Pathways to Better Prediction of the Madden-Julian Oscillation Over the Indian Ocean

Ajda Savarin

University of Miami, ajdas1@gmail.com

Follow this and additional works at: http://scholarlyrepository.miami.edu/oa_theses

Recommended Citation
http://scholarlyrepository.miami.edu/oa_theses/638

This Open access is brought to you for free and open access by the Electronic Theses and Dissertations at Scholarly Repository. It has been accepted for inclusion in Open Access Theses by an authorized administrator of Scholarly Repository. For more information, please contact repository.library@miami.edu.
PATHWAYS TO BETTER PREDICTION OF THE MADDEN-JULIAN OSCILLATION
OVER THE INDIAN OCEAN

By

Ajda Savarin

A THESIS

Submitted to the Faculty
of the University of Miami
in partial fulfillment of the requirements for
the degree of Master of Science

Coral Gables, Florida

December 2016
A thesis submitted in partial fulfillment of
the requirements for the degree of
Master of Science

PATHWAYS TO BETTER PREDICTION OF THE MADDEN-JULIAN OSCILLATION
OVER THE INDIAN OCEAN

Ajda Savarin

Approved:

Shuyi S. Chen, Ph.D.
Professor of Ocean Sciences

Chidong Zhang, Ph.D.
Professor of Ocean Sciences

Arnold L. Gordon, Ph.D.
Professor of Oceanography,
Department of Earth and
Environmental Sciences,
Columbia University

Guillermo Prado, Ph.D.
Dean of the Graduate School
SAVARIN, AJDA (M.S., Meteorology and Physical Oceanography)

Pathways to Better Prediction of the Madden-Julian Oscillation Over the Indian Ocean

(December 2016)

Abstract of a thesis at the University of Miami.

Thesis supervised by Professor Shuyi S. Chen.
No. of pages in text: (86)

The Madden-Julian Oscillation (MJO) is the leading source of predictability on seasonal and subseasonal scales in the tropics, and is one of the least understood phenomena of tropical meteorology. Although extensive research has been done on the topic of MJO initiation and its eastward propagation, there is not a single widely-accepted theory that explains the phenomenon. The lack of understanding is reflected in the poor representation of the MJO in global and numerical weather prediction models. In this study, a regional, atmosphere-ocean coupled model is used to perform a series of high- and low-resolution experiments in coupled and uncoupled configurations to address the effects of moist physics, resolution, and atmosphere-ocean coupling on the simulation of MJO. As a case study, we use the second MJO event (MJO2) observed during the Dynamics of the MJO (DYNAMO) field experiment (October 2011 - March 2012). MJO2 is currently the best observed MJO event on record, and the copious amount of data collected by multiple observational platforms is used for model evaluation.

We find that the MJO in the coupled model run at a relatively low resolution (12 km grid spacing) is very sensitive to the choice of convective parameterization, which does not only affect the amount and distribution of precipitation, but also influences the vertical structure of winds and relative humidity in the atmospheric boundary layer. Two convective parameterizations, with different triggering mechanisms for convection are considered, with one producing tropical cyclones, and the second one simulating an MJO, though not accurately representing individual features. When resolution is increased to a convection-permitting
grid spacing (4 km resolution) in the coupled configuration, the representation of individual convective features improves regardless of the convective parameterization outside the high-resolution domain. A high precipitation bias is present in all experiments and can be linked to a high bias in the surface-layer air-sea flux parameterization, and a positive bias in the ocean mixed layer depth. Reducing the air-sea flux bias through a modification of the buoyancy-driven turbulence parameterization for air-sea fluxes succeeds at reducing the precipitation bias and improves (weakens) the surface winds, and enhances the representation of MJO’s eastward propagation. With weaker surface winds, the ocean cooling is reduced, which slightly offsets the introduced modification. The high-resolution uncoupled experiment (atmosphere only) does not produce an MJO - it produces excessive precipitation that is present throughout the 15-day simulation and extends eastward from the Indian Ocean, but never propagates out of it. The study concludes that high (cloud-permitting) resolution is integral in accurately representing the precipitation features associated with the MJO, and the MJO-induced upper-ocean cooling in atmosphere-ocean coupled experiments is essential for the MJO’s eastward propagation.
Acknowledgements

I would like to thank my advisor, Dr. Shuyi Chen, for giving me the opportunity to work on this topic. She sparked my interest in tropical meteorology, provided help and guidance along the way, and supported my travel addiction by enabling me to present my research at scientific conferences across the United States. I also want to thank my committee members, Dr. Chidong Zhang and Dr. Arnold L. Gordon, for their insights into this work and helpful comments during the thesis writing process.

I would also like to thank all past and present members of the ‘Hurricanes and Coupled Atmosphere-Ocean Systems’ research group, especially Chia-Ying Lee, Falko Judt, Milan Curcic, Rachel Zelinsky, and Brandon Kerns for welcoming me into the world of graduate school and providing valuable suggestions during our traditional Friday-afternoon group meetings.

Finally, I would like to thank my family, my friends, and Philippe, for their support and encouragement. To my family, who provided me with countless opportunities that have led me here and valiantly supported my migration across the Atlantic Ocean. To my friends, for making Miami my home-away-from-home, and for all the adventures we shared. To Philippe, for his love, companionship, support, and long discussions about life, science, and puppies.

This research has been supported by grants from the National Oceanographic and Atmospheric Administration (NOAA, under the grant NA11OAR4310077), the National Science Foundation (NSF, under the grant AGS1062242), and the Office of Naval Research (ONR, under the grant N000141110562).
## Contents

**List of Figures** vi

**List of Tables** ix

### 1 Introduction

1.1 Motivation .................................................. 1
1.2 Background and Literature Review .......................... 3
   1.2.1 The MJO .............................................. 3
   1.2.2 Challenges in MJO Modeling .......................... 6
1.3 Science Objectives and Working Hypothesis ................. 8
1.4 Outline ...................................................... 8

### 2 Methodology

2.1 Observational Datasets ..................................... 10
   2.1.1 DYNAMO Field Observations .......................... 10
   2.1.2 Satellite Data ......................................... 12
   2.1.3 Global Model Analysis Products ....................... 13
2.2 An MJO Event Observed During DYNAMO ..................... 13
2.3 Numerical Model ........................................... 15
2.4 Model Experiments .......................................... 17

### 3 Effect of Model Resolution and Moist Physics on the MJO

3.1 MJO Precipitation and Surface Winds ....................... 21
3.2 Boundary Layer - Free Troposphere Interaction ............ 27
3.3 Precipitation Bias .......................................... 35
   3.3.1 Rainfall Rate Distribution .......................... 38
   3.3.2 Moisture Convergence ................................ 40
3.4 Summary .................................................... 42

### 4 Atmosphere-Ocean Coupling and the MJO

4.1 SST Cooling and MJO Eastward Propagation ................. 46
4.2 Air-Sea Fluxes and Their Parameterization ................. 51
   4.2.1 Bulk Air-Sea Fluxes and Turbulence Parameterization . 53
   4.2.2 Wind Speed, Temperature, and Moisture Disequilibria . 56
   4.2.3 Effect on Precipitation .............................. 60
4.3 Upper Ocean Structure ..................................... 63
   4.3.1 Initial State of the Upper Ocean in the Coupled Model . 64
   4.3.2 Evolution of the Upper Ocean During the MJO ........ 68
4.4 Summary .................................................... 73
5 Conclusions 75
  5.1 Summary of Key Results ........................................ 75
  5.2 Future Work .................................................... 79

References 80
List of Figures

1.1 Adapted from Madden and Julian (1972), Figure 16. Schematic depiction of time and space variations of the disturbance associated with the 40-50-day oscillation in the zonal (equatorial) plane. Light blue lines at the bottom of each panel show the mean pressure disturbance, with negative anomalies shaded in red, and positive anomalies shaded in blue. Circulation cells are based on the mean zonal wind disturbance, and highlighted in red, with arrows showing the direction of circulation. Regions of enhanced large-scale convection are indicated schematically by cloud symbols. The relative tropopause height is indicated in dark blue at the top of each panel. 4

2.1 Adapted from Yoneyama et al. (2013), Figure 1. Observational network for the DYNAMO field campaign. Dashed lines indicate the northern and southern sounding arrays, with the sites marked as Ma (Malé, Maldives), C (Colombo, Sri Lanka), G (Gan Island), R (R/V Revelle), D (Diego García, Addu Atoll), and Mi (R/V Mirai). 11

2.2 Hovmöller diagrams of 5°S-5°N averaged a) TRMM rainfall rate (in mm h⁻¹); b) ECMWF analysis surface zonal winds (in m s⁻¹); and c) OI SST (in °C). Time is increasing upward. Brown shading on the bottom of each panel indicates the maximum height of topography in the averaging region. Black contours in b) and c) outline the area over which the rainfall rate is ≥ 1 mm h⁻¹. 14

2.3 WRF nested domain configuration with grid resolutions of 36-, 12-, and 4-km over D01, D02, and D03, respectively. Colors indicate HYCOM SST at time of model initiation. Quadrilaterals in the IO show the DYNAMO sounding array observation sites, which are labeled as Ma (Malé, Maldives), C (Colombo, Sri Lanka), G (Gan Island), R (R/V Revelle), D (Diego García, Addu Atoll), and Mi (R/V Mirai). 16

3.1 5°S-5°N-averaged Hovmöller diagrams of rainfall rate (mm h⁻¹; a-e), surface zonal winds (m s⁻¹; f-j) and SST (°C; k-o) for observations (top) and model experiments. SST panels only include the IO area. Brown shading on the bottom panels indicates the height of maximum topography within the averaging region. 25

3.2 LPT evolution of the precipitation field for a) TRMM observations, b) AO12-KF, c) AO12-TK, d) AO4-KF, and e) AO4-TK. Colored contours outline the area where the accumulated daily precipitation amount exceeds 13 mm day⁻¹, and is sustained for at least three days, with the color indicating time. 26
3.3 A large equatorial convective event with multiple MCSs observed during the active phase of the MJO on November 24, 2011 (adapted from Chen et al. (2016), Figure 6). a) TPW (mm, color contour) and GridSat IR brightness temperature (grey, K), overlaid with the WP-3D aircraft track (black line). Circle markers show locations of released dropsondes. White circles mark the dropsonde profiles used in constructing the b) vertical cross-section of wind speed in m s\(^{-1}\) (color) with horizontal wind direction indicated by wind barbs (magnitude in kt). Black outline and shaded area indicate a RH \(\geq 85\%\), and white circle markers in b) indicate the top of the BL.

3.4 As in Figure 3.3 b, but for model experiments: a) AO12-KF; b) AO12-TK; c) AO4-KF; and d) AO4-TK. Black contours and opaque grey shading indicate areas where the RH \(\geq 85\%\), with the white dots indicating the top of the BL. e-h) show near-surface profiles of \(\theta_v\) (left) and RH (right) from R/V Revelle sounding observations (blue) and each UWIN-CM experiment (red).

3.5 Observed (left) and model-simulated soundings of RH (a-e) and zonal winds (f-j) taken from DYNAMOs equatorial observational sites Gan, and R/V Revelle. 195 soundings from the two sites are used to create the observed distributions (a, f), and compared to 240 model-simulated soundings. The colored area indicates the % of soundings in each bin. Shaded areas serve to highlight the low-level, mid-level, and upper-level features described in text.

3.6 Time series of rainfall rate averaged between 55 and 97\(^\circ\)E, 10\(^\circ\)S and 10\(^\circ\)N (the portion of the IO covered by D03). Thin lines indicate hourly (for model) and 3-hourly (for TRMM) rainfall rates, while thick lines show rainfall rates smoothed using a 24-hour running mean.

3.7 GridSat IR brightness temperatures showing two snapshots of the MJO2 event: a) MJO convection over the IO (November 26, 2011 at 18 UTC), and b) MJO convection over the MC and suppressed over the IO (December 4, 2011 at 18 UTC).

3.8 Distribution of rainfall rates within D03 for TRMM (black) and model experiments (color). Note that the x-axis is displayed logarithmically.

3.9 Time series of: a) instantaneous moisture convergence into D03; b) cumulative moisture convergence into D03; c) cumulative moisture convergence difference between \(AO4-TK\) and \(AO4-KF\) (blue minus brown); d) cumulative total rainfall; and e) cumulative total rainfall difference between \(AO4-TK\) and \(AO4-KF\) (blue minus brown).

4.1 5\(^\circ\)S-5\(^\circ\)N-averaged Hovmöller diagrams of rainfall rate (mm h\(^{-1}\); a-e), surface zonal winds (m s\(^{-1}\); f-j) and SST (\(^\circ\)C; k-o) for observations (top) and high-resolution coupled and uncoupled model experiments. SST panels only include the IO area. Brown shading on the bottom panels indicates the height of maximum topography within the averaging region.
4.2 LPT evolution of the precipitation field for a) TRMM observations, b) UA4-TK, c) AO4-TK, and d) AO4-TK-FLX. Colored contours outline the area where the accumulated daily precipitation amount exceeds 13 mm day$^{-1}$, and is sustained for at least three days, with the color indicating time. 50

4.3 a-c) Latent and d-f) sensible heat flux distributions from R/V Revelle (yellow), and UA4-TK (a, d), AO4-TK (b, e), and AO4-TK-FLX (c, f). Thick lines show the mean values of the flux at a given wind speed. 52

4.4 Magnitude and distribution of buoyancy-driven turbulence (CV) contribution to wind speed, as a function of wind speed in a) AO4-TK and b) AO4-TK-FLX. 55

4.5 Distributions of a-c) 10 m wind speed, d-f) moisture disequilibria, and g-h) temperature disequilibria between the ocean surface ($q_s$, $T_s$) and the air just above it ($q_a$, $T_a$). R/V Revelle observations are shown with yellow circles, with model distributions in contour for UA4-TK (a, d, g), AO4-TK (b, e, h), AO4-TK-FLX (c, f, i). Thick lines show the mean value at a given wind speed. 57

4.6 Time series of rainfall rate averaged between 55 and 97$^\circ$E, 10$^\circ$S and 10$^\circ$N (the portion of the IO covered by D03). Thin lines indicate hourly (for model) and 3-hourly (for TRMM) rainfall rates, while thick lines are smoothed using a 24-hour running mean. 61

4.7 SST difference between HYCOM and ECMWF global analyses. Red colors indicating that HYCOM SSTs are warmer than those in ECMWF analysis. 66

4.8 a) Temperature (in $^\circ$C), b) salinity (in psu), and c) potential density ($\sigma_0$, in kg m$^{-3}$) profiles from M1 (solid) and M2 (dashed); observations are shown in black, and the corresponding profiles at UWIN-CM initialization time are in red. 67

4.9 Distribution of UWIN-CM salinity with depth within 2$^\circ$ of M1 and M2. 68

4.10 Time evolution of temperature (in $^\circ$C; a-c), salinity (in psu; d-f), and potential density ($\sigma_0$ in kg m$^{-3}$; g-i) at M1. Mooring observations are shown on the left (a, d, g), followed by AO4-TK (b, e, h) and AO4-TK-FLX (c, f, i). Black dots indicate the mixed layer depth. 69

4.11 Frequency distributions of a-c) temperature (in $^\circ$C), d-f) salinity (in psu), and g-i) potential density ($\sigma_0$ in kg m$^{-3}$) combining profiles from M1 and M2. Mooring observations are shown on the left (a, d, g), followed by AO4-TK (b, e, h) and AO4-TK-FLX (c, f, i). The black lines show the range of the 10$^{th}$ and 90$^{th}$ percentile of observed values at a given depth. 72
List of Tables

2.1 Experiments used in this study. Coupling refers to atmosphere-ocean coupling. Resolution lists grid resolutions of all nested domains in the atmosphere component of UWIN-CM. CP marks the convective parameterization used. The modification column describes the way a model adjustment is implemented relative to the unmodified experiment. 19

3.1 Rainfall rates averaged over the IO portion of D03 (as in Figure 3.6). Total column shows average rainfall rate for the entire time series, while the Active and Suppressed columns refer to the time periods between November 22 and 30 (active) and December 1 and 6 (suppressed). The Difference column shows the difference between the average rainfall rates during the active and suppressed phases. 36

3.2 Same as Table 3.1, but showing rainfall rate bias compared to TRMM (in mm h$^{-1}$) followed by the % overestimation of TRMM rainfall rates in parentheses. A 100% overestimation indicates an average rainfall rate that is twice as high as in TRMM. 38

4.1 LHF and SHF biases compared to direct flux measurements from R/V Revelle. Values are given both for the full distribution and for the low winds only ($\leq$ 5 m s$^{-1}$). Biases for both TK and KF experiments are included for comparison. 53

4.2 Rainfall rates averaged over the IO portion of D03 (as in Figure 4.6). Total column shows the average for the entire time series, while the Active and Suppressed columns refer to the time periods between November 22 and 30 (active) and December 1 and 6 (suppressed). The Difference column shows the difference between the average rainfall rates during the active and suppressed phases. 62

4.3 Same as Table 4.2, but for rainfall rate bias in mm h$^{-1}$ followed by the % overestimation of TRMM rainfall rates in parentheses. A 100% rainfall rate bias indicates that model rainfall rates are twice as high as observed. 63

4.4 Ocean mixed layer depth (m) and the average bias in mixed layer depth given in m and % of observed value for different thresholds of density change from the surface. 71
List of Abbreviations

- AXBT - airborne expendable bathy-thermograph
- AXCTD - airborne expendable conductivity-temperature-depth probe
- BL - atmospheric boundary layer
- CE - convective envelope
- CP - convective parameterization
- DYNAMO - Dynamics of the Madden-Julian Oscillation
- ECMWF - European Centre for Medium-Range Weather Forecasts
- HYCOM - HYbrid Coordinate Ocean Model
- IO - Indian Ocean
- LHF - latent heat flux
- LPT - large-scale precipitation tracking
- MC - Maritime Continent
- MJO - Madden-Julian Oscillation
- MJO2 - second MJO event during DYNAMO
- MLD - mixed layer depth
- SHF - sensible heat flux
- SST - sea surface temperature
- TPW - total precipitable water
• TRMM - Tropical Rainfall Measurement Mission

• UWIN-CM - Unified Wave Interface - a Coupled Model

• WRF - Weather Research and Forecasting model
Chapter 1

Introduction

1.1 Motivation

The Madden-Julian Oscillation (MJO) is a large-scale atmospheric disturbance affecting the global tropics that was first described by Madden and Julian (1971). It is characterized by a slow eastward-propagating coupling between precipitation and large-scale circulation. Intense precipitation (strong convection) associated with the MJO typically initiate and builds over the Indian Ocean (IO), then moves eastward across the Maritime Continent (MC) and into the western Pacific. The large area of active deep convection (the active phase of the MJO) is surrounded to its east and west by relatively suppressed convection (the suppressed phase of the MJO). Near the surface, the active MJO is accompanied by anomalously strong westerly winds to its west, and anomalously strong easterlies to its east. Thus, at any one location, the passage of the MJO results in alternating phases of active and suppressed convection, with a shift between anomalously strong easterly and westerly winds (Zhang 2005). The MJO is a global phenomenon with a time period of 30-90 days, and it is the dominant mode of intraseasonal variability in the tropics. The changes in precipitation and atmospheric circulation due to the MJO impact both tropical and extratropical weather patterns on many different time scales (Zhang 2013).

In the tropics, it has been linked to the modulation of tropical cyclone formation in all ocean basins (Maloney and Harmann 2000), the onset and break phases of monsoons (Jones and Carvalho 2002), and even to the development of El Niño in the tropical Pacific Ocean (McPhaden et al. 2006). Outside the tropics, it has been linked to extreme weather events over North America, such as flooding, drought, and cold air outbreaks (Jones 2000; Bond and Vecchi 2003; Vecchi and Bond 2004).
Even though the global impacts of the MJO are well-established, there is no unified theory that could simultaneously explain its convective initiation, eastward propagation, variability, and structure. This lack of thorough understanding of the MJOs mechanisms and multi-scale interactions is reflected in the poor representation of MJO in numerical weather prediction and climate models. It is one of the leading sources of predictability on seasonal-to-subseasonal timescales, but most state-of-the-art climate models have difficulty representing it, which renders forecasts of future climate less reliable. Besides climate models, weather models also have a difficult time predicting and simulating MJO events.

Part of the reason why the study of the MJO has not advanced as far as that of other meteorological phenomena, such as tropical cyclones, is the lack of direct observations, especially those of MJO initiation. As it occurs over the IO, there are few long-term data records for atmospheric data, and even fewer for the ocean. But there has been a lot of research and progress in recent years, owing to an organized effort of four collaborative field campaigns: CINDY, DYNAMO, AMIE, and LASP (Cooperative IO Investigation on Intraseasonal Variability in the Year of 2011; Dynamics of the Madden-Julian Oscillation; ARM MJO Investigation Experiment, and Littoral Air-Sea Processes, respectively). With the goal of improving the understanding of key MJO initiation processes and model simulation and prediction of the MJO, many observations were collected over the IO and MC from land-, water-, and air-based platforms. Four MJO events were observed during the DYNAMO observation period between October 2011 and March 2012, and since then, the collected observations have inspired a multitude of observational and modeling studies.

In this study, we focus on trying to improve the understanding of some physical processes that are necessary for a model to represent in order for it to be capable of producing an MJO. In particular, we analyze the second MJO event that occurred during DYNAMO, which is so far the most well-observed MJO event in history, with land-based, ship, buoy, mooring, and aircraft observations all available to analyze and evaluate the models results. Using an atmosphere-ocean coupled model in a regional setting allows us to compare the
MJO simulations at different resolutions, ranging from low-resolution experiments with different convective parameterizations (common in most global and climate models) and from here on abbreviated as CPs, to high-resolution experiments in which convection can be resolved explicitly. In addition, the use of an atmosphere-ocean coupled model allows us to conduct coupled experiments, where the ocean responds to atmospheric forcing, and uncoupled experiments, in which the sea surface temperature (SST) is held constant. Since all of these experiments are simulating a real MJO event, the processes and areas in which the model is performing poorly can be identified by direct and statistical comparison to observations.

1.2 Background and Literature Review

1.2.1 The MJO

Though a lot of progress has been made in studying and understanding the MJO, the conceptual model of MJO evolution, initially proposed by Madden and Julian (1972), is still fundamentally valid today, though it serves more as an illustration of the phenomenon, and not as an explanation. That is an impressive accomplishment, as all they used to deduce the pattern was 5-10 years of station pressure and zonal winds at 850- and 150-hPa data from stations located in tropical regions around the world. The schematic of their conceptual model, though slightly adapted, is shown in Figure 1.1. The MJO initiates with a negative pressure anomaly over Africa and the IO, where convection begins to organize at or near the equator (phase 1). The pressure anomalies then amplify and spread over the entire IO, and deep convection cells develop in phase 2. Two anomalous circulation cells arise on each side of the active convection - surface westerlies at the surface with easterlies aloft to its west (behind active convection), and surface easterlies with westerlies aloft to its east (ahead of it). The cells result in low-level convergence and upper-level divergence at the location of active MJO convection, which begins to propagate eastward.
Figure 1.1: Adapted from Madden and Julian (1972), Figure 16. Schematic depiction of time and space variations of the disturbance associated with the 40-50-day oscillation in the zonal (equatorial) plane. Light blue lines at the bottom of each panel show the mean pressure disturbance, with negative anomalies shaded in red, and positive anomalies shaded in blue. Circulation cells are based on the mean zonal wind disturbance, and highlighted in red, with arrows showing the direction of circulation. Regions of enhanced large-scale convection are indicated schematically by cloud symbols. The relative tropopause height is indicated in dark blue at the top of each panel.
As the MJO propagates into the central Pacific, the associated convection weakens, and soon the signal in low-level winds disappears (phases 6, 7, and 8). In the last two phases, the MJO signal is decoupled from precipitation, and is mostly evident in upper-level winds as it travels around the globe (Madden and Julian 1972). The MJO features a multi-scale structure - the eastward-propagating convective envelope consists of multiple mesoscale convective systems that propagate westward, but with each new system forming to the east of the previous one. The average propagation speed of the MJO is $\sim 5 \text{ m s}^{-1}$, though the propagation is slower when the convection and large-scale circulation are coupled together, and faster after the convection dies. The MJO exhibits a strong seasonal cycle with a strong peak in the Northern Hemisphere winter (related to the Australian Monsoon), and a weaker peak in Northern Hemisphere summer (related to the Asian Monsoon; Zhang 2005). As the active MJO propagates across ocean, the associated surface winds, extensive cloud cover, and large amounts of precipitation help induce upper-ocean mixing and cool the ocean surface (de Szoeke et al. 2015; Moum et al. 2014).

Various indices have been developed to track the active MJOs position around the globe. Most rely on statistical analyses of global data and extended time records of anomalies, such as a combination of low- and upper-level zonal wind and outgoing longwave radiation (in the real-time multivariate MJO index - RMM; Wheeler and Hendon 2004), or of a single variable such as the outgoing longwave radiation (the outgoing longwave radiation MJO index OMI; Kiladis et al. 2014). Indices such as these are great for evaluating climate model simulations, with their global data coverage and many year integrations, but cannot be used in regional model simulations due to lack of data coverage. There is not yet a non-global way of tracking the MJO as a whole, including the zonal wind anomalies as well as precipitation, but there is a way of tracking precipitation envelopes produced by the MJO - the large-scale precipitation tracking (LPT; Kerns and Chen 2016). LPT uses spatially smoothed 3-day rainfall accumulations to identify and track precipitation features. The features need to occur on a large scale of at least $3 \cdot 10^5 \text{ km}^2$, exhibit an average eastward
propagation of at least 2 m s\(^{-1}\), and persist for at least 10 days in order to qualify. This is enough to track the convective envelopes of the MJO, and can be used to identify the evolution of individual MJO events in both time and space, which is not possible when using other indices.

Different MJO indices identify different components of the MJO, and are at times inconsistent with each other. This problem arises from the fact that there are many different hypotheses that try to explain the behavior of the MJO, and they often disagree on what constitutes an MJO event. So far, none of the theories are able to simultaneously explain the MJOs selection of (global) spatial scales, its intraseasonal and variable period, and its eastward propagation and speed. Some of the theories depend on the modification of linear equatorial waves by convective heating (wave-CISK and frictional wave-CISK; Lindzen 1974), moisture modes (Sobel and Maloney 2013), scale interaction (Krishnamurti et al. 2016), and air-sea interaction (DeMott et al. 2015). The lack of agreement and understanding of the phenomenon leads to an unsatisfying performance of global and regional models.

### 1.2.2 Challenges in MJO Modeling

Hung et al. (2013) showed that out of 20 state-of-the-art climate models (CMIP5) that were used in the fifth IPCC assessment report (AR5), only one third was able to capture the variability on MJO timescales, and even that was generally weaker than observed. Only one of the twenty models reproduced a realistic eastward propagation. This is an improvement over the previous generation of climate models (CMIP 3), in which precipitation features near the equator tend to persist in the same place for too long, and for which eastward propagation is not present, or is slower than that of the MJO (Lin et al. 2006).

Accurately modeling the precipitation (specifically precipitation associated with the MJO) has long been a challenge to the scientific community, and it is something that both climate and regional models struggle with. This predicament is generally attributed to the CP schemes that are used in models to represent sub-grid scale processes (Gustafson
and Weare 2004). In general, it seems that CPs in which convection is triggered by a process related to low-level moisture convergence seem to perform better than other types of CP (Hung et al. 2013). Other studies have shown that climate models that use super-parameterization, a technique by which a cloud-resolving model is inserted into each grid box of a global model for a better representation of cloud processes, tend to produce better MJOs - suggesting that cloud processes are important for MJO prediction (Randall et al. 2003; Benedict and Randall 2007; Stan et al. 2010). Other studies have shown that atmosphere-ocean coupling in climate models improves the representation of the MJOs eastward propagation, indicating that the changing ocean can modulate MJO activity (Zhang et al. 2006). However, even coupled climate models, such as those described in Hung et al. (2013) have a difficult time simulating an MJO.

Compared to climate models, using regional models to study the MJO has its advantages and disadvantages. Most often, increasing resolution to a grid spacing where convection is resolved explicitly results in an improvement of the MJO simulation, but at a cost of area coverage, and simulation lengths. High resolution experiments over large areas are computationally expensive for long time integrations. The added resolution can be offset by the need to provide initial and lateral boundary conditions that necessarily come from lower-resolution analysis or reanalysis data, which come with biases of their own. But even with these potential issues, regional models show a lot of promise for helping us learn about the MJO because of the possibility of direct comparison to observations, which can help reveal model biases and lead to potential improvements. Similar to results from climate models, regional models show improvement in MJO prediction when moist convection is explicitly resolved (Holloway et al. 2013).

More mixed views are expressed on the role of air-sea interaction. There are studies that claim that global models with cloud-system-resolving, atmosphere-only simulations are perfectly capable of reproducing an MJO (Miura et al. 2007; Liu et al. 2009), expressing the opinion that some MJO events are largely controlled by atmospheric internal
dynamics (Fu et al. 2015). However, the majority of studies indicate that the ocean and its modulation of the atmosphere play an important role and lead to improvements of MJO simulation in climate, and even more so, in regional models, as this study shows. Success has been obtained with implementing both a crude, 6-hourly updating of SSTs (Wang et al. 2015), as well as with simple mixed-layer ocean models and more complicated fully coupled atmosphere-ocean models (Woolnough et al. 2007).

1.3 Science Objectives and Working Hypothesis

The purpose of this study is to examine the effects that model physics and resolution have on the model simulation of an MJO event, with the goal of improving the model skill in simulating an MJO. The study highlights some physical processes that the model should be able to represent accurately in order to simulate an MJO, as well as draw attention to processes that are misrepresented in the model, and that can be improved in the future.

Specifically, we address the effects that different CPs have on the MJO simulation in a coupled model (at lower resolutions), and how explicitly resolved convection (and no CP) at higher resolutions changes or contributes to the MJO simulation. When considering different CPs, we focus on how the CPs interact with the atmospheric boundary layer, and how that interaction can be beneficial or detrimental to the simulation of the MJO. When using explicitly resolved convection at high resolutions, we investigate the effects of atmosphere-ocean coupling on the simulation of the MJO, comparing coupled and uncoupled (atmosphere-only) model experiments, focusing on improving the amount of precipitation that is produced. With a working hypothesis that high resolution, explicitly resolved convection, and interactive atmosphere-ocean coupling all contribute to a successful MJO, we can answer questions about whether these processes are really essential, or simply beneficial, to a simulation.
1.4 Outline

Chapter 1 gives an overview of the current standing in terms of numerical weather prediction of the MJO. We pose questions about the role of moist physics, convective parameterization, and air-sea interaction which address processes that a model should be able to represent in order to improve the simulation of the MJO-associated precipitation and eastward propagation. Our working hypothesis is presented. Chapter 2 outlines the methodology that was used to address the aforementioned intentions of this thesis. The MJO event that is simulated for this study is presented. The atmosphere-ocean coupled model is described, along with the experiments that we conduct in order to address the stated objectives. The data used to compare both model results and observations are described. In Chapter 3, we address the effects of moist physics and resolution on the MJO. Experiments with different CPs at low resolutions, and no CP at high resolutions, are compared to observations. A detailed analysis of the representation of convection and moisture will highlight the advantages and shortcomings of using different CPs, as well as those of explicitly resolving convection at higher resolutions. Chapter 4 addresses the effects of atmosphere-ocean coupling by comparing and contrasting coupled atmosphere-ocean, and uncoupled, atmosphere-only, simulations. Analysis of atmospheric moisture and the effects of the ocean state in the model illustrate the role that atmosphere-ocean coupling plays in the MJO. Chapter 5 concludes this thesis and discusses the implication of the findings presented in this study.
2.1 Observational Datasets

2.1.1 DYNAMO Field Observations

The DYNAMO field campaign in late 2011 and early 2012 observed four MJO events that initiated over the IO and propagated eastward, collecting an unprecedented amount of in-situ observations centered around the MJO. MJO2, the second observed event that occurred in late November and early December of 2011 was chosen as the case study for this investigation, as it coincided with the intense observation period during which many observational platforms were operating together. Figure 2.1, adapted from Yoneyama et al. (2013) shows the locations of DYNAMO observing platforms, with the six main sites located at the corners of two quadrilaterals in the central equatorial IO. The northern sounding array had its vertices in Malé, Maldives (Ma; 4.19°N, 73.53°E), Colombo, Sri Lanka (C; 6.91°N, 79.87°E), Gan Island, Addu Atoll (G; 0.69°S, 73.15°E), and R/V Revelle (R; 0°N, 80.50°E). The southern sounding array included Gan Island and R/V Revelle, as well as Diego García (D; 7.35°S, 72.48°E) and R/V Mirai (Mi; 8.00°S, 80.5°E). The R/Vs Mirai and Revelle were positioned at the indicated locations during the first part of MJO2, leaving for port on December 1, 2011, and December 3, 2011, respectively.

The MJO2 event coincided with the intense observing period of DYNAMO, during which atmosphere soundings were released every three hours at Gan, Diego García and R/Vs Revelle and Mirai, and every six hours at Malé and Colombo. The radiosonde sounding observations provide high-resolution vertical profiles of temperature, moisture, and winds, showing the evolution of those fields as the MJO develops and propagates away from the central IO. At R/Vs Revelle and Mirai (marked with stars in Figure 2.1), direct
covariance flux measurements were recorded in addition to the calculation of bulk air-sea fluxes. Covariance air-sea flux measurements are used to evaluate the model parameterization of air-sea latent and sensible heat fluxes over water.

Three DYNAMO moorings, located at the equator and 79°E, 1.5°S and 79°E, and 9.75°S and 78.5°E (marked with magenta squares in Figure 2.1), are used to examine and evaluate the upper-ocean structure in the coupled model. These moorings provide hourly temperature and salinity profiles, and are used in addition to the moorings in the Research Moored Array for African-Asian-Australian Monsoon Analysis and Prediction (RAMA) mooring array. RAMA moorings also provide temperature and salinity profiles in the upper ocean, though the data record for the period of interest is less comprehensive than that of DYNAMO moorings. At some sites, data is unavailable or intermittent during MJO2; at other sites, available data includes either salinity or temperature, but not both. RAMA moorings are located in the IO and are marked in Figure 2.1 with dark purple triangles, and are used to verify DYNAMO mooring observations, and not for in-depth analysis.
To get a better idea of the spatial extent of some observed features, we make use of data collected by the Lockheed WP-3D Orion aircraft which was operated by the National Oceanographic and Atmospheric Administration (NOAA) and was stationed at Diego García during DYNAMO. The WP-3D flew 12 missions between November 11 and December 13, 2011 which are summarized in (Chen et al. 2016), with six of the flights during the period of interest for this study. The aircraft provided continuous measurements at flight level, and periodically deployed dropsondes and airborne expendable bathy-thermographs and conductivity-temperature-depth probes (AXBTs and AXCTDs). The dropsondes measure vertical profiles from the flight level of the aircraft to the surface, and AXBTs and AXCTDs measure vertical profiles of ocean salinity (AXCTD only) and temperature starting at the ocean surface. At times, the dropsondes were deployed in a straight line to sample specific precipitation features, providing high-resolution vertical cross-sections of temperature, humidity, and winds.

2.1.2 Satellite Data

Satellite data is used to supplement the field observations during DYNAMO because it can provide data over a large area instead of at a single point, but often at a reduced temporal and spatial resolution. The reduced resolution is not a serious drawback when applied to the MJO because of its characteristically slow eastward propagation and large horizontal extent, and satellite data is used to qualitatively evaluate the model simulations in terms of large-scale precipitation, precipitable water, and SST evolution. The Tropical Rainfall Measurement Mission (TRMM) Multi-satellite Precipitation Analysis (TMPA) version 7, also known as TRMM 3B42 (from here on TRMM) provides a 3-hourly rainfall rate dataset at 0.25° spatial resolution (Huffman et al. 2007). It is useful for following the convective envelopes associated with the MJO, and is the basis of the LPT precipitation tracking algorithm (Kerns and Chen 2016). Observations of total precipitable water (TPW) are available hourly from the Morphed Integrated Microwave Imagery at CIMSS (MIMIC-TPW) product at a 0.25° resolution (Wimmers and Velden 2011), where high temporal resolution is a
result of blending swaths from many different satellites. The same is the case when using the Gridded Satellite (GRIDSAT-B1) infra-red brightness temperature (Knapp et al. 2011) that is used to demonstrate the spatial extent of high clouds during the observed MJO event. Satellite observations of SST are only available daily, at a 0.25° resolution, for which we use the Optimally Interpolated microwave SST (OISST) product.

2.1.3 Global Model Analysis Products

The European Centre for Medium-Range Weather Forecasts (ECMWF) global analysis is used to obtain large-scale surface wind fields at a 6-hourly temporal resolution, which is higher than the daily satellite surface wind products. The analysis uses the ECMWF model guidance, but assimilates both satellite and DYNAMO field observations for a more accurate representation of the large-scale environment with good vertical resolution. The data for the period of interest is available at 0.25° resolution and contains 25 vertical levels.

2.2 An MJO Event Observed During DYNAMO

This study focuses on a case study of a single MJO event - MJO2, which was the second MJO event observed during DYNAMO. The motivation behind using this event is the large amount of observations that were collected and are available to evaluate the models performance. Additionally, reproducing a specific MJO event is something that general circulation models (GCMs) often have a very difficult time with (Subramanian and Zhang 2014); instead, GCMs produce statistical composites of MJOs that dont necessarily reflect any individual case (Subramanian et al. 2011).

MJO2 occurred during the boreal winter in late November and early December 2011, and was preceded by an MJO event starting in late October 2011, and directly succeeded by a weak MJO event starting in late December 2011. The Hovmöller diagrams showing 5°S-5°N-averaged rainfall rates, surface zonal winds, and SST are shown in Figure 2.2, with time increasing upward. The fields are shown for fifteen days that correspond with
the model integration period. At the bottom of each panel, brown shading indicates the maximum height of topography within the averaging region, separating the longitudes of the IO (50-95°E) and MC (95-150°E), and emphasizing the height of topographical features of the MC.

Figure 2.2: Hovmöller diagrams of 5°S-5°N averaged a) TRMM rainfall rate (in mm h\(^{-1}\)); b) ECMWF analysis surface zonal winds (in m s\(^{-1}\)); and c) OI SST (in °C). Time is increasing upward. Brown shading on the bottom of each panel indicates the maximum height of topography in the averaging region. Black contours in b) and c) outline the area over which the rainfall rate is ≥ 1 mm h\(^{-1}\).

The eastward propagation of precipitation is separated into two distinct convective envelopes (CEs) that are separated by a local suppression of precipitation near the equator (Figure 2.2a). The first CE starts at around 65°E on November 22 and propagates eastward for 6 days before dissipating shortly after entering the MC. The second CE starts at around 75°E on November 27 and propagates eastward for 9 days before shifting north and off the equator near 130°E. Both CEs are accompanied by zonal surface wind convergence near the leading edge of precipitation, with westerly winds extending across the IO to the west, and easterly winds to the east, as seen in ECMWF analysis data (Figure 2.2b).
Intense precipitation, extensive cloud cover, and strong surface winds induce large-scale upper ocean cooling of about 1.5°C (Figure 2.2c), which is consistent to what de Szoeke et al. (2015) find from in-situ observations. After the MJO2 propagates into the MC, precipitation over the IO is strongly suppressed, and surface westerly winds covering the entire IO. In this period, the upper ocean gradually starts to recover, and SST begins to increase.

### 2.3 Numerical Model

The model used for this study is the Unified Wave Interface a Coupled Model (UWIN-CM; Chen et al. 2013; Chen and Curcic 2015). It is an atmosphere-surface waves-ocean coupled model developed at the University of Miami, and it allows great flexibility in coupling as any of the three components can be combined as desired. For this study, only the atmosphere and ocean components are used, as including surface waves would require a lot more computational resources to execute. The components of UWIN-CM are the Weather Research and Forecasting (WRF version 3.6.1) with the Advanced Research WRF (ARW) dynamical core (Skamarock et al. 2008) for the atmosphere, and the Hybrid Coordinate Ocean Model (HYCOM version 2.2.98; Wallcraft et al. 2009) for the modeling of ocean circulation. The components are coupled together every two minutes, exchanging information about radiation, air-sea fluxes, and SST.

WRF is configured with three nested grids of 36-, 12-, and 4-km horizontal grid spacing for D01, D02, and D03 shown in Figure 2.3. The outermost domain (D01) extends from 16°-174°E and 32°S-32°N, with the inner nests focusing on the equatorial region between 36°-165°E, 15°S-15°N, and 55°-154°E, 10°S-10°N for D02 and D03, respectively. The chosen horizontal configuration allows for coverage of an area that exceeds the region if interest (equatorial IO and MC), assuring that the boundaries of the domain are removed from the area we study. At the same time, the increase in resolution from 36- to 12- and 4-km, focused in the region of interest allows for experiments that require a CP (needed for resolutions of D01 and D02), as well as one where such parameterization is not necessary.
(in D03). There are 36 sigma levels in the vertical, with about 8 levels in the atmospheric boundary layer (BL). The BL parameterization we use is the Yonsei University scheme (YSU; Hong et al. 2006), and the surface layer uses a parameterization based on the Monin-Obukhov similarity theory.

Figure 2.3: WRF nested domain configuration with grid resolutions of 36-, 12-, and 4-km over D01, D02, and D03, respectively. Colors indicate HYCOM SST at time of model initiation. Quadrilaterals in the IO show the DYNAMO sounding array observation sites, which are labeled as Ma (Malé, Maldives), C (Colombo, Sri Lanka), G (Gan Island), R (R/V Revelle), D (Diego Garcíá, Addu Atoll), and Mi (R/V Mirai).

An explicit WRF single-moment microphysics scheme (WSM5; Hong et al. 2004) is used in all domains to resolve grid-size moisture processes, and in D01 and D02, a CP is used in addition to microphysics to represent sub-grid vertical motions. To examine how using different CPs affects the MJO simulation, two different CPs are tested. The Tiedtke CP (Tiedtke 1989) is a version of the CP used in the ECMWF global model, which was shown to outperform the Global Forecast System (GFS) for the MJO2 event during DYNAMO (Kerns and Chen 2014a). To trigger convection, it relies on low-level moisture convergence; schemes with a similar triggering mechanism were found to improve MJO precipitation in climate models (Hung et al. 2013). In the model experiment names, the use
of the Tiedtke CP in D01 and D02 will be indicated by including \textit{TK} in the name. To test a CP that makes use of a different triggering mechanism (and a closure scheme based on the presence of CAPE), we chose the Kain-Fritsch CP (Kain and Fritsch 1990). The use of the Kain-Fritsch CP is indicated by including \textit{KF} in the model experiment name.

HYCOM is configured with a single domain with 32 vertical levels and a grid spacing of 0.08°, which translates to about 8.9 km at the equator and about 7.7 km at the domains northern and southern boundaries. The domain covers an area slightly smaller than that of D01 in WRF, extending from 30 to 172°E, and from 33°S to 33°N. HYCOM is a hydrostatic ocean circulation model with a hybrid vertical coordinate system - the vertical coordinates transition between isopycnal layers (of constant density) in the open ocean, terrain-following layers in shallow coastal regions, and z-level coordinates in the mixed layer. The top three (z-level) layers are centered at depths of 0.5, 2, and 4.8 m. A non-local K-profile scheme is used for the parameterization of vertical mixing (Large et al. 1994).

2.4 Model Experiments

To address the effects of resolution, model physics, and atmosphere-ocean coupling on the MJO simulation in UWIN-CM, a set of eight experiments is designed and executed. We vary the CP used in D01 and D02 to address the effects of different CPs using either the Tiedtke or Kain-Fritsch CPs that have different mechanisms for triggering convection. We vary the resolution by looking at experiments in which the high-resolution D03 is excluded, and ones where it is included; besides the effects of resolution, this also illustrates how the MJO changes when no CP is used. And we vary the atmosphere-ocean coupling by looking at high-resolution experiments that include atmosphere-ocean (from here on referred to as coupled), and atmosphere-only (uncoupled) simulations. In the uncoupled simulations, SST is constant and shown by color shading in Figure 2.3. In the coupled simulations, the depicted SST pattern represents the initial state, but SST evolves in time in response to both ocean dynamics and atmospheric forcing.
As has been shown in previous studies (e.g. Holloway et al. 2013), high-resolution experiments with explicitly resolved convection tend to overproduce the amount of precipitation when compared to observations; this is also the case in high-resolution UWIN-CM experiments. To try and improve on that, we designed an extra experiment in which we modify the parameterization of air-sea latent and sensible heat fluxes by adjusting the parameterization of sub-grid turbulence in the surface layer of WRF. The adjustment is implemented only over water, where we have direct observations of air-sea fluxes for comparison. To adjust the sub-grid turbulence parameterization, we simply halve the magnitude it would have had in the unmodified case; for a given temperature and moisture difference across the air-sea interface, and the same surface wind speed, the latent and sensible heat fluxes would be slightly reduced. The adjustment is described in more detail in Section 4.2.1, and the model experiments with the implemented modification include FLX in their names.

All experiments are displayed in Table 2.1, with their main characteristics and differences listed. The naming convention adheres to the following pattern. The first two letters indicate the state of atmosphere-ocean coupling: UA for an uncoupled, atmosphere-only experiment, and AO for an atmosphere-ocean coupled experiment. Following is a number indicating the highest used grid resolution: 12 for experiments where only the low-resolution D01 and D02 are used with CP, and 4 for experiments that include the high-resolution, cloud-permitting D03. The subsequent letters indicate the choice of CP that was used in D01 and D02: TK for the Tiedtke CP, and KF for the Kain-Fritsch. An appended FLX at the end indicates the modification to the surface sub-grid turbulence parameterization in relation to the calculation of air-sea latent and sensible heat fluxes. For example, AO4-KF is an experiment that includes atmosphere-ocean coupling at high-resolution (incorporating D03), and with the Kain-Fritsch convection parameterization applied over D01 and D02.

All model experiments are initialized at 00Z on November 22, 2011, which coincides with the time that MJO2 begins to organize over the equatorial IO. The simulations are
Table 2.1: Experiments used in this study. Coupling refers to atmosphere-ocean coupling. Resolution lists grid resolutions of all nested domains in the atmosphere component of UWIN-CM. CP marks the convective parameterization used. The modification column describes the way a model adjustment is implemented relative to the unmodified experiment.

<table>
<thead>
<tr>
<th>Experiment</th>
<th>Coupling</th>
<th>Resolution</th>
<th>CP (D01, D02)</th>
<th>Modification</th>
</tr>
</thead>
<tbody>
<tr>
<td>AO12-KF</td>
<td>Yes</td>
<td>36-, 12-km</td>
<td>KF</td>
<td>No</td>
</tr>
<tr>
<td>AO12-TK</td>
<td>Yes</td>
<td>36-, 12-km</td>
<td>TK</td>
<td>No</td>
</tr>
<tr>
<td>AO4-KF</td>
<td>Yes</td>
<td>36-, 12-, 4-km</td>
<td>KF</td>
<td>No</td>
</tr>
<tr>
<td>AO4-TK</td>
<td>Yes</td>
<td>36-, 12-, 4-km</td>
<td>TK</td>
<td>No</td>
</tr>
<tr>
<td>AO4-KF-FLX</td>
<td>Yes</td>
<td>36-, 12-, 4-km</td>
<td>KF</td>
<td>Air-Sea Flux</td>
</tr>
<tr>
<td>AO4-TK-FLX</td>
<td>Yes</td>
<td>36-, 12-, 4-km</td>
<td>TK</td>
<td>Air-Sea Flux</td>
</tr>
<tr>
<td>UA4-KF</td>
<td>No</td>
<td>36-, 12-, 4-km</td>
<td>KF</td>
<td>No</td>
</tr>
<tr>
<td>UA4-TK</td>
<td>No</td>
<td>36-, 12-, 4-km</td>
<td>TK</td>
<td>No</td>
</tr>
</tbody>
</table>

integrated in time for 15 days (360 hours), at which point the MJO2 has propagated most of the way across the MC and is being steered northward and away from the equator. The initial and boundary conditions for the atmosphere come from ECMWF operational analysis, with the boundary conditions are updated every 12 hours throughout the simulation. The ocean and SST initial and boundary conditions come from HYCOM global analysis, with the boundary conditions updated every 24 hours.
Chapter 3

Effect of Model Resolution and Moist Physics on the MJO

Studies evaluating climate models have shown that different CPs can alter the model-simulated MJOs (Hung et al. 2013), linking CPs that directly rely on moisture-convergence to trigger precipitation to produce better MJOs. However, climate models run at very coarse resolutions, with grid spacings of 0.5, 1, or even 2°, and the representation of the MJO is evaluated in a statistical way, and not on a case-to-case basis. It has also been shown that the use of super-parameterization, a process by which convection is resolved explicitly within each grid cell of a climate model, improves the representation of the MJO (Randall et al. 2003; Benedict and Randall 2007). In this chapter, we test whether the same holds when resolution is increased from climate- to regional-model scales. The low resolution experiments are run at a grid spacing of 12 km, which corresponds to \( \sim 0.11° \) at equator, and the high resolution experiments are run at a grid spacing of 4 km (\( \sim 0.036° \) at equator).

At low resolutions, we test the performance of two CP schemes, the Kain-Fritsch in \textit{AO12-KF}, and the Tiedtke in \textit{AO12-TK}, with this being the only difference between the two experiments. Both CPs are mass flux schemes, which means the amount of precipitation produced is based on the calculation of the mass that is entering and leaving a cloud. They both include the parameterization of both shallow and deep convection. The KF scheme’s convective trigger function relies on adding a temperature perturbation to a parcel of air closest to the surface; if the addition of the temperature perturbation creates a buoyant parcel of air that is able to ascend to its lifting condensation level, convection is triggered. The KF schemes closure relies on the removal of convective available potential energy, or CAPE. If convection is present, the vertical temperature and moisture profiles will be modified by an adjustment of updraft and downdraft velocities until 90\% of CAPE is removed.
(Kain 2004). In the TK scheme, deep convection is triggered when there is sufficient low-level environmental convergence, or when the environmental moisture is greater than 90%. Shallow convection occurs under large-scale suppressed conditions and is proportional to surface evaporation.

At high resolution, we examine what happens when convection is resolved explicitly and not parameterized. The grid spacing is assumed to be small enough to begin resolving convective updrafts and downdrafts, and the generation of precipitation relies on the parameterization of microphysical processes. This results in a more realistic distribution of heat and moisture, and enables horizontal and vertical wind components to be directly modified by convection. Main differences between AO4-KF and AO4-TK are caused by different CPs used in D01 and D02; they produce different environments outside of D03, which are then used as boundaries for the high-resolution domain.

### 3.1 MJO Precipitation and Surface Winds

Due to the MJO’s slow propagation and large horizontal extent, time-longitude diagrams of observed fields are useful for examining the large-scale characteristics of the MJO precipitation and wind patterns. They can be very informative when comparing the observed features to those produced by different model experiments for a simple and intuitive evaluation of the model’s performance. Figure 3.1 shows Hovmöller diagrams of rainfall rate (in mm h\(^{-1}\); a-e), surface zonal winds (in m s\(^{-1}\); f-j) and SST (in °C; k-o) for observations (top) and model experiments as follows from top to bottom: AO12-KF, AO12-TK, AO4-KF, and AO4-TK. The latitudinal area averaging is done between 5°S and 5°N, and time in the panels increases upward. The observations shown are identical to those shown in Figure 2.2, and are included for ease of comparison.

The model simulations reproduce the observed features with varying degrees of success. In AO12-KF (Figure 3.1 b, g, l), each intense precipitation event near the equator triggers the formation of a tropical cyclone that propagates westward, with no signal of coherent...
eastward propagation in either precipitation or surface wind fields. The tropical cyclone
signatures cannot be seen clearly as the averaging is limited to 5° from the equator, but
the associated precipitation occurs on November 23-25, as well as on November 28-30,
starting near the equator, then slowly propagating poleward. Two tropical cyclones were
indeed observed spinning off the tail ends of MJO2, but they occurred later than those in
AO12-KF, and farther away from the equator. What cannot be seen here, but is described in
more detail in section 3.3 is that the precipitation outside the produced tropical cyclones in
AO12-KF is concentrated at low rainfall rates that tend to persist throughout the simulation.

On the other hand, AO12-TK (Figure 3.1 c, h, m) produces a strong and coherent
eastward-propagating feature in precipitation, surface winds, and SST. However, instead of
two distinct precipitation features with suppressed precipitation near the equator between
them, the model only produces one (with some equatorial suppression after intense pre-
cipitation. The CE propagates across the MC earlier than was observed. Surface westerly
winds are not as strong or as expansive as observed, and that is reflected in weaker-than-
observed upper-ocean and SST cooling. In AO12-TK, large-scale equatorial suppression of
precipitation over the IO does not persist for more than 3 days after the MJO has passed,
and new precipitation features begin to develop after December 1.

The differences between AO4-KF (Figure 3.1 d, i, n) and AO4-TK (Figure 3.1 e, j, o)
are subtler, with both experiments producing an MJO event, indicating that once convection
is no longer parameterized, the lower-resolution environment only has a small influence
on what goes on in the high-resolution region. Both AO4-KF and AO4-TK produce two
distinct, eastward-propagating CEs with accompanying surface westerly winds and SST
cooling. AO4-KF generally produces less rain than AO4-TK, with a stronger suppression
between the two CEs. Surface winds in AO4-KF are weaker, and consequently, the SST
cools to a lesser extent, and not as rapidly. In AO4-TK, the surface winds are stronger
and the westerly winds are more persistent over the IO, though the signal over the MC is
somewhat more scattered. Compared to AO12-KF and AO12-TK, the precipitation maxima
are amplified, and the amount of precipitation is increased, which is an expected result of increased resolution.

Overall, the experiments that include higher resolution and explicitly resolved convection (AO4-KF and AO4-TK) are able to represent the large-scale features of the MJO, such as two distinct convective envelopes and strong, expansive surface westerly winds. The SST cooling is also stronger, and can partly be attributed to stronger surface winds and increased amount of precipitation. One of the lower-resolution, parameterized-convection experiments (AO12-TK) is also able to reproduce an MJO, though those same large-scale features are not as well represented. None of the model experiments cool the SST at the rate that was observed, which is related to all of them (except AO12-KF) producing too much precipitation.

The high bias in precipitation is further examined in section 3.3, but excessive precipitation is also evident in the LPT tracking of the MJO in the model. In TRMM observations, a daily precipitation accumulation of 13 mm captures the top 5% of intense precipitation events over the IO and MC (Kerns and Chen 2016), while the area corresponding to the same precipitation threshold in the model is much larger. The evolution of the LPT tracker in both time and space is shown in Figure 3.2 - the colors indicate time progression, while the contours outline the area over which \( \geq 13 \) mm of daily rainfall accumulation has been sustained for at least three days. The track for the observed LPT (from TRMM data) is shown on top (3.2, followed by UWIN-CM experiments as follows: AO12-KF, AO12-TK, AO4-KF, and AO4-TK. The LPT tracker allows us to compare the spatial and temporal evolution of precipitation features in both model and observations, and it can illustrate the eastward propagation of large-scale convective elements. Kerns and Chen (2016) implement a set of criteria to discriminate MJO events from other large-scale equatorial convective events such as tropical cyclones, Kelvin waves, etc. In order for an event to qualify as an MJO, the the LPT track of any precipitation event needs to satisfy two criteria: (1) time continuity of at least ten days - to eliminate synoptic scale events, and (2) average eastward
propagation speed of at least $2 \text{ m s}^{-1}$ to eliminate stationary and westward propagating systems.

In TRMM data, the propagation is very smooth, with the precipitating area alternating between the Southern and Northern Hemispheres, and both CEs seen in Figure 3.1 a tracked as a single feature. With a time continuity of 528 hours (22 days), and a mean eastward propagation speed of $3.15 \text{ m s}^{-1}$, it qualifies as an MJO according to the criteria of Kerns and Chen (2016). At the initial time of the model simulations (00Z on November 22, 2011), the observed LPT area is localized over the equatorial western IO, but none of the UWIN-CM experiments able to capture the initiation accurately. Model experiments initially produce a much larger area of precipitation that encapsulates most of the equatorial IO east of $55^\circ$E, even extending over the MC, which can partly be attributed to the model spinning up and adjusting the physics and dynamics into a balanced state.

As expected from Figure 3.1, no precipitation features in $AO12-KF$ qualify as an MJO, failing both the time continuity and eastward propagation criteria, and thus does not produce an MJO. The $AO12-KF$ LPT lasts for 186 hours (7.75 days) and exhibits an average westward propagation of $1.19 \text{ m s}^{-1}$. In sharp contrast, $AO12-TK$ exhibits a very smooth eastward propagation up until three days before the end of the simulation. At that point, the precipitation over the IO overpowers the MJO-associated eastward-propagating precipitation, and the tracker is pulled back over the IO. The overall mean propagation speed is $1.88 \text{ m s}^{-1}$, which is slow due to the fact that the LPT propagates westward for the last three days; discarding that data and only looking at the portion of the LPT that propagates eastward, the eastward propagation speed increases to $4.5 \text{ m s}^{-1}$. The north-south shifts in the model LPTs are not as clear as what can be seen in TRMM data, though they are present, and the precipitating area tends to be larger than observed, as is the case in the MJO-producing experiments. Both $AO4-KF$ and $AO4-TK$ LPTs qualify as MJO events, though their propagations are quite different. In $AO4-KF$, the LPT propagates out of the IO and across the MC, but the transition is not smooth, with a mean eastward propagation
Figure 3.1: 5°S-5°N-averaged Hovmöller diagrams of rainfall rate (mm h⁻¹; a-e), surface zonal winds (m s⁻¹; f-j) and SST (°C; k-o) for observations (top) and model experiments. SST panels only include the IO area. Brown shading on the bottom panels indicates the height of maximum topography within the averaging region.
Figure 3.2: LPT evolution of the precipitation field for a) TRMM observations, b) AO12-KF, c) AO12-TK, d) AO4-KF, and e) AO4-TK. Colored contours outline the area where the accumulated daily precipitation amount exceeds 13 mm day$^{-1}$, and is sustained for at least three days, with the color indicating time.
speed is 5.8 m s\(^{-1}\), with convection stalling over the eastern IO, and then jumping across the MC towards the western Pacific. In AO4-TK, which produces more rain compared to AO4-KF, the LPT covers a larger area. A smooth propagation from the western to the central IO is followed by the LPT extending to the east, with the tail end never leaving the IO completely.

The difference between AO4-KF and AO4-TK can in part be attributed to AO4-TK producing more (and more intense) precipitation, so the LPT captures more than just the most intense 5% of sustained convection. If a higher precipitation accumulation threshold is used for tracking the MJO, the extent of convective areas becomes smaller, and the propagation in UWIN-CM becomes smoother. For example, when a 16 mm day\(^{-1}\) precipitation accumulation threshold is used to define LPT contours, the AO12-TK MJO shows clear eastward propagation and the dissipates between Sumatra and Borneo on December 2, while the AO4-KF MJO is pulled back over the IO and does not connect with precipitation in the western Pacific. The AO4-TK MJO propagates farther east than when a lower precipitation threshold is used for tracking, but still does not move out of the IO. This indicates that the precipitation and eastward propagation in AO4-TK are more robust than in other experiments displayed.

### 3.2 Boundary Layer - Free Troposphere Interaction

In this section, we address the differences between how model experiments deal with the interactions between the BL and the free troposphere above it, as the model simulations are strongly affected by the choice of the CP. In analyzing a large convective system that is a part of the large-scale MJO convective envelope using aircraft-released dropsondes, we identify and bring to light problems with how the two CPs deal with the interaction, which is especially detrimental when the KF CP is used.

The observed convective system that occurred on November 24, 2011, is depicted in Figure 3.3 and has been previously described in Chen et al. (2016). Figure 3.3 a shows
the large-scale TPW environment in color, overlaid with IR brightness temperature in greyscale. The central IO equatorial region is shown, and there is a lot of environmental moisture is present near the equator, with and drier air around 7°S. Black line indicates the flight track that the WP-3D aircraft took during the November 24 mission, with the starting and ending positions at Diego García located at 7.35°S and 72.48°E. Circle markers show the locations of every dropsonde released during the flight, and the larger white dots indicate the dropsondes that trail along the equator and sample the vertical cross-section shown in Figure 3.3 b. The vertical cross-section shows wind speed in m s⁻¹ (color), with the horizontal wind speed (in kt) and wind direction shown by wind barbs. Following traditional convention, a short flag indicates 5 kt, and a long one, 10 kt; a flag indicates a wind speed magnitude of 50 kt. The vertical cross-section shows a strong jet of winds descending from west to east with predominantly westerly winds, especially near the surface. The jet reaches the surface between 77 and 79°E, producing strong winds of almost 20 m s⁻¹ right at the surface, and has a structure similar to that of a rear-inflow jet feeding into a mesoscale convective system, though on a larger scale. In Figure 3.3 b, the black outlined and shaded areas show the near-saturated air with a RH ≥ 85%, and the white circle markers show the top of the BL. The BL top is defined as the first height at which the difference between the surface value of virtual potential temperature (θ_v) reaches 0.5°C.

The equatorial vertical cross-section of winds for the four UWIN-CM experiments is shown in Figure 3.4 a-d, with experiments arranged from top to bottom as before: AO12-KF, AO12-TK, AO4-KF, and AO4-TK, with the legend following that of 3.3 b. The right panels (Figure 3.4 e-h) show vertical profiles of θ_v and RH from R/V Revelle sounding observations in blue, and model-simulated soundings at the same location in red. R/V Revelle was at that time located at the equator and 80.5°E, and observed the passage of the same convective system.

The difference among the model experiments is striking - especially that between AO12-KF, AO12-TK, and the AO4 experiments. All of them produce strong winds, though their
Figure 3.3: A large equatorial convective event with multiple MCSs observed during the active phase of the MJO on November 24, 2011 (adapted from Chen et al. (2016), Figure 6). a) TPW (mm, color contour) and GridSat IR brightness temperature (grey, K), overlaid with the WP-3D aircraft track (black line). Circle markers show locations of released dropsondes. White circles mark the dropsonde profiles used in constructing the b) vertical cross-section of wind speed in m s$^{-1}$ (color) with horizontal wind direction indicated by wind barbs (magnitude in kt). Black outline and shaded area indicate a RH $\geq$ 85%, and white circle markers in b) indicate the top of the BL.

location in the vertical varies. In $AO12$-$KF$, the jet is stronger than observed, and has greater horizontal and vertical extent. Westerly winds are strong, but do not reach the surface; instead, winds above 13 m s$^{-1}$ seem to sit above the top of the BL. The surface winds do not exceed 10 m s$^{-1}$, while the winds aloft exceed 25 m s$^{-1}$. Interestingly, the top of the boundary layer is often located above a layer of saturated air for which the RH exceeds 85%, highlighted by the shaded contour and outlined in black. This case is clearly illustrated when looking at how model-simulated soundings compare to the observed sounding profile at R/V Revelle on the right. The RH profile in $AO12$-$KF$ exhibits a sharp local max-
Figure 3.4: As in Figure 3.3 b, but for model experiments: a) $AO12-KF$; b) $AO12-TK$; c) $AO4-KF$; and d) $AO4-TK$. Black contours and opaque grey shading indicate areas where the RH $\geq$ 85%, with the white dots indicating the top of the BL. e-h) show near-surface profiles of $\theta_v$ (left) and RH (right) from R/V Revelle sounding observations (blue) and each UWIN-CM experiment (red).
imum at 500 m, followed by a rapid decrease from ~92 to 60%. And the top of the BL coincides with the height at which the RH reaches a local minimum.

The pattern of a sharp local maximum, followed by a local minimum is not as drastic in other experiments, though the top of the BL generally coincides with a slight decrease in RH. In AO12-TK, the local maximum is much less prominent, with a RH profile that remains nearly uniform even above 2000 m. In some ways, the vertical cross-section of winds seems like the reverse of that produced by AO12-KF, with stronger winds near the surface, and weaker winds aloft. For this particular cross-section, winds above 1000 m rarely exceed 10 m s$^{-1}$, while surface winds reach up to ~15 m s$^{-1}$. The descending jet is weak compared to observations, and has a smaller vertical extent. Increasing resolution improves the representation of the jet for both CPs. In AO4-KF, the jet now reaches to the surface, with a vertical extent comparable to observations. The surface winds are stronger, and the sharp local maximum in RH is replaced by a more gradual decrease above the BL, which is quite close to observations. In AO4-TK, the vertical extent of the jet is improved, as is the wind speed. In both AO4-KF and AO4-TK, the surface extent of the jet is greater than that observed.

Though Figure 3.3 is a single case of a spatially well-observed rear-inflow jet along the equator from the DYNAMO field campaign, we can generalize our findings by turning to sounding observations from the equatorial DYNAMO sites - Gan and R/V Revelle, located at 0.69°S, 73.15°E and 0.1°N, 80.5°E, respectively. Sounding observations at those locations were taken every 3 hours during the modeled period. 120 soundings from Gan and 75 soundings from R/V Revelle (which left its listed location on December 3) are combined into a sample size of 195 soundings that cover the modeled period and contain data of vertical profiles of atmospheric moisture, temperature, and winds. The distributions of RH and zonal winds with height are shown in Figure 3.5, with RH shown on top (Figure 3.5 a-e), and zonal winds on the bottom (U, Figure 3.5 f-j). The distributions of combined sounding observations are shown on the left. Color shading indicates the percentage of all soundings
that fall into a certain RH range at a given height. The height ranges highlighted in grey draw focus to three areas in which the model experiments differ from each other, and from observations.

Figure 3.5: Observed (left) and model-simulated soundings of RH (a-e) and zonal winds (f-j) taken from DYNAMOs equatorial observational sites Gan, and R/V Revelle. 195 soundings from the two sites are used to create the observed distributions (a, f), and compared to 240 model-simulated soundings. The colored area indicates the % of soundings in each bin. Shaded areas serve to highlight the low-level, mid-level, and upper-level features described in text.

At high levels (between 11,200 and 12,500 m), the observed distribution is reproduced by \textit{AO12-TK}, \textit{AO4-KF}, and \textit{AO4-TK}, though the observed spread is better represented in the high-resolution experiments. \textit{AO12-KF} produces a very concentrated distribution with
no drier soundings, and the $AO12-TK$ exaggerates the frequency of saturated profiles.

In the mid-levels near the melting layer (between 5,000 and 7,000 m), observations show a large concentration of saturated soundings, but there's also a large spread in RH, with some soundings with $\sim 20\%$ RH representing the mid-latitude dry air that intruded into the equatorial region from the west (summarized in Kerns and Chen 2014b). $AO12-KF$ shows very little spread, and the vertical RH profiles are generally very moist and fairly uniform throughout the simulation, showing that there is indeed no large-scale regime change occurs during the 15-day simulation. There is also pronounced evidence of a melting level. $AO12-TK$ shows very good spread, with some very dry soundings and some evidence of saturation near the melting level, though it is not as noticeable as in observations. Between $AO4-KF$ and $AO4-TK$, the TK shows better spread with some RH values below 40\% (not in in $AO4-KF$), and a more pronounced frequency of a saturated profile and melting level.

At low levels (between 200 and 1,500 m), observations exhibit a local maximum in RH, which the $AO12-KF$ exaggerates, while it is mostly smoothed out in $AO12-TK$. This is also reflected in the profiles of zonal wind (Figure 3.5 f-j). In sounding observations, westerly winds gradually increase from surface values centered near 5 m s$^{-1}$, and reach a maximum $\sim 2,000$ m. Above 2000 m, westerly winds weaken, turning to easterlies in mid-upper levels. In $AO12-KF$, westerly winds at the surface show a uniform profile of winds $\sim 3$ m s$^{-1}$ that persists throughout the BL; at the top of the BL, RH reaches a local minimum and above that, winds begin to intensify. Above the BL, westerly winds reach peak intensity $\sim 3,500$ m, which coincides with the height at which the jet in Figure 3.4 is strongest, indicating that it is a common occurrence. Mid-level westerly winds in $AO12-KF$ are actually stronger than in any other simulation. In $AO12-TK$, the low-level westerlies do not intensify much with height, instead remaining between 0 and 10 m s$^{-1}$ up to $\sim 4000$ m, indicating a very-well mixed layer extending well above the BL.

In $AO4-KF$ and $AO4-TK$, the low-level local maximum in RH is not as pronounced as in $AO12-KF$, and more pronounced than in $AO12-TK$, which are both improvements
compared to observations. Between the two, AO4-KF exhibits a sharper gradient in RH near the top of the BL with a slightly sharper gradient evident in AO4-KF. Surface westerly winds in high-resolution experiments are generally slightly stronger than observed, with magnitude increasing up to \( \sim 1,500 \) m. Peak wind intensities are not as high as observed, and occur lower in the atmosphere, but are in general an improvement over the lower-resolution experiments.

These comparisons clearly identify an issue in how the model boundary layer and convective parameterizations interact, with the KF and TK CPs seemingly standing on the opposite side of the spectrum. On one side, there is the KF CP, where the constant saturated, or nearly-saturated top of the BL presents an obstacle to the mixing and entrainment across the BL top. The presence of the saturated layer essentially decouples the boundary layer from the free troposphere above it, preventing mid-level jets from mixing down into the BL and to the surface. On the other side, the TK CP tends to over-mix across the top of the BL, with the surface values varying very little with height over the lowest \( \sim 2,000-4,000 \) m. Experiments with explicitly resolved convection and no convective parameterization lie somewhere in between and better reflect the collected observations.

It is unclear whether the problems with mixing across the top of the BL in WRF (and consequently UWIN-CM) only arise in the case where the YSU planetary BL scheme is coupled to either of the two CP schemes we use in this study. But as the scenarios that result from the two CPs are very different, we can definitely conclude that the accurate representation of mixing and entrainment across the top of the BL is important for processes that contribute towards the MJOs initiation and eastward propagation. Between the two CPs, it seems that overly-active mixing across the BL top is more beneficial to the simulation of the MJO, than is under-active mixing. These results agree with previous studies on climate models (e.g. Hung et al. 2013) in that the use of a CPs in which the convective trigger relies on low-level moisture convergence (here the Tiedtke CP) tend to produce better MJOs.
3.3 Precipitation Bias

Precipitation is the main feature of the MJO, and predicting it accurately is one of the most difficult aspects of MJO modeling. As could be seen in Figures 3.1 and 3.2, all experiments that simulate an MJO produce too much precipitation. Time series of rainfall rates comparing TRMM observations (black) and UWIN-CM experiments (color) is shown in Figure 3.6. The time series show the hourly (thin lines), and 24-hour smoothed (thick lines) rainfall rates averaged across the portion of the IO covered by D03, which covers the area between 55 and 97°E, 10°S and 10°N. The signal of MJO precipitation in the IO is seen in TRMM observations between November 22 and 30, when the rainfall rates are relatively high with an average of 0.614 mm h⁻¹, followed by a suppressed period between December 1 and 6, during which the average rainfall rate is 0.162 mm h⁻¹.

![Figure 3.6: Time series of rainfall rate averaged between 55 and 97°E, 10°S and 10°N (the portion of the IO covered by D03). Thin lines indicate hourly (for model) and 3-hourly (for TRMM) rainfall rates, while thick lines show rainfall rates smoothed using a 24-hour running mean.](image)

Average rainfall rates for the entire time series, as well as the averages for the active and suppressed periods of the MJO, are shown in Table 3.1, followed by the average rainfall
rate biases in Table 3.2. The switch between active and suppressed phases is based on the
time at which the observed MJO propagates out of the IO and into the MC on December
1. Though not simulating an MJO, the \textit{AO12-KF} shows the least overall precipitation
bias, for two reasons. One, it does not capture the increased rainfall intensity caused by
the MJO meaning that the precipitation bias is negative during the active MJO period,
offsetting the positive bias during the suppressed period. Two, it produces the least amount
of precipitation overall compared to the experiments that produce an MJO and exaggerate
the precipitation amount and intensity. The other experiments all have positive rainfall
biases, as is expected from looking at the Hovmöller analysis and LPT tracks, with the
lowest one in \textit{AO12-TK} which also produces less rain overall than the high-resolution
experiments. The difference in average rainfall rates between the experiments is due to how
precipitation is being resolved, and is explained in more detail in section 3.3.1. Between
\textit{AO4-KF} and \textit{AO4-TK}, the TK produces more rain everywhere, which has to do with the
amount of moisture entering D03 through the boundaries and is addressed in section 3.3.2.

Table 3.1: Rainfall rates averaged over the IO portion of D03 (as in Figure 3.6). Total col-
umn shows average rainfall rate for the entire time series, while the Active and Suppressed
columns refer to the time periods between November 22 and 30 (active) and December
1 and 6 (suppressed). The Difference column shows the difference between the average
rainfall rates during the active and suppressed phases.

<table>
<thead>
<tr>
<th>Experiment</th>
<th>Total</th>
<th>Active</th>
<th>Suppressed</th>
<th>Difference</th>
</tr>
</thead>
<tbody>
<tr>
<td>AO12-KF</td>
<td>0.5009</td>
<td>0.5619</td>
<td>0.4094</td>
<td>0.1525</td>
</tr>
<tr>
<td>AO12-TK</td>
<td>0.5344</td>
<td>0.6322</td>
<td>0.3877</td>
<td>0.2445</td>
</tr>
<tr>
<td>AO4-KF</td>
<td>0.6170</td>
<td>0.7053</td>
<td>0.4846</td>
<td>0.2207</td>
</tr>
<tr>
<td>AO4-TK</td>
<td>0.7142</td>
<td>0.8137</td>
<td>0.5648</td>
<td>0.2489</td>
</tr>
<tr>
<td>TRMM</td>
<td>0.4325</td>
<td>0.6137</td>
<td>0.1619</td>
<td>0.4531</td>
</tr>
</tbody>
</table>

The difference between rainfall rates during the active and suppressed MJO phases
is one way of illustrating the intensity of the precipitation suppression over the IO after
the passage of the MJO. It is shown in the right-most (Difference) column of Table 3.1.
The larger the magnitude of the difference, the more suppressed the precipitation over the IO is during the suppressed phase, relative to what it was during the active phase. The suppression in TRMM observations, with a rainfall rate difference of 0.453 mm h$^{-1}$ is very strong, and for an illustration of the difference, Figure 3.7 shows two GridSat infrared brightness temperature (Knapp et al. 2011) snapshots of the MJO2: Figure 3.7 while the MJO convection is active over the IO, and Figure 3.7 b while the MJO convection is active over the MC. During the active phase, the IO is covered with large-scale convective systems, while in the suppressed phase, all the precipitation is concentrated over the MC, with a single tropical cyclone present in the southern IO. The rest of the IO shows almost no convective activity at that time.

Figure 3.7: GridSat IR brightness temperatures showing two snapshots of the MJO2 event: a) MJO convection over the IO (November 26, 2011 at 18 UTC), and b) MJO convection over the MC and suppressed over the IO (December 4, 2011 at 18 UTC).
Table 3.2: Same as Table 3.1, but showing rainfall rate bias compared to TRMM (in mm h\(^{-1}\)) followed by the % overestimation of TRMM rainfall rates in parentheses. A 100% overestimation indicates an average rainfall rate that is twice as high as in TRMM.

<table>
<thead>
<tr>
<th>Experiment</th>
<th>Total</th>
<th>Active</th>
<th>Suppressed</th>
</tr>
</thead>
<tbody>
<tr>
<td>AO12-KF</td>
<td>0.068 (33.6%)</td>
<td>-0.053 (-6.5%)</td>
<td>0.250 (243.8%)</td>
</tr>
<tr>
<td>AO12-TK</td>
<td>0.101 (35.1%)</td>
<td>0.017 (3.4%)</td>
<td>0.227 (232.7%)</td>
</tr>
<tr>
<td>AO4-KF</td>
<td>0.183 (72.6%)</td>
<td>0.089 (16.4%)</td>
<td>0.324 (307.0%)</td>
</tr>
<tr>
<td>AO4-TK</td>
<td>0.280 (115.9%)</td>
<td>0.196 (33.8%)</td>
<td>0.405 (389.2%)</td>
</tr>
</tbody>
</table>

None of the UWIN-CM experiments reproduce the amount of precipitation suppression over the IO that is shown in TRMM observations (and in Figure 3.7). However, all the experiments that simulate an MJO reproduce at least some suppression, with it being most prominent in AO4-TK, followed by AO12-TK and AO4-KF. Interestingly, though AO4-TK has the highest overall precipitation bias, it also produces the most suppression - relative to its IO active phase. Some reasons as to why the model rainfall rates, and rainfall rate biases, are they way they are, are addressed in the next two sections, touching upon the different rainfall rate distributions that are produced by UWIN-CMs parameterizations, and the convergence of moisture into D03.

### 3.3.1 Rainfall Rate Distribution

Figure 3.8 shows the distributions of 3-hourly precipitation rates for TRMM (black) and UWIN-CM experiments. The model-produced precipitation has been re-gridded to the resolution and frequency of TRMM data (3-hourly, 0.25\(^\circ\)) over the high-resolution domain. To create the precipitation rate distribution, fifty logarithmically-spaced bins are created, capturing rainfall rates between 0.1 and 100 mm h\(^{-1}\). All the distributions have been normalized so that the total area under the distribution curve equals one, making the comparison between them possible. In TRMM data, the median value, with 50% of the rainfall rates lying below, and 50% above it, is at a rainfall rate of 1.03 mm h\(^{-1}\). The 10% of most extreme rainfall rates fall above 4.20 mm h\(^{-1}\). 50% of all rainfall occurs between
0.44 and 2.39 mm h$^{-1}$. These numbers might not be representing the actual distribution of precipitation rates, as TRMM has a difficult time accurately representing intense, and very weak rainfall rates due to 3-hourly time averaging (Prakash et al. 2012). It is used here as a general guideline of what the rainfall rate distributions should look like outside the extrema.

![Rainfall Rate Distribution](image)

Figure 3.8: Distribution of rainfall rates within D03 for TRMM (black) and model experiments (color). Note that the x-axis is displayed logarithmically.

Compared to TRMM, the biggest difference in the model experiments is between $AO12$-$KF$ and $AO12$-$TK$, while distributions of $AO4$-$KF$ and $AO4$-$TK$ are nearly identical. The reason for the difference is the way that precipitation is generated in the model. In high resolution experiments, where convection is resolved explicitly, all precipitation is produced by the microphysics parameterization scheme. It is responsible for producing rain on a grid scale, representing processes that describe the formation, growth, and sedimentation of hydrometeors. Since both $AO4$-$KF$ and $AO4$-$TK$ use the same microphysics parameterization scheme, the resulting distributions look very similar, with the differences likely arising from different amounts of moisture available, or from different vertical velocities.

In the low resolution experiments, the microphysics parameterization scheme is responsible for producing precipitation on a grid scale, but not many strong updrafts that help the
formation of precipitation can be resolved at a resolution of 12 km x 12 km, where each grid cell represents an area of 144 km$^2$. Most of the precipitation in $AO12-KF$ and $AO12-TK$ is produced by the CPs which are responsible for parameterizing the sub-grid precipitation (75.8%, and 80.9%, respectively), and it shows that the two CPs produce rainfall in different ways. In $AO12-KF$, 50% of the rainfall rates lie between 0.44 and 1.57 mm h$^{-1}$, which is a much narrower range than in TRMM, with values concentrated around 1.2 mm h$^{-1}$. So $AO12-KF$ produces more rain rates $\leq$ 2 mm h$^{-1}$, less moderate rain rates between 2 and 11 mm h$^{-1}$, and then more extreme rain rates $\geq$ 11 mm h$^{-1}$. In $AO12-TK$, the distribution is smoother, with rainfall rates better distributed among low-moderate values. The middle 50% of rainfall rates are contained between 0.33 and 2.75 mm h$^{-1}$, which yields a slightly wider range than what is seen in TRMM. In general, $AO12-TK$ overproduces rainfall rates $\geq$ 4 mm h$^{-1}$, and under-produces less intense precipitation.

This can explain the difference between the average rainfall rates produced by the $AO12-KF$ and $AO12-TK$ experiments shown in Figure 3.6. $AO12-KF$ produces a lot of low rainfall rates (drizzle), which helps keeps the overall average rainfall rate low, while $AO12-KF$ produces more moderate rainfall rates. $AO4-KF$ and $AO4-TK$ produce a lot of high and extreme rainfall rates, which is responsible for their high averages. However, small differences in the rainfall rate distributions do not explain the difference in the amount of precipitation produced. That can be partly attributed to the amount of moisture that enters the high-resolution domain through its borders.

3.3.2 Moisture Convergence

Moisture convergence into a region is only of the many parts of the complicated moisture budget equation. However, in this case, a simple analysis can illustrate a reason for the difference between the amounts of rainfall produced in $AO4-KF$ and $AO4-TK$ experiments. The analysis is shown in Figure 3.9.

To approach the problem, a simplified moisture flux ($\nabla \cdot q \vec{v}$) was calculated at all the boundaries of D03, with care being taken to use only points that fall outside D03. The
Figure 3.9: Time series of: a) instantaneous moisture convergence into D03; b) cumulative moisture convergence into D03; c) cumulative moisture convergence difference between AO4-TK and AO4-KF (blue minus brown); d) cumulative total rainfall; and e) cumulative total rainfall difference between AO4-TK and AO4-KF (blue minus brown).

Influx of the moisture flux was summed up along the longitudes and latitudes of every D03 border, and vertically throughout the troposphere (between 1,000 and 200 hPa). This was done for every hour of model output to yield a time series of instantaneous moisture convergence into D03, which is shown in Figure 3.9 a, with positive values showing the influx of moisture into D03. Thin lines represent the hourly moisture flux convergence into D03, and they exhibit a lot of high-frequency variability. The hourly time series are filtered using a 24-hour running mean, which is shown with thicker lines. In general, AO4-TK tends to have more moisture converging into D03 (or less moisture leaving it), though the relationship is difficult to discern from instantaneous time series.
A better representation is shown in Figure 3.9 b, where the moisture convergence into D03 is shown cumulatively from the beginning of the simulation. At times when the time series show a positive gradient (increasing with time), moisture is converging into D03, while moisture is leaving D03 when the gradient is negative (decreasing with time). This shows that a lot more moisture is being brought into D03 in AO4-TK, with the difference between AO4-TK and AO4-KF shown in Figure 3.9 c. Figure 3.9 d and e show the cumulative accumulated rainfall in D03 for both AO4-KF and AO4-TK in Figure 3.9 d, and their difference in Figure 3.9 e. There is a positive relationship between the difference in moisture convergence and the difference in amount of total precipitation. The difference between total accumulated rainfall in the two experiments (Figure 3.9 d) increases faster when there is a bigger difference in the amount of moisture flux entering D03 (Figure 3.9 e). This can, at least in part, explain why AO4-TK produces more precipitation than AO4-KF over the entire D03, and over each individual part of it, such as the IO (see difference between AO4-KF and AO4-TK in Figure 3.6).

3.4 Summary

In this chapter, we addressed the effects that resolution, and different CPs have on the representation of the MJO2 event in UWIN-CM. We have done so by comparing two sets of model experiments - two at low (12 km) resolution where a CP is necessary to represent the effects of convection, and two at high (4 km) resolution, where it is not. We have tested two different CP schemes - the Kain-Fritsch (KF) and the Tiedtke (TK) in the low resolution cases; in the high-resolution cases, the CP schemes are still used in the low-resolution parent domains, leading to two high-resolution experiments, one for each choice of CP.

As previous studies have shown for MJO simulations with climate and global models (Hung et al. 2013; Pilon et al. 2016), the results of our study confirm that the choice of CP makes a large difference in whether the model is even capable of simulating an
MJO, though it is not the important factor. Similar to previous studies such as Hung et al. (2013) for climate model applications, the CPs that use low-level moisture convergence as a triggering mechanism for precipitation tend to be more successful in producing an MJO. Though our low resolution is much higher than the resolution of climate models, our results are consistent with theirs. The \textit{AO12-TK} produces a convective envelope that exhibits strong and coherent eastward propagation in both precipitation and surface zonal winds, until a few days before the end of the simulation. Though the representation of the precipitation feature is not very accurate to observations and high resolution experiments, it is encouraging that low-resolution experiments are able to represent at least some part of the MJO successfully, as high-resolution experiments can be very computationally demanding.

Every time large-scale convection begins to organize in \textit{AO12-KF}, it spins up a into a tropical cyclone that propagates westward, instead of an eastward-propagating precipitation envelope of the MJO. However, even though the experiment does not produce an MJO, further analysis of the two experiments side by side identified an important flaw in how the CPs interact with the model parameterization of the BL. The low-resolution experiments with different CPs produce significantly different vertical structure in the BL. In \textit{AO12-KF}, we identified the presence of a very moist, often saturated layer just under the top of the BL, which either acts as a barrier to vertical mixing and entrainment from the free troposphere above, or is a result of the lack of vertical mixing (Chandra et al. 2015). Instead of strong winds above the BL entraining momentum downwards, creating jets that descend all the way to the surface, strong jets in \textit{AO12-KF} sit on top of the BL. Compared to sounding observations, \textit{AO12-KF} actually produces westerly winds that are stronger than in any other model experiment, but they are confined above the BL, so there is very little trace of surface westerly winds. This lack of mixing through the BL top in \textit{AO12-KF} could be related to the experiment overproducing tropical cyclones (Mallard et al. 2013). The mixing across the top of the BL that is under-represented in \textit{AO12-KF} is exaggerated in \textit{AO12-TK}, with surface properties being mixed up well above the BL. In that respect, \textit{AO12-KF} and \textit{AO12-}
TK seem to fall on the opposite ends of a spectrum, with the high resolution experiments sitting somewhere in between them, and better reflect observations.

Regardless of the difference between the low-resolution experiments, both \textit{AO4-KF} and \textit{AO4-TK} produce an MJO with strong eastward propagation of zonal winds and precipitation that is separated into two distinct precipitation features separated by a precipitation suppression near the equator. Both experiments produce too much precipitation that is too intense, compared to both TRMM observations and lower resolution experiments, with \textit{AO4-TK} producing more precipitation than \textit{AO4-KF}. The difference is linked to a weaker influx of moisture through the boundaries of D03 in the latter. Generally \textit{AO4-TK} performs better than \textit{AO4-KF} in terms of ocean cooling, the surface westerly wind signal, the MJOs eastward propagation, and the amount of suppression that occurs after the passage of the active MJO, though the precipitation bias is greater.

In the next chapter, we discuss the effects of atmosphere-ocean coupling on the representation of the MJO in UWIN-CM, where we analyze the precipitation bias in the high-resolution experiments in more detail, and attempt to reduce it.
Atmosphere-Ocean Coupling and the MJO

The atmosphere and ocean are parts of a coupled system, continuously interacting in through the exchange of heat, moisture, salt, and momentum. Though there are cases in which either component can act nearly independently, that is not the case for phenomena that occur in the global tropics - such as the MJO. During its passage, the atmospheric patterns associated with the MJO, such as strong surface winds, intense precipitation and extensive cloud cover, influence upper-ocean mixing and cool the SST over the IO (de Szoeye et al. 2015). In the western Pacific, the westerly wind bursts from an active MJO help push warm water toward the central Pacific and reduce the equatorial gradient in SST, which has been linked to onset of El Niño (Moore and Kleeman 1999; McPhaden et al. 2006). In turn, oceanic phenomena such as the phases of ENSO and the Indian Ocean Dipole modulate MJO activity (Tam and Lau 2005).

The two-way interaction between the ocean and atmosphere is critical to the eastward propagation of the MJO, but it is also what makes coupled models unpredictable to work with. With an incomplete understanding of how the atmosphere and ocean interact in reality, and lack of observational data over oceans, accurate representation of air-sea interaction processes in models is difficult at best. In addition, the atmosphere and ocean models used in coupling have their own biases and parameterizations, and have most often been tuned to perform well independently, and over a region that does not necessarily match the region they are applied to. One example of a biased parameterization is described in section 4.2, and involves an inaccurate parameterization of air-sea latent and sensible heat fluxes.

Some studies have shown that a coupled model that includes an ocean component that responds to atmospheric forcing is not necessary for a successful MJO simulation (Miura
et al. 2007; Kim et al. 2009). Commonly, those studies use models without a convective parameterization, which is done either by using high resolution, or by using superparameterization. Superparameterization is a method by which a cloud-resolving model is inserted into each grid box of a low-resolution model (Randall et al. 2003; Benedict and Randall 2007). Other studies that use DYNAMO observations show that the interaction between the MJO and its underlying ocean does not follow a single pattern, and that the ocean plays a more important role in some events than it does in others (Fu et al. 2015). Nonetheless, studies that highlight the importance of air-sea coupling in the MJO are more common, and illustrate that the use of atmosphere-ocean coupling improves the representation of the MJO and its eastward propagation.

In this chapter, we address the role of atmosphere-ocean coupling in the simulation of the MJO2 event by using three high-resolution, cloud-permitting model experiments. Each set includes two model experiments with identical configurations, except for the choice of CP that is used in the outer domains - one uses the Kain-Fritsch (KF), and the other one the Tiedtke (TK) CP. The three sets of experiments are the uncoupled, atmosphere-only UA4-KF/TK, the atmosphere-ocean coupled AO4-KF/TK, and the atmosphere-ocean coupled AO4-KF/TK-FLX with a modified air-sea flux parameterization. The AO4-KF/TK are the same experiments that were used in Chapter 3. The flux modification in AO4-KF/TK-FLX was implemented to try and reduce the precipitation bias present in AO4-KF/TK that has been described in Chapter 3. AO4-TK produced better results than AO4-KF in terms of upper-ocean cooling and the robustness of MJO propagation, though it produced a higher precipitation bias. Only the results from TK experiments are shown in this chapter, though the presented results and differences between the simulations are consistent regardless of which CP scheme is used.
4.1 SST Cooling and MJO Eastward Propagation

Figure 4.1 shows Hovmöller diagrams of rainfall rate (in mm h\(^{-1}\); a-d), surface zonal winds (in m s\(^{-1}\); e-h) and SST (in °C; i-l) for observations (top) and model experiments as follows from top to bottom: \textit{UA4-TK}, \textit{AO4-TK}, and \textit{AO4-TK-FLX}. The latitudinal area averaging is done between 5°S and 5°N, and time increases upward. The observations shown are identical to those shown in Figure 2.2, and are included for easier comparison. As described in Chapter 3, \textit{AO4-TK} produces two distinct eastward-propagating precipitation features that are separated by relative precipitation suppression near the equator, with accompanying surface westerly winds and SST cooling. Compared to observations, the amount and intensity of precipitation are overestimated in all experiments. Ocean cooling is present in both coupled experiments, though it is not as spatially extensive, and does not occur as rapidly as observed.

In \textit{UA4-TK}, the SST is held constant from initial conditions, and thus no ocean cooling is simulated. Surface winds are stronger than both observations and those in \textit{AO4-TK}, and the combination of strong winds blowing over a warm ocean results in strong evaporation. The influx of moisture that results from strong evaporation can sustain intense and long-lasting precipitation, especially over the IO. It is difficult to distinguish any eastward propagation in the noisy precipitation field, and the signal in surface westerly winds does not reach as far east as in the coupled experiments or in observations. In \textit{AO4-TK-FLX}, where we attempt to reduce the precipitation bias by modifying the air-sea flux parameterization, the amount of precipitation is indeed reduced compared to \textit{AO4-TK}. The precipitation field is less noisy, and the two convective envelopes appear more distinct, though they are also less coherent due to the overall reduction of precipitation. Zonal westerly winds are generally weaker over the IO than in \textit{AO4-TK}, but the surface westerly winds that extend over the MC during the second half of the simulation remain strong. Less, (and less intense) precipitation, less extensive cloud cover, and weaker surface winds induce less upper-ocean mixing in \textit{AO4-TK-FLX} than in \textit{AO4-TK}, and the SST in the IO remains warmer.
The suppression of precipitation in the OP after the MJO does not appear to be greatly reduced in the Hovmöller diagram shown in Figure 4.1 d, but the convective envelope in *AO4-TK-FLX* propagates farther east than in any other experiment (except *AO4-KF*, where it jumps to the western Pacific), and that can be seen more clearly in the LPT tracks shown in Figure 4.2. In *AO4-TK-FLX* (Figure 4.2 d), the convective envelope continuously propagates eastward from November 22 until December 5. After that, the LPT tracker jumps to the warm seas north of New Guinea, indicating that the precipitation over the IO has been sufficiently suppressed. When a higher precipitation threshold is used, the *AO4-TK-FLX* propagates smoothly eastward and out of the IO, but the convective envelope dissipates as it reaches Borneo on December 3. Until then, the time evolution of the convective envelope in *AO4-TK-FLX* is most similar to the observed one, with an average eastward propagating velocity of $3.7 \text{ m s}^{-1}$, though the LPT areas are still larger than observed. As is described in Chapter 3 (Figure 4.2 c), the *AO4-TK* LPT exhibits a smooth propagation from the western to the central IO, but the LPT never completely leaves the eastern IO, because the IO precipitation is never suppressed enough.

In *UA4-TK* (Figure 4.2 b), the LPT initially picks up the entire IO as a region of intense sustained precipitation. The LPT slowly grows and extends eastward into the western Pacific, without its tail end propagating eastward or diminishing in size over the IO. With the LPT centroid propagating eastward at an average of $1.48 \text{ m s}^{-1}$, the *UA4-TK* precipitation features do not qualify as an MJO event according to the criteria of Kerns and Chen (2016). The intense precipitation over the IO never stops, and this persistence of precipitation is a common feature found in many climate models, regardless of whether they are coupled to an ocean model (Hung et al. 2013). Previous studies have shown that atmosphere-ocean coupling improves the eastward propagation of MJO convection in climate models (Zhang et al. 2006). This study shows that, at cloud-permitting resolutions, atmosphere-ocean interaction is an integral factor in the MJOs propagation. The upper-ocean cooling that is induced as the MJO propagates over the ocean helps create an en-
Figure 4.1: 5°S-5°N-averaged Hovmöller diagrams of rainfall rate (mm h⁻¹; a-e), surface zonal winds (m s⁻¹; f-j) and SST (°C; k-o) for observations (top) and high-resolution coupled and uncoupled model experiments. SST panels only include the IO area. Brown shading on the bottom panels indicates the height of maximum topography within the averaging region.
Figure 4.2: LPT evolution of the precipitation field for a) TRMM observations, b) UA4-TK, c) AO4-TK, and d) AO4-TK-FLX. Colored contours outline the area where the accumulated daily precipitation amount exceeds 13 mm day$^{-1}$, and is sustained for at least three days, with the color indicating time.
environment that is unfavorable for sustaining precipitation, pushing the MJO eastward and over the warm waters of the MC. In the uncoupled UA4-TK, no upper-ocean cooling occurs, and consequently precipitation over the IO is sustained through the entire integration period.

4.2 Air-Sea Fluxes and Their Parameterization

The atmosphere and ocean interact through air-sea fluxes of momentum, trace elements such as sea-salt, and sensible and latent heat. Of those, the sensible and latent heat fluxes are important sources of sensible heat and moisture for the atmosphere, and a misrepresentation of those is partly responsible for the precipitation bias present in UWIN-CM. After initial analysis of air-sea fluxes, it was noted that the sensible and latent heat fluxes in AO4-TK and UA4-TK had a significant high bias (Figure 4.3). The AO4-TK-FLX experiment was designed to reduce the high bias in air-sea fluxes, with the goal of reducing the precipitation bias in UWIN-CM.

Figure 4.3 shows the sensible (SHF) and latent (LHF) heat fluxes directly measured by instruments mounted on R/V Revelle, that was located at the equator and 80.5°E for the first ten days of the model integration period. R/V Revelle flux measurements were taken every ten minutes, and each measurement is represented by a yellow circle. The observations are binned according to the value of wind speed at which a given flux value occurs, so that the mean value for a given wind speed can be calculated - it is indicated by the yellow line. The distribution is normalized by the most frequent value so the distributions are easier to compare. Colored contours indicate the distribution of LHF (Figure 4.3 a-c) and SHF (Figure 4.3 d-f) produced in UWIN-CM, with the thicker line indicating the model mean for each experiment. Since the model output frequency is hourly, the UWIN-CM data is taken within a radius of 1° from the position of R/V Revelle to increase sample size and improve the representativeness of the data. The trend of air-sea fluxes increasing with increasing wind speed is reproduced, though the actual flux values are generally overestimated.
Figure 4.3: a-c) Latent and d-f) sensible heat flux distributions from R/V Revelle (yellow), and \textit{UA4-TK} (a, d), \textit{AO4-TK} (b, e), and \textit{AO4-TK-FLX} (c, f). Thick lines show the mean values of the flux at a given wind speed.

The difference between experiments and observations is not large when considering SHF, for which the observed values range between -5 and 100 W m\(^{-2}\), and the average positive bias has magnitudes of 11.8, 6.0, and 3.9 W m\(^{-2}\) in \textit{UA4-TK}, \textit{AO4-TK}, and \textit{AO4-TK-FLX}, respectively. The bias is higher than average at low wind speeds (\(\leq 5 \text{ m s}^{-1}\)), though the difference is marginal compared to the bias in the amount of LHF. The LHF bias of 83.12 W m\(^{-2}\) in \textit{UA4-TK} translates to LHF values that are twice, and at times even three times as high as those measured at R/V Revelle, while the lower range of observed values is never reproduced in the model. Atmosphere-ocean coupling improves the mean LHF bias by \(\sim 35\%\), though values at low wind speeds remain high. The implemented reduction of air-sea flux magnitude, described in more detail in section 4.3.1, results in a further reduction of the average bias by \(\sim 15\%\), and almost \(\sim 20\%\) when only considering low wind conditions.
The LHF and SHF biases are summarized in Table 4.1, with values from both the KF and TK experiments for comparison, though only the TK experiments are shown. Biases are given in $\text{W m}^{-2}$, with the % change from the observational mean given in parentheses next to every entry. The biases are calculated for both the entire distribution, and also the distribution limited to low wind speeds, where low wind speeds include surface winds $\leq 5 \text{ m s}^{-1}$.

Table 4.1: LHF and SHF biases compared to direct flux measurements from R/V Revelle. Values are given both for the full distribution and for the low winds only ($\leq 5 \text{ m s}^{-1}$). Biases for both TK and KF experiments are included for comparison.

<table>
<thead>
<tr>
<th>Experiment</th>
<th>LHF bias (W m$^{-2}$ (%))</th>
<th>LHF bias (low winds) (W m$^{-2}$ (%))</th>
<th>SHF bias (W m$^{-2}$ (%))</th>
<th>SHF bias (low winds) (W m$^{-2}$ (%))</th>
</tr>
</thead>
<tbody>
<tr>
<td>UA4-TK</td>
<td>98.6 (116.6%)</td>
<td>84.2 (181.8%)</td>
<td>12.7 (156.5%)</td>
<td>14.2 (291.9%)</td>
</tr>
<tr>
<td>AO4-TK</td>
<td>63.9 (72.47%)</td>
<td>59.5 (130.1%)</td>
<td>6.9 (86.3%)</td>
<td>6.8 (139.1%)</td>
</tr>
<tr>
<td>AO4-TK-FLX</td>
<td>53.7 (54.0%)</td>
<td>39.6 (83.0%)</td>
<td>4.7 (68.70%)</td>
<td>4.9 (101.7%)</td>
</tr>
<tr>
<td>UA4-KF</td>
<td>93.5 (97.9%)</td>
<td>73.8 (159.4%)</td>
<td>10.2 (109.5%)</td>
<td>8.3 (170.8%)</td>
</tr>
<tr>
<td>AO4-KF</td>
<td>65.1 (73.3%)</td>
<td>59.1 (128.6%)</td>
<td>5.5 (68.1%)</td>
<td>4.7 (97.3%)</td>
</tr>
<tr>
<td>AO4-KF-FLX</td>
<td>52.8 (55.1%)</td>
<td>42.4 (89.3%)</td>
<td>4.7 (60.7%)</td>
<td>4.1 (84.8%)</td>
</tr>
</tbody>
</table>

Surface winds are not the only things that affect the magnitude of air-sea fluxes, and the variation of other parameters that play important roles, such as the wind speed magnitude and the temperature and moisture disequilibria across the air-sea interface, are described in section 4.2.2. The effects of atmosphere-ocean coupling the air-sea flux modification on the amount and intensity of precipitation are described in section 4.2.3.

### 4.2.1 Bulk Air-Sea Fluxes and Turbulence Parameterization

Air-sea interaction consists of multiple processes that are not well understood, and their parameterizations remain mostly empirical - from transfer coefficients for heat and moisture, to the effects of turbulence. The effect of buoyancy-induced sub-grid turbulence (CV) on air-sea fluxes in UWIN-CM is represented by a factor they called convective velocity, which attempts to represent the temperature and moisture instability of the surface layer. In
the surface layer parameterization chosen for this study, which is based on Monin-Obukhov similarity theory (WRF sf_sfclay\_physics option 1), different relationships are used to parameterize the convective velocity over land and over water. The modification to the formulation of CV is only implemented over water, as we have direct flux observations from R/V Revelle to evaluate the different model distribution.

The buoyancy-driven turbulence parameterization is added to the surface wind speed in UWIN-CM in order to prevent the model latent and sensible heat fluxes from approaching zero in low-wind regimes. From observations we know that that is not the case, so the reason for the existence of the parameterization is sound, though the parameterization itself is not perfect. The formulation of LHF or SHF in WRF and UWIN-CM is illustrated in equations 4.1 - 4.3:

\[ LHF / SHF = f(WSP, \Delta T, \Delta q) \]  
\[ WSP = \sqrt{WSP_{sfc}^2 + CV^2} = \sqrt{U_{sfc}^2 + V_{sfc}^2 + CV^2} \]  
\[ CV = \begin{cases} 
2\sqrt{\theta_{v,o} - \theta_{v,a}} & \text{when } \theta_{v,o} > \theta_{v,a} \\
0 & \text{otherwise.} 
\end{cases} \]

LHF and SHF depend mainly on wind speed, and temperature and moisture disequilibria \((\Delta T, \Delta q, \text{respectively})\) across the air-sea interface. WSP is the effective wind speed used in the LHF or SHF calculation, and is comprised of surface winds \((U_{sfc}, V_{sfc})\), and the buoyancy-driven turbulence \(CV\). The CF factor is present when the surface conditions are unstable with respect to buoyancy, which happens when the virtual potential temperature at the surface \(\theta_{v,o}\) is greater than it is just above the surface \(\theta_{v,a}\). This ensures that the effective wind speed \(WSP\) is never negative, and that when calculating the magnitude of LHF (and SHF), the effective wind speed is always greater than the surface wind speed in unstable conditions, and equal to the surface wind speed when conditions
are stable. The CV parameterization was created from data collected over Kansas, USA (Businger et al. 1971).

The only difference between \textit{AO4-TK} and \textit{AO4-TK-FLX} is that in \textit{AO4-TK-FLX}, we halve the magnitude of CV, effectively removing the two in front of the square root in Equation 4.3. Physically, that means that for the same temperature and moisture disequilibria across the air-sea interface, and the same surface winds, the SLF and LHF will be lower in \textit{AO4-TK-FLX} than in \textit{AO4-TK}. The distribution of CV magnitudes in \textit{AO4-TK} and \textit{AO4-TK-FLX} is shown in Figure 4.4 as a function of wind speed. In the original formulation used in \textit{AO4-TK}, the magnitude of convective velocity ranges between 2 and 5.5, which are reduced to a range between 1 and 3 in \textit{AO4-TK-FLX}. The frequencies displayed in Figure 4.4 are normalized by the total number of model points used, so the color indicates the percentage of points in any given wind speed and convective velocity. The fact that the magnitude of convective velocity in \textit{AO4-TK-FLX} scales according to the implemented modification implies that changing the parameterization of CV does not have a significant effect on near-surface stability, making the interpretation of the differences between \textit{AO4-TK} and \textit{AO4-TK-FLX} more easily attributable to the reduction in air-sea fluxes.

The VC distributions in Figure 4.4 do not show any dependence of CV on wind speed, though the distribution becomes more restricted at wind speeds $\geq 15$ m s$^{-1}$. This, the addition of CV to wind speed is most effective at low winds, where the buoyancy-driven turbulence contribution to wind speed can be many times greater than the wind speed itself; in high-wind regimes, its addition is not felt as strongly. This can be seen in the difference between the SHF and LHF distributions of \textit{AO4-TK} and \textit{AO4-TK-FLX} shown in Figure 4.3. Compared to \textit{AO4-TK}, the average LHF bias in \textit{AO4-TK-FLX} is reduced by $\sim 15\%$, but is improved by $\sim 35\%$ when only considering wind speeds $\leq 5$ m s$^{-1}$. The average SHF bias in \textit{AO4-TK-FLX} is reduced by $\sim 15\%$ compared to \textit{AO4-TK}, but almost 30$\%$ in low winds - though a reduction of SHF by 30$\%$ corresponds to a mean SHF reduction of less than 2 W m$^{-2}$ at low winds.
4.2.2 Wind Speed, Temperature, and Moisture Disequilibria

Besides being proportional to surface winds, the air-sea heat fluxes are also proportional to the temperature (SHF) and humidity (LHF) difference across the air-sea interface, as well as moisture and heat exchange coefficients. The moisture and heat exchange coefficients follow empirical parameterizations that depend on the surface wind speed and are not evaluated here. The distributions of surface winds, and those of moisture and temperature disequilibria across the air-sea interface are shown in Figure 4.5. The measurements shown in yellow used are from R/V Revelle and are taken from the same time as the data used to calculate the air-sea heat flux distributions.

Figure 4.5 a-c shows the frequency distribution of wind speed for R/V Revelle in yellow and model experiments in blue (UA4-TK), purple (AO4-TK), and red (AO4-TK-FLX). R/V Revelle wind speed observations exhibit a peak at \( \sim 3 \text{ m s}^{-1} \), and a secondary peak \( \sim 7 \text{ m s}^{-1} \); the frequency of stronger winds falls off sharply, especially above \( \sim 13 \text{ m s}^{-1} \). The observed pattern is best reproduced in AO4-TK-FLX, with a bimodal distribution with peaks \( \sim 4 \) and \( \sim 10 \text{ m s}^{-1} \), though the peaks are farther apart and both low winds and very high
winds are slightly underrepresented. Both $AO4-TK$ and $UA4-TK$ exhibit unimodal distributions, with peaks at $\sim 8$ and $\sim 7$ m s$^{-1}$, respectively. Though the wind speed distribution in $AO4-TK$ peaks at a higher value than that of $UA4-TK$, the peak in $AO4-TK$ is sharper, with high winds still being under-represented, along with winds $\leq 5$ m s$^{-1}$. In $UA4-TK$ the
frequency of low wind speeds occurrence is rarest, but high winds occur much more frequently, with winds $\geq 6 \text{ m s}^{-1}$ being over-predicted. The biases identified throughout this section are all calculated relative to wind speed, which means that the distribution of wind speed itself has no bearing on them. What is affected are the amounts of SHF and LHF actually entering the atmosphere at any one time - since the wind speed distributions in UWIN-CM are skewed toward high values, the high values of SHF and LHF occur more often than observed and this excessively warm and moisten the surface layer.

Figure 4.5 d-f shows the distribution of the moisture disequilibrium across the air-sea interface - the water vapor mixing ratio difference between the ocean surface ($q_s$) and the air just above it ($q_a$) in g kg$^{-1}$. When the difference is positive, which is the case throughout the modeled period, it means that water is evaporating from the ocean, and water vapor is being added to the atmosphere - the larger the difference is, the more evaporation occurs. The ocean surface is assumed to be saturated, and the mixing ratio of the ocean surface is equal to the saturation mixing ratio calculated using the SST, which means its solely a function of SST. This explains the discrepancy between the UA4-TK and the AO4-TK and AO4-TK-FLX experiments. In UA4-TK, the SST does not cool in response to strong winds and evaporation, so at a given location, $q_s$ is constant, and the magnitude of the moisture disequilibrium only depends on the amount of water vapor in the surface layer of the atmosphere. Regardless of wind speed, the mean moisture difference remains nearly constant, dropping from $\sim 8$ to $\sim 7.5$ g kg$^{-1}$, which is a difference of less than one part per thousand. In the coupled experiments, the ocean cools from initial conditions, and the $q_s$ decreases accordingly. The mean magnitude of the moisture difference decreases as wind speeds increase, implying that stronger winds induce more upper-ocean cooling (reducing $q_s$) and also cause more evaporation (increasing $q_a$), moistening the surface layer and reducing the moisture disequilibrium across the air-sea interface.

R/V Revelle observations show a large spread at low winds, with a mean $\sim 8$ g kg$^{-1}$, that drops sharply to below 7 g kg$^{-1}$ around 4-5 m s$^{-1}$, with some values as low as 5 g kg$^{-1}$. 
This cluster of data points comes from November 29 and 30, which is just after the MJO has passed to the east of R/V Revelle. Though the range of observed values are generally captured in the coupled experiments, the timing does not match. In UA4-TK, the moisture disequilibrium remains nearly constant relative to wind speed, and does not capture the reduction that was observed at wind speeds between 3 and 10 m s\(^{-1}\), but captures the mean observed values near the beginning and end of the wind speed distributions. In AO4-TK, the reduction \(\sim 3\) m s\(^{-1}\) is reproduced, but the moisture disequilibrium at higher wind speeds is lower than observed; the same is true in AO4-TK-FLX, though to a slightly lesser extent. In the coupled experiments, the average moisture disequilibrium bias is negative, while it is positive in the uncoupled experiment.

The distributions of the temperature disequilibria across the air-sea interface for model experiments and observations are shown in Figure 4.5 g-i. The temperature disequilibrium is the temperature difference between the ocean surface (SST, or \(T_s\)) and the air just above it (\(T_a\)) in °C. When the difference is positive, sensible heat is being added into the atmosphere, warming the atmosphere; the larger the difference, the more sensible heat can be added. There is no clear relationship between the magnitude of the temperature disequilibrium and wind speed, except that low temperature disequilibrium values do not occur at high wind speeds. There are some small negative values observed, and those indicate that the surface temperature in the atmosphere is warmer than SST, and sensible heat is leaving the atmosphere and entering the ocean. However, there are very few such cases at the R/V Revelle location during the modeled period, and when they occur, the temperature difference is small.

UA4-TK shows a large positive bias in the distribution of temperature disequilibria, producing consistently high values and indicating that the atmosphere is on average 2 - 4°C cooler than the ocean. Part of that is due to the \(T_s\) remaining constant since SST never changes. Another possible reason is that the atmosphere cools down, and that can be attributed to either cold air advection, or the production of convective cold pools from
excessive precipitation. In the coupled experiments, the biases are still positive, but of lower magnitude, and interestingly, the temperature differences bias in \textit{AO4-TK-FLX} is slightly larger than in \textit{AO4-TK}. That can be attributed to a weaker SST cooling in \textit{AO4-TK-FLX} compared to \textit{AO4-TK}.

In all three experiments shown, the biases in the temperature disequilibrium are positive, meaning that the temperature disequilibrium in the model is greater than in observations, which works to increase the SHF bias. The magnitude of the SHF bias itself is on the order of 1-10 W m\(^{-2}\), and does not seem to be greatly affected by the magnitude of the temperature disequilibrium. This can be deduced from the fact that the bias in temperature disequilibrium is higher in \textit{AO4-TK-FLX} than it is in \textit{AO4-TK}, though the SHF flux is decreased, indicating that wind speed plays a more important role. In \textit{AO4-TK-FLX}, the buoyancy-driven turbulence is reduced for a given temperature and moisture difference across the air-sea interface, and that change overpowers the positive bias in the temperature disequilibrium. A similar conclusion can be drawn for the moisture disequilibria, where \textit{AO4-TK} and \textit{AO4-TK-FLX} both show a negative mean bias, though the LHF bias is positive. Between the two coupled experiments, \textit{AO4-TK-FLX} has a smaller negative bias, and would this have a larger bias if the moisture disequilibrium was the determining factor.

The flux distributions shown in Figure 4.3 show a consistent high bias at all wind speeds (though the magnitude has been reduced both by coupling and the reduction in CV), indicating that the relationship between wind speed and air-sea flux is not well represented. Considering that wind speed seems to be the main contributor to the magnitude of SHF and LHF (and their biases in UWIN-CM), further investigation should focus on the parameterizations of moisture and temperature transfer coefficients and their role in reducing SHF and LHF biases, though that lies outside the scope of this study.

**4.2.3 Effect on Precipitation**

In previous sections, we have shown that consistent positive biases in LHF and SHF are produced in part by the surface layer parameterization that we use to run UWIN-CM.
The SHF biases may appear large in percentage, but are really not significant compared to the biases in LHF, as they are on average an order of magnitude smaller. SHF contributes to warming (or cooling) of the surface layer by its contact with the ocean surface, while LHF contributes to moistening (or drying) of the surface layer through evaporation, and thus provide moisture that helps produce or enhance precipitation.

The reduction of the LHF bias by $\sim 10 \text{ W m}^{-2}$ (more at low wind speeds) from $AO4-TK$ to $AO4-TK-FLX$ (Figure 4.3 b, c; Table 4.1) is qualitatively associated with the reduction in the amount of precipitation produced in UWIN-CM (Figure 4.1 c, d), and a smoother eastward propagation of the LPT track (Figure 4.2 c, d). In this section, we describe the effects on the amount and intensity of precipitation more quantitatively, by looking at time series of rainfall rates averaged over the area of the equatorial Indian Ocean that is covered by D03 ($55 - 97^\circ E$, $10^\circ S$ - $10^\circ N$) shown in Figure 4.6. It is analogous to Figure 3.6 shown in Chapter 3, but for model experiments analyzed in this chapter - $UA4-TK$, $AO4-TK$, and $AO4-TK-FLX$. As before, thin lines show hourly (UWIN-CM) and 3-hourly (TRMM) rainfall rates, and thick lines are smoothed using a 24-hour running mean to more clearly display the differences. The signal of the MJO in the IO is seen in TRMM observations between November 22 and 30, when the rainfall rates are relatively high with an average of 0.614 mm h$^{-1}$, followed by the suppressed period with an average rainfall rate of 0.162 mm h$^{-1}$.

The average rainfall rates for the entire time series, as well as the averages for the active and suppressed periods of the MJO are summarized in Table 4.2, followed by the average rainfall rate biases in Table 4.3. As before, the split between active and suppressed phases is based on the time that the observed MJO propagates out of the IO and into the MC on December 1. The highest rainfall rates are found in $UA4-TK$, the uncoupled experiment with constant SST, with average rainfall rates twice as high as those observed by TRMM. In terms of the strength of suppression, which is simply the difference between the average rainfall rates during the active and suppressed phases, the rainfall rates are actually
marginally enhanced when the observed MJO2 is in its suppressed phase (refer back to Figure 3.7, bottom), though the difference is negligible. We can conclude that the UA4-TK produces no large-scale suppression, and consequently no eastward propagation of precipitation (Figure 4.2), and is thus unsuccessful at simulating an MJO, which was already concluded based on the eastward propagation speed of the UA4-TK LPT.

Table 4.2: Rainfall rates averaged over the IO portion of D03 (as in Figure 4.6). Total column shows the average for the entire time series, while the Active and Suppressed columns refer to the time periods between November 22 and 30 (active) and December 1 and 6 (suppressed). The Difference column shows the difference between the average rainfall rates during the active and suppressed phases.

<table>
<thead>
<tr>
<th>Experiment</th>
<th>Total</th>
<th>Active</th>
<th>Suppressed</th>
<th>Difference</th>
</tr>
</thead>
<tbody>
<tr>
<td>UA4-TK</td>
<td>0.9195</td>
<td>0.9167</td>
<td>0.9235</td>
<td>-0.0068</td>
</tr>
<tr>
<td>AO4-TK</td>
<td>0.7142</td>
<td>0.8137</td>
<td>0.5648</td>
<td>0.2489</td>
</tr>
<tr>
<td>AO4-TK-FLX</td>
<td>0.6168</td>
<td>0.7221</td>
<td>0.4590</td>
<td>0.2631</td>
</tr>
<tr>
<td>TRMM</td>
<td>0.4325</td>
<td>0.6137</td>
<td>0.1619</td>
<td>0.4531</td>
</tr>
</tbody>
</table>
The coupled experiments \((AO4-TK\) and \(AO4-TK-FLX)\) both produce significantly less (and less intense) precipitation, as well as a distinct period of suppression after the main body of the model MJO has passed out of the IO, and into the MC, so both of them produce MJO-like convective envelopes. The precipitation bias is still high, especially in the suppressed phase, but it is much improved from the bias shown in \(UA4-TK\). The modification of the buoyancy-driven turbulence in \(AO4-TK-FLX\) reduces the precipitation bias of \(AO4-TK\) by over 50\% when considering the IO region, with a greater improvement during the suppressed phase, showing that the suppression is more intense in \(AO4-TK-FLX\).

The way precipitation is generated in UWIN-CM could explain some of the differences described in in Chapter 3 - between low resolution experiments with convective parameterization, and high resolution experiments with explicitly resolved convection. But the distribution of rainfall rates is not the cause of the differences in the precipitation amount among the high-resolution coupled and uncoupled experiments. The experiments exhibit nearly identical rainfall rate distributions, with \(UA4-TK\) producing marginally less low rainfall rates \((\leq \sim 0.5 \text{ mm h}^{-1})\), and marginally more rainfall rates between \(\sim 1\) and \(12 \text{ mm h}^{-1}\) compared to \(AO4-TK\) and \(AO4-TK-FLX\). Instead, the differences in the amount of precipitation in the high resolution experiments can be directly related to the reduction of the positive LHF bias, and consequently the amount of moisture that enters the atmosphere through evaporation from the ocean surface.

Table 4.3: Same as Table 4.2, but for rainfall rate bias in \(\text{mm h}^{-1}\) followed by the \% overestimation of TRMM rainfall rates in parentheses. A 100\% rainfall rate bias indicates that model rainfall rates are twice as high as observed.

<table>
<thead>
<tr>
<th>Experiment</th>
<th>Total</th>
<th>Active</th>
<th>Suppressed</th>
</tr>
</thead>
<tbody>
<tr>
<td>UA4-TK</td>
<td>0.482 (292.4%)</td>
<td>0.297 (52.8%)</td>
<td>0.760 (801.7%)</td>
</tr>
<tr>
<td>AO4-TK</td>
<td>0.280 (115.9%)</td>
<td>0.196 (33.8%)</td>
<td>0.405 (389.2%)</td>
</tr>
<tr>
<td>AO4-TK-FLX</td>
<td>0.183 (59.9%)</td>
<td>0.106 (18.6%)</td>
<td>0.299 (271.8%)</td>
</tr>
</tbody>
</table>
4.3 Upper Ocean Structure

We have so far mostly shown analyses of the atmospheric variables related to the simulation and eastward propagation of the MJO. In this section, we address the state of the upper ocean, and its evolution throughout the simulation. We look at the HYCOM global analysis fields that are used to initialize the ocean component of UWIN-CM, and how they compare to observations in section 4.3.1. We analyze the evolution of the upper ocean in terms of temperature, salinity, and density to see how the passage of the MJO affects the upper ocean. Only the AO4-TK and AO4-TK-FLX experiments are compared, since UA4-TK does not include an ocean component. The upper-ocean structure and the effect of the MJO forcing on upper-ocean cooling are discussed, with a focus on the analysis of mixed-layer depths in UWIN-CM and linking them to why the SST does not cool as much as it does in observations.

The ocean state analysis is more complicated than the analysis of the atmospheric fields due to a lack of observations in the ocean. The DYNAMO field campaign included three surface and sub-surface moorings (M1, M2, and M3), located at equator and 79°E, 1.5°S and 79°E, and 9.75°S and 78.5°E, respectively. During MJO2, the surface sensors measuring SST were not operating on M1, and only collected data from M2 for half of the model period, so the sub-surface analysis of the upper ocean fields only includes values starting at a depth of 1 m. Besides the moorings, the ship sites R/V Revelle, located at the equator and 80.5°E, and R/V Mirai, located at 8.0°S, 73.15°E collected some ocean data, though they only record data at the surface and at depths of 5- and 10-m. Observations from the RAMA mooring array in the IO are also available, though the salinity data is only recorded by six of the eighteen stations during the modeled time period. Besides the stationary observation sites that provide data throughout the modeled period, there are some AXBT and AXCTD data available. These were dropped by the WP-3D aircraft, and provide high-resolution data in the vertical. These data are used together to gain an insight of how the UWIN-CM initial state compares to observations, and evolved during the modeled period.
4.3.1 Initial State of the Upper Ocean in the Coupled Model

The passage of MJO2 over the Indian Ocean induced a large-scale SST and upper ocean cooling of almost 2°C over the course of 13 days. Though the coupled model experiments reproduce the cooling trend, they underestimate the magnitude of the cooling and its rapid onset. The warm bias in SST produced by the model can, through increased evaporation, contribute to the positive bias in precipitation that has been described in Section 4.2. Because of the ocean’s high density compared to that of the atmosphere, adjustments to atmospheric forcing take longer, and the effect of initial conditions, whether accurate or inaccurate, can be present for days. It is therefore important to examine the conditions that are used to initialize the ocean component of UWIN-CM. Figure 4.7 shows the difference between SSTs of HYCOM global analysis ocean initial condition (shown in Figure 2.3) and the SST obtained from ECMWF analysis that is used as an initial condition for the atmosphere. Positive values are indicated in red and show where HYCOM analysis is warmer than the ECMWF analysis.

The average difference between the SST in HYCOM and ECMWF analyses is small, indicating that HYCOM global analysis shows almost no overall bias in the region we are considering, though there are local regions where biases are present. For example, the SSTs in the MC and west of Sumatra are warmer in HYCOM, as are the seas west of Australia, but the Bay of Bengal and the seas east of the Philippines are cooler. We initialize the ocean component of UWIN-CM with HYCOM global analysis SST values so that the dataset for the entire ocean comes from the same source. This might have caused a short period of disequilibrium and initial adjustment in the lower levels of the atmosphere, as the given lower boundary condition source is different from the rest of the atmosphere data, but the atmosphere adjusts to disequilibria much faster than the ocean. Since we are taking the ocean initial conditions (including the SST) from HYCOM global analysis without changing the resolution, the model fields should be in equilibrium and there should be no adjustment period for the ocean.
Figure 4.7: SST difference between HYCOM and ECMWF global analyses. Red colors indicating that HYCOM SSTs are warmer than those in ECMWF analysis.

Though there is no average bias in the SST data when comparing the SSTs of the atmosphere and ocean initial conditions, the state of the upper ocean is more difficult to verify. Since there is no comprehensive dataset for upper-ocean data, the only way to evaluate the accuracy of the ocean initial condition data is by using point-by-point comparison. The initial profiles of temperature, salinity, and density (which is mostly determined by temperature and salinity) from DYNAMO moorings M1 and M2 are shown in black in Figure 4.8, with the profiles at the nearest UWIN-CM points in red. The solid lines in both cases show the profiles of M1, while the dashed lines show those of M2. In the observed profiles, the thermocline region (in which temperature decreases by $\sim9^\circ$C over a depth change of $\sim50$ m) starts at $\sim60$ m in M1 and $\sim40$ m in M2. Both M1 and M2 exhibit an increase in salinity down to a depth of $\sim40$ m, below which salinity decreases, but the salinity values near the surface are quite different, with M1 having much higher salinity than M2. This is probably due to precipitation over M2 freshening the surface layer. The density profiles seem to be more sensitive to the changes in temperature than they are to changes in salinity, where the
M1 profile is denser at the surface than M2, with both higher temperature and salinity. M2 has higher density below 40 m, at depths at which the temperature is lower than in M1, and the salinity does not decrease as fast.

![Profile comparison](image)

**Figure 4.8:** a) Temperature (in °C), b) salinity (in psu), and c) potential density ($\sigma_0$, in kg m$^{-3}$) profiles from M1 (solid) and M2 (dashed); observations are shown in black, and the corresponding profiles at UWIN-CM initialization time are in red.

The shapes of the temperature and density profiles are generally reproduced in UWIN-CM, though with some minor differences, but the initial salinity profiles are unusual and very different from observations. Both UWIN-CM salinity profiles have a very sharp salinity maximum between 70 and 100 m, followed by a salinity minimum between 100 and 160 m. In the model M1, the minimum is followed by a second salinity maximum below 160 m. This representation of the salinity profile is not limited to the two profiles shown here, but can be seen in the shape of the salinity distribution shown in Figure 4.9. The distribution takes into account all the model points that fall within 2° of either of the mooring locations M1 and M2. A trace of the salinity pattern described above can be seen even when the profiles from the entire IO are considered, though the distinct pattern is smoothed out.
The UWIN-CM temperature profile at M1 is more than 1°C cooler than observed, and the profile remains cooler at depths below 120 m, but in between the model temperature profile is warmer. The temperature over the top 60 m changes little, with an isothermal surface layer much thicker than was observed. At M2, the SST is very close to the observed SST, but decreases constantly and gradually throughout the column. The unusual salinity profile seems to enhance the vertical gradients in density, though the shape of the density profile seems to mostly depend on temperature. Though the UWIN-CM profiles at M1 and M2 are initially cooler than mooring observations, the large-scale initial SST shows only small differences from observed values.

### 4.3.2 Evolution of the Upper Ocean During the MJO

Relatively strong surface winds, extensive cloud cover and intense precipitation associated with the MJO all contribute to the cooling of the SST and the upper ocean. This cooling, that is generally reproduced in the coupled experiments, helps create an environ-
ment that is unfavorable for sustained precipitation, and improves the eastward propagation of the MJO in UWIN-CM. The upper-ocean cooling is limited by the fact that the mixed layer in the upper ocean is almost twice as deep as what was observed - a problem stemming from a bad initial condition. Figure 4.10 shows the time evolution of temperature (in °C; a-c), salinity (in psu; d-f) and density (in kg m\(^{-3}\); g-i) in the upper ocean at the M1 location. The left column shows the observed evolution, followed from left to right by AO4-TK and AO4-TK-FLX. Black circles indicate the mixed layer depth (MLD) that is chosen at a point where the density difference from the surface value first reaches 0.05 kg m\(^{-3}\). The value is estimated for both model and observations so that the MLD is calculated using the same method in all datasets. The 0.05 kg m\(^{-3}\) threshold was chosen based on criteria from Thomson and Fine (2003), though it lies on the high end of suggested values. It was chosen so that the calculated MLD is close to the model-calculated MLD, though the one produced by the model is still deeper than the calculated one.

Figure 4.10: Time evolution of temperature (in °C; a-c), salinity (in psu; d-f), and potential density (\(\sigma_0\) in kg m\(^{-3}\); g-i) at M1. Mooring observations are shown on the left (a, d, g), followed by AO4-TK (b, e, h) and AO4-TK-FLX (c, f, i). Black dots indicate the mixed layer depth.
Before the winds and precipitation associated with the MJO reach the locations of DY-NAMO moorings on November 24, there is little change in the vertical structure of the upper ocean. As the first wave of intense MJO convection passes over, the SST cools by $\sim 0.8^\circ$C and the mixed layer deepens, while the high salinity layer is pushed deeper. During the passage of the second wave of MJO convection on November 27, SST cools by additional $\sim 0.3^\circ$C and the mixed layer deepens even further. The layer of high salinity is either diffused or advected away from M1, though a local salinity maximum at a depth of $\sim 60$ m persists throughout. After the MJO-associated precipitation has moved east of the moorings, the SST slowly recovers, the mixed layer shoals, and the surface salinity rapidly decreases. The same pattern in SST is present at M2 (not shown), but the mixed-layer salinity that is initially lower than at M1 actually increases during the passage of the first wave of convection, then the mixed layer freshens and stays fresh until the end of the modeled period.

$AO4-TK$ and $AO4-TK-FLX$ both start off $\sim 1$C cooler than observed, and cool by $\sim 0.4^\circ$C during the passage of the first convective envelope, and an additional $0.2^\circ$C during the passage of the second one; UWIN-CM starts off cooler than observed, and ends up warmer. As mentioned in previous sections, UWIN-CM does not cool the ocean as extensively, or as rapidly as what was observed. Similar to observations, the upper-ocean profiles do not change much over the first two days, after which the ocean MLD starts to steadily increases until December 2, after which it remains nearly constant $\sim 90$ m. The differences between $AO4-TK$ and $AO4-TK-FLX$ are small. Both model experiments show a positive bias in the depth of the mixed layer, with an average bias of 35.7 m in $AO4-TK$ and 41.7 m in $AO4-TK-FLX$, which is an overestimation of the observed MLD by 50 and 80%, respectively. These biases are calculated with the MLD density threshold of 0.05 kg m$^{-3}$; the biases for other thresholds are listed in Table 4.4 along with mean MLD values for each of the thresholds.

The positive bias is present regardless of what density threshold is used for calculating the MLD, and helps to explain why the SST in UWIN-CM does not cool as rapidly and as
Table 4.4: Ocean mixed layer depth (m) and the average bias in mixed layer depth given in m and % of observed value for different thresholds of density change from the surface.

<table>
<thead>
<tr>
<th>Threshold (kg m(^{-3}))</th>
<th>Observed (m)</th>
<th>AO4-TK</th>
<th>AO4-TK-FLX</th>
<th>AO4-TK bias</th>
<th>AO4-TK-FLX bias</th>
</tr>
</thead>
<tbody>
<tr>
<td>0.01</td>
<td>23.83</td>
<td>51.85</td>
<td>50.00</td>
<td>27.53 (43.8%)</td>
<td>25.44 (28.1%)</td>
</tr>
<tr>
<td>0.02</td>
<td>26.63</td>
<td>60.23</td>
<td>62.60</td>
<td>33.00 (56.0%)</td>
<td>35.29 (69.8%)</td>
</tr>
<tr>
<td>0.03</td>
<td>28.78</td>
<td>63.83</td>
<td>69.54</td>
<td>34.77 (55.6%)</td>
<td>40.19 (84.8%)</td>
</tr>
<tr>
<td>0.04</td>
<td>30.48</td>
<td>65.91</td>
<td>72.81</td>
<td>35.20 (51.2%)</td>
<td>41.73 (85.9%)</td>
</tr>
<tr>
<td>0.05</td>
<td>32.03</td>
<td>67.66</td>
<td>74.17</td>
<td>35.69 (50.0%)</td>
<td>41.73 (79.6%)</td>
</tr>
</tbody>
</table>

The cooling induced by strong surface winds is mixed down by the ocean model, but it is mixing down to 50 - 100 m, and the surface wind forcing necessary to induce a 1°C cooling over such a deep layer is much greater than what was produced in UWIN-CM. The excessively deep ocean mixed layer prevents the ocean surface from cooling to the extent, and at the rate, that is observed, creating SSTs that are warmer than observed. The presence of overly warm SSTs means that \( q_s \) is higher, which contributes to evaporation and helps to sustain precipitation by supplying moisture to the atmosphere.

The bad initial condition in the salinity field is propagated throughout the simulation in both coupled experiments. The tongue of high salinity that is present at ~50 m depth in the observations is shifted downward, with maximum salinity values ~80 m, and it slowly gets diffused and/or advected away from the positions of M1 and M2, but the fresh layer below it persist almost throughout the simulation. Besides the representation of salinity under the surface, UWIN-CM doesn't seem to represent the surface salinity well either, with the observed trend of freshening as time progresses being reversed. The model starts
with a relatively fresh surface that then increases in salinity throughout the period. The temperature, salinity, and density distributions are depicted in Figure 4.11, and they more clearly show the differences between the observations and model experiments, though the time information is lost. The black lines indicate the $10^{th}$ and $90^{th}$ percentiles of observed profiles at any given depth, and help illustrate the difference between modeled and observed profiles.

### 4.4 Summary

In this chapter, we addressed the effects that different types of atmosphere-ocean coupling have on the representation of the MJO2 event in UWIN-CM. We have done so by comparing an uncoupled, atmosphere-only experiment, in which the SST is held constant from initial conditions, to two atmosphere-ocean coupled experiments where the atmosphere and ocean components are coupled together every two minutes. The difference between coupled experiments is that in one of them, we attempt to reduce the air-sea latent and sensible heat fluxes by modifying the parameterization of buoyancy-driven turbulence (CV) in the surface layer of UWIN-CM. The results are robust regardless of what type of convective parameterization is used in the coarser domains, so only the results from experiments using the Tiedtke convective parameterization (TK) are shown.

We have found that in the uncoupled simulation ($UA4-TK$), there is no MJO and associated eastward propagation; instead the precipitation is produced nearly constantly and over the entire equatorial IO, and expands eastward over the MC. In $UA4-TK$, strong surface winds blow over an ocean that never cools in response to surface winds, which produces a lot of evaporation; the moisture influx into the atmosphere promotes and sustains the precipitation over the IO and never produces a suppression. In the coupled experiments, the MJO-induced SST and upper-ocean cooling reduce the amount of air-sea fluxes being imported into the atmosphere, creating an environment unfavorable for sustained precipitation, contributing to the suppression of precipitation behind the MJO. Though the pre-
Figure 4.11: Frequency distributions of a-c) temperature (in °C), d-f) salinity (in psu), and g-i) potential density ($\sigma_0$ in kg m$^{-3}$) combining profiles from M1 and M2. Mooring observations are shown on the left (a, d, g), followed by AO4-TK (b, e, h) and AO4-TK-FLX (c, f, i). The black lines show the range of the 10$^{th}$ and 90$^{th}$ percentile of observed values at a given depth.
Precipitation is suppressed after the MJO relative to the amount and intensity of precipitation produced during the active MJO convection over the IO, it is nowhere near as suppressed as what was observed. Parts of that can be attributed to the high LHF bias and the fact the upper ocean in UWIN-CM does not cool at the rate that was observed. Evaporation rates remain relatively high, providing moisture to sustain precipitation. The lack of ocean cooling is linked to the overly deep mixed layer in the ocean model - any cooling of the surface that occurs through evaporation is being mixed down into a layer that is between 50 and 100 m thick, reducing the effect of surface cooling due to the large volume of warm water it is being mixed with.

The lack of ocean cooling is not the only reason why UWIN-CM produces excessive precipitation. We identify a large positive bias in the parameterization of air-sea fluxes at all wind speeds, that arises from the surface layer parameterization in WRF (and consequently, UWIN-CM). The mean latent heat flux bias of 71%, or 83.1 W m$^{-2}$ in the uncoupled configuration is reduced to 43%, or 46.47 W m$^{-2}$ with the addition of atmosphere-ocean coupling and the cooling SSTs it brings. However, the bias is still high, and is increased to 82% when only wind speeds $\leq$ 5 m s$^{-1}$ are considered. In the simulation where we reduce the magnitude of CV, which is added to wind speed when air-sea fluxes are calculated, the air-sea flux bias is further reduced, especially at low wind speeds, and the amount of precipitation is further reduced. AO4-TK-FLX also produces weaker surface winds and less SST cooling. This effect counteracts the change we implemented in the model, as the higher SSTs allow for higher air-sea fluxes regardless of the change in CV that was implemented.

The experiments and analysis in this chapter show that atmosphere-ocean coupled models perform better with respect to the MJO. But no matter how good or relevant the corrections that we make to the atmosphere model are, including an ocean component in the coupled system adds another layer of complexity to interpreting results, and many more possibilities for introducing errors into the forecast. From the results of this study, we find
that the presence of a dynamic ocean model strongly contributes to the successful simulation of the MJO, though there are many areas of the coupled modeling system that can be improved upon.
Chapter 5

Conclusions

5.1 Summary of Key Results

The aim of this study was to identify and better understand a few of the important processes that contribute to the formation and eastward propagation of the MJO. Atmosphere-ocean (coupled) and atmosphere-only (uncoupled) UWIN-CM simulations were analyzed to investigate the effects of convective parameterization, resolution, and atmosphere-ocean coupling on the simulation of the MJO. Not all experiments successfully simulated the observed MJO event, but the high resolution, atmosphere-ocean coupled experiments generally performed better, simulating two distinct eastward-propagating convective features that were consistent with DYNAMO observations. The combination of detailed observations taking measurements of the MJO2 event over the central IO and the high-resolution UWIN-CM data provided a great opportunity to examine how the processes that seem critical for a successful simulation of the MJO are represented in numerical models. Detailed analysis of the model results, their differences, and deviations from observations, have led to the following conclusions:

1. UWIN-CM experiments at relatively low (12 km) grid resolutions are sensitive to the choice of convective parameterization, which affects the ability to simulate an MJO, the amount of precipitation, and the boundary layer structure. No MJO occurs when the Kain-Fritsch convective parameterization is used (in \textit{AO12-KF}), but one with strong eastward propagation is produced with the Tiedtke convective parameterization (in \textit{AO12-TK}). This agrees with previous climate model studies, in which the convective parameterization that use a moisture convergence-related convective trigger (like in TK CP) tend to produce better MJOs (Hung et al. 2013). But even in \textit{AO12-TK}, the precipitation pattern associated
with the observed MJO is not reproduced; the two distinct convective envelopes features observed during MJO2 (Figure 2.2) are merged into one continuous feature of intense precipitation near the equator in \textit{AO12-TK} (Figure 3.1 c). The amount of precipitation and its intensity both have a positive bias, and UWIN-CM is not able to suppress the precipitation over the IO to the extent that it is suppressed in observations.

The choice of convective parameterization also have an effect on the mixing and entrainment that occur across the top of the boundary layer. In \textit{AO12-KF}, the mixing is very weak, and the westerly jets that descend all the way to the surface in observations remain confined above the boundary layer, with mostly easterly winds at the surface. According to Mallard et al. (2013), this could be linked to the overproduction of tropical cyclones in simulations that use the KF CP, which is what occurs in \textit{AO12-KF}. In \textit{AO12-TK}, the opposite seems to happen, with mixing being too strong, and the moisture and winds between from the surface up to 2 - 4 km often remain homogeneous, and the descending westerly wind jets are confined close to the surface. The observations lie somewhere in between what was produced in low resolution experiments, and is better reproduced when the resolution is increased to a point where convection no longer needs to be parameterized.

(2) Increasing resolution to a cloud-permitting (4 km) resolution improves simulations of MJO2 regardless of the convective parameterization used, with an MJO present even in the case where is no MJO low-resolution counterpart experiment. When convection is resolved explicitly, differences between experiments are subtler compared to the differences between low- and high-resolution experiments. The MJO at high resolution becomes more comparable to observations in terms of surface winds, precipitation, and upper-ocean cooling. Mixing across the top of the boundary layer is more realistic when no convective parameterization is used, and better reproduces the observed vertical profiles of temperature, relative humidity, and winds. Though the coupled high-resolution experiments produce MJO-like precipitation and eastward propagation, there is a high precipitation bias present, and the large-scale suppression over the IO is not as intense as observed.
(3) The representation of the MJOs eastward propagation strongly depends on whether the model can reproduce the period of large-scale suppression that occurs after an MJO event passes through. If that suppression is not present, then even if some part of convection propagates east, it would be considered an extension of the precipitating feature, rather than a propagation, as its western edge remains anchored over the IO. The eastward extension of precipitation occurs in uncoupled experiments, and we can clearly see it when using LPT tracking (Figure 4.2 b). It is also a common feature in many climate models (Lin et al. 2006; Hung et al. 2013), where the precipitation persisting over the IO causes the MJOs in climate models to only exhibit a weak eastward propagation. In UA4-TK, the passage of the MJO does not lead to a cooling response in the upper ocean - strong winds blowing over constantly warm water provide a continuous source of moisture to the atmosphere. The influx of moisture can sustain precipitation over the IO, as the environment never becomes unfavorable for development of intense convection.

The moisture flux into the atmosphere is exaggerated not only due to strong winds and warm water, but also because of how air-sea fluxes are parameterized in the surface layer of the atmosphere model. In coupled model experiments, even when modifying the parameterization in order to reduce the latent heat flux bias, the result is a consistent positive bias that is present at all wind speeds. The reduction of the latent heat flux bias results in a reduction of the amount and intensity of precipitation, and a smoother and more extensive eastward propagation (due to a clearer suppression of precipitation over the IO). However, the bias in precipitation and latent heat fluxes persists, and is most likely related to the inaccurate parameterizations in the air-sea moisture and heat transfer coefficients.

Atmosphere-ocean coupled models such as UWIN-CM have a great advantage when it comes to modeling of tropical phenomena such as the MJO, as they are able to resolve the interactions between the two components of the Earth system. When atmosphere-only models are used to try an simulate an individual event, they are either forced to keep a
constant SST, or they need to update it from either analysis data or observations. We have shown in this study that in the case where SST is held constant, a lot of rain is produced over entire high resolution domain (D03) and through the entire 15-day simulation period. Though there have been studies that show that the role of the ocean and air-sea coupling is not consistent for all MJO events, with some MJOs being mostly controlled by atmospheric dynamics, while others are strongly coupled to the underlying ocean (Fu et al. 2015; Wang et al. 2015). We speculate that the ocean and air-sea interaction play a role in all MJO events, though the magnitude and extent of upper-ocean cooling are not always as strong as those caused by MJO2; uncoupled models may have better skill reproducing the cases in which the upper-ocean cooling is weak or nonexistent.

Based on this study, however, we believe that when trying to predict MJO events that have not yet occurred, atmosphere-ocean coupled models should be used for their prediction. Partly because we have no observations of future SSTs that we could use as the lower boundary conditions for an uncoupled model, and partly because the extent of ocean cooling that an MJO event is going to induce is not known in advance. However, with coupled models, two sets of initial conditions are needed, one for the atmosphere, and one for the ocean. Large amounts of observations of the atmosphere are collected daily, and they are being incorporated into models to produce initial and boundary conditions that are very good, compared to the quantity of observations that go into improving ocean models. Oceans have a higher density than the atmosphere, so the motions are relatively slow, and take a longer time to adjust to forcing. We see an example in this study, where a bad initial condition in salinity persists through most of the 15-day simulation. Thus we believe that in order to improve the prediction of the MJO in high-resolution numerical models, resources should be devoted to the understanding and better representation of the ocean conditions and processes of air-sea interaction over the IO and MC.
5.2 Future Work

This study showed that in a regional model, atmosphere-ocean coupling and explicitly resolved convection are necessary to realistically simulate the MJO precipitation features and its eastward propagation. With successful MJO simulations from the \textit{AO4-TK} and \textit{AO4-TK-FLX} experiments, we would like to focus on the MC barrier effect - the role that the complicated topography of the MC plays in the eastward propagation of the MJO. Specifically, we would like to address the question of how the diurnal cycle of precipitation over the MC islands is affected by the passage of the MJO, and what impact the changes have on the MJO propagation itself. Kerns and Chen (2016) find that about one third of the MJO events initiated over the IO do not propagate all the way across the MC and into the western Pacific, and this trend is further exaggerated in numerical models (Inness and Slingo 2006).

Studies have shown that in climate models, it is the orography of the islands, and not their presence, that blocks the MJO signal from propagating eastward across the MC (Inness and Slingo 2006), especially considering the high mountainous terrain of Sumatra, Borneo and New Guinea islands. It has been suggested that resolving the detailed topography of the region produces more realistic MJO simulations (Wu et al. 2009). As the convective envelope of the MJO is propagating through the MC, it interacts with the diurnal cycle of precipitation over islands (Peatman et al. 2014) and the waters that separate them (Napitu et al. 2015). Using a coupled model and modifying the land distribution and topography over the MC provides a unique tool that can be used to study the MC barrier effect and changes to the Indonesian throughflow that provides inter-ocean exchange pathways between the Pacific and Indian Oceans (Gordon 2005).
References


Madden, R. A. and Julian, P. R. (1972). Description of global-scale circulation cells in the tropics with a 40-50 day period.


