging temperature (humidity) only produced the weakest (strongest) amount of precipitation, and their best MJO simulation was obtained when only the winds were nudged. The differences between their and our results may come from many factors, such as $T G$, model configurations, especially physics, nudging tactics, and intrinsic differences between MJO events.

The nudging method allows us to quantify the dependence of MJO simulations on the vertical structure of humidity by specifying the vertical profile of the nudging
weighting function (Figure 5.1). Correcting the humidity field only in the upper troposphere (GU) did not improve the MJO simulation much from CTRL (Figure 5.3, upward-pointing gray triangle). Correcting humidity in the lower or mid troposphere alone led to substantial improvement in U850, but insufficient improvement for precipitation (Figure 5.3, downward and right-pointing gray triangles). There are two lessons here. First, humidity in both the lower- and mid-troposphere is important. Second, it is much easier to get corrected MJO signals in U850 than in precipitation. We will revisit this again later in this study.

Three additional grid nudging experiments were conducted, in which water vapor nudging was performed for its time mean (Gtm), zonal mean (Gxm), and time and zonal mean (Gtxm). In Gtm, the humidity field is forced toward the spatial distribution in ERAI averaged over the modeling period. In Gxm, nudging reduces the zonal mean error in humidity. None of these produced any MJO signals (Table 5.1). These results suggest that it is the humidity errors in tempo-spatial perturbations, not only in the mean, that are detrimental to MJO simulations. The role of the zonal scale of humidity in MJO simulations was investigated through spectral nudging discussed next.

5.5. Zonal Modes of Moisture Corrections

We used spectral nudging in the zonal direction to evaluate the dependence of MJO simulations on the zonal scale of water vapor. In all simulations with spectral nudging, the default vertical profile of the weighting function was used (Figure 5.1).

First, correcting errors in planetary-scale perturbations in water vapor of a single zonal wavenumber k= 0, 1, 2, or 3 (S0, S1, S2, S3) did not make MJO simulations
realistic except for U850 in S1 ($D^U=0.9$), in which simulated precipitation is still unrealistic ($D^{PCP}=1.7$, Table 1). Correcting water vapor perturbations of a single zonal wavenumber along with the mean ($S0k. k = 1, 2, \ldots, 6$) did not produce a realistic MJO simulation either, with $D > 1$ for both precipitation and U850 (Table 5.1). These results appear to be counterintuitive. The MJO is a zonal wavenumber 1 phenomenon, and it is expected that its associated $k = 1$ signals in water vapor alone should help the MJO simulation. Apparently, the role of water vapor in MJO simulations has a multi-scale characteristic.

To further quantify the role of zonal scales of water vapor in MJO simulations, we made another set of spectral nudging experiments by nudging humidity of zonal wavenumbers 0 to $k$ inclusively, with $k = 1, 2, \ldots, 18$ in $S0tok$. MJO simulations were improved in both precipitation and U850 when more planetary scale perturbations of humidity are nudged up to $k = 3$, as indicated by the reduction of the normalized error distance $D$ (Figure 5.4a). Including higher zonal wavenumbers did not further improve the simulation drastically but did show a convergence to benchmark values of $D$, which are from GQ. This is expected. The non-monotony is considered noise, which is expected to reduce or even disappear in ensemble simulations. It is also important to notice that the decrease of the error in both precipitation and zonal wind is more drastic from $Sk0to2$ to $Sk0to3$. Overall, Figure 5.4a suggests that the fidelity of MJO simulations depends mainly on planetary-scale, not synoptic-scale, perturbations in humidity.
Figure 5.4. Normalized error distance ($D_{\text{PCP}}, D_{\text{U}}$) for experiments of (left panel) $S0\text{tok}, k = 0, 1, 2, \ldots, 18$ and (right) $\text{Snok}, k = 0, 1, 2, \ldots, 6$. The black line and circles is for precipitation and the gray for $U850$. $D = 1$ corresponds to the circle in Figure 5.3.

In the final set of spectral nudging experiments, humidity of all zonal wavenumbers were nudged except for a single wavenumber $k$ in $\text{Snok}, k = 0, 1, 2, \ldots, 6$. These experiments explore how MJO simulations would degrade because of errors in perturbations of a particular zonal scale. With nudging for all zonal wavenumbers the result is practically the same as that from GQ, which is used here to compare to these nudging experiments. In all these simulations, $D < 1$ for $U850$ (Figure 5.4b). For precipitation, $D < 1$ except $k = 0, 1, \text{and} 3$. With the exception of $k = 3$, $D$ decreases with increasing $k$ and converges to the value for GQ. This suggests again that water vapor perturbations of synoptic scales are less important than those of planetary scales.
Figure 5.5. Scatter diagram of normalized tracked (a) magnitude and (b) speed for anomalies in precipitation and zonal wind at 850hPa. Dashed vertical and horizontal lines mark the values of TRMM precipitation and ERAI zonal wind. The solid vertical and horizontal lines represent ± one standard deviation from the TRMM and ERAI values for the precipitation and zonal wind amplitude (a) or speed (b), respectively. The simulations plotted are the same as in Figure 5.3.

One can hardly fail to get an impression from Figures 5.3 and 5.4 that when model errors in humidity are corrected (via nudging), simulated U850 is far more likely than simulated precipitation to be improved. In addition, regardless of their exact values, D for U850 are smaller than for precipitation in 27 out of the 39 simulations (Table 5.1). This is not surprising, because precipitation is known to be one of the most difficult fields to be reproduced well by numerical models (Arakawa, 2004; Kim et al., 2009; Molinari and Dudek, 1992). On the other hand, correcting errors in water vapor would not directly affect U850 unless through precipitation. The MJO is commonly perceived as a convection-circulation coupled phenomenon. It is therefore expected that simulations of
MJO signals in precipitation and U850 are improved together when errors in humidity are corrected. This is indeed the case for their amplitudes (Figure 5.5a). From all nudging experiments, amplitudes of MJO U850 and precipitation are correlated (coefficient=0.78, significant at the 95% confidence level), meaning they are improved or degraded proportionally. For the 22 observed MJO events, the amplitudes of their precipitation and U850 are also correlated (coefficient = 0.73, significant at the 95% confidence level). However, their speeds are correlated in neither observations nor our simulations. So in both observations and simulations there is a certain degree of decoupling between MJO U850 and precipitation manifested in their different propagation speeds.

5.6. Simulation of a Phantom MJO Event

The role of humidity in MJO simulations is further examined in an extreme nudging experiment. In this experiment, the initial and lateral boundary conditions were from a period (February - April 2009) without any MJO in observations (Figure 5.6a). No MJO was simulated for this period (in NoMJO), as expected (Figure 5.6b). In another simulation (NoMJOGQ), we used the same initial and lateral boundary conditions as in NoMJO but nudged the full humidity field from the period of CTRL (October – December 2009, Same as in GQ). Surprisingly, a phantom MJO event was produced (Figure 5.6c). Its signal in precipitation is close to the observed MJO in that period with D < 1 (left pointing open triangles in Figure 5.3). This strongly suggests that simulated MJO precipitation solely depends on humidity in the model, regardless of other fields that are constrained through the initial and the boundary conditions. The importance of other fields to simulated MJO precipitation must come in through their influences on humidity.
While the critical role of humidity in MJO simulations has been previously recognized (Hannah and Maloney, 2011; Kim, 2014; Thayer-Calder and Randall, 2009), it has never been demonstrated to this extreme level as by this phantom MJO simulation. The simulated MJO signal in U850 is a different story. In this case the wind amplitude is too large (a different kind of failure from the previous simulations) and D > 1 (see Table 5.1 and Figure 5.3b). (D > 1) This suggests that the MJO circulation is slaved by precipitation up to certain extent because both winds amplitude and speed are improved in NoMJONQ (left pointing black triangle in Figure 5.3b) from NoMJO (left pointing open triangle in Figure 5.3b). However, the phantom MJO signal in precipitation is insufficient to force a realistic MJO signal in U850 close enough to the observed case.

**Figure 5.6.** Time – Longitude diagrams of the 5-day running mean anomalies in precipitation averaged over 10°N-10°S from (a) TRMM, (b) NoMJO, and (c) NoMJOGQ. Black lines are tracks from the tracking method. Table 5.1 lists their speeds and magnitudes. Day 1 is February 1, 2009.
5.7. Mechanisms for Humidity Effects

The results from the nudging experiments described in the previous section demonstrate the critical role of humidity in MJO simulations. The exact reason for the improvement of MJO simulations by correcting errors in humidity is explored here. Previous studies have suggested several factors important to MJO simulations. They include the mean state (Kim et al., 2011; Ray and Li, 2013; Slingo, 1996), sensitivity of precipitation to humidity (Hannah and Maloney, 2011; Thayer-Calder and Randall, 2009), vertical structures of diabatic heating (Li et al., 2009; Lin et al., 2004), and zonal asymmetry in tropospheric humidity (Hsu and Li, 2012; Hsu et al., 2014). These and other factors are examined in this section.

5.7.1. Mean State

CTRL produced obvious mean dry biases in both precipitation (Figure 5.7, dashed gray line) and column-integrated water vapor (not shown) in comparison to TRMM and ERAI, respectively. When errors in humidity were fully corrected by nudging in GQ, the dry biases in CTRL were corrected in certain regions but wet biases were introduced in others (not shown). The mean state of humidity is unlikely to be a critical factor in MJO simulations discussed here, as also demonstrated by the S0 tests. We have made 16 additional WRF simulations for the same MJO event, each using a different combination of 8 cumulus and 3 planetary boundary layer schemes. Some of these simulations produced mean dry biases in tropical precipitation and tropospheric water vapor, others wet and moist biases (Ulate et al., 2014). None of these simulations produced the MJO signal (D > 1 for all these simulations). When simulations suffer from both an absence of realistic MJO signals and large biases in mean precipitation and/or
water vapor, the two may not be causally related, as in the case discussed in this study. The reason for WRF’s failure of producing MJO signals resides in the spatiotemporal variability of water vapor.

**Figure 5.7.** Pressure-Time mean moistening nudging tendencies (contours, gKg\(^{-1}\)day\(^{-1}\)) from GQ, mean precipitation (mmday\(^{-1}\)) from GQ (solid gray line), CTRL (dashed gray line), and TRMM (black line) averaged over the Indian Ocean (10°N-10°S, 65 - 95°E).

### 5.7.2. Large-Scale Patterns of Water Vapor

Planetary-scale patterns of water vapor play critical roles in this MJO event in both observations and simulations. During the period of this event, the atmosphere is dry over most of the equatorial Indian Ocean, Maritime Continent, and Pacific, according to vertically integrated water vapor (IWV) from ERAI (Figure 5.8a, contours). An exception is a “moisture corridor” over central and eastern Indian Ocean in the early part of the period (days 10 – 30), Maritime Continent briefly in the middle of the period (around day 30), and western Pacific later in the period (days 30 – 60). Heavy rainfall occurs in this moisture corridor (Figure 5.8a, colors), which is the manifestation of this MJO event in precipitation. There is an apparent jump of heaviest rain as it moves from the eastern
Indian Ocean to western Pacific across the Maritime Continent (Figure 5.8a). Such jumpy behavior over the Maritime Continent in the eastward propagation of MJO rainfall is common (Hsu and Lee, 2005; Peatman et al., 2014).

**Figure 5.8.** Time-longitude diagrams of the total rain rate (colors) overlaid with integrated wave vapor (IWV, contours, mm) averaged over 10°S – 10°N from (a) TRMM and ERAI, (b) CTRL, and (c) GQ.
In CTRL (Figure 5.8b), the atmosphere is too dry over the central and eastern Indian Ocean during the early part of the period, over the Maritime Continent during almost the entire period, and over the western Pacific early in the period; it is too moist over the western Pacific during the entire period except the early part. In consequence, there is perpetual heavy rain locked over the western Pacific and no rain over the Maritime Continent. This destroys the MJO signal in precipitation in this simulation. The role of water vapor nudging in GQ (Figure 5.8c) is to introduce more water vapor over the central and eastern Indian Ocean and less over the western Pacific in the early part of the period, and less moisture over the western Indian Ocean and more moisture over the Maritime Continent between days 20 and 30 (Figure 5.8c and 5.7). These moisture corrections help to restore the moisture corridor shown by ERAI but missed in CTRL, which induces rainfall to move from the Indian Ocean to the Pacific across the Maritime Continent, hence the MJO signal in precipitation.

For example, nudging tendencies \( \left( \frac{\partial a}{\partial t} \right)_{\text{Nudging}} \) in GQ over the Indian Ocean (IO, 10°N-10°S, 65 - 95°E) indicate that when MJO convection is active there (days 10 – 25) the largest dry bias in CTRL is in the lower troposphere (Figure 5.7), which is apparently the main reason for the failure of CTRL to produce MJO signals in precipitation. When this lower-tropospheric dry bias was corrected by nudging in GQ, a very clear MJO signal in precipitation close to TRMM was produced.

Figure 5.8 clearly shows the critical role of large-scale patterns of water vapor, with moisture and dryness in the right location at the right time, in MJO precipitation. It also suggests that the failure of CTRL to produce the observed MJO event is not in the response of its parameterized rainfall to environmental moisture or in its dry biases in
moisture. It is the model’s inability to produce the correct large-scale pattern of water vapor. This explains why our other simulations of WRF also fail to produce the MJO when they suffer from moist biases: the needed dry episodes are missing. The inability of WRF to produce correct moisture distribution and variability in the tropics is rooted in the deficiencies of its parameterization schemes for not only cumulus, but also the boundary layer and microphysics, and their improper combinations (Ulate et al., 2014).

5.7.3. Contributions from Planetary-Scale Perturbations

The spectral nudging experiments revealed that every single component of planetary-scale \( (k = 1 - 3) \) perturbations in water vapor is essential for realistic MJO simulations. The reason for this is not immediately obvious. Figure 5.9 shows time-longitude diagrams of ERAI IWV of \( k = 1, 2, 3 \) and their combinations. The eastward propagation of total IWV from the Indian to Pacific Ocean seen in Figure 5.8a is mostly provided by zonal wavenumber \( k = 1 \) (Figure 5.9a). It alone can help produce correct propagation speed of MJO precipitation but not its amplitude (Figure 5.3a). Perturbations of higher planetary zonal wavenumbers in water vapor are needed for realistic amplitude in simulated MJO precipitation. Although they are all quasi-stationary, temporal fluctuations in their amplitudes strengthen the MJO signals produced by the \( k = 1 \) perturbation. IWV perturbation of \( k = 2 \) show perpetual positive anomalies over the Maritime Continent and western Pacific, with its amplitude increasing around day 30 (Figure 5.9b) when positive anomalies in IWV of \( k = 1 \) pass over there.
**Figure 5.9.** Time-Time-Longitude diagrams of vertically integrated water vapor (mm) of zonal wavenumber (a) $k=1$, (b) $k=2$, (c) $k=3$, (d) $k=4$, (e) $k=1-3$, and (f) $k=1-4$ averaged over 10°S–10°N from ERAI. The land mask distribution is shown below (e) and (f).

Positive anomalies in IWV perturbation of $k = 3$ (Figure 5.9c) mainly help eastward movement in total IWV over the Indian Ocean; its negative anomalies over the Maritime Continent help suppress rainfall there in the early and later stages of the MJO event (Figure 5.9c). MJO signals in IWV would be too weak over the Maritime Continent and western Pacific without its component of $k = 2$ (Figure 5.9b), and too weak over the Indian Ocean without its component of $k=3$. In combination, perturbations of $k = 1-3$
(Figure 5.9e) make the eastward moving signal of IWV much stronger and sharper than that of k = 1 alone (Figure 5.9a, e), which is needed for a realistic amplitude of MJO precipitation (Figure 5.4). It is interesting to note that IWV perturbation of k = 4 (Figure 5.9d) helps to generate the jump from the Indian to western Pacific Oceans across the Maritime Continent seen in total IWV (Figure 5.8f) by reducing IWV over the Maritime Continent. But without this IWV jump, simulation of MJO precipitation can be realistic (in S03) as measure by the normalized error distance.

The vertical structure of anomalies in humidity for this MJO event shows a sharp zonal gradient in the lower troposphere including the boundary layer (Figure 5.10, left column). Positive anomalies in low-level humidity move eastward with negative anomalies preceding to the east (before day 20) and trailing to the west (after day 20). In the regions of maximum rainfall, large positive anomalies extend upward to above the 500 hPa level due to transport by deep convection. East of it, they are confined to the lower troposphere. The sources for such low-level moisture, often referred to as low-level moistening, have been a point of debate. Possibilities include frictional moisture convergence (Benedict and Randall, 2007; Maloney and Hartmann, 1998; Wang and Li, 1994), shallow and congestus clouds (Del Genio et al., 2012; Hagos and Leung, 2011), and moisture convergence by the large-scale circulation (Majda and Stechmann, 2009), and synoptic-scale eddies (Majda and Stechmann, 2009; Maloney, 2009; Sobel and Maloney, 2013). The deepening of the low-level moist layer has been previously observed (Johnson et al., 1999; Kiladis et al., 2005; Tian et al., 2006) and perceived as a necessary condition for the development of MJO deep convection (Johnson and Ciesielski, 2013; Zhang, 2005) and its role in the MJO is a main target of the recent
DYNAMO field campaign (Yoneyama et al., 2013). It has been suggested that the eastward propagation of the MJO is facilitated by the negative anomalies in moisture east of an MJO convection center as well as west of it (Hsu and Li, 2012; Maloney, 2009; Zhao et al., 2013).

**Figure 5.10.** Longitude-pressure diagram of (left column) anomalies in ERAI water vapor mixing ratio $q$ (gkg$^{-1}$, colors) and TRMM precipitation (mmday$^{-1}$, top of each panel) and (right) $q$ nudging tendencies (gkg$^{-1}$day$^{-1}$ colors) and precipitation anomalies (top of each panel) in S0to3, all averaged over 10$^\circ$S – 10$^\circ$N and over 5 days centered at days (from top row down) 10 (a and b), 20 (c and d), 30 (e and f), and 40 (g and h).
This structural evolution in moisture anomalies is not reproduced in CTRL. After day 10 when influences of the initial conditions are gone, anomalies in low-level moisture scatter over the Indian and Pacific Oceans without large positive or negative centers (not shown). With humidity nudging, structures of moisture anomalies in GQ (Figure 5.7) becomes very similar to those in ERAI seen in Figure 5.10. Because CTRL suffers from a mean dry bias, the total nudging tendencies in GQ are dominantly positive. The mean dry bias is corrected mainly by nudging of the $k = 0$ component. Nudging of the $k = 1$, 2, and 3 components yields positive low-level tendencies in a shallow layer that gradually deepens from east to west with the maximum tendencies located where maximum anomalies in $q$ are (Figures 5.9, 5.10). This structure is well maintained when water vapor components of higher zonal wavenumbers are included in the spectral nudging, as they add more details to the general structure of the nudging tendencies and improve MJO simulations incrementally. A realistic MJO simulation can, however, be made by nudging water vapor components of only $k = 0 – 3$ (Figure 5.4).

5.8. Chapter Summary

We attempted to simulate an observed MJO event using a tropical channel configuration of the WRF model but failed completely no matter what parameterization schemes for cumulus convection and the boundary layer available to WRF were used. We then made several attempts to correct errors in the simulations through nudging. The realism of MJO simulations was quantitatively assessed by a new MJO metric in terms of a normalized error distance that measures combined errors in simulated propagation speed and amplitude of MJO precipitation and U850 against TRMM observations and
ERAI products. With the caution that effects of water vapor nudging may depend on parameterization schemes, the main results, their implications and interpretations can be summarized as the following:

(a) The deepening of the low-level moist layer leading to the MJO convective center, a repeatedly observed feature of the MJO, is achieved primarily by water vapor perturbations of planetary scales \((k = 1 - 3)\). Synoptic-scale perturbations in water vapor may improve, but are not essential to, a realistic MJO simulation.

(b) MJO precipitation is much more difficult to be reproduced realistically than its circulation, such as zonal wind at 850 hPa (U850). Simulated MJO U850 can be close to the observed, while simulated MJO precipitation is not. In this sense, using precipitation against observations to evaluate MJO simulations is of a higher standard than using the circulation alone and should be applied to future numerical studies of the MJO.

(c) There might be two reasons for a realistic simulation in MJO circulation but not in its precipitation. Deficient cumulus parameterization schemes may work against MJO dynamics set by initial and lateral boundary conditions that help produce MJO signals in the circulation in a regional model (Ray and Zhang, 2010; Ray et al., 2009). When a reduction in errors in precipitation by water vapor nudging is insufficient to make simulated precipitation realistic, it might be sufficient to minimize the negative effects to allow the simulated circulation to be realistic under the influence of MJO dynamics provided by the lateral boundary conditions.

(d) The second reason is a degree of decoupling between the MJO circulation and precipitation. While amplitudes of MJO precipitation and U850 are highly correlated in both observations and simulations, their propagation speeds are not. While the
decoupling might be counterintuitive based on the perception that MJO is a product of convection-circulation interaction (Rui and Wang, 1990), it may not be so surprising if one accepts the possibility that MJO dynamics are determined by its intrinsic structure that is independent of convection although the two actively interact. Such signals of MJO intrinsic structure not directly associated with convection have begun to be revealed (Ling et al., 2013; Matthews, 2008; Straub, 2013). The MJO circulation appears to be determined by such an MJO dynamic structure, not solely by MJO precipitation. In the period of February-March 2009, nudging of MJO signals in humidity can fake phantom MJO signals in precipitation but not in the circulation (in NoMJOQ).

In summary, this study introduces a new method of spectral nudging of water vapor and a new MJO diagnostic metric that quantifies errors in simulations of individual MJO events. The results from this study highlight the critical role of planetary-scale (zonal wavenumber 1 – 3) perturbations of water vapor in the MJO and demonstrate possible decoupling of MJO convection and circulation. We recommend our new metric to all future numerical studies of the MJO. We desire to further investigate the issue of convection-circulation decoupling of the MJO.
Chapter 6

Role and Limitations of the Boundary Conditions in MJO Simulations

6.1. Motivation and Background

Regional Model simulations are frequently-used tool in meteorology. Their resolution can be rapidly increased at a lower cost than running a global model at the same resolution. Additionally, the simulations (theoretically) will be under a more controlled environment. This is because IC, LBC, and bottom conditions (BTC, usually SST) are provided for the RM, rather than sea surface temperature (SST) alone as for the GCMs. However, the RMs typically fail in successfully reproducing the MJO. On the other hand, some GCM are able to reproduce the MJO by starting only from SST as initial conditions in forecast mode (Woolnough et al., 2001). The reasons why the RMs are not able to capture the MJO, even when the IC, and LBC have the MJO signal in them (u, v, q, t) have not been studied yet.

The RM has only one IC, one BTC, and 4 LBC: North, South, East and West; unless the RM has a channel configuration, in which case the East and West boundaries are periodic (Ray et al., 2009). The top of the model is not considered a BC because this
one has filters to avoid the reflection of waves. The IC are present only at time 0, whereas
the LBC and BTC can either change in time or remain constant, depending on the type of
simulation.

The importance of LBC and BTC is to transfer information from the large scale
into the domain. Then, the RM will “feel” the large scale and, ideally, will be constrained
by it (Rinke and Dethloff, 2000). Despite the decrease in degrees of freedom from the
GCMs to the RMs, the RMs struggle to obtain a successful MJO simulation without
special treatment such as prescribing daily varying sea surface temperature (Hagos and
Leung, 2011) and nudging certain variables toward reanalysis data (Hagos and Leung,
2011). It has been widely studied that the RM may reproduce MJO signals in zonal wind
but not in precipitation (Gustafson and Weare, 2004; Monier et al., 2009; Ray and Zhang,
2010).

Numerical models such as ECMWF are able to obtain better MJO simulations at
coarse resolution than RM numerical simulations where the IC, LBC, and BTC were
taken from the global climate numerical model (Bechtold et al., 2014). This suggests that
in the limited area models, the information provided by the BC is somehow lost during
the simulation period.

Some studies have shown that the model internal dynamics dominate during
simulations after the “spin up” time (first 5 to 15 days) has passed (Giorgi and Bi, 2000;
Giorgi et al., 1993; Jones et al., 1995; Wu, 2005). Thereafter, the model will develop its
own circulations. However, this subject is an ongoing debate. Wu (2005) found that RMs
do not have sensitivity to LBC or IC, even when the IC are perturbed inside a specific
area in the domain. However, Giorgi and Bi (2000) argued that the differences in the
results when LBC and IC are changed are based on model configuration, physics, synoptic conditions, and region. These results demonstrate how complicated and non-linear the interactions between the model internal variability, LBC, BTC, and IC are.

The IC and LBC interaction has been widely studied in the context of ensemble forecasting methods (Christensen, 2001; Nutter et al., 2004; Wu, 2005). Many of these studies tend to be specific to mid latitudes forecasts (Giorgi and Bi, 2000; Gustafsson et al., 1998; Liang et al., 2001). This is because when the tropics are included in the domain, they generate more uncertainty (Liang et al., 2001). It has also been shown that the convective regime is largely influenced by the type of BC used. Salzmann (2004) show that when periodic LBC were used, the convection organizes first in squall lines and into small single clouds afterwards. On the other hand, the simulations with time dependent LBC show the presence of both squall lines and small clouds simultaneously. Only one study has tested different LBC to MJO simulations: (Ray and Zhang, 2010), which showed the importance of the time dependent LBC in order to achieve the wind transition to MJO initiation stages. However, the effect on the MJO convection was not studied.

Since the MJO interacts with the ocean by forcing surface fluxes and precipitation, which lead to SST changes (Park et al., 2005; Wang and Xie, 1998; Woolnough et al., 2001; Zhang, 2003), the BTCs play an important role on RM simulations of the MJO. The MJO and ocean interaction is described as the decrease in the SST behind and below the convective center of the MJO, by adding fresh water and blocking the shortwave radiation absorption by the ocean. On the other hand, the SSTs to the east of the convective center are associated with positive anomalies since they received more insolation (Maloney and Sobel, 2004). This lag between negative
anomalies-convective center, and positive anomalies is important to be captured during MJO simulations. However, the best way to approach how to more-realistically simulate the MJO-Ocean interaction has not reached a consensus (Bernie et al., 2005; Hendon, 2000; Kim et al., 2010; Subramanian and Zhang, 2013; Vintzileos and Gottschalck, 2013; Vitart et al., 2007). There is a lack of studies where the bottom conditions are examined in RMs. Seo et al. (2014) studied the coupled and uncoupled simulations in MJO forecasts. Their results show the improvement of the MJO in the fully coupled simulation, similar to the results reported by climate simulations in GCMS (Kim et al., 2008; Waliser and Lau, 1999).

There is also a debate within the scientific community about the optimal (if any) RM configuration that one should use when studying the MJO. Presently, different studies use different LBC, IC, and BTC (SSTs) configurations. This leads to a discussion about the most appropriate way to configure the RM so the simulated MJO (if any) will be produced by the correct mechanisms, and not forced by erroneous configurations. Usually RM studies obtain somewhat successful MJO when the SSTs are updated in time. The main improvement is in the eastward propagation of the MJO (Hagos and Leung, 2011). The main implications of this assumption need to be studied.

In summary, the use of RM in order to simulate specific MJO events has not been studied to the fullest. The scientific community needs to reach a base level where the principal advantages and deficiencies of RMs should be evaluated so that we can begin to answer many of these open questions involving the MJO. The research presented in this chapter seeks to investigate these issues and fill in the knowledge gap.
6.1.1. The Decoupling Problem in MJO Simulations

Another problem that has been observed in both GCM and RMs is the decoupling between the wind and the precipitation associated to the MJO. The reason why the numerical models are not able to reproduce the “convective MJO” while obtaining the “dynamical MJO” characteristics has not yet been explained. However, this suggest that the lateral boundary conditions may influence the circulation in the model interior but are not sufficient to affect simulated precipitation.

This decoupling has also be documented in observations when eastward propagating signals resembling those of the MJO exist in precipitation but not in wind (Gottschalck, 2013), and in MJO prediction when skill remains high for the wind but low for precipitation (Ling et al., 2014). This decoupling of the MJO is contradictory to the common perception that the MJO is rooted in coupling of the circulation and convection (Khouider et al., 2012; Kim et al., 2008; Zhang, 2005). For this reason, the possible MJO decoupling of the circulation and convection should not be left without further investigation.

This chapter will first quantify, using the MJO tracking method described in chapter 3, the speed, amplitude, start day and scale associated with the MJO on each simulation. These values are then compared with the values obtained from reanalysis (ERAI) and satellite observations (TRMM).

6.2. Methods

The simulations in this chapter aim to reproduce the MJO_DY case, which took place between November and December 2011 (see Chapter 2). The first set of numerical
simulations constitute those listed in Table 6.1. This Table lists 28 numerical simulations with 7 physics configurations (see Table).

**Table 6.1.** Numerical Simulations: The start day for all the simulations is November 10 2011. All the simulations listed have 28 vertical levels, used the WSM6 microphysics scheme, and were run in IO_50 (see Figure 2.3).

<table>
<thead>
<tr>
<th>Name</th>
<th>LBC</th>
<th>SST</th>
<th>Cumulus</th>
<th>Planetary Boundary Layer</th>
<th>Radiation</th>
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<td>Mod ZM PB</td>
<td>RRTM</td>
<td>CAM</td>
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<td>Update</td>
<td>MYJ</td>
<td>RRTM</td>
<td>CAM</td>
</tr>
<tr>
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<td>Constant</td>
<td>UW</td>
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<td>CAM</td>
</tr>
<tr>
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<td>Constant</td>
<td></td>
<td>RRTM</td>
<td>RRTMG</td>
</tr>
<tr>
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<td>TK</td>
<td>YSU</td>
<td>RRTMG</td>
</tr>
<tr>
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<td>Update</td>
<td>MYJ</td>
<td>RRTMG</td>
<td>RRTMG</td>
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<td>RRTMG</td>
</tr>
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</tr>
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<td>RRTMG</td>
<td>RRTMG</td>
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<tr>
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<td>Update</td>
<td></td>
<td>RRTMG</td>
<td>RRTMG</td>
</tr>
<tr>
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<td>Constant</td>
<td>Constant</td>
<td></td>
<td>RRTMG</td>
<td>RRTMG</td>
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<tr>
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<td></td>
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<td>RRTMG</td>
<td>RRTMG</td>
</tr>
<tr>
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<tr>
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<tr>
<td>SAS2 SST</td>
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<tr>
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<td>RRTMG</td>
<td>RRTMG</td>
</tr>
<tr>
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<td></td>
<td>RRTMG</td>
<td>RRTMG</td>
</tr>
</tbody>
</table>

The main differences between the simulations is the cumulus, planetary boundary layer, and radiation schemes. The simulations were chosen to have different physics configurations in order to compare if the changes associated with the type of IC, LBC, or BTC (which will be refer as SST from now on) are of the same order as those by the physic choices. Per each unique physics configuration 4 simulations were made by using
different IC, LBC, or BTC modalities: (1) the control simulation: LBC(t,x,y,z) and SST(x,y), (2) LBC (x,y,z) and SST(x,y), (3) LBC(x,y,z) and SST(t,x,y), and (4) LBC(t,x,y,z) and SST(t,x,y).

The control simulation was chosen to be LBC(t,x,y,z) and SST(x,y), since in this study we consider LBC(t,x,y,z) and SST(t,x,y) to be the wrong approach when reproducing MJO events. Other studies have shown that most of the simulations in which the BTC (SSTs) are updated present a clear, eastward propagating MJO structure. However, as was stated previously, we argue that this method is wrong. This because different investigations show the clear influence of the MJO on the ocean, particularly the cooling of SSTs due to cloudiness and fresh water input. The passage of the MJO over the ocean can clearly be detected by looking at the SST changes during that period. In numerical simulations, when the BTC are updated at a particular Δt, the precipitation will develop by following the positive SST anomalies to the east of the MJO convective center in the Indian Ocean, instead of to the west as has been shown in observations. We argue that because the MJO signal is in the SSTs, the precipitation will trigger and follow the positive anomalies with an eastward propagation. Therefore, the precipitation associated with the MJO will be over the warmer SSTs, rather than behind it (Figure 6.1).
Figure 6.1. Schematic of the precipitation and SST anomalies (SST') relationship in observations (a and b) and numerical simulations where the SSTs are updated (c and d). \( t_0=0 \) and \( t_1=t_0+\Delta t \).

The quantification of the MJO is made by the calculation of MJO tracks, following steps in Chapter 3. The tracks are calculated for the precipitation and U850. The decoupling of wind and precipitation will be measured by using the MJO tracking method for the winds and the precipitation. However, instead of comparing against reanalysis, the comparison will be done between simulations. This way, we can obtain insight about how fast or strong the changes in the coupling are between simulations. This method is a gross quantification of the MJO.

In order to study in depth the evolution of the simulations with respect to the control simulation, we calculate the root mean square difference (RMSD) with respect to the control (CTRL) simulation. For example, the root mean square difference of a
variable \(a = a(x, y, p, t)\) over an area \(A\) of \(M \cdot N\) grid points in the zonal and meridional direction for a simulation SIM will be:

\[
\text{RMSD}_{t_k}^{a,\text{SIM}} = \sqrt{\frac{1}{M \cdot N} \sum_{i=0}^{q} \sum_{j=r}^{s} (a_{ijtk}^{\text{SIM}} - a_{ijtk}^{\text{CTRL}})^2}
\]

(6.1)

where \((i=0, j=r)\) and \((i=q, j=s)\) are the start and end coordinates of the area of interest \(A\). The difference (RMSD), then may be RMSD=RMSD(t,k) (as in Equation 6.1), RMSD=RMSD(x,t,k), RMSD=RMSD(y,t,k), or RMSD=RMSD(k) depending on the spatial evolution that one is interested in\(^{18}\). The areas where the difference will be calculated are the totality of the domain (IOMC), the Indian Ocean (IO), and the Maritime Continent (MC). These regions are the same as the specified in Chapter 4 (see Figure 4.1).

The organization will be measured by the evolution of the spectral density of different zonal scales. The 2-dimentional \((x,y)\) Fourier analysis in different selected areas at specific times will be calculated. This will give insight to the organization of the convection and other quantities in different MJO simulations and the pattern of propagation of the precipitation and moisture (up-scaling vs down-scaling).

6.3. MJO Precipitation and Winds

Similarly to Chapter 5, the success or failure of a simulation was obtained by the closeness of each simulations to the TRMM (precipitation) and ERAI (U850) values of the amplitude and speed values from applying the tracking method (see chapter 3). In the case of the precipitation, the track amplitudes were less than half of the TRMM amplitude (5.54 mmday\(^{-1}\) see Fa and Table 6.2). The majority of the speeds range between

\(^{18}\) The sums in Equation 6.1 will depend on the space and time evolution that one wants to study.
3 and 5 m s\(^{-1}\), for all CTRL simulations with the exception of ZMPB1 and ZMPB2 (orange and pink asterisk in Figure 6.2). The track speed in these simulations is larger than 18 m s\(^{-1}\), which indicates that there is not a dominant eastward moving feature during the simulation. If the criteria mentioned in Chapter 3 section 3.2.1 is applied, none of the simulations listed in Table 6.1 reproduces the precipitation pattern associated to the MJO_DY case. Moreover, if the results from the tracking method are normalized by the climatological values and compared with the results from applying the method to TRMM and ERA data, it is clear that none of the simulations is within the climatological standard deviation circle for precipitation (see Figure 6.3a). This is true for the two longitudinal ranges: L=40°E-160°E and L=50°E-160°E tested (see Figure 6.4a).

**Figure 6.2.** Scatter diagram of amplitude vs speed of precipitation (a) and U850 (b). Each color represents a group of simulations (see Table 6.1). Each shape represent CTRL (asterisk), SST updates (square), CBC (triangle), or SST CBC (circle) simulations. Gray symbols are the mean of the different types of simulations (for example, the gray asterisk is the mean of all CTRL simulations values). Squares around the TRMM/ERAi symbol (star) are 1 standard deviation from the climatological values of the amplitude and speed at the same longitudinal range L=40°E -160°E.
TABLE 6.2. MJO tracking results for the simulations listed in Table 6.1. The longitudinal range is L=40°E-160°E. OBS indicate the results from TRMM or ERAI for precipitation or wind. TRMM and ERAI data was regrided to the simulations’s grid resolution (50Km).

![Table Image]

Figure 6.3. Scatter diagram of normalized amplitude vs normalized speed of precipitation (a) and U850 (b). Each color represents a group of simulations (see Table 6.1). Each shape represent CTRL (asterisk), SST updates (square), CBC (triangle), or
SST CBC (circle) simulations. Gray symbols are the mean of the different types of simulations (for example, the gray asterisk is the mean of all CTRL simulations values). Circles around the TRMM/ERAI or CLIM symbols (star and plus sign, respectively) are 1 standard deviation from the climatological values of the amplitude and speed at the same longitudinal range L=40°E-160°E.

In the case of the zonal winds at 850hPa, the simulations have more realistic values, especially in amplitude (Figure 6.3b). However, for some simulations, there is not clear propagation of the signal (ZMPB2 CTRL, and TK1CTRL). From all 28 simulations, only ZMPB2 CBC (pink triangle on Figure 6.3b) is within the climatological values of ERAI U850 (see Figure 6.4b). This is also true for when L=50°E-160°E (Figure 6.4b). In this case, when the longitudinal range (L, See chapter 3) shortens from 40°E-160°E to 50°E-160°E, ZMPB2 SSTCBC also classifies as a successful U850 MJO simulation (pink circle, Figure 6.4b). ZMPB2 SST (ZMPB2 SSTCBC) have a speed of 8.0 (7.5) ms\(^{-1}\) in precipitation, however, the magnitude of this moving feature is not large enough to be considered an MJO. In this case, the decoupling between convection and dynamics within numerical simulations can be seen.

If the tracking method results are group by cumulus and planetary boundary schemes, it is clear that both precipitation and wind behave differently depending on the type of scheme used. Figure 6.5 shows the MJO tracking results for amplitude, speed, and scale\(^{20}\) for all the simulations. In this case, these figures allow one to compare BC (lateral and bottom) as well as physics configuration. Nonetheless, neither of the simulations in

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\(^{19}\) The of the tracking method are shown for L=40°E-160°E (Figure 6.3) and L=50°E-160°E (Figure 6.4). We consider L=40°E-160°E to be the best tracking range, since L is the largest.

\(^{20}\) The scales is obtained based on the mean number of consecutive positive anomalies and the grid size.
Table 6.1 satisfies $D<1$ for both precipitation and U850, the results from the tracking method highlight some common behaviors between simulations:

The mean amplitude and speed values of the CTRL, SST, and SST CBC simulations (gray asterisk, triangle and circle in Figure 6.3b), are within the $D=1$ circle (square in Figure 6.2) in U850. Nevertheless the number of members is small (7). This also emphasizes the importance of diversity in physics within ensemble systems.

**Figure 6.4.** Same as Figure 6.3 but for $L=50^\circ$E-160$^\circ$E.
**Figure 6.5.** Precipitation (left) and U850 (right) amplitude (a and b), speed (c and d), scale (e and f), and start day (g and h) for 28 simulations (see Table 6.1). Each color represents a group of simulations (see Table 6.1). Each shape represents CTRL (asterisk), SST updates (square), CBC (triangle), or SST CBC (filled circle) simulations. Gray symbols are the mean of the different types of simulations (for example, the gray asterisk is the mean of all CTRL simulations values). Horizontal black solid (dashed) lines are the TRMM/ERAi (climatology) values per each variable calculated at the same longitudinal range L=40°E -160°E (see Table 6.2).

For almost all physics except SAS2 (SAS2 and SAS1), the differences in precipitation (U850) amplitude between simulations with the same physics are smaller
than those between simulations with the same BC and different physics (see Table 6.3). This implies that the precipitation associated with the MJO changes the most by changing the physics rather than the BC on a within group of simulations.

Major differences in precipitation amplitude are linked to those simulations on which SST are updated and have different physics configurations (0.71 mm day$^{-1}$). This corroborates that different physics schemes will act differently with similar changes in the surface conditions.

The variations in precipitation speed between simulations with same physics and different BC and those with different physics are similar. It is worthy to note that SAS simulations (SAS1 and SAS2) have the smallest differences in precipitation speed, which makes their result more consistent between BC configurations. This result is also similar for the precipitation amplitude on SAS1. However, the standard deviation between the simulations with different BC using SAS2 is the largest of all the physics configuration (0.46 mm day$^{-1}$, Table 6.3).

For all simulations with the exception of SAS2 (469.35 Km) and ZMPB2 (273.10Km), the differences in the zonal scale of the most prominent east-moving precipitation feature is smaller in simulations with the same physics but different BC than those with different physics and the same BC (Table 6.3). This means that the scale-pattern of the precipitation is likely defined by the physics rather than the BC. This is true for the precipitation amplitudes as well (with the exception of SAS2).

**Table 6.3.** Standard deviation of the amplitude, speed and scale obtained from the tracking method (Chapter 3) for precipitation (left 3 columns) and U850 (right, 3 columns) for each group of simulations. The cell's fill portion is proportional to the order
of the cells (vertically), where a full colored cell represents the highest value of each column.

<table>
<thead>
<tr>
<th></th>
<th>Precipitation</th>
<th></th>
<th></th>
<th></th>
<th>Zonal Wind Speed 850hPa</th>
</tr>
</thead>
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<tr>
<td></td>
<td>Amplitude</td>
<td>Speed</td>
<td>Scale</td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td>(mm/day⁻¹)</td>
<td>(ms⁻¹)</td>
<td>(Km)</td>
<td></td>
<td></td>
</tr>
<tr>
<td>TK1 (CTRL, SST, CBC, SSTCBC)</td>
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<td>204.32</td>
<td>0.48</td>
<td>0.43</td>
</tr>
<tr>
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<td>250.91</td>
<td>0.31</td>
<td>8.25</td>
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<td>0.75</td>
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<td>273.10</td>
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<td>405.33</td>
<td>0.97</td>
<td>6.94</td>
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<td>0.71</td>
<td>6.96</td>
<td>455.05</td>
<td>0.94</td>
<td>5.64</td>
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<td>381.14</td>
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<td>3.62</td>
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<td>2.81</td>
<td>3540.05</td>
<td>0.80</td>
<td>3.66</td>
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<tr>
<td>CLIM</td>
<td>0.61</td>
<td>7.34</td>
<td>405.33</td>
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<tr>
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<td>CLIM</td>
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<td>2.81</td>
<td>3540.05</td>
<td>0.80</td>
<td>3.66</td>
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Neither changes in the physics nor the BC configuration have greater deviations than the climatological values of the precipitation amplitude and zonal scale (1.73 mm/day⁻¹ and 3540.05 Km, respectively). However, none of the simulations are close to the precipitation scale in TRMM (13050 Km, See Table 6.2). From the zonal scales calculations, one can conclude that the lack of the MJO signal is due to the lack of convective organization in the large scale, an improvement of which should ultimately improve the amplitude. Larger amplitudes in the precipitation are linked to larger zonal scales (TK2 triangle in Figure 6.5.a, b, and c). The improvements of ZMPB2 CTRL are linked to the increase of zonal scale. It is also worthy to note that the representation of the
large scales in the wind patterns appears to not be as limited as in the precipitation (Figure 6.5f).

For SAS and ZMPB, the precipitation amplitude increases (decreases) from the CTRL when SST are updated (BC are kept constant, CBC). TK2 and TK3 simulations show the same behavior when SSTs are updated (asterisk to square)\(^{21}\). The ensemble mean of the simulations’ BC groups (gray symbols in Figure 6.5) does not show any clear change. Most of the simulations have small increases or decreases, which are still not enough to reach the strength of the climatology or the TRMM value for this MJO event (dashed and solid black lines in Figure 6.5a, respectively).

Most of the simulations in which there is some eastward propagating precipitation feature\(^{22}\), with the exception of ZMPB1 and ZMPB2, show an increase\(^{23}\) (decrease) of the precipitation speed as the SST are updated (BC are kept constant, CBC). The change between CTRL and the simulations where the SSTs are updated and the BC are kept constant (CBC) is not clear. However, in the TK simulations SSTCBC, there is an increase in the precipitation speed.

Regardless that ZMPB1 and ZMPB2 show unrealistic values of the precipitation speed in CTRL (orange and pink asterisk in Figure 6.5b), there are more realistic values in ZMPB2 SST, ZMPB1 and 2 CBC, and ZMPB1, and ZMPB2 SSTCBC. Note that there is not an eastward moving precipitation feature in ZMPB1 CTRL nor ZMPB1 SST. However, this appears once the BC are kept constant: ZMPB1 CBC and ZMPB1 SSTCBC (orange triangle and circle in Figure 6.5c).

\(^{21}\) TK1 is not included in this statement since there is no clear eastward moving feature (speed is 20ms\(^{-1}\)). see Figure 6.5 b.
\(^{22}\) Speed less than 20ms\(^{-1}\).
\(^{23}\) In SAS1 it remains the same, see Table 6.2.
By updating the SSTs in the numerical simulations, the precipitation will start earlier. The start day of the precipitation measured using TRMM is day 7 (day 1 is November 10, 2011), while in the simulations the mean start of CTRL simulations is day 18, whereas in SST simulations it is day 5 (see Table 6.2). When the tracking method is applied to the SST anomalies with the same resolution, over the same period of time, and for the same longitudinal extent as the TRMM, ERAI, and the WRF simulations, the results are: 0.20°C amplitude, 5.5ms\(^{-1}\) speed, 3.0 start day, and 2260 Km zonal scale (see Figure 6.6). As mentioned in the previous paragraphs, when SSTs are updated, the start day decreases to values closer to the SST start day (day 5 for SST versus day 18 for CTRL, See Table 6.2). The precipitation speed tends to adjust to the SSTs speed, however, the biggest change is observed in the time that the precipitation is triggered. This supports our idea of the precipitation to be triggered over the SSTs (see Figure 6.1), and the wrong in numerous modeling studies.

**Figure 6.6.** Time-Longitude diagram of SST anomalies (°C). The black line is the best track resulting from applying the tracking method to the SST anomalies data. This track has 0.20°C amplitude, 5.5ms\(^{-1}\) speed, 3.0 start day, and 2260 Km of scale. Day 1 is Nov 10, 2011.
The winds speed and the wind amplitude do not show a clear trend when we look at the different BC updates (see Figure 6.5). The zonal scale of the wind anomalies of those simulations with realistic (close to ERAI MJO) speeds is a different story. The zonal scale of the main U850 east-moving feature increases (from the CTRL) when the SSTs are updated (TK2, TK3, SAS1, SAS2). The U850 amplitude also seems to decrease, with the exception of SAS2.

The simulations where the SSTs were updated appears to show unrealistically fast eastward moving waves. There is however, one special case: ZMPB2. In this case CTRL does not get any MJO wind signal, however, in ZMPB2: SSTs, CBS and SSTCBC both the amplitude and the speed improves, to the point of being within the climatological values for ZMPB2 SST. These improvements are also associated with an improvement in the zonal scale of the east moving feature.

6.4. Evolution of the Differences Between Simulations

The evolution of the differences between simulations with different BC will be accomplished by looking at a group of simulations with the same physics. First, we are going to focus on simulations TK1 (see Table 6.1). Figure 6.7 shows the evolution of the RMSD for the IO and MC region for TK1. When the SSTs are updated, the water vapor field starts to change after the first 6h period. These changes first occur closer to the surface, and as the interaction time increases they penetrate to upper levels. From the separation between lines in Figure 6.7a, c and e, it is clear that the change of difference in time is greater in the first days, and rapidly saturates. However, in the IO the change in differences remains the same at the end of the simulation, while in the case of the MC
there is surprisingly a decrease of difference in the last 5 to 10 days (see Figure 6.7.b). This means that the differences over the Ocean are rapidly growing and leave no room for improvement, whereas over land the saturation occurs more slowly. These differences occur in all TK simulations (TK1 and TK2) in the humidity fields.

**Figure 6.7.** Domain averaged vertical of water vapor mixing ratio (gKg\(^{-1}\)) RMSD in the IO (left) and MC (right) for TK1 (a and b) SST-CTRL (c and d) CBC-CTRL, and (e anf
f) SSTCBC-CTRL (right). Colors represent different times. The black line is the time mean.

In CBC simulations, the magnitude of the differences is larger than in SST during the first hours. This highlights the rapid propagation of these boundary differences to the interior of the domain over the ocean (Figure 6.7c) and land (Figure 6.7d). Both IO and MC domains are located close to 20 degrees (~2000 Km) away from the boundaries.

Interestingly, the differences within the first 150hPa are greater on CBC than SST in IO (Figure 6.7c), which also indicates the effect of the LBC close to the surface and the vulnerability of the local humidity in the numerical model to the large scale patterns (information in the LBC). Similarly to SST, the differences over land for CBC seems to monotonically increase up to day 40 and then decrease with the final forecast times (see Figure 6.7d).

SST CBC differences seem to be dominated by the CBC more than SST. Similarly to CBC, the differences in the surface humidity are larger than in SST in the IO. The MC also shows greater moisture differences closer to the surface; however, this occurs during the last hours of the simulation. Differences decrease at both the middle and upper levels, but most drastically at the mid levels. This decrease in the differences will reach values close to the first 10 days in the last 5 to 10 days of the simulations (see Figure 6.7f, red vs blue lines). While it is not known exactly why the differences decrease at later times, they are at a maximum when the actual MJO event occurred and decrease once the MJO has moved east. Despite the fact that none of these simulations successfully simulate an MJO during this time period, the BC are trying to communicate information to the RM domain that an MJO exists when they do not have one, leading to
particularly large differences. This can also be observed in the differences between zonal wind, especially in CBC simulations (see Figure 6.8 c and d). In this case the major differences between CTRL and CBC is the middle levels during the MJO period in the IO (days 15 to 20), which later decrease. Similar progression in the difference patterns can be observed for relative humidity (Figure 6.9), temperature (Figure 6.10), and omega (Figure 6.11).

The differences with respect to CTRL for temperature show that the rate of change at which the difference decreases is greater over the land (MC) vs the ocean (IO) (see Figure 6.10). This in the CBC (Figure 6.10c and d) and SSTCBC (Figure 6.10e and f). The RMSD in omega are interesting, since this variable characterize the large scale vertical transport and the heating. Based on these differences it is clear that the modality of the LBC setup (time dependent vs independent) affects the precipitation over land and ocean. However, it seems that the IO omega profile shape changes more than in the MC. In the IO there is an increase in the transport in the low and mid levels (Compare Figure 6.11 a and c), this is enhanced in CBC SST.

Figures 6.7 to 6.11 give a glance on how the difference changes in general. However, it is important to study where within the domain the differences are occurring, and how this would evolve. In order to do this we look at the location of the maximum top 1% of this differences throughout the simulation. We focus on the location of the maximum difference in precipitable water, rain, and wind. These three variables were chosen because both the precipitable water (PW) and wind (U850) are updated in the domain LBC, however, the precipitation is a product of the model physics.

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24 The top 1% of the total domain corresponds to the 234 greatest values, since the domain has 266x88 grid points a every vertical level, and we are considering only three dimensional variables.
**Figure 6.8.** Same as Figure 6.7 but for Zonal Wind RMSD.

**Figure 6.9.** Same as Figure 6.7 but for Relative Humidity RMSD.
**Figure 6.10.** Same as Figure 6.7 but for Temperature RMSD.

**Figure 6.11.** Same as Figure 6.7 but for Omega RMSD.
It is clear from Figure that for SST most of the differences are located in the south, close to the IO in the first 20 days, and move to the east, as time evolves. It is not completely surprising that the greatest differences are over the Ocean when CTRL and SST are compared. CBC differences are greater in the east-west and north-south boundaries at the beginning of the simulation, and later they penetrate more inland (from the north and form the west), especially from the IO.

**Figure 6.12.** Top 1% (maximum differences) precipitable water difference distribution locations in latitude (row 1) and longitude (row 2), and RMSD distribution (row 3) for SST (a), CBC (b), and CBCSST (c) for TK1.

Similar results are observed in CBC SST, once more, emphasizing the effect of the LBC in the simulation results. Particularly, differences in precipitable water are maximized with respect to control in the IO where SST changes (see Figure 6.6). In CBC and CBCSST, the differences are maximized closer to North/South boundaries where LBC differs from the control simulation the most. The value of the differences between SST, CBC, and CBC SST tends to increase with time, as it can be seen from their histogram (Figure 6.12, row 3), this is especially noticeable in SST. That said, the value of the differences in CBC and CBC SST are greater than in SST, even during the first 10
days of the simulation. This highlights the important role of the LBC on regional modeling. In this case, this seem to be more dominant than the bottom conditions.

The geographical distribution of the wind differences (Figure 6.13) show a different picture than those in precipitable water. In this case, it seems that greatest differences are shifted to the northeast in SST (southeast in CBC and CBC SST). Differences in U850 show more of a bimodal distribution in the IO and WP, especially CBC and CBC SST. It is also important to notice that the distribution of differences in water vapor mixing ratio (Figure 6.12) show a much different distribution than those in U850. This is interesting, since, both fields (wind and water vapor) are updated within the boundary conditions (lateral and bottom) directly. The greatest differences in precipitable water are located at the edges of the domain whereas in wind the are mostly inside the domain in both latitude and longitude (Figure 6.13). This is more evident in CBC and CBC SST.

**Figure 6.13.** Same as Figure 6.12 but for Zonal Wind at 850hPa (U850).

Similarly to the precipitable water differences, the values of the U850 increase during the simulation period. However, it seems that the values are noticeable greater
around day 30 in CBC and CBC SST, and then there is a decrease in those values, as it has been pointed out previously.

**Figure 6.14.** Same as Figure 6.12 but for precipitation.

Despite that the differences in precipitable water are maximized near the north and south boundary, the precipitation differences are maximized close to the domain center (see Figure 6.14). In this case this is because the value of the precipitation is linked to the physics rather than the LBC. However, the displacement of the precipitable water and the rain maximum differences locations confirm the fact that the errors or the differences in the BTC and LBC propagate faster on parameterized processes. This is especially evident in when the LBC configuration changes (Figure 6.14b and c). The differences are also noticeable when the mean water vapor mixing ratio and the precipitation are compared (see Appendix F).

The value of the precipitation differences among simulations, similarly to other variables, seems to increase with time during the simulation. However, in this case the increase is monotonic. For example if the rain histograms in Figure 6.14 are compared with the precipitable water histograms in Figure 6.12 it is clear that the differences in the precipitation increase during the simulations, whereas in the precipitable water there is a
decrease on the value of the differences close to day 30 (red bars versus green bars in Figures 6.12 and 6.14, row 3).

6.5. Precipitation Organization

The precipitation organization is measured by the changes in the zonal and meridional spectrum during the simulation. In CTRL the precipitation organizes mostly north of the equator in zonal scales between domain wavenumber 1 and 2, which corresponds to 13300Km and 6650Km. SST simulations seem to follow similar meridional organization, with the main difference being increased large-scale precipitation south of the equator (Figure 6.15). This follows the idea that the precipitation will tend to organize by following SST changes when the bottom conditions are updated. However, when the LBCs are constant the precipitation will organize to larger zonal scales, almost like a double ITCZ (but not MJO). Now most of the precipitation moves south of the domain and there is an increase of organization of the large scales (Figure 6.16). There is also more high wavenumber / scattered precipitation near the boundary with CBC, further from CTRL, especially north of the equator (Figure 6.16).

By updating the BTC and by keeping constant the LBC appears to organize the convection into larger scales than in CTRL (Figure 6.16). CTRL, however shows a localized maximum in domain zonal wavenumber 3 (~4433 Km) not present in SST, CBC, nor CBC SST.
**Figure 6.15.** Meridional distribution of time mean zonal spectrum for CTRL (a), SST (b), CBC (c), and CBC SST (d) TK1 simulations. The wavenumber is relative to the domain size: ~13300Km in the zonal direction and ~4400Km in the meridional direction.

**Figure 6.16.** Time and meridional mean zonal spectrum for CTRL, SST, CBC, and CBC SST TK1 simulations.
Most of the convective activity is located in the IO and WP regions (Figure 6.17). By updating SSTs, the large scale convection tends to move to the east, especially east of the west Pacific. The large scale convection activity in the WP decreases with respect to CTRL. In the case of the CBC and CBC SST, the effect of the time independent LBC appears to dominate over the SST changes in terms on how the convection organizes (Figure 6.18). If we look at the changes in the IO versus the MC (Figure 6.18) it is clear that updating SSTs helps organize the precipitation into larger scales. Moreover, domain meridional wavenumber 2 and 3 (~2200Km and ~1466Km) increase between 1 to 2% between CTRL and SST (Figure 6.18) in the IO region.

**Figure 6.17.** Zonal distribution of time mean meridional spectrum for CTRL (a), SST (b), CBC (c), and CBC SST (d) TK1 simulations. The wavenumber is relative to the domain size: ~13300Km in the zonal direction and ~4400Km in the meridional direction.
6.6. Chapter Summary

The precipitation amplitude is more determined by physics (cumulus scheme chosen) rather than the boundary conditions. The greatest differences in precipitation amplitude are in those simulations in which SSTs updated with different physics configurations, which indicates the difference responses between different closures to BTC. The precipitation speed changes are associated with both BC and physics. Differently from the precipitation amplitude, which can more or less be determined (in terms of strong or weak precipitation) by the cumulus scheme chosen rather than to the BC; the speed of the most prominent eastward-moving feature seems to be the result of a combination of factors.

The scale of the most prominent eastward-moving feature does not reach the climatological nor TRMM value (13 times smaller on average). This highlights the lack of organization in the large scale in all the numerical simulations. This scale seems to be...
determined by physics rather than the BC. The effect of the bottom conditions or BTC in
the precipitation is to trigger convection earlier. In simulations where the SSTs are
updated the convection started on average 4 days earlier than in simulations where the
bottom conditions remain constant (day 3 vs day 7). The solution to this premature
triggering of convection is to use a fully coupled ocean-atmosphere regional model
setting rather than just update the SSTs. This will ensure a more realistic behavior of the
ocean and the atmosphere as a whole system.

It is also important to mention that the simulation with best results in precipitation
is our version of ZMPB scheme (ZMPB2). This scheme is used in the CAM model. The
code for this scheme was provided by Dr. Guang Zhang at the Scripps Institution of
Oceanography, UC San Diego and it was implemented in version 3.3 of WRF during the
development of this research. The main difference between ZMPB (original form WRF)
and ZMPB2 is the dependence of the latter to the large-scale moisture (see Zhang (2002)
and description in Chapter 2).

In the case of the wind, similarly to the precipitation, the wind amplitude and
scale are more sensitive to model physics than BCs or SSTs. Further, the response to
different BC appear similar between similar schemes. For example, TK and SAS show a
decrease in the precipitation amplitude in CBC simulations. Another important result is
that the speed of the most prominent eastward feature is too high in many simulations –
almost a “band” of westerly wind develops over entire MJO region then rapidly
dissipates.

The evolution of the differences between simulations with similar physics initially
occur closer to the surface, and as simulation time increases they penetrate to upper
The change of the differences in time is greatest in the first two weeks, then saturates beyond week 3. In the IO the change in the differences remains the same at the end of the simulation, while in the case of the MC there is surprisingly a decrease in the difference values in the last 10 to 15 days. The greatest differences occur between days 25-35. This is when the real MJO is over MC. BCs in control are trying to communicate that there is an MJO while the RM does not have one. This leads to greater differences between control and CBC simulations (see Figure 6.7).

CBC simulations show a faster difference growth in water vapor mixing ratio in the first 10 days of the simulation (Figure 6.19). The differences are greater early on in the IO, greater later on in the MC region. The change in LBC leads to more rapid/early changes from the control than change in SST. Updating SST alone cannot compensate the water vapor mixing ratio differences due to constant boundary conditions (CBC).

**Figure 6.19.** Vertical profile of Water Vapor Mixing ratio RMSD Growth ($\frac{D}{Dt}(RMSD)$) in the IO (a) and MC (b) for TK1. The tick solid lines are for SST, the thin solid line are for CBC, and the dashed lines are for CBC-SST. Each color represents an interval of time during the simulation period. Units are in gKg^{-1}6h^{-1}. 
Unlike for water vapor, CBC and CBC with updated SST are not similar for wind after day 10 (see Figure 6.20). Updated SST appears to constrain error in winds (starting from surface and working upwards) given CBC, but does not for water vapor mixing ratio. This could be because differences in water vapor get amplified by physics and updating the SSTs cannot compensate for early divergence from the control simulation.

**Figure 6.20.** Same as Figure 6.19 but for Zonal Wind.

The differences from the physics are greater than the information carried out by the LBC, IC, and BTC’s, and the model biases. The physics differences from the water vapor mixing ratio and temperature are larger than the differences associated with the winds and their respective ensemble mean. The different location of the maximum differences between precipitation and precipitable water (see Figure 6.21) indicate that the parameterized process will carry the errors or differences faster than the model dynamics will carry the information from the boundary conditions. This is important
because it emphasizes the importance and the non-linearity between the process that are being parameterized within the model. It may not be trivial, but errors within the boundary will carry out to the interior of the domain by the small scale processes that are being parameterized. Therefore the importance of evaluating and the danger of just choosing any scheme without further investigating the most fitting scheme for the problem that needs/wants to be solved must be emphasized.

Zonal wavenumbers vary more than meridional wavenumbers for precipitation (and also for water vapor mixing ratio and wind). The changes in wavenumber 1 from run to run considerably. Updating SST increases large-scale organization but decreases small-scale organization with respect to CTRL. CBC and CBC SST tend to also increase large scale organization, however, this is more localized in certain regions. The time evolution of the spatial spectrum is associated with the physics schemes which have the greatest differences, and the growth of the spectral density values start from the small scales and gradually propagate to the large scales. This will have to hold for the majority of the cumulus physics used.
**FIGURE 6.21.** Location of the top 1% greatest precipitable water (left) and precipitation (right) differences between CTRL and SST (a and b), CTRL and CBC (c and d), and CTRL and CBC SST (e and f). Each symbol and color represent a period of time during the simulation.
Chapter 7

Conclusions

This study has been, in simplest terms, an exhaustive exercise in trying to successfully simulate several real MJO cases using many configurations of the WRF model, most of which were quite unsuccessful. However, we have managed to develop a number of useful tools for quantifying and analyzing the MJO, and comparing the MJO signal in an array of varying data sets. Understanding of the Indian Ocean / Maritime Continent water cycle has also been expanded upon, as has the understanding of the roles of different physical quantities (wind, moisture, temperature) at varying spatial scales in initiating and maintaining the MJO. Lastly, as boundary conditions are believed to be a significant source of error in MJO simulations, the effect and propagation of errors in the boundary conditions has been explored.

The first contribution of this work is the development of a new simple technique in order to quantify MJO simulations. This was done by tracking the most prominent eastward moving feature in each simulation and calculating its magnitude, propagation speed, and start day. In chapter 3, we show how this method can be used as a way of obtaining the MJO climatological variations in precipitation and wind propagation speed between different MJO cases. This information later was used in order to discriminate between successful and unsuccessful MJO simulations. The tracking method was also
used to measure the forecast skill at different lead times for a particular GCM deterministic and ensemble forecasts.

During the development of this work we have learned that there is not a parameterization recipe for success when one tries to reproduce an MJO event. However, we have shown that the main processes that can greatly affect the success or not of an MJO simulation usually depends upon how the convective processes are parameterized (including the interaction between hydrometeors, microphysics) through the entire atmospheric column (including the surface). Additionally, in the transition between the surface to low levels to middle levels, coupling plays an important role, especially in correctly resolving the moisture and during the first stages of the MJO initiation.

It has been shown that a successful MJO simulation is linked to the mean behavior of the numerical model environment, especially the water vapor. We found that there is no single type of parameterization scheme that can be identified as the main source of the biases in the simulated water cycle. It is the combination of errors from the parameterized cumulus, PBL, and microphysics processes that allow the simulated water cycle deviate from that in YOTC.

A clear example of the impact of the choice of cumulus and planetary boundary layer schemes in shown in Figure 7.1. This Figure shows the moisture flux added within 3 different layers in the vertical: 1000 to 900, 900 to 700, and 700 to 500 hPa. The effect of how the cumulus processes are parameterized is going to affect the direction and the strength of the circulation pattern between simulations, especially over the ocean. Where the moisture transport and availability are linked to the surface (Figure 7.1a). In the MC region, the direction of the moisture flux is the same in all the chosen boundaries. In the
IO however, 3 schemes (TK-MYJ, NAS-MYJ, and NSAS-YSU) show erroneous circulation in the west boundary of the Indian Ocean. In this case they show moisture leaving the IO through the west boundary. Closer to the surface the planetary boundary layer scheme seems to dominate the direction of the moisture flux (Figure 7.1a).

There is an over moistening linked to the south boundary in the IO region between 900 and 700 hPa. This over moistening from the south seems to be compensated by a lack of moisture transport from the North boundary in the IO (Figure 7.1b). The MYJ boundary layer is linked to loss of moisture flux from the west IO boundary. This loss of moisture is present in the simulations with the greatest dry biases. In the case of the MC, there is a lack of moisture from the north MC in all the simulations shown in Figure 7.1b. Lastly, the erroneous circulation (see Figure ) can be seen in Figure 7.1c. In this case all the numerical simulations except TK-YSU show disagreement with YOTC on both the strength and direction of the moisture flux in the IO. The major disagreement is on the west IO boundary, where the simulations show an easterly flow, whereas, in YOTC the moisture flow is westerly. Therefore, randomly combining off-the-shelf parameterization schemes for simulations in diabatically-dominated regions such as the tropics is dangerous. The risk is that an apparent satisfactory simulation might simply result from cancellation of error from different parameterization schemes and the underlying physics they present is wrong. The tropical water cycle should be correctly simulated only as parameterization schemes of different processes are improved all together.
FIGURE 7.1. Mean, vertically integrated water flux across the IO and MC domains ($\langle F_E \rangle$, $\langle F_W \rangle$, $\langle F_S \rangle$, $\langle F_N \rangle$) integrated from (a) 1000–900 hPa, (b) 900 to 700 hPa, and (c) 700 to 500 hPa (time mean of Equations 4.6a – d). Note that in Equations 4.6 a – 6 the vertical integration range is from $k=1$ to $k_{top}$ and in this case
they change to to k(p=1000) to k(p=900) (a), k(p=900) to k(p=700) (b), and k(p=700) to k(p=500) (c). Each color represents a numerical simulation or reanalysis. The size of the arrows is scaled. The scale is presented in the bottom-right corner of each map.

Capturing the MJO precipitation constitutes a much more difficult task than obtaining the circulation patterns within the MJO. This may be related to the fact that the wind signal is a more large scale structure, which is directly present in the U850, and therefore it is easier for the numerical models to follow such structure. Whereas, in the case of the precipitation, the signal is present in the water vapor and the model needs to do more work in order to accurately trigger the precipitation at a required location and specific time that would constitute a successful MJO simulation.

The reproduction of the planetary scales of moisture during the regional MJO simulation are key to obtained a successful MJO simulation, especially zonal wave number 1, 2, and 3. The errors associated with the physical parameterizations in the RM dominate the simulation, up to the point where the information provided by the LBC, IC, and BTC is no longer useful. The errors in the physics as a whole conglomerate affects the mean climate during the simulation, and this limits the capacity of reproduce MJO events. The decoupling in the RM then occurs mainly because the “dynamic” MJO information is translated by the lateral boundary conditions more efficiently than the thermodynamic information. The RMs dynamic errors are overshadowed by the frequent update of the LBC whereas the humidity field would adjust faster to the model’s error, since the moisture and temperature tendencies feedback into the momentum Equations.

In this study, we also attempted to correct the errors in unsuccessful channel simulations of the MJO through nudging. It was found that the deepening of the low-
level moist layer leading to the MJO convective center is achieved primarily by water vapor perturbations of planetary scales \((k = 1 - 3)\), while synoptic-scale perturbations improve the simulation but are not essential. It was also shown that MJO precipitation is much more difficult to reproduce than its circulation (e.g. U850). Therefore, evaluating MJO simulations with respect to their precipitation depiction is a higher standard than simply evaluating based on winds alone.

Two reasons why the simulated MJO circulation is more realistic than the modeled precipitation are proposed. The first is that deficient cumulus parameterization schemes may work against MJO dynamics set by initial and lateral boundary conditions that help produce MJO signals in the circulation in a regional model (Ray and Zhang, 2010; Ray et al., 2009). When a reduction in errors in precipitation by water vapor nudging is insufficient to make simulated precipitation realistic, it might be sufficient to minimize the negative effects to allow the simulated circulation to be realistic under the influence of MJO dynamics provided by the lateral boundary conditions. The second reason is a degree of decoupling between the MJO circulation and precipitation. While this may sound counterintuitive based on the perception that MJO is a product of convection-circulation interaction (Rui and Wang, 1990), it may not be so surprising if one accepts the possibility that MJO dynamics are determined principally by its intrinsic structure rather than the accompanying convection. The MJO circulation appears to be determined by such an MJO dynamic structure, not solely by MJO precipitation. There has been some support for this idea in the recent literature (Ling et al., 2014; Matthews, 2008; Straub, 2013). Similarly, it was shown that nudging of MJO signals in humidity during the February-March 2009 period can fake phantom MJO signals in precipitation.
but not in the circulation (in NoMJOBQ). The newly-proposed method of spectral nudging of water vapor and a new MJO diagnostic metric that quantifies errors in simulations of individual MJO events highlight the critical role of planetary-scale (zonal wavenumber 1 – 3) perturbations of water vapor in the MJO and demonstrate possible decoupling of MJO convection and circulation.

Finally, the role and limitations of the boundary conditions in MJO simulations was investigated. It was found that precipitation magnitude as well as spatial pattern is more sensitive to changes in physics than changes in boundary conditions, both lateral and bottom. Major differences in precipitation amplitude are linked to those simulations in which SST are updated and have different physics configurations, corroborating with the fact that different physics schemes will behave differently with similar changes in the surface conditions. Additionally, a lower variability amongst the simulations than the climatological variability suggests some predictability throughout the simulated MJO timescales. However, these simulations would be most useful if bias-corrected, as all significantly under-predict precipitation magnitude.

The evolutions of errors and its dependence upon boundary conditions was also examined. When SSTs are updated, these changes are first felt in the lowest levels of the atmosphere, then gradually propagate to the upper levels over a period of a few days. Once errors over the IO saturate, they remain fairly steady. On the other hand, errors over the MC fluctuate with time, with errors at a maximum when the true MJO fails to be simulated, and then decrease thereafter. This is perhaps due at least in part to the BCs trying to communicate information to the regional domain that an MJO exists when it doesn’t have one, leading to “confusion” in the model that enhances errors during this
timeframe, then decreases once the BCs are no longer communicating information about
a present MJO.

The differences in LBC and BTC are spread by the parameterized processes to the
interior of the domain. This is done by the moisture and temperature tendencies which
feedback into the large scale. The differences from the physics are greater than the
information carried out by the LBC, IC, and BTC’s, and the model biases in humidity and
temperature. This is observed especially over the Ocean.

The poor representation of the MJO in a regional model setting will more likely to
continue to be a problem, even if LBC, IC and BTC are improved upon model initiation
or as they are updated in time. The key to solve the problem is in the model physics.
Perhaps we need to erase the board and start over.
Appendix A: Tracking Method Results for TRMM MJO Cases Listed in Table 3.1
**Figure A.1.** Time-Longitude (left column) and Time-Speed (3 right-most columns) of TRMM precipitation anomalies (same as Figure 3.2a) and tracks magnitude (same as...
Figure 3.2b) in mmday⁻¹ for MJO cases 1 to 5 (rows 1 to 5). The track magnitudes for different Ls: 40°E-160°W, 40°E-160°E, 50°E-160°E are shown in columns 2, 3, and 4 respectively. The start day (horizontal gray line), speed (vertical gray line), and magnitude (black circle) of the track with the greatest magnitude for each L are highlighted on columns 2, 3, and 4. The track with the greatest magnitude for each L are also highlighted in the time-longitude diagram: 40°E-160°W (black line), 40°E-160°E (dashed black line), and 50°E-160°E (dashed gray line). See Table for details on the MJO cases.
FIGURE A.2. Same as Figure A.1 but for cases 6 to 10.
**Figure A.3.** Same as Figure A.1 but for cases 11 to 15.
FIGURE A.4. Same as Figure A.1 but for cases 16 to 20.
Figure A.5. Same as Figure A.1 but for cases 21, 22, and 23.
Appendix B: Tracking Method Results for MJO Cases Listed in Table 3.1.
FIGURE B. 1. Time - Longitude (left column) and Time-Speed (3 right-most columns) of U850 anomalies (same as Figure 3.3a) and tracks magnitude (same as Figure 3.3b) in ms⁻¹
for MJO cases 1 to 5 (rows 1 to 5). The track magnitudes for different Ls: 40°E-160°W, 40°E-160°E, 50°E-160°E are shown in columns 2, 3, and 4 respectively. The start day (horizontal gray line), speed (vertical gray line), and magnitude (black circle) of the track with the greatest magnitude for each L are highlighted on columns 2, 3, and 4. The track with the greatest magnitude for each L are also highlighted in the time-longitude diagram: 40°E-160°W (black line), 40°E-160°E (dashed black line), and 50°E-160°E (dashed gray line). See Table for details on the MJO cases.
Figure B.2. Same as Figure B.1 but for cases 6 to 10.
Figure B.3. Same as Figure B.1 but for cases 11 to 15.
FIGURE B.4. Same as Figure B.1 but for cases 16 to 20.
FIGURE B.5. Same as Figure B.1 but for cases 21, 22, and 23.
Appendix C: Tracking Method Results for ECDY Deterministic Forecast
**Figure C.1.** Same as Figure 3.2 (top) and Figure 3.3 (bottom) for ECMWF deterministic forecast at lead time = 1 day.
**Figure C.2.** Same as Figure C.1 but for lead time = 2 days.
FIGURE C.3. Same as Figure C.1 but for lead time = 3 days.
FIGURE C.4. Same as Figure C.1 but for lead time = 4 day.
**Figure C.5.** Same as Figure C.1 but for lead time = 5 days.
Figure C. 6. Same as Figure C.1 but for lead time = 6 days.
FIGURE C.7. Same as Figure C.1 but for lead time = 7 days.
**Figure C.8.** Same as Figure C.1 but for lead time = 8 days.
Figure C.9. Same as Figure C.1 but for lead time = 9 days.
Figure C.10. Same as Figure C.1 but for lead time = 10 days.
Appendix D: Finding MJO Cases Using the Tracking Method with OLR Data

Twenty-nine years of OLR data were used to try to identify MJO events using the tracking method. The annual cycle and interannual anomalies (pervious 120 days) were removed from the 29 years of data. The mean of the anomalies over 10S –10N was obtained and a 5 day running mean filter was applied to them. The tracks were drawn in the longitude-time diagrams by follow the methodology explained in chapter 3:

(i) S number of tracks were drawn for each day, where S corresponds to the slope of the track. The slopes range from 3 to 20 ms-1 on 0.5 ms-1 increments (34 tracks with different slopes, see Figure 3.1)

(ii) Each track has a magnitude assign to them (see Equation 3.1). However, since the negative anomalies of OLR are associated with convection, the magnitude of each track is based on the addition of the negative anomalies:

(iii) The longitudinal extent used was 40°E – 160 °W.

The 3 steps described above were for each day of the 29 year period. The distinction between possible MJO cases was done by first selecting the track with the minimum (since negative OLR anomalies are related to convection) amplitude for each
day. If the magnitude of the minimum track in the following 5 days is also negative, the
third day of the 5 days is considered a possible MJO day.

The MJO periods were obtained by reducing the possible MJO days to those days
in which the track number of consecutive negative anomalies is analogous to half L,
where for this case L=40°E – 160°W. Additionally, a period was considered an MJO
period if there where 3 or more continuous possible MJO days. The list MJO events is
presented in Table D.1. Histograms of MJO amplitude and speed are shown in Figures
D.1 and D.2. Figures D.3 to D.17 show the time-longitude diagram of OLR anomalies
and the time-speed diagram of track amplitudes.
**TABLE D.1.** MJO cases found by applying the tracking method to OLR data. Dates highlighted with gray colors are those in which the real multivariate MJO index (RMM) has a magnitude greater than 1.

<table>
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<th>Start day</th>
<th>Amplitude Wm(^2)</th>
<th>Speed ms(^{-1})</th>
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<td>-9.48</td>
<td>6.00</td>
</tr>
<tr>
<td>3 11/8/86</td>
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<td>5.50</td>
</tr>
<tr>
<td>4 12/9/90</td>
<td>-11.94</td>
<td>10.50</td>
</tr>
<tr>
<td>5 9/2/95</td>
<td>-16.18</td>
<td>9.00</td>
</tr>
<tr>
<td>6 9/28/00</td>
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<td>4.50</td>
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<tr>
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<td>7.00</td>
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<tr>
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<td>8.00</td>
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<tr>
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<tr>
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<tr>
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<tr>
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<td>45 10/14/66</td>
<td>-13.27</td>
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**Figure D. 1.** Histogram of MJO tracking amplitudes.

**Figure D. 2.** Histogram of MJO speed.
**Figure D. 3.** Time-Longitude diagram (a and b) of OLR anomalies (Wm$^{-2}$) and Time-Speed diagram of track amplitudes (Wm$^{-2}$) (c and d) for years 1980 (a and b) and 1981 (c and d). The straight solid black line in the time-longitude diagrams correspond to the MJO event found by applying the tracking method to the OLR data. The white stars in the time-speed diagrams highlight the MJO case start day and speed in the time-speed diagram of the MJO track which best describes the MJO event.
Figure D.4. Same as Figure D.3 but for 1982 and 1983.
**Figure D. 5.** Same as Figure D.3 but for 1984 and 1985.
FIGURE D. 6. Same as Figure D.3 but for 1986 and 1987.
**Figure D. 7.** Same as Figure D.3 but for 1988 and 1989.
**Figure D.8.** Same as Figure D.3 but for 1990 and 1991.
Figure D. 9. Same as Figure D.3 but for 1992 and 1993.
**Figure D. 10.** Same as Figure D.3 but for 1994 and 1995.
FIGURE D. 11. Same as Figure D.3 but for 1996 and 1997.
**Figure D. 12.** Same as Figure D.3 but for 1998 and 1999.
**FIGURE D. 13.** Same as Figure D.3 but for 2000 and 2001.
Figure D. 14. Same as Figure D.3 but for 2002 and 2003.
FIGURE D. 15. Same as Figure D.3 but for 2004 and 2005.
Figure D.16. Same as Figure D.3 but for 2006 and 2007.
Figure D. 17. Same as Figure D.3 but for 2008 and 2009.
**Appendix E: MJO-09 Reanalysis and Analysis Time-Longitude Diagrams**

**Figure E.1.** Time – Longitude diagram of precipitation anomalies (mm day$^{-1}$) for reanalysis and analysis used during MJO – 09 case: TRMM (a), YOTC (b), MERRA (c), and CRSF (d). The vertical black lines indicate the west and east boundaries of the IO and MC domains.
Appendix F: Mean Water Vapor Mixing Ratio and Rain from TK1

**FIGURE F.1.** Time mean water vapor mixing ratio (gKg$^{-1}$) at 850 hPa for CTRL (a), SST (b), CBC (c), and CBC SST (d). The blue (gray) contours are the time mean of cloud (rain) water mixing ratio. The arrows are the time mean wind direction at 850 hPa.
**FIGURE F. 2.** Time mean precipitation (mm day$^{-1}$) for CTRL (a), SST (b), CBC (c), and CBC SST (d). The arrows are the time mean wind direction at 850hPa.
References


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