On the Relative Role of Shallow Convection in Tropical Variability

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UNIVERSITY OF MIAMI

ON THE RELATIVE ROLE OF SHALLOW CONVECTION IN TROPICAL VARIABILITY

By

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ON THE RELATIVE ROLE OF SHALLOW CONVECTION IN TROPICAL VARIABILITY

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Tropical and subtropical clouds have a large influence in shaping the observed patterns of wind, temperature, and moisture in many time scales and different regions of the world. While shallow clouds have been traditionally investigated in trade wind regimes, shallow clouds in the deep tropics have recently gained attention because of their possible importance in the development of deep convection. Shallow clouds are ubiquitous in the tropics and connect the boundary layer and the lower troposphere. Difficulties to observe and model shallow clouds have resulted in unclear aspects about their interactions with the other elements of the tropical variability. It is a challenging task to address such interactions, but the objective is well worth to pursue.

This study takes a three-pronged approach to advance the understanding of the role of shallow convection in the variability of the deep tropics. The first part of this study investigates the variability and feedbacks of shallow clouds in their large-scale environment from more than ten years of observations of a vertically pointing cloud radar and soundings over Manus Island, in the tropical western Pacific. To do so, it is developed a method to estimate bulk shallow cloud moistening due evaporation of cloud condensed water from observations of liquid water content, precipitation, and temporal cloud fractions. The second part of this study combines the Manus and reanalysis data to investigate the role of shallow convection in the evolution of large-scale convective
events associated and not associated to Madden-Julian Oscillation (MJO). The third part of this work concerns the role of shallow convection in driving large-scale circulation. To investigate this problem, the region of study is focused on the tropical Atlantic. Surface equatorial winds over the equatorial Atlantic in models are notorious for their westerly biases and their association to Amazonian rainfall. This part of the study investigates the connection between the westerly biases and shallow convection over the Amazonia using simulations of several atmospheric global climate models.

Our results indicate that shallow clouds provide a non-negligible amount of moisture to the lower troposphere. They, however, lack coherency with perturbations in low-tropospheric moisture, temperature, and wind circulation in synoptic time scales. In longer time scales, anomalies in their depth, occurrence, and estimated moistening can be, to a certain degree, related to those in lower tropospheric stability. Consistently, during the five to ten days prior to the rainfall peaks of MJO and non-MJO large-scale convective events at Manus, anomalous increases in low-level moisture are evident, but they cannot be attributed to moistening by shallow clouds. During this period, shallow clouds provide background moistening. The characteristic low-level moistening prior the rainfall peaks of MJO events is mainly caused by anomalous nonlinear zonal advection.

The analysis about the role of shallow convection in the MJO revealed MJO signals that have not been observed before. They include an ultra slow (2.8 m s\(^{-1}\)) structure in mid to upper tropospheric temperature and geopotential height anomalies that propagates eastward over the central and eastern Pacific with no discernable anomalies in precipitation associated to it. These results suggest the existence of a intrinsic structure of
the MJO governed by dry dynamics. Different process might act to energize this structure. Moist convection, including shallow convection, can be an effective one.

The importance of shallow convection in the tropical variability and circulation seems to be case-dependent and subject to the interaction with other factors. This is the case in the equatorial Atlantic. Our results indicate that westerly biases over this region can be associated to weak or absent Amazonian shallow convection, but they can also be associated to weak boundary layer entrainment.
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CHAPTER 1: Introduction

The tropical western Pacific, Amazonia, and West Africa are among the most important regions of atmospheric moist convection (convection hereafter) in the world. These tropical centers of convection, and the circulation they produce, shape the mean tropical circulation (Gill 1980). Understanding the causes of the variability in convection is important not only for scientific reasons, but also for socio-economic ones. Atmospheric and oceanic phenomena in the tropics, such as the Madden Julian oscillation (e.g. Zhang 2013) or ‘El Niño’ the Southern Oscillation (ENSO, Battisti and Sarachik 1995), have a large influence in global weather forecasts. Global climate models (GCMs) however, have severe difficulties not only to produce reliable forecasts, but also to reproduce the mean tropical climate. For example, they have historically failed to reproduce the Amazonian and West African rainfall and the equatorial easterlies along the Atlantic Ocean (Richter and Xie 2008). One common weakness in models is their representation of convection, and particularly its representation of shallow convection (Tiedke 1989). Models that adequately represent shallow convection have shown significant improvements in many aspects (Tiedke 1989; Bretherton and Park 2008), including the MJO.

Shallow cumulus clouds (shallow clouds hereafter) are the most populous group of clouds in tropical and subtropical regions. They are responsible for about 20% of the total tropical rainfall (Short and Nakamura 2000; Lau and Wu 2003; Masunaga and Kummerow 2006). The importance of low-level clouds has been subject of many studies in the context of the subtropical climate and variability (Klein and Hartman 1993; Bony
and Dufresne 2009), and to lower extend, in the deep tropics. Subtropical low-level clouds thrive over the eastern ocean basins, where relatively cold ocean surface temperatures and warm sinking air results in a large vertical temperature inversion. Subtropical shallow clouds are key to conserve the energy balance of the earth because they reflect back to the space shortwave radiation, and they moisten the trades to sustain a moisture flux towards the Inter-Tropical Convergence Zone (ITCZ).

Shallow clouds in the deep tropics have recently gained attention because of their possible importance during the development of tropical perturbations. The cloud evolution in tropical perturbations typically follows a pattern characterized by an initial shallow-cloud stage, which is followed by a deep-cumulus stage and a stratiform stage (Mapes et al. 2006). Yanai et al. (1973) first explicitly pointed out the possible importance of moistening of the lower troposphere by shallow cumulus clouds in the development of deep convection. The connection of shallow clouds with the boundary layer makes these clouds capable of modulating low-tropospheric moistening. Positive low and mid-level moisture variations diminish the detrimental effects of dry air in buoyant air parcels. Shallow convection through their diabatic heating effects drives low-level and surface winds that may promote moisture convergence and enhance surface fluxes (Wu 2000, 2003; Zhang and Hagos 2009; Bretherton and Park 2008). In addition to the importance of the moistening and heating effects of shallow clouds, radiative cooling from these clouds is crucial to explain convective aggregation in idealized numerical simulations (Bretherton et al. 2005; Muller et al. 2013).

The role of shallow cloud moistening and heating in tropical perturbations is a hypothesis, not an observed fact, however. Such hypothesis is based mainly on observed
coherent increases in low-level moisture and shallow clouds (Kemball-Cook and Weare 2001), but it has been supported also by numerical studies (Benedict and Randall 2007). For example, previous studies found that adequate MJO simulations require sufficient low-level diabatic heating (Li et al. 2009; Zhang and Song 2009; Cai et al. 2013; Lappen and Schumacher 2014). There is no doubt that the moistening and heating effects from shallow clouds exist and are important to the low tropospheric environment. But it is unclear if increased shallow convection is a cause to promote deep convection, or an effect of other mechanisms, for example, the large-scale circulation. Moistening by clouds and advection needs to be carefully scrutinized before a final conclusion can be unambiguously supported on this matter. This study investigates the role of shallow clouds in the variability of the deep tropics, and it aims to better understand the importance of their moistening and heating effects with respect to that of other processes in these regions.

There are several challenges in quantifying the effects of shallow clouds either from observations or numerical simulations. Observations of heating and moistening by shallow clouds are limited to a few cases (Nita and Esbensen 1979; Johnson and Lin 1997; Bellenger et al. 2014) derived from comprehensive field campaigns. Simulations of shallow clouds are computationally expensive (Zhao and Austin 2001; Khairoutdinov et al. 2006) and may still suffer from erroneous microphysics parameterizations (Stevens and Seifert 2008). The effects of shallow clouds in models not able to resolve their scale are parameterized at best (Tiedke 1989; Albrecht et al. 1979; Betts 1973; Bretherton and Park 2010). However, the lack of sensitivity to atmospheric water vapor, and faulty closures (Tokioka et al. 1988; Slingo 1996; Maloney and Hartman 2001; Arakawa et al.
2004; Thayer-Calder and Randall 2008; Kim et al. 2011; Park and Bretherton 2008) results in a poor representation of the effects of shallow and congestus clouds (Slingo et al. 2003; DelGenio 2012).

Observing shallow clouds faces several difficulties. The detection of shallow cumulus clouds from space is challenging due to partial beam filling conditions, the proximity of these clouds to the surface, and weak reflectivity from the non-precipitating or weakly precipitating shallow cumulus clouds (Sassen and Wang 2008; Zuidema and Mapes 2008). Shallow non-precipitating clouds warm and cool through condensation and detrainment at their lower and upper levels, respectively (Nita and Esbensen 1974; Johnson and Lin 1997; Schumacher et al. 2009). The moistening and heating effects of non-precipitating and precipitating clouds may differ. Non-precipitating shallow clouds do not dry the atmosphere by rainfall, and their diabatic heating profiles might be bottom-heavier than those of precipitating shallow clouds.

The full spectrum of shallow clouds through the MJO life cycle was observed during the Dynamics of the MJO (DYNAMO) field campaign (Yoneyama et al. 2013) with, however, limited record length. The observational site at Manus Island (Long et al. 2013) from the U.S. Department of Energy (DOE) Atmospheric Radiation Measurements (ARM) provides unique long-term (1996-2014) observations to investigate clouds in a tropical deep convective regime. This site in located in the heart of the Pacific warm pool. The dataset of Manus Island comprises observations from vertically pointing cloud radars, radiosondes, rain gauges, and other instruments. These data have been used to evaluate satellite observations (Hollars et al. 2004) and model simulations (Chen and DelGenio 2009), to estimate cloud radiative heating rates (McFarlane et al. 2007; Mather
and McFarlane 2009; Wang et al. 2010), and to document the cloud evolution during the MJO (Deng et al. 2013) and the seasonal cycle (Mather 2005) at Manus.

This study is divided in three parts. The first part of this study (Chapter 3) investigates the variability and feedbacks of shallow clouds in their large-scale environment at Manus Island. In this part we introduce a new method to estimate bulk shallow cloud moistening from observations available at Manus (liquid water content, precipitation, and temporal cloud fractions). This method is rather rudimentary and suffers from several limitations that will be discussed later. The moistening estimates from this method, however crude they might be, contrast to other estimates (e. g., Bellenger et al. 2015) based on variations of coherent increases in clouds and moisture. The second part of this study (Chapter 4) combines the Manus and reanalysis data to investigate the role of shallow convection in the evolution of large-scale convective events associated and not associated to the MJO. The third part (Chapter 5) concerns the role of shallow convection in driving large-scale circulation. The region of study in this part of the study focuses on the tropical Atlantic. Surface equatorial winds over the tropical Atlantic in models are notorious for their westerly biases and their association to Amazonian rainfall. This part of the study investigates the connection between shallow convection over the Amazonia and the westerly biases from simulations of several atmospheric global climate models.

All the data used in this work is presented in Chapter 2. The particular questions addressed, methods, results, and conclusions, of the present dissertation in the present dissertation are explained in Chapters 3-5. Concluding remarks from this study are given in Chapter 6.
CHAPTER 2: Data Sets

The data used in this study include ground and satellite observations (subsection 2.1), assimilation products (subsection 2.2), and model outputs (subsection 2.3).

2.1 Observations

Observations (Table 2.1) used include data from vertically pointing 35 GHz (Ka-band) radars at the ARM site at Manus Island (2 ° 3’S, 147° 25’ E), which have been collected since 1996 (Long et al. 2013), and from Gan Island (0 ° 42’S, 73° 9’ E) of Addu Atoll, which were collected during DYNAMO (October 2011- January 2012, Yoneyama et al. 2013). The radar at the ARM Manus site is the millimeter cloud radar (MMCR, Kollias et al. 2007). The one at Gan is the Ka-band zenith pointing radar (KAZR). Data from these radars were processed by the Active Remotely-Sensed Cloud Locations algorithm (ARSCL, Clothiaux et al. 2000), which provides a best estimate of the vertical distribution of hydrometeors and cloud boundaries (see Appendix D for details). Data from Manus also included observations from a microwave radiometer (MWR), upper-air soundings, and optical rain gauges. Other observed data used are rainfall estimates from the Tropical Rainfall Measuring Mission (TRMM 3B42v7, Kummerow et al. 2000). The period used of these data is April 03, 2001 - March 07, 2011, during which the Manus-ARSCL and sounding data overlapped. Data gaps, however, exist (Fig. 2 of Kalesse and Kollias 2013).

Soundings at Manus are available at 11 and 23 UTC (9 PM and 9AM local time). Specific humidity, temperature, pressure, and zonal and meridional wind from the two soundings per day was averaged to represent local-time daily means. Daily means were
treated as missing when either the 11 or 23 UTC sounding was unavailable. Soundings with the 0 ° C level outside the observed range of 3.5 to 5.5 km (Geerts and Dawei 2004) were considered unreliable and excluded from our analysis. Boundary-layer heights from the 11 and 23 UTC soundings derived from four methods (Sivaraman et al. 2012) were averaged to provide the best estimate of daily means.

There are six MMCR profiles per minute in the ARSCL data over different vertical ranges with a 45 m resolution during the period covered by this study. Most of the observations over Manus, however, have a temporal resolution of one minute (Table 2.1). To homogenize these data and for computational convenience, the ARSCL reflectivity was averaged into one minute profiles in a vertical range of 100 m to 22 km, with 100 m resolution. Before the averaging, missing value flags were placed over levels where no hydrometeors were detected. Daily means of upper-level soundings were also interpolated to a 100 m vertical resolution. This grid spacing is close to that of the MMCR boundary layer mode (90 m, Moran et al. 1998). The data described above is used in chapters 3 to 5.

Data used in this study also includes surface wind and sea level pressure (SLP) observations from the International Comprehensive Ocean Atmosphere Dataset (ICOADS, Worley et al. 2005), and precipitation from the Climate Prediction Center (CPC) Merged Analysis of Precipitation (CMAP, Xie and Arkin 1997) dataset in monthly resolution were used. $Q_1$ was also estimated from radiosonde data of the Large-Scale Biosphere Atmosphere experiment (LBA, Silva Diaz et al. 2002).
Table 2.1: Observations.

<table>
<thead>
<tr>
<th>Source and variables</th>
<th>Location and Horizontal Resolution</th>
<th>Vertical Resolution and Coverage</th>
<th>Temporal Resolution (overlapping period: 3 April 2001 - 7 March 2011, except *)</th>
</tr>
</thead>
<tbody>
<tr>
<td>ARSCL-MMCR: reflectivity (Clothiaux 2000)</td>
<td>Manus</td>
<td>Every 0.1 km (binned), from 0.1 km to 22 km</td>
<td>1 minute</td>
</tr>
<tr>
<td>ARSCL-KAZR: reflectivity</td>
<td>Gan</td>
<td>Every 0.1 km (interpolated), 0.1 km to 22 km</td>
<td>1 minute (*8 November 2011 - 8 February 2012)</td>
</tr>
<tr>
<td>KAZR-SPOL-SMARTR: reflectivity (Feng et al. 2013)</td>
<td>Gan</td>
<td>Every 0.1 km (interpolated), 0.1 km to 22 km</td>
<td>1 minute (*8 November - 2011 - 8 February 2012)</td>
</tr>
<tr>
<td>Soundings: Specific humidity, Temperature, Pressure, and zonal and meridional wind.</td>
<td>Manus</td>
<td>Every 0.1 km (interpolated), 0.1 km to 22 km</td>
<td>11 and 23 UTC* (Every 4 hours during 24 September 2011 – 30 March 2012)</td>
</tr>
<tr>
<td>Sounding boundary layer height estimates (Sivaraman et al. 2012).</td>
<td>Manus</td>
<td>1 level</td>
<td>11 and 23 UTC</td>
</tr>
<tr>
<td>Optical rain gauge: rain rate</td>
<td>Manus</td>
<td>1 level, at surface</td>
<td>1 min</td>
</tr>
<tr>
<td>Optical rain gauge: rain rate</td>
<td>Gan</td>
<td>1 level, at surface</td>
<td>1 min (*3 October 2011 - 9 February 2012)</td>
</tr>
<tr>
<td>Ceilometer: first cloud base</td>
<td>Manus</td>
<td>1 level</td>
<td>1 min</td>
</tr>
<tr>
<td>Micro-Pulse Lidar (MPL): first cloud base</td>
<td>Manus</td>
<td>1 level</td>
<td>1 min</td>
</tr>
<tr>
<td>Microwave Radiometer (MWR): Liquid water path (LWP) and integrated water vapor (IWV)</td>
<td>Manus</td>
<td>1 level</td>
<td>1 min</td>
</tr>
<tr>
<td>TRMM 3B42 rainfall</td>
<td>From 50° S to 50° N, and from 0° to 360° E, in a 0.25° x 0.25° grid</td>
<td>Surface</td>
<td>Daily Accumulated</td>
</tr>
<tr>
<td>LBA sounding</td>
<td>Amazonia</td>
<td>22 levels</td>
<td>1 month (*Nov. 1998 – Feb. 1999)</td>
</tr>
<tr>
<td>CMAP</td>
<td>Global, 2.5 X 2.5</td>
<td>1 level</td>
<td>1 month (*Nov. 1998 – Feb. 1999)</td>
</tr>
<tr>
<td>ICOADS</td>
<td>Global, 1 X 1</td>
<td>1 level</td>
<td>1 month (*1979-2002)</td>
</tr>
</tbody>
</table>
2.2 Reanalysis

Reanalysis data used (Table 2.2) include rainfall, specific humidity, temperature, geopotential height, and horizontal and vertical wind in varying periods from the operational analysis of the European Center for Medium Range Weather Forecasts (ECMWF) prepared for the Manus site (EC-ARM hereafter), global ECMWF Interim reanalysis (ERAI, Dee et al. 2011), Climate Forecast System Reanalysis (CFSR, Saha et al. 2010), and Modern Era Retrospective-Analysis for Research and Applications (MERRA, Rienecker et al. 2011) products. Physical tendencies of temperature from EC-ARM, CFSR and MERRA, and those of specific humidity from EC-ARM were also used. EC-ARM data for Manus were interpolated to a 100 m vertical resolution to match the interpolated resolution of the Manus soundings.

2.3 Model Outputs

Simulations of rainfall, temperature, and zonal and vertical wind from GCMs of the Atmospheric Model Inter-comparison Project phase 2 (AMIP2, Glecker et al. 1996), their coupled versions of the Coupled Model Inter-comparison Project phase 3 (CMIP3, Meehl et al. 2005), and the AMIP models of CMIP5 (Meehl et al. 2009) were used Table (2.3).
**Table 2.2: Reanalysis.**

<table>
<thead>
<tr>
<th>Source and variables</th>
<th>Location and Horizontal Resolution</th>
<th>Vertical Resolution and Coverage</th>
<th>Temporal Resolution (overlapping period: 3 April 2001 - 7 March 2011, except *)</th>
</tr>
</thead>
<tbody>
<tr>
<td>ECMWF operational analysis: specific humidity, temperature, pressure, zonal and meridional wind, and temperature and humidity tendencies</td>
<td>Interpolated to Manus (0.56 X 0.56 original)</td>
<td>Every 0.1 km (interpolated), 0.1 km to 22 km</td>
<td>1 hr</td>
</tr>
<tr>
<td>ERA Interim reanalysis: specific humidity, temperature, geopotential height, and zonal, meridional, and vertical wind</td>
<td>Global, 1.5° X 1.5°</td>
<td>Irregular grid, 27 levels 1000 - 100 hPa</td>
<td>1 day, and 1 month (*Nov. 1998 – Feb. 1999)</td>
</tr>
<tr>
<td>MERRA: Subgrid temperature tendencies and precipitation</td>
<td>Global, 0.5 X 0.67</td>
<td>Irregular grid, 25 levels 1000 – 100 hPa</td>
<td>1 month (*Nov. 1998 – Feb. 1999)</td>
</tr>
<tr>
<td>CFSR: Subgrid temperature tendencies and precipitation</td>
<td>Global, 0.75 X 0.75</td>
<td>Irregular grid, 27 levels 1000 – 100 hPa</td>
<td>1 month (*Nov. 1998 – Feb. 1999)</td>
</tr>
</tbody>
</table>
Table 2.3: Model outputs.

<table>
<thead>
<tr>
<th>Data</th>
<th>Period</th>
<th>Temporal Resolution</th>
</tr>
</thead>
<tbody>
<tr>
<td>#cnrm_cm3</td>
<td>1979-2000</td>
<td>Monthly</td>
</tr>
<tr>
<td>#gfdl_cm2_1</td>
<td>1980-1999</td>
<td>Monthly</td>
</tr>
<tr>
<td>#giss_model_e_r</td>
<td>1979-2002</td>
<td>Monthly</td>
</tr>
<tr>
<td>#ipsl_cm4</td>
<td>1979-2002</td>
<td>Monthly</td>
</tr>
<tr>
<td>#miroc3_2_hires</td>
<td>1981-2002</td>
<td>Monthly</td>
</tr>
<tr>
<td>#miroc3_2_medres</td>
<td>1981-2002</td>
<td>Monthly</td>
</tr>
<tr>
<td>#mri_cgm2_3_2a</td>
<td>1979-2002</td>
<td>Monthly</td>
</tr>
<tr>
<td>#ukmo_hadgem1</td>
<td>1979-2000</td>
<td>Monthly</td>
</tr>
<tr>
<td>^CNRM-CM5</td>
<td>1979-1999</td>
<td>Monthly</td>
</tr>
<tr>
<td>^GFDL-HIRAM-C180</td>
<td>1979-1999</td>
<td>Monthly</td>
</tr>
<tr>
<td>^GFDL-HIRAM-C360</td>
<td>1979-1999</td>
<td>Monthly</td>
</tr>
<tr>
<td>^GISS-E2-R</td>
<td>1979-1999</td>
<td>Monthly</td>
</tr>
<tr>
<td>^HadGEM2-A</td>
<td>1979-1999</td>
<td>Monthly</td>
</tr>
<tr>
<td>^IPSL-CM5A-LR</td>
<td>1979-1999</td>
<td>Monthly</td>
</tr>
<tr>
<td>^MIROC5</td>
<td>1979-1999</td>
<td>Monthly</td>
</tr>
<tr>
<td>^MPI-ESM-LR</td>
<td>1979-1999</td>
<td>Monthly</td>
</tr>
<tr>
<td>^MRI-AGCM3-2H</td>
<td>1979-1999</td>
<td>Monthly</td>
</tr>
<tr>
<td>^MRI-AGCM3-2S</td>
<td>1979-1999</td>
<td>Monthly</td>
</tr>
<tr>
<td>^MRI-CGCM3</td>
<td>1979-1999</td>
<td>Monthly</td>
</tr>
<tr>
<td>^NorESM1-M</td>
<td>1979-1999</td>
<td>Monthly</td>
</tr>
</tbody>
</table>

*Diabatic heating available as temperature tendency terms.
# AMIP2 and CMIP3 models.
^AMIP models of CMIP5.
CHAPTER 3: Variability of Shallow Clouds in Deep Convective Regimes

3.1 Background

Shallow clouds in regions of deep convection have recently gained attention because of their possible ‘preconditioning’ role in their environment (Yanai et al. 1979; Kemball-Cook and Weare 2001; Benedict and Randall 2007). There are relatively few studies that have investigated the characteristics of low-level clouds in the deep tropics (Malkus and Riehl 1964; Nita and Esbensen 1979; Nichols and LeMone 1980; Johnson and Lin 1997; Bellenger et al. 2013) in comparison to those focusing on subtropical regimes clouds (e.g. Klein and Hartman 1993; Klein 1997; Bony and Dufresne 2005; Stevens 2007; Wood and Bretherton 2007). Studies on subtropical clouds have found useful empirical relations between characteristics of low-level clouds and their environment in different time scales (Klein and Hartman 1993; Nuijens et al. 2009; Brueck et al. 2014). The most important of them is that between low-level cloudiness and lower tropospheric stability in monthly to seasonal time scales (Klein and Hartman 1993; Klein 1997; Wood and Bretherton 2007; Sun et al 2011). Empirical relations of shallow clouds and their environment provide a simple framework to evaluate model parameterizations (Slingo 1987; Zhang et al. 2011), and to evaluate their feedbacks (e.g. Miller 1997). The stark difference in the environment of the deep tropics compared to that of the trade wind regimes, suggests important differences in the relations between low-level clouds and their environment (cloud-environment relations hereafter) between these regions. This Chapter investigates the variability of shallow clouds in two regimes of deep convection, Manus Island in the tropical western Pacific, and Gan Island of Addu Atoll in the Indian Ocean (Fig. 3.1).
There are two objectives behind this Chapter. The first (subsection 3.3) is to document the frequency distributions, macrophysical characteristics (i.e., occurrence, height, liquid water content, and rainfall), and diurnal and seasonal cycles of shallow clouds in deep convective regimes. The second objective (subsection 3.4) is to diagnose cloud-environment relations in these regions, and their possible feedbacks. Cloud-environment relations were investigated using linear correlation matrixes statistically filtered in different time scales. This approach was applied only to the Manus data because of the long record length and comprehensive observations.

We defined shallow clouds in the radar observations at this site as time-height continuous radar echoes with their bases below 2 km and top below the freezing level (approximately 4.5 km height). The definition of shallow clouds in this study includes warm-phase clouds with only liquid droplets. These clouds can be referred to as “low convective clouds”, as opposed to deep convective clouds, to be distinguished from shallow (thin) clouds in the mid and upper troposphere. We attempted to develop a method to estimate moistening by shallow cloud through evaporation of cloud liquid water using observations of liquid water path (LWP), cloud depth and temporal fraction, and the surface rain rate. This method is rudimentary and suffers from several limitations as will be discussed. However, to our best knowledge, this study is the first attempt to estimate cloud moistening directly using field observations. Previous diagnosis of shallow cloud moistening are based on coherent relations between cloud fractions and low tropospheric variations (e.g., Bellenger et al. 2015). Given the simplicity of our method, it needs to be considered as a first step towards more realistic approaches.
The methods used are described in section 3.2. The results are presented in sections 3.3 and 3.4. A discussion and conclusions are provided in section 3.5.

3.2 Methods

The methods we used in this Chapter are divided in four steps. First, we excluded echoes that were considered unreliable because of water vapor attenuation (3.2.1). Second we applied a simple cloud classification method to identify shallow clouds (3.2.2). Third, we develop a method to estimate bulk shallow cloud moistening by evaporation of condensed water (3.2.3). Fourth, we constructed correlation matrixes to identify relations between macrophysical characteristics of shallow clouds and environmental parameters (3.2.4).

3.2.1 Exclusion of Radar Profiles Strongly Attenuated by Rain

Echoes from the vertically pointing 35-GHz radar are susceptible to attenuation due to absorption by water and water vapor, and especially blocking by heavy rainfall (Matrosov 2005; Kollias et al 2007a; Feng et al 2009; Feng et al 2014). To avoid these errors, MMCR observations were excluded when the surface rain rate exceeded a threshold of 25 mm hr\(^{-1}\). The determination of this threshold is explained in Appendix A.

3.2.2 Shallow Cloud Classification

The MMCR can only detect clouds passing over or standing above. We treated temporarily adjacent hydrometeor echoes as objects that represent clouds. Cloud objects in the first cloud layer from the ground with their bases below 2 km (close to the maximum height of the hourly boundary layer, Fig. 3.2d) and their tops below the
freezing level were classified as shallow clouds. These classification criteria are consistent with previous studies (Kollias et al. 2007b; Riley et al. 2011; Deng et al. 2013; Feng et al. 2014). From this approach, we aim to detect isolated the shallow clouds that did not transit to deep. Fig. 3.2d and g shows all (adjacent and not adjacent to deep profiles) 1-min echo-top echoes that were classified as shallow (blue), and those that were classified as parts of a shallow cloud object (not adjacent to deep). The difference between them is small (compare blue, and green and orange in Fig. 3.2d and g).

A given shallow cloud was classified as precipitating if one of the following conditions was met in any minute: reflectivity greater than 0 dBZ below 4 km (Mather et al. 2007), gauge rain rate greater than 0.1 mm hr\(^{-1}\), wet MWR window. Otherwise the shallow cloud was classified as non-precipitating (see Appendix E for a discussion about the sensitivity to these thresholds). Over Manus (see Feng et al. 2014 for an analysis of precipitating shallow clouds over Gan), ground precipitation was detected in 46% of the shallow clouds, of which 7% had rain rates larger than 25 mm hr\(^{-1}\) (red in Fig. 3.1d), our threshold for intolerable radar echo attenuation. About 78% of the shallow non-precipitating clouds exhibited virga. Virga was identified when the cloud based detected by the MPL or ceilometer was higher than the first MMCR echo base (separated by at least one vertical grid point, 0.1 km). In this study shallow echoes with virga are considered non-precipitating because they do not remove moisture from the atmosphere.

3.2.3 Shallow Cloud Moistening Estimates

The continuity equation of water vapor, \( q \), can be represented by:

\[
\frac{\partial q}{\partial t} = - \nabla \cdot (Vq) + e - c
\]  

(3.1)
where $\nabla$ is the three dimensional gradient operator, $\mathbf{V}$ the three-dimensional wind vector, $e$ evaporation, and $c$ condensation. Eq. 3.1 neglects the conversion of ice (water vapor) to water vapor (ice) (e.g. Yanai et al. 1973; Johnson et al. 1976; Johnson et al. 2015). Two processes determine the local moisture tendency, $\partial q/\partial t$: moisture flux divergence by the circulation, $-\nabla \cdot (\mathbf{V} q)$, and liquid-vapor conversion, $e-c$. Both processes can, to a certain degree, be estimated using the observations and reanalysis products described in section 2. The role of shallow clouds in $\partial q/\partial t$ from soundings at Manus through $c-e$ is discussed in this subsection. A large-scale perspective the circulation in $\partial q/\partial t$ in reanalysis products over the Manus domain ($\sim10^o \times 10^o$, centered on Manus) is presented in Chapter 4. As it will be shown latter, the moisture variability observed by the soundings at Manus is very close to that interpolated at Manus from EC-ARM, and that of ERAI over the Manus domain. This suggests that the 5-day running mean sounding tendencies can be fairly diagnosed by reanalysis products.

Our method aims to estimate bulk (daily mean, averaged over the lower troposphere) moistening by shallow clouds due to evaporation of condensed water. All water vapor that enters a cloud through its base may condense or not. The condensed water would rain out or evaporate. Condensed water can, to a certain extent, be measured. Detrainment of uncondensed vapor (Langhans et al. 2015), moistening by the circulation in response to cloud diabatic heating (Chikira 2014; Janiga and Zhang 2015), and other factors (e.g., updraft strength) cannot be estimated from the Manus observations and are not included in our approach.
Without turbulent fluxes and flux divergence by the circulation (to be considered in Chapter 4) from Eq. (3.1), the water vapor tendency is determined solely by the liquid-vapor conversion $e-c$:

$$\frac{\partial q}{\partial t} = e - c \quad \text{(3.2)}$$

We explore the bulk moistening effect of shallow clouds through their vertically integrated evaporation and condensation effects:

$$< \frac{\partial q}{\partial t} > = < e - c > \quad \text{(3.3)}$$

where $< > = - \int_{h}^{H} ( \cdots ) dp$, and $h$ and $H$ are the levels of the cloud base and top respectively. Over a life span of a shallow cloud, $\Delta t$, the net effect of $< e - c >$ is moistening through evaporation of the portion of the cloud liquid water content (LWC) that is not removed by precipitation. If that portion of LWC can be approximated by mean LWC and precipitation over $\Delta t$, then the moisture tendency over $\Delta t$ of a shallow cloud would be:

$$< \frac{\partial q}{\partial t} > \approx < \frac{\Delta q}{\Delta t} > \sim \frac{< \text{LWC}>}{\Delta t} - P \quad \text{(3.4)}$$

where $P$ is the mean rain rate over the period $\Delta t$. We could only observe the vertical profile of a cloud or a portion of it as it passed over the vertically pointing radar. The time series of cloud radar observations is therefore in a time-height domain (Fig. 2). In this time-height domain, we divided the daily observations into fixed $\Delta t$ bins. We then evaluated the mean LWC over the sub-period $\tau$ during which a shallow cloud was
observed. Notice that the anvil part of a cloud was not counted as a shallow cloud based on our definition.

After these approximations, we defined a daily moisture tendency, $M$, due to evaporation of the portion of the mean LWC that was not precipitated:

$$M = \sum_{i=1}^{n} \left( \frac{\langle LWC \rangle}{\Delta t} - P \right) \Delta t$$  \hspace{1cm} (3.5)$$

where $n$ is the number of $\Delta t$ in a day (e.g., $n = 24$ for $\Delta t = 60$ min). In this method, with a given LWC, a relatively short $\Delta t$ implies fast cloud dissipation that produces a large tendency of moistening (if evaporation dominates) or drying (if precipitation dominates). We estimated $M$ above the boundary layer. By assuming linear increase of LWC with height (Han et al. 1995; Wood and Taylor 2001) in shallow clouds, it can be shown
(Appendix B) that \(<LWC>\) is related to the liquid water path (LWP) as
\[
<LWC> = \left(\frac{\text{LWP}}{H^2}\right)(H-h_b)^2,
\]
where \(h_b\) is the height of the boundary-layer top. Thus from Eq. (3.5),
\[
M = \sum_{i=1}^{n} \frac{(\frac{\text{LWP}_i}{H^2})(H-h_b)^2}{\Delta t} - P_i \Delta t
\]

At Manus, LWP is available from MWR measurements (Table 1). \(H\) was obtained from the MMCR (section 3.1), \(h\) from the soundings, and \(P\) from optical rain gauges. Numerical studies (e.g. Zhao and Austin 2005) have suggested that the typical life time of shallow non-precipitating and precipitating clouds is about 18 and 25 min, respectively. These studies have typically simulated shallow clouds with their top heights close to 2 km. A longer duration is expected for deeper clouds that can reach the freezing level. In this chapter, we choose \(\Delta t = 60\) min. Results are qualitatively the same by increasing or decreasing this period 30 min (see Chapter 4). \(M\) was estimated separately for shallow precipitating and non-precipitating clouds.

We also estimated \(<LWC>\) using two LWC-reflectivity relations for stratocumulus clouds (Liao and Sassen 1994; Frisch 1995). However, due to the fact that reflectivity increases as a power of six with the diameter of the cloud particles, these estimates suffered from errors with orders of magnitude larger than those based on LWP. These results are not reported here.

### 3.2.4 Correlation Matrixes

Empirical relations between anomalies in shallow cloud characteristics and those in their environment were diagnosed using Pearson’s correlation matrixes. Daily anomalies of all
data were calculated by removing their mean seasonal cycle. Variability frequencies lower than 5, 14, and 30 days, and in the band between 1 and 5 days, 5 and 14 days, and between 5 and 30 days in all data (Table 3.1) were filtered. Running-mean time filters were used instead of the more commonly used spectral filters to avoid data interpolation. Also, daily anomalies were binned in periods of 5, 7, and 30 days to analyze their correlation at different time lags, from 1 to 5, in each temporal resolution. Correlations not significant at the 95% confidence level according to the Student’s t-test (using n-1 degrees of freedom) were set to zero.

Observations in the matrix include: occurrence, depth, liquid water content (LWC), and rainfall rate, of shallow cumulus (SC) precipitating (P) and non-precipitating (NP) clouds, their local tendencies, sounding specific humidity ($q$), temperature ($T$), zonal ($u$) and meridional ($v$) wind, and TRMM 3B42 and Manus gauge rainfall (see Table 3.1). In this study, we defined cloud occurrence as the total time that cloud echoes were observed over the radar during a fixed period of time. We used 1 day as such fixed period of time. The estimated shallow cloud moistening $M$ (subsection 3.2.3) was also included in the correlation matrices. Observed sounding derived parameters include: the lower tropospheric stability (LTS, Klein and Hartman 1993), the estimated inversion strength (EIS, Wood and Bretherton 2007), convective available potential energy (CAPE, estimated assuming an undiluted lifted surface parcel), and convective inhibition (CIN). Reanalysis data in the matrix include: sea surface temperature (SST), and advection of humidity and temperature, and $Q_1$ and $Q_2$ (Yanai et al. 1979) using ERAI.
### Table 3.1: Variables in the correlation matrixes.

<table>
<thead>
<tr>
<th></th>
<th>Variables</th>
</tr>
</thead>
<tbody>
<tr>
<td>1</td>
<td>Deep Cloud Occurrence</td>
</tr>
<tr>
<td>2</td>
<td>Clear Sky Occurrence</td>
</tr>
<tr>
<td>3</td>
<td>NP SC Occurrence</td>
</tr>
<tr>
<td>4</td>
<td>P SC Occurrence</td>
</tr>
<tr>
<td>5</td>
<td>NP SC Thickness</td>
</tr>
<tr>
<td>6</td>
<td>P SC Thickness</td>
</tr>
<tr>
<td>7</td>
<td>M NP SC</td>
</tr>
<tr>
<td>8</td>
<td>M P SC</td>
</tr>
<tr>
<td>9</td>
<td>SC rain rate</td>
</tr>
<tr>
<td>10</td>
<td>NP SC LWP</td>
</tr>
<tr>
<td>11</td>
<td>P SC LWP</td>
</tr>
<tr>
<td>21</td>
<td>LTS</td>
</tr>
<tr>
<td>31</td>
<td>ERAI $\nu$-advection of temp</td>
</tr>
<tr>
<td>22</td>
<td>EIS</td>
</tr>
<tr>
<td>32</td>
<td>ERAI $\omega$-advection of temp</td>
</tr>
<tr>
<td>23</td>
<td>T in the low trop</td>
</tr>
<tr>
<td>33</td>
<td>ERAI $-Q_s/L$</td>
</tr>
<tr>
<td>24</td>
<td>$u$-wind in the low-trop.</td>
</tr>
<tr>
<td>34</td>
<td>ERAI $Q/I/C_p$</td>
</tr>
<tr>
<td>25</td>
<td>$v$-wind in the low-trop.</td>
</tr>
<tr>
<td>35</td>
<td>ERAI SST</td>
</tr>
<tr>
<td>26</td>
<td>Wind speed in the low-trop.</td>
</tr>
<tr>
<td>36</td>
<td>Height of 80% rel. hum.</td>
</tr>
<tr>
<td>27</td>
<td>PBL height</td>
</tr>
<tr>
<td>37</td>
<td>Height of 60% rel. hum.</td>
</tr>
<tr>
<td>28</td>
<td>ERAI $v$-advection of humidity</td>
</tr>
<tr>
<td>38</td>
<td>Height of 40% rel. hum.</td>
</tr>
<tr>
<td>29</td>
<td>ERAI $u$-advection of humidity</td>
</tr>
<tr>
<td>39</td>
<td>El Nino Index 3</td>
</tr>
<tr>
<td>30</td>
<td>ERAI $u$-advection of temp</td>
</tr>
<tr>
<td>40</td>
<td>El Nino Index 4</td>
</tr>
</tbody>
</table>

SC: Shallow Cumulus, NP: Non-Precipitating, P: Precipitating, low troposphere: 0.1 to 4.5 km. $M$ was estimated using $\Delta t = 60$ min. See text for more details.

### 3.3. Characteristics of Shallow Clouds at Manus and Gan Islands

#### 3.3.1 Cloud Frequency Distributions

The vertically pointing MMCR can only detect clouds passing over or standing above. Most echoes over Manus and Gan were observed continuously for less than 5 min (Fig 3.2a and e). Over Manus, lower (below 4 km height) and upper (above 10 km height)
tropospheric echoes have longer durations than those in the middle troposphere (Fig. 3.2a). These characteristics are not evident over Gan (Fig. 3.2e) possible due the relatively short period of these observations. In a joint probability distribution of echo top height and echo depths (Fig. 3.2b and e), there are four distinct groups of shallow clouds. One is along the diagonal line in Fig. 3.2b. Clouds on this line have their bases close to 0.1 km, which is indicative of precipitation, so their apparent depths are roughly equal to their top heights. The other groups are composed of clouds with depths less than 2 km. They are clustered in the upper, mid, and lower troposphere, respectively. The three clusters in Figs. 3.2c and g must not be mistaken as observational support of trimodal tropical convection discussed by Johnson et al (1999). The middle frequency distribution peak is mainly composed of alto clouds (Riley et al. 2009). One possible cause for the absence of the congestus or middle level peak in the frequency distributions is rooted in the land effect. However, such effect must be much more pronounced in Manus compared to that in Gan (Fig. 3.3), but the peak is missing in both frequency distributions. The clouds with echo-top heights below 5 km are the primary targets of this study. Fig. 3.2e shows the total echo-top population based on one-minute observations and their fractions that were classified as shallow. The fraction of shallow clouds with rain rates greater than 25 mm hr\(^{-1}\) and excluded from this study is small. Varying the rainfall threshold slightly does not affect the subsequent results.
Figure 3.2 Frequency distributions of (a, e) echo duration vs. height, (b, f) echo depths vs. top heights, (c, g) echo-top heights, and (d, h) 11 and 23 UTC and boundary layer heights vs. echo object base heights of non-precipitating shallow cloud from the MMCR at (upper panel) Manus Island during 03 April 2001 to 07 March 2011 and KAZR and (lower panel) Gan during 08 October 2011 to 08 February 2012.

Figure 3.3 The locations of the (red stars) ARM sites in (a) Addu Atoll in the Indian Ocean and (b) Manus Island in the tropical western Pacific. Note the longitudinal and latitudinal scales are different in each panel.
3.3.2 Macrophysical Characteristics and Moistening

The frequency distributions of shallow non-precipitating and precipitating cloud depths show several differences (Fig. 3.4b). The range of depths of non-precipitating shallow clouds spans from 0.1 to 2 km, and more than 70% of their depths are shorter than 0.5 km. Shallow precipitating clouds are not generally larger than 3.0 km depth (95% of their sample), but they can reach 4 km depth. The depths of shallow precipitating clouds are more normally distributed than those of the non-precipitating clouds. The median of the former is close to 1.5 km and that of the latter is about 0.4 km. Shallow precipitating clouds are taller but shallow non-precipitating clouds are more abundant (Fig. 3.4a); the mean daily occurrence of non-precipitating clouds are larger than those of the precipitating ones. This feature is particularly evident over Gan, were most of the shallow clouds were non-precipitating during the DYNAMO period (Fig. 3.2g). Over Manus, the daily mean occurrence of shallow non-precipitating (precipitating) clouds is 1 hr and 15 min (57 min) (Fig. 3.4a). For both types of clouds the frequency distribution of occurrence, LWP (Fig. 3.4c), moistening (Fig. 3.4d), and rain rates for the precipitating ones (Fig. 3.4e) shape gamma distributions with, approximately, 0.2 mm day$^{-1}$ as median value. The mean LWP for non-precipitating (precipitating) shallow clouds is 0.2 mm (0.6 mm), but there is almost no difference between their estimated moistening, $M$. The mean moistening of both types of clouds is about 1 mm day$^{-1}$, and their frequency distributions are statistically the same. Shallow non-precipitating clouds (Fig. 3.5a and d) and the non-precipitating sections of the precipitating ones (Fig. 3.5b and e) have reflectivity profiles that increase with height. Consistently with the adiabaticity assumption (Appendix A), these profiles suggest a linear increase with height of LWC. Reflectivity profiles of
precipitating clouds are not-linear however. They tend to have quasi-symmetric parabolic vertical distributions with low reflectivity values at their ends, particularly at the higher one. These reflectivity distributions manifest growing and dissipation of cloud drops by accretion close to the reflectivity peak, and evaporation towards the top and bottom. More specific details of the reflectivity distributions under precipitating conditions are considered unreliable without further treatment (see Feng et al. 2008) in this study.

Figure 3.4: Frequency distributions of shallow (black) non-precipitating and (red) precipitating clouds in (a) occurrence, (b) depth, (c) liquid water path, (d) moistening $M$, and (e) rainfall at Manus during 03 April 2001 to 07 March 2011.
Figure 3.5: Reflectivity-height frequency distributions of (a, d) non-precipitating shallow clouds, and the (b, e) non-precipitating and (c, f) precipitating profiles of precipitating shallow clouds over (upper panels) Manus during 03 April 2001 to 07 March 2011 and (lower panels) Gan during 08 October 2011 to 08 February 2012.

The LWP associated to shallow non-precipitating clouds shorter than 1 km is typically lower than 2 mm for shallow (Fig. 3.6a). In contrast, typical LWP values of shallow precipitating clouds shorter than 1 km height can be as high as 4 mm. There is a weak tendency of LWP to decrease with depth (Fig. 3.6a) in the presence of shallow non-precipitating clouds, and to increase with depth (Fig. 3.6b) in the presence of the precipitating ones. There are weak increases in surface rain rate with as a function of cloud depth (Fig. 3.6c).

The cloud base of both, precipitating and non-precipitating shallow clouds ranges typically from 700 to 900 m height (Fig. 3.7). This value is consistent in observations from ceilometer and micro pulse lidar, and in the ARSCL value added product. The frequency peak almost at the surface in ARSCL (3.7b) simply manifests precipitation
conditions. Some weaknesses of millimeter vertically pointing radars and their products during precipitating conditions are discussed in Appendix A.

Figure 3.6 Two dimensional frequency distributions of cloud depth against liquid water path of (a) non-precipitating (NP) and (b) precipitating (P) shallow clouds (SC), and against (c) rain rate of P SC over Manus during 03 April 2001 to 07 March 2011.

Figure 3.7 Frequency distributions of cloud base heights from (a) ceilometer, (b) ARSCL, and micro-pulse lidar observations of non-precipitating (NP) and precipitating (P) shallow clouds (SC) over Manus during 03 April 2001 to 07 March 2011.
3.3.3 Diurnal Cycle

The temperature closest to the surface in EC-ARM and ERAI is 1000 hPa. We used temperature at 100 m from soundings as a proxy for that at 1000 hPa. EC-ARM upper air data did not reproduced well the diurnal cycle in 100 m temperature and relative humidity over Manus during the DYNAMO period (September 24th, 2011 to March 30th, 2012). EC-ARM was therefore were not used to evaluate the diurnal cycle over the full period of Manus observations.

The abundance of shallow non-precipitating clouds over Manus peaks close to 10 AM (local time), starts decreasing slowly afterwards, and reaches a minimum by 6 AM. From 6 AM to 10 AM the population of shallow non-precipitating clouds grows the fastest. During this period, lower-tropospheric relative humidity decreases (Fig. 3.8a) while the LTS (Fig. 3.8e), and the LCL deepens the fastest (Fig. 3.8f). Shallow precipitating clouds are more abundant (Fig. 3.8a) after the peak in shallow non-precipitating clouds, from 10 AM to 2 PM, and before the abundance peak of deep clouds, approximately from 2 to 6 PM (Fig. 3.8a). The diurnal cycle in CAPE (Fig. 3.8d) seems unrelated with the occurrence of deep clouds. CAPE almost monotonically increases during the night, from 8 PM to 4 AM, and decreases during the day, from 4 AM to 8 PM. During the peak of deep clouds, approximately between 2-6 PM, 100 m temperature, the IWV (Fig. 3.8g), the liquid water path (LWP, Fig. 3.8h), and the hourly-accumulated rainfall (Fig. 3.8i) have their largest values, and the LTS its lowest values.

The diurnal cycle in the cloud population over Manus is consistent with a diurnal cycle over land. Over tropical land areas, non-precipitating cumulus clouds develop first
in the early morning, they are followed by shallow precipitating and deep clouds hours later. Shallow clouds in the diurnal cycle facilitate the vertical flux transport to the free troposphere by deepening and destabilizing a saturated boundary layer from previous late afternoon rainfall (Betts and Jakob 2002a, 2002b).

Over Gan, no distinct diurnal cycle is observed in shallow clouds is observed from September 24th of 2011 to March 30th of 2012. During January 2012, shallow clouds tended to peak in the early morning, close to 4 to 6 AM (Fig. 3.9). More data over Gan is needed to reach a more complete diagnostic. The seasonality of the diurnal variations over Manus is discussed next.

Figure 3.8: Diurnal cycle in cloud occurrence of non-precipitating (NP) and precipitating (P) shallow cumulus (SC) and deep clouds, and (black) several parameters derived from observations at Manus during 03 April 2001 to 07 March 2011. The diurnal cycle for the (blue) DYNAMO period over Manus, 0 August 2011 to 0 March 2012, is also shown.
3.3.4 Seasonal Cycle

The seasonal cycle in shallow and deep cloud occurrence is similar is several aspects (Fig. 3.10). During January to March (~ days 1-120, Fig. 3.10a), both types of clouds exhibit a strong diurnal cycle, and there are one period of large cloud occurrence, July to October (~ days 191-300, Fig. 3.10a), one of low occurrence, during May to June (~ days 121-190, Fig. 3.10a), during which the diurnal cycle is relatively weak. In contrast, the seasonal cycle in the size of shallow clouds is weak, with few salient features (Fig. 3.11). During July to October, there are abundant short shallow non-precipitating clouds (Fig. 3.10a) and there are relatively weaker variations in the rest of the year. There is no evident seasonal cycle in the depth of shallow precipitating clouds, but they tend produce more rainfall during the first part of the year compared to the second part (Fig. 3.10b). Cloud depths of deep clouds are considered unreliable and are therefore not analyzed (see Appendix A). However, consistent with the abundance peak of deep clouds July to
October (Figs. 3.10b), accumulated rainfall from these clouds is larger during these months (Fig. 3.10e) than during the other ones.

Figure 3.10: Seasonal and diurnal cycle of occurrence of shallow cumulus (SC) (a) non-precipitating (NP) and (b) precipitating (P) clouds, and deep clouds over Manus during 03 April 2001 to 07 March 2011. The hourly mean occurrence is shown in the right panel. The daily mean accumulated rainfall of (c) shallow and (d) deep clouds is also shown. 5 day running means were applied to these data.
Figure 3.11: Seasonal cycle of cloud occurrence by height of shallow (a) non-precipitating and (b) precipitating clouds and their (c) accumulated rainfall over Manus during 03 April 2001 to 07 March 2011. 5 day running means were applied to these data.

While there is no evident relation between the seasonal cycle in shallow non-precipitating or precipitating clouds and that in wind, temperature, or specific humidity from the soundings, we highlight two aspects of their behavior. First, the seasonal peak of occurrence in shallow non-precipitating clouds, during July to October (~ days 191-300, Fig. 3.10a), have common features with that in lower tropospheric zonal and meridional winds (Figs. 3.12a-c), and to a lower extend with zonal wind and tropospheric temperature. Such seasonal peak in non-precipitating shallow occurs in a relatively cold regime imbedded in northwesterly winds. Second, the lack of consistency between shallow cloud occurrence, and lower tropospheric temperature (Fig. 3.12c) and humidity (Fig. 3.12d) might be seen as a first indication of a relatively weak role of shallow clouds in driving environmental conditions.
Figure 3.12: Seasonal cycle in mean (a) zonal and (b) meridional wind, (c) specific humidity, (d) temperature, and (e) lower tropospheric stability from soundings at Manus during 03 April 2001 to 07 March 2011. The mean accumulated rainfall is shown below (f). 5 day running means were applied to these data. The annual mean was subtracted to specific humidity and temperature.
Lower tropospheric stability presents a weak seasonal cycle, with relatively high values during the first part of the year compared with that in cloud occurrence (Fig. 3.12e). We found no correlation between the seasonal evolution of LTS and cloud occurrence. However, as it will be shown in the next section, anomalies in shallow cloud occurrence and LTS present relatively high correlations under certain conditions. In summary, the seasonal cycle in low-tropospheric winds, particularly its meridional wind component, is apparently a large influence to that in cloud occurrence over Manus. We tried to identified a correlation from this qualitative diagnosis in the seasonal cycle but found none.

3.4. Cloud-Environment Relations

The time series of shallow cloud occurrence of the total period of observations at Manus shows variability in several time scales, from synoptic time scales to inter-annual time scales (Fig. 3.13a). Both, precipitating and non-precipitating shallow clouds show similar long-term variability (Fig. 3.13 c and d). Such fluctuations do not quantitatively dependent on the rainfall identification method we applied (see Appendix E). In this section, we explore the possible connections of cloud anomalies with those in their environment.
The correlation between shallow cloud occurrence (variables 3 and 4 in Fig. 3.14), depth (vars. 5 and 6 in Fig. 3.14), liquid water content (vars. 10 and 11 in Fig. 3.14), and their rainfall (var. 11 in Fig. 3.14) with LTS (var. 21 in Fig. 3.14) and EIS (var. 22 in Fig. 3.14) is larger than that with other environmental variables (Fig. 3.14 a). And they are larger in low frequencies (monthly, Fig. 3.14 c and f) compared than those in high frequencies (daily and synoptic, Fig. 3.14 a, b, d, e). In general, there are more (less) abundant precipitating (non-precipitating) clouds with larger (shorter) depths in periods of high LTS values (Fig. 3.15). However, the anomalies in occurrence and depth have no evident connection with LTS during periods of positive anomalies of LTS (to right of the black line in Fig. 3.15). The relative high correlations between occurrence and depth of shallow clouds with LTS in monthly time scales is consistent with the well documented relation of LTS and cloudiness of trade wind cumuli (Klein and Hartmann 1993; Klein 1997; Wood and Bretherton 2007; Zhang et al. 2011). More observations are required to

Figure 3.13 Time series of occurrence (min) in (a) shallow clouds (SC), (b) only the non-precipitating (NP) SC, and only the precipitating (P) SC.
confirm our diagnosis however. The sample size of our data is relatively small, about 90 - 120 samples, or 3 to 4 different-month data in 30-day low pass filtered data. Despite the large influence of the low frequency variability, the ‘El Nino’ the southern oscillation (ENSO) ‘3’ (central to eastern Pacific region) and ‘4’ (central to western Pacific region) indexes (vars. 39 and 40 in Fig. 3.14) are only weakly correlated to the shallow cloud perturbations in occurrence and depth.

Figure 3.14: Correlation matrixes of daily anomalies in Manus and ERAI observations (Table 3.1 for details of the variables) using (a) no frequency filters, a (b) 5 and (c) 30-day low-pass filter, (d) a 5-day hi-pass filter, and a 5 to 30-day and a 1 to 3-month band pass filter from Manus during 03 April 2001 to 07 March 2011. Only statistically significant correlations are shown. The black lines enclose cloud macrophysical characteristics.
Figure 3.15: Scatter diagrams of 30-day low-pass filtered daily anomalies in shallow cumulus (SC) (a, b) precipitating (P) and (c, d) non-precipitating (NP) (a, c) cloud depth and (b, d) occurrence against lower tropospheric stability at Manus during 03 April 2001 to 07 March 2011. Correlation lines are shown dashed and their coefficients are shown in their respective colors. One negative and positive standard deviation lines are blue. The zero LTS anomaly line is black.

To further explore the apparent independence of between anomalies in the depth and occurrence of shallow clouds with lower tropospheric wind speed, and humidity, the data of the daily and 5-day high-pass filtered correlation matrixes were analyzed in terciles of precipitation, LTS, and lower tropospheric RH separately. The data on each tercile represents different regimes: below normal, normal, and above normal. Results in this analysis consistently show the independence of the cloud occurrence and depth with the other variables. The local tendencies in cloud occurrence and depth were analyzed but
no relation was found. The correlation matrix presents many interesting relations among only environmental variables (for example, among vars. 12-16 of Fig. 3.10), but a detailed diagnosis of them is out of the scope of the present work.

The depth and occurrence of the shallow clouds are directly related to their estimated shallow cloud moistening $M$ (Eq. 3.6). Tall shallow clouds have larger liquid water contents, rain rates, and moistening effects (Fig. 3.12d) than short ones. When shallow non-precipitating (precipitating) clouds are more abundant they tend to be shallower (deeper) (Fig. 3.16a-c). These results suggest that for a given period with scarce (abundant) shallow clouds, most of them would be relatively tall non-precipitating (precipitating) clouds. More and deeper shallow clouds, either non-precipitating or precipitating, would introduce more water into the troposphere. But abundance and depth of shallow non-precipitating clouds is anti-correlated. Consequently, larger abundance of shallow non-precipitating clouds is anti-correlated ($r=-0.54$) with $M$ (Fig. 3.16b). That is, larger sizes of these clouds would produce larger moistening, $M$, compared to larger abundances. However, a weakness in our approach to estimate $M$ is that it does not takes into account differences in cloud life spans as a function of their depth, as it would be expected in a more realistic approach.
Figure 3.16: Scatter diagrams of 30-day low-passed filtered daily anomalies in (a) occurrence of precipitating against that of non-precipitating shallow clouds, occurrence of (b) non-precipitating and (c) precipitating shallow clouds against their depth, and (d) and depth of shallow precipitating clouds against their estimated moistening $M$ at Manus during 03 April 2001 to 07 March 2011. The correlation line is red and zero LTS anomaly line is black. One negative and positive standard deviation lines are blue.

The relation between anomalies in moistening effects from shallow clouds and those in LTS is different for precipitating and non-precipitating shallow clouds. The correlation between non-precipitating shallow cloud moistening and LTS is -0.42 (0.41) (vars. 5 and 22 and Fig. 3.14). These relations however should be ignored for positive anomalies of LTS. Figures 3.15 a and b and 3.16d suggest that there are relatively deep but scarce shallow precipitating clouds in stable regimes (large LTS), and during these
regimes, the shallow cloud moistening $M$ tends to be relatively high compared to less stable (lower LTS) (Fig. 3.17a). Moistening from shallow non-precipitating clouds is also relatively large during stable regimes (Fig. 3.17b), but this characteristic for these clouds is mainly due a larger cloud occurrence (Figs. 3.16d), not depth (Figs. 3.16c).

Figure 3.17 Scatter diagrams of 30-day low-pass filtered daily anomalies in shallow cumulus (SC) (a) precipitating (P) and (b) non-precipitating (NP) moistening $M$ against lower tropospheric stability at Manus during 03 April 2001 to 07 March 2011. The correlation line is red. One negative and positive standard deviation lines are blue.

We examined the correlations between macrophysical characteristics of shallow clouds (depth: variables 5 and 6, moistening: variables 7 and 8, and precipitation: variable 9, in Fig. 3.14), with anomalies in low-level moisture, temperature, and convergence (variables 23-38 in Fig. 3.14) to investigate the possible role of shallow clouds in driving large-scale environmental perturbations. The correlations between these macrophysical characteristics of shallow clouds with sounding low tropospheric specific humidity (variables 13 and 14, Fig. 3.14), temperature (variables 23, Fig. 3.14), wind
(variables 24-25, Fig. 3.14), and advection (variables 27-32, Fig. 3.14) were found weak to non-existent. However, these macrophysical characteristics of shallow clouds have a relatively high correlation with the depth of a given relative humidity threshold (variables 36-38 in Fig. 3.14) in monthly time frequencies (variables 36 to 38 in Fig. 3.14c). This relatively high correlation suggests shallow clouds could be ‘cause for’ or ‘consequence of’ increases in low tropospheric moisture. If shallow cloud moistening were the cause for the anomalies in the depth of the cloud layers, then cloud occurrence, depth, and moistening should also be correlated with wind (variable 24 in Fig. 3.14), integrated water vapor, sounding humidity, and total precipitation (variables 12-16 in Fig. 3.14), because the anomalies in the moist layer are correlated with these large-scale variables. But the cloud characteristics are not correlated with them. These simple associations suggest that shallow cloud moistening cannot be the cause for the positive anomalies in low tropospheric moisture.

3.5 Summary and Discussion

This Chapter presents two sets of analyses to the observed characteristics of shallow clouds from more than 10 years of observations over the ARM Manus site in the tropical western Pacific. First, this study documents the climatological macro-physical characteristics of these clouds from frequency distributions, and their diurnal and seasonal cycles. This analysis was also briefly made to observations of shallow clouds at Gan in the equatorial Indian Ocean, during DYNAMO. Second, this study investigates their environmental controls and possible role of their feedbacks to their environment.
The diurnal cycles in the cloud population over Manus, and that in temperature, relative humidity, CAPE, and other variables, are largely modulated by the land effect. However, there is a clear distinction in the peak times in shallow non-precipitating, precipitating, and deep clouds. Shallow non-precipitating clouds tend to peak in the morning, from 10 to 12 AM, shallow precipitating in the early afternoon, from 12 to 2), and deep clouds in the afternoon, from 2 to 4 PM. The peak in occurrence of shallow clouds coincides with a fast deepening of the boundary layer. This evolution is consistent with that over tropical land regions (Betts and Jakob 2002b). Shallow non-precipitating and precipitating, and deep clouds have their minima in the early morning, from 6 to 8 AM. There is relatively weak seasonal cycle of the shallow clouds over Manus, which has similarities with that in low tropospheric wind speed, particularly by its meridional component. However, we found they are not statistically correlated.

We found statistically significant correlations to describe shallow cloud occurrence and depth anomalies at Manus in relatively long time scales. Particularly, shallow cloud occurrence and depth have relatively high correlations with LTS in monthly time scales. There tend to be more shallow non-precipitating clouds during relatively stable regimes (positive anomalies in LTS), but their sizes are relatively short. The behavior of shallow precipitating clouds is different in regard of the depth of the clouds. There tend to be more and relatively big shallow precipitating clouds during relatively stable regimes. Although the cloud characteristics were analyzed by means of linear correlations, they are more likely to behave in at least two modes, with no smooth transition between them, as suggested by the clustered scatter diagrams of Figs. 3.11. These results can be exemplified in the following idealization of two regimes, one
relatively unstable and moist (A), and the other highly stable and dry (B). The depth and strength of convection would be discouraged in the second regime. There, small and weak convective clouds would tend to be abundant in the second regime compared to the first one. Without including the moistening effects due diabatic-heating induced convergence, more abundant shallow clouds would produce larger moistening in the low-levels in the second regime.

Difficulties in estimating shallow cloud moistening are many, but a better characterization of the effects of shallow precipitating and non-precipitating clouds is crucially needed to reduce many uncertainties in climate sensitivity to these clouds, particularly those in subtropical regions (Bony and Dufresne 2005). Shallow precipitating clouds are taller but shallow non-precipitating clouds are more abundant. Taller shallow clouds have larger liquid water contents, rain rates, and moistening effects. The moistening effects of relatively large populations of small non-precipitating shallow clouds are smaller than those of less abundant but taller shallow clouds. From our results, it is difficult to diagnose if shallow precipitating or non-precipitating clouds produce larger moistening to the lower troposphere (Rapp et al. 2009). Both types of clouds have similar moistening tendencies associated to their evaporation of condensed water. However, shallow precipitating clouds may be produce larger moistening by their dynamical effects.

Bellenger et al. (2015), using shipborne lidar and sounding data over the central equatorial Indian Ocean, estimated the magnitude of moistening by shallow clouds to be 10-20 g kg$^{-1}$ day$^{-1}$ (25 - 50 mm day$^{-1}$) on the scale of a few tens of minutes and between 1.5 - 4 km and 1-4 g kg$^{-1}$ day$^{-1}$ (2.5 - 10 mm day$^{-1}$) on the scale of a few hours. Their
estimated moistening tendency on the order of a day would approach the same order of magnitude of our estimates: 1-3 mm day$^{-1}$. There are, however, several differences between their and our estimates. Our estimates include only moistening by evaporation of condensed water as shallow clouds completely dissipate. Clouds can moisten the environment by detrainment of uncondensed water vapor (Langhans et al. 2015) without changing the LWC, which is not included in our estimates. Also not included in our estimates is moistening by the circulation in response to diabatic heating of precipitating shallow clouds (Chikira 2014; Janiga and Zhang 2015). On the other hand, the estimate of shallow cloud moistening by Bellenger et al. (2015) is based on coherent variations of shallow cloud occurrence and low-tropospheric moisture. It includes all moistening effects of turbulent transport, convergence induced by diabatic heating of shallow clouds, and large-scale advection. These differences highlight the difficulty of adequately estimating moistening by shallow clouds and uncertainties in the current knowledge on this subject.

The nature of the point measurements and the instrumentation available at Manus forced us to make several assumptions, which greatly constrain our diagnostics. Our assumptions are:

(a) Averaged time series of vertical profiles observed at a point may represent three-dimensional characteristics because of ergodicity. Point measurements cannot, however, distinguish clouds that remain shallow from those that grow deep. This limitation can be alleviated by using data from scanning cloud radars or satellites. Such data are not yet available from tropical deep convective regions.
(b) MMCR echoes are reliable when the rain rate is lower than 25 mm hr\(^{-1}\). This is a conservative approach that may have led to an underestimate of total amount of shallow clouds. This choice of the threshold is based on a relatively small number of observations (Appendix 1). The issue of attenuation by rain in cloud radar echoes needs more rigorous treatment (Feng et al. 2009).

(c) Precipitation from shallow clouds is adequately measured by collocated rain gauges. This assumption neglects the drying effect of shallow-cloud rain evaporated within the boundary layer and errors of single rain gauge measurement (Ciach 2003). A relationship between Ka-band reflectivity and the rain rate (Matrosov 2005; Leon et al 2008; Deng et al 2014) for shallow clouds, after validated, might be used to complement the rain gauge measurements.

(d) Precipitating shallow clouds can be defined by thresholds in MMCR reflectivity (0 dBZ exists below 4 km height) and collocated rainfall (larger than 0.1 mm hr\(^{-1}\) or wet MWR window). This is adapted from previous studies (Mather et al. 2007) but needs to be verified.

(e) A typical life span of shallow clouds is 30 – 90 min. The life span of shallow clouds, however, depends on conditions of the environment. Detailed observations from scanning cloud radars are needed to assess life spans of shallow clouds.

(f) The bulk (daily mean, lower troposphere) moistening by shallow clouds is mainly determined by evaporation of condensed water. This assumption neglects moistening by detrainment of uncondensed vapor and the eddy fluxes associated to shallow clouds. Also not included in our moistening estimates is indirect moistening by the circulation in
response to diabatic heating of precipitating shallow clouds (Chikira 2014; Janiga and Zhang 2015).

(g) LWC increases linearly with cloud depth (adiabaticity). This was adapted from previous studies (Han et al 1994; Wood and Taylor 2001). More realistic estimate of LWC in shallow clouds is needed. We also calculated the shallow cloud moistening by including the boundary layer portion into $M$ (Eq. 7) without applying this assumption and found qualitatively the same results. LWC has been estimated using cloud radar data for marine stratus (Atlas 1954; Liao and Sassen 1994; Frisch 1995; McFarlane et al 2002). It is unclear whether this is applicable to precipitating shallow cumuli in tropical deep convective regions.

These assumptions could have led to systematic biases in our estimates of moistening tendency of shallow clouds, but would not affect their temporal variability, which comes mainly from much more reliable occurrence of shallow clouds observed by the cloud radar (see Chapter 4, subsection 4.3.3 for more details). The method we present to estimate shallow cloud moistening is nevertheless primitive, and can even be flawed due to its unrealistic assumptions. But it might be the first attempt to estimate shallow-cloud moistening in a tropical deep convective region directly using field observations. We encourage its validation or invalidation, particularly using more comprehensive observations and more reliable assumptions.
CHAPTER 4: Shallow Convection and the MJO

4.1 Background

The diabatic heating and moistening effects of shallow cumulus clouds have been frequently hypothesized crucial to the oscillation first documented by Madden and Julian (1971, 1972), or MJO. Adequate MJO simulations require sufficient low-level diabatic heating (Li et al. 2009; Zhang and Song 2009; Cai et al. 2013; Lappen and Schumacher 2014). Part of this hypothesis is based on observed coherent increases in low-level moisture and shallow clouds leading to MJO rainfall peaks (Kemball-Cook and Weare 2001; Benedict and Randall 2007). Other part has come from model simulations. Yanai et al (1973) first explicitly pointed out the importance of moistening of the lower troposphere by shallow cumulus clouds in the development of deep convection. Such a moistening role of shallow cumulus clouds has subsequently been supported by observations (Nitta and Esbensen 1974; Johnson and Lin 1997) and more recently has been assumed partially responsible for the observed increase in low-level moisture leading to convective peaks of the MJO (Johnson et al. 1999; Kemball-Cook and Weare 2001; Benedict and Randall 2007). The importance of shallow clouds during the MJO is consistent with the cloud evolution during these events documented from satellite observations (Lau and Wu 2003; Masunaga et al. 2008; DelGenio et al. 2012; Riley et al. 2013; Barnes and Houze 2013). However, arguments for the idea that moistening and heating effects of shallow clouds is crucial to the MJO have mostly been based on the observed or simulated simultaneous increases in low-level moisture, surface convergence, and abundance in shallow clouds.
In this study, we evaluate the possible moistening and heating role of shallow cumulus clouds (hereafter shallow clouds) in the MJO in comparison to that of the circulation. We proposed a null hypothesis that shallow cumulus clouds are the cause for the commonly observed increase in low-level moisture and diabatic-heating induced low-level convergence leading to convective peaks of the MJO. To falsify this null hypothesis, we used the observations from the ARM site at Manus Island to estimate moistening from shallow clouds on observed moisture tendencies and compare them to moistening by the circulation derived from reanalysis products.

As part of the hypothesis testing, fluctuations in moisture and shallow clouds during large-scale convective events associated with the MJO and not associated with the MJO (non-MJO events hereafter, Ling et al 2013) over Manus are compared. This comparison is motivated by our conviction that mechanisms for the MJO cannot be fully understood by studying the MJO alone; any MJO mechanism must be unique only for the MJO and not applicable to other large-scale convective events. If increasing shallow clouds and their moistening were observed prior to rainfall peaks of both MJO and non-MJO events, they might be important to deep convection in general but cannot be considered a specific mechanism for the MJO.

First, this study documents the evolution in moisture, temperature, wind, and clouds of MJO and non-MJO large-scale convective events at Manus Island using the observations at this site. Second, it explores the global nature of these events using global reanalysis. The full methodology of this Chapter is described next (subsection 4.2). The method to estimate shallow cloud moistening from the available radar observations described in the previous Chapter (subsection 3.2.3) was also used here. Results are
divided in three sections: The first two sections document the moistening (subsection 4.3) and heating (subsection 4.4) evolution during MJO and non-MJO events from observations and reanalysis at Manus. In these chapters, reanalysis are first validated against sounding observations before they are used as proxies for unavailable observations. The third section of results (subsection 4.5) presents the global perspective of the MJO from reanalysis. A summary is given after that (subsection 4.6). Our results from these sections lead us to propose a new hypothesis about the MJO. This is presented at the end of this Chapter (subsection 4.7).

**4.2 Methods**

The general strategy of this study is to compare the local and advective moisture and temperature anomalies and their tendencies over Manus with the evolution of shallow clouds and their estimated tendencies in these fields in composites of MJO and non-MJO large-scale convective events. A two-tailed t-student test was applied to all composites of anomalies to mark results significant at the 95% confidence level.

In this part of the study we used the data from the Manus radar, soundings, and gauge processed from Chapter 3. Also, consistent with that chapter, daily anomalies to were calculated by removing their respective climatological daily means. After that, a 5-day running mean was applied to the total and anomalies data. Local and advective tendencies were calculated after the smoothing. All derivatives were computed with a centered difference scheme. The 5-day smoothed climatological daily mean was used to fill missing observations of boundary layer or melting level heights in the soundings. Geopotential height from ERAI was derived from geopotential (Φ) as $Z=\Phi/g$. The ERAI
and EC-ARM data were first validated against the Manus soundings before they were used for global diagnostics.

### 4.2.1 Event Identification

Large-scale convective events over Manus were identified using the TRMM 3B42 precipitation data. Events that showed propagation from the Indian Ocean to the tropical western Pacific were classified as MJO events, the others as non-MJO events. Each event was considered from days -35 to 15 relative to its rainfall peak (day 0). Details of this procedure are described in Appendix C. A total of 22 MJO events and 15 non-MJO events were identified, of which 13 MJO events and 11 non-MJO events were covered by MMCR and sounding observations (Table 2.1). These events were covered on average 91% of the time by the MMCR and 80% by the soundings. We used the 13 MJO events and 11 non-MJO events in our composite analysis in section 4.3, where the evolution of clouds is examined in detail, but we used the 22 MJO events and 15 non-MJO events in sections 4.4 and 4.5 were, except for subsection 4.4.4, only soundings were used.

The composite of rainfall anomalies of the 13 MJO events (Fig. 4.1a) shows eastward propagation at a mean speed of 7.1 m s$^{-1}$ from the Indian Ocean towards the central Pacific. The amplitude of the RMM index associated to this composite is always larger than 1 after MJO initiation over the Indian Ocean (Fig. 4.1c). As expected, the propagation of the positive rainfall anomalies is accompanied by anomalous low-level (850 hPa) westerlies to the West and easterlies to the East. The composite of rainfall anomalies for the 11 non-MJO events shows an isolated convective center over Manus (Fig. 4.1b) with no sign of zonal propagation, and its associated RMM index composite is
always less than 1 (Fig. 4.1d). At day 0, the composite of rainfall anomalies over the tropical western Pacific are similar in spatial scale and strength for both types of events (Figs. 4.1e and f). The main difference is that westerly anomalies west of the convective center at Manus extend through the Maritime Continent to the eastern Indian Ocean only in MJO events. We also made composites using the 22 MJO events and 15 non-MJO events and found essentially no qualitative or quantitative differences with the ones discussed here.

Table 4.1: MJO and non-MJO events over Manus during the months of October to April. See Appendix C for details of C1 and C2.

<table>
<thead>
<tr>
<th>MJOs events</th>
<th>Non-MJOs events</th>
</tr>
</thead>
<tbody>
<tr>
<td><strong>Day 0</strong></td>
<td><strong>Conditions Satisfied</strong></td>
</tr>
<tr>
<td>20011013</td>
<td>C1 and C2</td>
</tr>
<tr>
<td>20011208</td>
<td>C1</td>
</tr>
<tr>
<td>20021126</td>
<td>C1 and C2</td>
</tr>
<tr>
<td>20030402*</td>
<td>C1</td>
</tr>
<tr>
<td>20031219*</td>
<td>C1 and C2</td>
</tr>
<tr>
<td>20040207*</td>
<td>C1 and C2</td>
</tr>
<tr>
<td>20040314*</td>
<td>C1</td>
</tr>
<tr>
<td>20041009*</td>
<td>C1 and C2</td>
</tr>
<tr>
<td>20050408</td>
<td>C1 and C2</td>
</tr>
<tr>
<td>20060121*</td>
<td>C1</td>
</tr>
<tr>
<td>20060330*</td>
<td>C1</td>
</tr>
<tr>
<td>20070103*</td>
<td>C1 and C2</td>
</tr>
<tr>
<td>20070128*</td>
<td>C1</td>
</tr>
<tr>
<td>20080102*</td>
<td>C1 and C2</td>
</tr>
<tr>
<td>20090422*</td>
<td>C1 and C2</td>
</tr>
<tr>
<td>20091120*</td>
<td>C1 and C2</td>
</tr>
<tr>
<td>20100111*</td>
<td>C1</td>
</tr>
<tr>
<td>20110326</td>
<td>C1</td>
</tr>
<tr>
<td>20120319</td>
<td>C1 and C2</td>
</tr>
<tr>
<td>20121106</td>
<td>C1</td>
</tr>
<tr>
<td>20130107</td>
<td>C1</td>
</tr>
<tr>
<td>20130225</td>
<td>C1 and C2</td>
</tr>
</tbody>
</table>

* Covered by the ARM observations at Manus.
Figure 4.1: Composites of (upper panels) longitude-time diagrams of daily anomalies in rainfall (colors) and 850 hPa zonal wind (contours with interval of 1 m s$^{-1}$, solid contours for westerlies and dashed for easterlies, thick solid for zeros) for MJO (left) and non-MJO (right) events over Manus Island, and their corresponding (middle panels) RMM phase diagrams (thick black lines for the composites, purple for individual events) for days -35 to 25 and (lower panels) spatial rainfall anomalies (colors) and 850 hPa zonal wind vectors at day 0. Tracking lines are in orange; red dots correspond to day 0 at Manus and blue dots to the end point of the track (see Appendix 3). Only results significant at the 95% confidence level are shown.
4.2.2 Budget Analysis

Temperature and moisture anomalies were diagnosed using ERAI over the Manus domain (Appendix C). The equations for potential temperature ($\theta$) and specific humidity ($q$) tendencies can be written as (Yanai and Tomita 1998):

$$\frac{\partial \theta}{\partial t} = (-V \cdot \nabla \theta) + \frac{Q_1}{c_p}(p_s/p)^R/c_p$$

(4.1)

$$\frac{\partial q}{\partial t} = -V \cdot \nabla q - \frac{Q_2}{L}$$

(4.2)

where $R$ is the gas constant, $c_p$ the specific heat at constant pressure, $L$ the, $\nabla$ the three dimensional gradient operator, $V$ the three dimensional wind vector, $p$ pressure, and $p_s$ a reference pressure at 1000 hPa, and $Q_1$ and $Q_2$ the apparent heat and moisture sources respectively (Yanai et al. 1973). For consistency with previous studies (e.g., Johnson et al. 2015) that diagnosed $Q_1$ and $Q_2$ we assumed an uncompressible flow ($\nabla \cdot V = 0$).

From Yanai et al (1973), $Q_1$ represents the combined effects of heating by evaporation $e$, condensation $c$. and radiation, $Q_R$. Similarly, the moisture sink $Q_2$ represents the combined effects of evaporation $e$, condensation $c$, and vertical turbulent fluxes, $\partial \{ \omega^* q^* \} / \partial p$. That is, $Q_2/L = c - e + \partial \{ \omega^* q^* \} / \partial p$, where $\{ \}$ denotes areal mean, and * deviation from the mean. In Eqs. (4.1) and (4.2), $T$, $q$, $V$, $c$ and $e$ are all areal means but $\{ \}$ is not used for simplicity. For gridded products (e.g. reanalysis and model simulations), the area for $\{ \}$ is the grid size. For observations, it is usually the area covered by a sounding array (Yanai et al. 1973; Johnson et al. 1996; Johnson and Ciesielski 2013). It is not immediately clear what the area for $\{ \}$ should be for point measurements such as those from Manus. It will be shown that the moisture variability observed by Manus soundings is comparable to that from EC-ARM interpolated to
Manus, and from ERA-I averaged over a domain of 10° x 10° centered at Manus. The connection between the point measurement and areal means partially comes from the temporal smoothing of Manus time series (section 3). This suggests Eqs. (4.1) and (4.2) can be applied to temporally smoothed observations at Manus to represent a horizontal area of 10.5° X 10.5° area (142.5E, 153E, 7.5S, 3N) centered at Manus (Appendix C).

We used this domain to estimate the tendencies for both MJO and non-MJO convective events.

\(Q_1\) and \(Q_2\) were computed as the residual of (4.1) and (4.2) respectively. \(q, \theta, V\) and \(Q_1\) were decomposed into their daily mean seasonal cycle components, \(\bar{q}, \bar{\theta}, \bar{V}, \bar{Q}_1\), and \(\bar{Q}_2\), and their anomalies, \(q', \theta', V', Q'_1\), and \(Q'_2\) (hereafter referred to as mean and anomalous components). A 5-day running mean was applied after that. By removing the mean, \(\bar{\theta}_t = -V \cdot \nabla \bar{\theta} - V' \cdot \nabla \bar{\theta}' + \frac{\bar{Q}_1}{c_p} (p_s/p)^{R/c_p}\) from (4.1) and \(\bar{q}_t = -V \cdot \nabla \bar{q} - V' \cdot \nabla \bar{q}' - \frac{\bar{Q}_2}{L}\) from (4.2), the anomalous temperature and moisture tendency equations are

\[
\frac{\partial \theta'}{\partial t} = -V' \cdot \nabla \bar{\theta} - V' \cdot \nabla \theta' - (V' \cdot \nabla \theta' - V' \cdot \nabla \bar{\theta}') + \frac{\bar{Q}_1}{c_p} (p_s/p)^{R/c_p}\]

\[
= -\left( u' \frac{\partial \bar{\theta}}{\partial x} + v' \frac{\partial \bar{\theta}}{\partial y} + \omega' \frac{\partial \bar{\theta}}{\partial p} \right) - \left( \bar{u} \frac{\partial \theta'}{\partial x} + \bar{v} \frac{\partial \theta'}{\partial y} + \bar{\omega} \frac{\partial \theta'}{\partial p} \right) - \left( u' \frac{\partial \bar{\theta}}{\partial x} - u' \frac{\partial \theta'}{\partial x} + v' \frac{\partial \theta'}{\partial y} \right) - \left( \bar{v} \frac{\partial \bar{\theta}}{\partial y} + \omega' \frac{\partial \bar{\theta}}{\partial p} - \bar{\omega} \frac{\partial \theta'}{\partial p} \right) + \frac{\bar{Q}_1}{c_p} (p_s/p)^{R/c_p}\]

\[
\frac{\partial q'}{\partial t} = -V' \cdot \nabla \bar{q} - V' \cdot \nabla q' - (V' \cdot \nabla q' - V' \cdot \nabla \bar{q}') - \frac{\bar{Q}_2}{L}\]
\[
\begin{align*}
&= -\left( u' \frac{\partial q}{\partial x} + v' \frac{\partial q}{\partial y} + \omega' \frac{\partial q}{\partial p} \right) - \left( \bar{u} \frac{\partial q'}{\partial x} + \bar{v} \frac{\partial q'}{\partial y} + \bar{\omega} \frac{\partial q'}{\partial p} \right) - \left( u' \frac{\partial q'}{\partial x} - u \frac{\partial q'}{\partial x} + v' \frac{\partial q'}{\partial y} - v \frac{\partial q'}{\partial y} \right) \\
&\quad - \frac{v' \partial q'}{\partial y} + \omega' \frac{\partial q'}{\partial p} - \omega' \frac{\partial q'}{\partial p} - \frac{q'_l}{L} \\
&\quad - \frac{(p/p_s)}{R/C_p} \frac{\partial T'}{\partial t} - \frac{(p/p_s)}{R/C_p} \frac{\partial \theta'}{\partial t} - \frac{(p/p_s)}{R/C_p} \frac{\partial q'}{\partial t} - \frac{q'_l}{L} \\
&\quad + \left( \frac{(p/p_s)}{R/C_p} \frac{\partial T'}{\partial y} \right) + \left( \frac{(p/p_s)}{R/C_p} \frac{\partial \theta'}{\partial y} \right) + \left( \frac{(p/p_s)}{R/C_p} \frac{\partial q'}{\partial y} \right) - \frac{q'_l}{L} \\
&\quad + \frac{RT}{\theta} \frac{\partial Z'}{\partial t} = \frac{-RT}{\theta} \frac{\partial Z'}{\partial t} \\
&\quad \text{(4.4)}
\end{align*}
\]

where \( u, v, \) and \( \omega \) are the zonal, meridional, and vertical wind components, respectively.

The first terms in the right hand side of (3) and (4) respectively, \(-\mathbf{V} \cdot \nabla \bar{\theta} \) and \(-\mathbf{V} \cdot \nabla \bar{q} \), represent the advection of mean temperature and moisture by anomalous wind. The second terms, \(-\mathbf{V} \cdot \nabla \theta' \) and \(-\mathbf{V} \cdot \nabla q' \), represent advection of temperature and moisture anomalies by mean winds. The third terms, \(-(\mathbf{V} \cdot \nabla \bar{\theta} - \mathbf{V} \cdot \nabla \bar{\theta}') \) and \(-(\mathbf{V} \cdot \nabla q' - \mathbf{V} \cdot \nabla q') \), represent anomalous nonlinear temperature and moisture advection, which will hereafter be expressed as \(-(\mathbf{V} \cdot \nabla \theta)') \), and \(-(\mathbf{V} \cdot \nabla q)') \).

We followed the standard practice to estimate \( Q_1 \) as a residual of Eq. 4.2 (Yanai and Tomita 1998), but to diagnose \( T' \) we multiplied each of the terms of (4) by \((p/p_s)^{R/C_p} \) which yields to:

\[
\begin{align*}
\frac{\partial T'}{\partial t} &= -\left( u' \frac{\partial T'}{\partial x} + v' \frac{\partial T'}{\partial y} + (p/p_s)^{R/C_p} \omega' \frac{\partial \theta'}{\partial p} \right) - \left( \bar{u} \frac{\partial T'}{\partial x} + \bar{v} \frac{\partial T'}{\partial y} + (p/p_s)^{R/C_p} \bar{\omega} \frac{\partial \theta'}{\partial p} \right) \\
&\quad + \left( \frac{(p/p_s)}{R/C_p} \frac{\partial T'}{\partial p} \right) + \left( \frac{(p/p_s)}{R/C_p} \frac{\partial \theta'}{\partial p} \right) + \left( \frac{(p/p_s)}{R/C_p} \frac{\partial q'}{\partial p} \right) - \frac{q'_l}{L} \\
&\quad + \left( \frac{(p/p_s)}{R/C_p} \frac{\partial T'}{\partial y} \right) + \left( \frac{(p/p_s)}{R/C_p} \frac{\partial \theta'}{\partial y} \right) + \left( \frac{(p/p_s)}{R/C_p} \frac{\partial q'}{\partial y} \right) - \frac{q'_l}{L} \\
&\quad + \frac{RT}{\theta} \frac{\partial Z'}{\partial t} = \frac{-RT}{\theta} \frac{\partial Z'}{\partial t} \\
&\quad \text{(4.5)}
\end{align*}
\]

In addition to diagnosing \( T' \) and \( q' \), we also diagnosed anomalies in geopotential height, \( Z' \). As it will be shown in sub-section 4.5, signals in \( Z' \) and \( \partial Z'/\partial t \) are stronger than \( T' \) and \( \partial T'/\partial t \) respectively. \( Z' \) is in hydrostatic balance with \( T' \):

\[
\frac{\partial Z}{\partial \ln(p)} = \frac{-RT}{\theta} \frac{\partial T}{\partial t} \\
\text{(4.6)}
\]
4.2.3 Diabatic Heating Profiles of Shallow Clouds

Diabatic heating profiles from shallow precipitating clouds were estimated following the method of Schumacher et al. (2009). This method prescribes a condensation profile typical to the type of clouds under analysis. The one we used for shallow clouds is shown in Fig. 4.2a. This profile is consistent with that of Schumacher et al. (2009) (their Fig. 1). The mean rainfall and echo top height of these clouds determines the magnitude and height of the idealized profile for a time period. In this study, diabatic heating profiles from shallow non-precipitating clouds were estimated by prescribing an evaporation profile in addition to that of condensation. Both profiles are symmetric opposites (Fig. 4.2a). The magnitude of each profile was determined by the mean liquid water content (LWC, Appendix B) of the non-precipitating shallow clouds. The sum of both evaporation and condensation profiles results in the typical shallow non-precipitating clouds profile (Fig. 4.2b), with positive and negative heating at the bottom and top respectively (Nita and Esbensen 1979; Lin and Johnson 1996). Both of these methods are highly idealized. Their major weakness is the arbitrary shape prescribed to the profiles. We estimated daily heating profiles from precipitating and non-precipitating shallow clouds by integrating their 30, 60, and 90 min averaged profiles. Only results using hourly profiles are reported here, but they are virtually the same for the other temporal resolutions.
4.3 Results Part I: Evolution of Moisture Anomalies from Manus

This subsection of results investigates the relation of shallow cloud moistening and the low tropospheric increases in moisture prior rainfall peaks of MJO and non-MJO events. Time goes from the right to left in temporal evolution composites at Manus for a direct comparison to longitude-height snapshots of the evolution of the MJO.

4.3.1 Moisture and its Tendencies

The MJO composite of Manus soundings reveals positive anomalies in low-level moisture (Fig. 4.3a) that emerge on day -10 and rapidly deepen towards day 0 (rainfall peak, Fig. 4.3i). Many studies have documented such moisture evolution leading to MJO rainfall peaks (Jones and Weare 1996; Johnson et al. 1999; Myers and Waliser 2003; Kiladis et al. 2005; Benedict and Randall 2007; Johnson and Ciesielski 2013; Yoneyama
et al. 2013). It occurs also prior to the rainfall peak of non-MJO events\(^1\), although less prominently (Fig. 4.3b). Immediately after the rainfall peak (day 1-3) of both types of convective events (Figs. 4.3i and j), the largest positive moisture anomalies reach up to 5-6 km height. There is almost no statistically significant signal of negative anomalies in either type of event (Figs. 4.3a and b), but negative tendencies are significant for MJO events during days 1-5 from 0 to 8 km (Fig. 4.3c). These tendencies are much weaker for non-MJO events (Fig. 4.3d). EC-ARM reproduces these features well (Figs. 4.3 e-h). The stepwise increase in the moist layer towards an MJO rainfall peak documented by Johnson and Ciesielski (2013) is more evident in EC-ARM than in soundings here.

Composites of integrated water vapor (IWV) estimated from the MWR and soundings are virtually the same for MJO events, and to a lesser degree for non-MJO events (Figs. 4.4a and b). IWV starts increasing 20 and 15 days prior to the rainfall peaks of the MJO and non-MJO events respectively. Positive IWV tendencies reach their maxima near day -5 and become negative near day 0 (Figs. 4.4c and d). Low-level (below the freezing level, ~4.5 km) IWV tendencies exhibit a pattern similar to those in total IWV, except that they reach their maxima earlier, around day -3, consistent to the increase in the depth of the moisture layer (Fig. 4.4a).

\(^{1}\)The secondary rainfall peak of non-MJO events near day -30 (Fig. 4j) is related to both preceding MJO and non-MJO events. See Appendix 3 for details.
Figure 4.3: Time-height composites of (a, b and e, f) daily mean water anomalies in vapor mixing ratio and (c, d, and g, h) their local tendencies from (a-d) soundings and (e-h) EC-ARM for (left) MJO and (right) non-MJO events at Manus Island. Composites of anomalies in daily gauge rainfall are shown in (i) and (j). Contours and circles mark results significant at the 95% confidence level.
Figure 4.4: Composites of daily (a,b) total integrated water vapor mixing ratio (IWV), (c,d) their local tendencies estimated from the microwave radiometer (MWR, black lines) and soundings (blue lines), and (e,f) daily gauge rain rates for (left) MJO and (right) non-MJO events at Manus Island. IWV over the lower troposphere (0.1 - 4.5 km, red lines) was also included in (a) and (b) with their scale given at the right ordinate. Circles and squares mark results significant at the 95% confidence level.

4.3.2 Shallow Clouds

For both MJO and non-MJO events, the frequency of occurrence (Figs. 4.5a – d) is higher for non-precipitating than for precipitating shallow clouds before and after the rainfall peaks (Figs. 4.5i and j). Both types of clouds undergo synoptic fluctuations and there tend to be more frequent non-precipitating shallow clouds prior to rainfall peaks of MJO events compared to non-MJO events. Neither of them shows a clear increase in their occurrence or depth (echo-top height) in total (Figs. 4.5a – d) and anomalies (Figs. 4.5e –
h) before the rainfall peaks of either MJO or non-MJO events. The anomalies in the daily frequency of non-precipitation shallow clouds during both types of events are qualitatively related to those in boundary layer moisture (compare blue and red lines of Figs. 4.6 a-d): positive (negative) anomalies in shallow non-precipitating clouds tend to concur with negative (positive) boundary layer moisture anomalies. A dry (moist) boundary layer is detrimental (nourishing) for convection to grow deep. Consequently, there are relatively more (less) shallow non-precipitating clouds when the boundary layer moisture is relatively low (high). This is particularly evident during days -25 to -20 of MJO events. This result is consistent with previous studies that have shown abundance of small shallow clouds during suppressed phases of the MJO or after its rainfall peak (Johnson and Lin 1997; Riley et al 2011; Barnes and Houze 2013). While this result makes physical sense, there is no statistically significant correlation between shallow clouds and boundary-layer moisture. We found no evident relation between shallow cloud occurrence and other environmental parameters, such as the boundary-layer height and lower-tropospheric stability (Fig. 4.6 e and f). Without clearly identified large-scale mechanisms, shallow clouds are regarded as random or stochastic phenomena. These results imply that, even if shallow clouds can effectively moisten the lower troposphere, they do not play a significant role in the observed low-level moisture increases leading to the rainfall peaks of large-scale convective events. The role of shallow clouds in moistening the lower troposphere, however, needs to be quantified. This is discussed in the next subsection.
Figure 4.5 Time-height composites of daily occurrence frequency of (a, b) non-precipitating (NP) shallow clouds (SC), (c, d) precipitating (P) SC, and (e-h) their anomalies for (left) MJO and (right) non-MJO events at Manus Island. SC daily accumulated gauge rain rates are added to (c) and (d) and their anomalies to (g) and (h) with their scales given at the right ordinates. Total daily gauge rain rates (k, l) are also shown. Contours mark results significant at the 95% confidence level.
Figure 4.6: Composites of anomalies in daily (a, b) occurrence of non-precipitating (NP) and precipitating (P) shallow clouds (SC), (c, d) mean water vapor mixing ratio in the boundary layer (BL, from 0.1 to 2 km), (e, f) BL height (BL') and lower tropospheric stability (LTS'), and (g, h) rain rates for (left) MJO and (right) non-MJO events at Manus Island. Circles mark results significant at the 95% confidence level.

4.3.3 Shallow Cloud Moistening

Composites of estimated daily moistening tendency by shallow clouds $M$ are shown in Figs. 4.7a-d, where the largest and smallest values correspond to $\Delta t = 30$ min and 90 min in Eq. (3.6), respectively, and the red line to $\Delta t = 60$ min. Because the occurrence of non-precipitating shallow clouds is slightly higher than that of the precipitating ones during both types of events, moistening by the former (0.9 mm day$^{-1}$ on average) is also slightly
larger than that of the latter (0.6 mm day\(^{-1}\)). However, moistening per cloud occurrence is larger (1.4 mm day\(^{-1}\) hr\(^{-1}\)) for precipitating than non-precipitating shallow clouds (1 mm day\(^{-1}\) hr\(^{-1}\)). The combined moistening by total shallow clouds (1.5 mm day\(^{-1}\)) is similar during MJO and non-MJO events. The magnitude of the moistening tendency \(M\) is comparable to the sounding-based low-level IWV tendencies (Fig. 4.7i and j) prior to rainfall peaks of both types of events (up to 1.2 mm day\(^{-1}\)). However, neither the total nor anomalies of \(M\) increases toward the rainfall peaks; they seem to fluctuate randomly in both types of convective events (Fig. 4.7 a-h) as does the cloud occurrence (Figs. 4.5a – d). The daily mean LWC of both types of clouds is nearly invariable during both types of convective events (Figs. 4.7g and h). The temporal fluctuations in shallow cloud moistening \(M\) mainly come from the cloud occurrence.

In summary, the estimated moistening tendencies of precipitating and non-precipitating shallow clouds do not vary in concert with the observed lower-tropospheric moisture tendency in either type of convective events. Shallow clouds are a non-negligible moisture source for the lower troposphere, but they do not explain the observed low-level moisture increases leading to the rainfall peaks in MJO and non-MJO events. We next explore what subgrid physical moisture tendency terms from EC-ARM might reveal.
Figure 4.7: Composites of daily moistening ($M$) from (a, b) non-precipitating (NP) and (c, d) precipitating (P) shallow clouds (SC), (e, f) tendencies of water vapor mixing ratio (IWV) from the soundings integrated (denoted by []) over the lower troposphere (from 0.1 to 4.5 km), (g, h) daily mean liquid water content (LWC) of SC, and (i, j) gauge rain rates for (left) MJO and (right) non-MJO events at Manus Island. Redlines in a-h (shaded areas) mark $M$ using $\Delta t = 60$ min (30-90 min).

4.3.4 Moisture Tendencies in EC-ARM

The physical moisture tendencies from EC-ARM, $\partial q/\partial t_{\text{physics}}$, are a combination of contributions from parameterization schemes for cumulus, turbulence, and microphysics. Their values below the freezing level cannot be attributed solely to shallow clouds.
Subgrid moisture tendencies above the boundary layer in EC-ARM are always negative during MJO and non-MJO events (Fig. 4.8 a and b). They are dominated by condensation and precipitation. There are two semi-perpetual peaks below and above the freezing level, which have been observed at many tropical locations (Yanai et al. 1973; Johnson et al. 1996). During MJO events, the anomalous moisture tendencies, \( \left( \frac{\partial q}{\partial t} \right)_{\text{physics}} \)', show drying that progressively deepen from the low to the mid troposphere at the same time that the rain rate increases toward its peak (Fig. 4.8 c). This implies a growth in the depth of precipitating clouds. For MJO events, there is an apparent deepening in positive \( \left( \frac{\partial q}{\partial t} \right)_{\text{physics}} \)', starting at the low levels (2 – 4 km) on days -20 to -25 days and reaching 8 – 10 km on day -5 days (Fig. 4.8c). But this repeats again immediately after the rainfall peak with greater strength. So it cannot be instrumental to the development of deep convection of the MJO. Anomalous drying and moistening are less significant during non-MJO events (Fig. 4.8 d). These results imply that shallow-cloud moistening in EC-ARM, if any, is overwhelmed by the drying from precipitation processes and, consequently, it cannot be detected from the total physical tendency term. It is interesting that moistening tendencies are limited to the boundary layer in the total tendencies (Fig. 4.8 a-b), but it extends into the lower troposphere in the anomalies. Since these characteristics are more notable following the rainfall peaks, a possible explanation is that they are caused by the evaporation of stratiform precipitation after the rainfall peaks.
The physical moisture tendencies of EC-ARM integrated in the lower troposphere (1 - 4.5 km, hereafter denoted by $[\partial q/\partial t_{\text{physics}}]$) are compared to the shallow-cloud moistening tendency $M$ in Fig. 4.9. $[\partial q/\partial t_{\text{physics}}]$ includes effects from shallow and deep clouds, turbulence, and microphysical processes, whereas $M$ includes only evaporation of condensed water in shallow clouds. Even during the suppressed phase of the MJO (days - 20 to -10), deep clouds contribute significantly to total rainfall (Riley et al 2011; DelGenio 2012; Barnes and Houze 2013). The observed total daily rainfall during the suppressed phase of the MJO in our composites is close to 10 mm day$^{-1}$ (Fig. 4.9 c). This value is well captured in EC-ARM and is approximately one order of magnitude larger.
than that from precipitating shallow clouds. The dominance of drying by precipitation in \( \frac{\partial q}{\partial t_{\text{physics}}} \) is the main reason for its difference from \( M \). The other possible causes for this difference however. One of them includes possible cause is errors in the physical moisture tendencies (Mapes and Bacmeister 2012). Figure 4.9 clearly demonstrates that neither \( M \) nor \( \frac{\partial q}{\partial t_{\text{physics}}} \) can explain \( \frac{\partial q}{\partial t} \) observed in the soundings at Manus or in EC-ARM. Next, we explore if moistening by the circulation can.

Figure 4.9: Composites of daily (a, b) moistening (\( M \)) from non-precipitating and precipitating shallow clouds combined (redlines for \( \Delta t = 60 \) min, shaded areas for \( \Delta t = 30-90 \) min), the physical moisture tendency from EC-ARM (\( \frac{\partial q}{\partial t_{\text{physics}}} \)) and the (c, d) moisture tendencies from the soundings (\( \frac{\partial q}{\partial t} \)) vertically integrated (denoted as []) over the lower troposphere (0.1 to 4.5 km), and (b, c) total rain rates from gauges and EC-ARM for (left) MJO and (right) non-MJO events at Manus Island.
4.3.5 Moisture Tendencies in ERAI

Local moisture tendency anomalies from ERAI over the Manus domain (Fig. 4.10a) are similar to those observed by the soundings (Fig. 4.4 c-d) and produced by EC-ARM (Fig. 4.4 g-h) at Manus during MJO and non-MJO events. ERAI subgrid tendency anomalies, \(-\frac{Q_2}{L}\) (Fig. 4.10 e-f), and the moisture anomalies related to physical processes in EC-ARM (Fig. 4.10 c-d) have also the same quantitative and qualitative characteristics. These comparisons lend confidence to use ERAI to diagnose the advection terms of Eq. (4.4) that cannot be made using the observations.

The largest terms of Eq. (4.4) are \(-\frac{Q_2}{L}\) (Fig. 4.11 e-f) and \(\omega' \frac{\partial q}{\partial p}\) (Fig. 4.11 g-h). These two terms vary in tandem but with the opposite signs. Their combination produces a small but persistent positive moistening in the lower troposphere with relatively small fluctuations before and after both types of events (Fig. 4.11 i-j). For MJO events, among the terms of the right hand side of Eq. (4.4), only the anomalous nonlinear zonal moisture advection, \(-\left(\mathbf{u}' \cdot \frac{\partial q'}{\partial x}\right)\) (Fig. 4.11c), can quantitatively and qualitatively explain the local low-level tendency anomalies (Fig. 4.11a).

The we presented so far in this subsection are more evident and easy to quantify by vertically integrating the terms of Eq. (4.4) over the lower troposphere, from 1000 to 600 hPa (approximately from the surface to 4.5 km), and by analyzing them in groups (Fig. 4.11). The combination of \(-\frac{Q_2}{L}\) and \(-\mathbf{V}' \cdot \nabla \bar{q}\) (blue and red lines in Fig. 4.11 a-b respectively) is positive and remains approximately constant with its values ranging from 0.3 to 0.8 mm day\(^{-1}\) during both types of events (green in Fig. 4.11 a-b). These two terms in combination provide ‘background’ moistening. The vertical component of \(-\mathbf{V}' \cdot \nabla \bar{q}\), as
shown in Fig. 4.10 g-h, is much larger than the horizontal ones. During both types of events, the combination of drying by the anomalous nonlinear advection, \(- (V' \cdot \nabla q')'\) (blue in Fig. 4.11 c), and the background moistening by \(- Q_2'/L - V' \cdot \nabla q\) almost matches the local moisture tendency (magenta in Fig. 4.9c). But \(- Q_2'/L - V' \cdot \nabla q\) and drying by anomalous nonlinear meridional advection \(- v' \partial q'/\partial y'\), red in Fig. 4.11e and f) almost cancel each other. In consequence, the anomalous nonlinear zonal advection alone \(- u' \partial q'/\partial x'\), orange in Figs. 4.11e and f) can almost match the positive moisture tendency anomalies before rainfall peaks, especially during MJO events.

Advection of moisture anomalies by the seasonal mean circulation, \(- \nabla \cdot V q'\), apparently does not play any role in the increase of the total low-level tendency prior day -5. But this term is needed to explain the sharp drying after day -5. Although the balance of moisture anomalies in MJO and non-MJO events is qualitatively similar, there is one major difference. Negative anomalies in the low-level local moisture tendencies near the rainfall peaks are stronger in MJO events than non-MJO events (black in Fig. 4.11c-f). This difference is consistent with the observed boundary layer drying in MJO events (Figs. 4.4c and d).
Figure 4.10: Time-height composites of terms from the budget equation of moisture anomalies using ERAI for (left column) MJO and (right) non-MJO events at Manus Island (a – j). Contours mark the 95% confidence level. Composites of daily rainfall anomalies at Manus Island from rain gauges (black) and over the Manus domain from TRMM (purple) are shown in (k) and (l).
Figure 4.11: Composites of terms in the budget equation of moisture anomalies integrated from 1000 to 600 hPa (denoted by []) for (left) MJO and (right) non-MJO events averaged over the Manus domain based on ERAI. Rainfall anomalies at Manus Island from rain gauges (black) and over the Manus domain from TRMM (purple) are shown in (g) and (h).

In addition to diagnosing the total moisture anomalies, we also diagnosed their intraseasonally filtered components (20-100 day, Kim et al. 2009). Applying this procedure altered structurally the moisture budget. As shown in Fig. 4.12, by removing high frequency variability the local moistening tendency from days -20 to -15 appears to
be caused by subgrid processes (blue in Fig. 4.12 c-d). In this sense, the budget diagnostics solely based on intraseasonal anomalies are incomplete and misleading.

Figure 4.12 same as Fig. 4.11 but (a-f) for intraseasonally (20-100 days) filtered anomalies.
4.4 Results Part II: Evolution of Wind and Diabatic Heating from Manus

This section of results investigates the anomalies in horizontal wind, pressure, and temperature at Manus and the diabatic heating of shallow clouds during both, MJO and non-MJO events. Anomalies in \( u \), \( v \) and \( \omega \) wind in the composites were divided by their standard deviations.

4.4.1 Horizontal Wind

Zonal wind anomalies of MJO events (Fig. 4.13a) exhibit a typical first baroclinic structure with low-level (below 10 km) easterlies and upper-level (above 10 km) westerlies that start roughly on day -15 and switch its signs on day -5. The strong low-level westerlies, commonly referred to as westerly wind bursts (WWBs, Luther et al. 1983), are a robust feature of MJO events. For non-MJO events, there is no low-level easterly prior to day 0 and low-level westerlies are much weaker and shorter in duration (Fig. 4.13b) than those of MJO events.

MJO events are also characterized by low-level southerly wind anomalies extending from the surface up to 3 km before the rainfall peak, approximately during days -20 to -5, and northerly wind anomalies after the onset (Fig. 4.13g). Considering the latitude of Manus Island (2° S), the low-level southerly wind anomalies ahead of local MJO rainfall peak can be part of the frictional convergence of the MJO (Wang 1988). These low-level meridional wind anomalies are mostly absent from non-MJO events (Fig. 4.13h). The stark contrast in the zonal wind anomalies between MJO and non-MJO events cannot be explained in terms of circulation responses to their rainfall peaks, which are similar for both types of events (Figs. 4.13m and n).
Composites using EC-ARM (Figs. 4.13c, d, i and j) and ERAI (Figs. 4.13e, f, k and l) tend to underestimate the magnitudes of the horizontal wind anomalies. But in general they capture the basic characteristics (structure, evolution, and their statistical significance) of both MJO and non-MJO events.

Figure 4.13: Time-height composites of daily anomalies in (a - f) zonal ($u'$) and (g - l) meridional wind ($v'$) from soundings, EC-ARM, and ERAI for (left) MJO and (right) non-MJO events at Manus Island. Daily rainfall is shown at the bottom. The 95% confidence level is marked by contours in (a) – (l) and by circles in (m) – (n).
4.4.2 Pressure and Geopotential Height

Pressure anomalies in MJO events also exhibit a first baroclinic structure, with negative anomalies in the lower troposphere (below 8 km) and positive anomalies aloft (Fig. 4.14a) that start on day -10, peak around day -5, and end at day 2. These anomalies are roughly in quadrature with the zonal wind anomalies (Fig. 4.14a). Earlier, weaker anomalies with a similar structure but opposite signs are present around day -30 (near the surface) and -22 (upper level). After the rainfall peak, high-pressure anomalies at low levels reemerge at day 10. The pressure anomalies in non-MJO events have a baroclinic structure only during the rainfall peak, approximately from days -5 to 5, but this is much weaker compared to that during MJO events (Fig. 4.14b).

Anomalies in pressure from EC-ARM (Figs. 4.14c-d) reproduced adequately the observed characteristics of the structure and evolution of the observed (Figs. 4.14a and b). Similarly, the anomalies in geopotential height from ERAI (Figs. 4.14e-f) are consistent with the observed ones. The (re)analysis data (Figs. 4.14c-f) show that during MJO events the largest anomalous upward motion occur approximately 5 (2 to 3) days after the largest lower (upper) tropospheric negative (positive) pressure/geopotential height anomalies.
4.4.3 Temperature

The main temperature signature in MJO events is a vertical boomerang-shaped positive anomaly that peaks near day 0, at about 10 km (Fig. 4.15a). The peak (> 1°C) occurs almost simultaneously with that in rainfall (Fig. 4.15g), but the positive anomalies in
temperature start on day -10 and end on day 5. There are weaker (~0.5°C) but significant negative temperature anomalies in the mid-upper troposphere on days -30 to -15 and after day 5. These cold anomalies before and after MJO rainfall peaks exist also over the Indian Ocean (Ling et al. 2012). They are absent in non-MJO events. Both the upper level warm and cold anomalies are topped with anomalies of the opposite signs above 15 km. Positive mid-tropospheric anomalies also exist for non-MJO events (Fig. 4.15b) during the rainfall peak, but they are much weaker (≤ 0.5°C) and vertically (7 - 11 km) and temporally (<5 days) confined than those of MJO events (4 - 14 km, 5 - 10 days). This large difference in temperature anomalies cannot be explained by the small difference in rainfall between the two types of events (Figs. 4.15g and h).

EC-ARM (Figs. 4.15c and d) and ERAI (Figs. 4.15e and f) capture well the observed characteristics of temperature anomalies of both MJO and non-MJO events, with one exception: the boundary layer warmth during MJO events from days -20 to -5. This is a relatively small feature compared to the other ones discussed above. Also, this feature is the only one that could be associated to shallow convection. This can be seen as an indication that shallow convective heating is not crucial to the MJO over Manus.
Figure 4.15: same as Fig. 4.11 but for temperature ($T'$).

4.4.4 Local Temperature Tendencies

Positive mid to upper tropospheric anomalies in temperature, $T'$, and geopotential height, $Z'$, occur 10 days prior to rainfall peaks of MJO events (Figs. 4.14 and 4.15). Their local tendencies, $\partial T'/\partial t$ and $\partial Z'/\partial t$, first appear on day -15 (Fig. 4.16a) and reach their maxima on day -7. Negative tendencies start near the rainfall peaks (day 0), reach their maxima on days 3 – 4, and last for about 10 days. In contrast, for non-MJO events, positive
tendencies do not appear until 5 days before rainfall peaks and there is hardly any following negative tendency anomaly (Fig. 4.16b).

Both EC-ARM and ERAI reproduce well the characteristics of $\partial T'/\partial t$ and $\partial Z'/\partial t$ observed by the soundings for both types of events (Figs. 4.16 c-f). There are no significant temperature anomalous tendencies associated to the temperature anomalies in the boundary layer during days -20 to -5 (Fig. 4.15a).

![Figure 4.16: Time-height composites of (a – f) $\partial T'/\partial t$ (shaded), (c and d) $\partial p'/\partial t$, and (e-f) $\partial Z'/\partial t$ (contour intervals shown in left panels, negative dashed) from soundings, EC-ARM and ERAI over Manus for (left) MJO and (right) non-MJO events. Daily rainfall is shown at the bottom. Only results significant at the 95% confidence level are shown for $\partial T'/\partial t$; those in $\partial p'/\partial t$ and $\partial Z'/\partial t$ are hatched.](image)
4.4.5 Diabatic Heating Profiles of Shallow Clouds

The estimated diabatic heating profiles of precipitating (Fig. 4.17a and c) and non-precipitating (Fig. 4.17b and d) shallow clouds have synoptic variations with no systematic increase before or after the rainfall peaks of both, MJO and non-MJO events. This result is consistent with our diagnosis of section 4.3.2 and 4.3.3 (e.g. Fig. 4.5). Also, the temperature anomalies in the lower troposphere during MJO events (Fig. 4.15a) are confined to the boundary layer, and their tendencies are statistically zero (Fig. 4.16a). The evolution of these shallow-cloud heating profiles contrasts with that in pressure (Fig. 4.14a) and zonal wind (Fig. 4.13a) that display well defined baroclinic structures.

Figure 4.17: Time-height composites of daily diabatic heating from (a, b) non-precipitating, (c, d) precipitating shallow clouds for (left) MJO and (right) non-MJO events at Manus Island. Composites of total rainfall are shown in (i) and (j). Daily mean rainfall from shallow clouds is added to (c) and (d) and its anomalies to (i) and (j) with their scales given at the right ordinates.
4.4.6 Heating Tendencies in ERAI and EC-ARM

Anomalies in the temperature tendencies due to parameterized physical processes from EC-ARM (Figs. 4.18a and b) are very similar to the apparent heating source, $Q_1/C_p$, estimated as a residual of the thermodynamic equation using ERAI (Figs. 4.18c and d). These similarities and those in the total anomalies (Fig. 4.15) lend confidence to using ERAI data in our following diagnosis. The apparent heating source, $Q_1/C_p$ (Fig. 4.18c), alone cannot explain the tropospheric signals in $\partial T'/\partial t$ and $\partial Z'/\partial t$ of MJO events (Fig. 4.16e) for two reasons. $Q_1/C_p$ are out of phase with $\partial T'/\partial t$, and $Q_1/C_p$ is almost identical in both MJO and non-MJO events but $\partial T'/\partial t$ are very different.

The observed $\partial T'/\partial t$ of both MJO and non-MJO events cannot be solely attributed to a single tendency term of the right hand side of Eq. (4.5). However, $\partial T'/\partial t$ can be almost entirely reconstructed without the meridional advection terms (compare Figs. 4.18g and h to Figs. 4.18e and f), which are negligibly small. The largest terms on the right hand side of Eq. (4.5) for both type of events are $Q_1/C_p$ and $-(p/p_s)^{R/C_p} \omega \partial \bar{\Theta}/\partial p$ (Figs. 4.18e and f). These terms nearly cancel each other, but their imbalance is comparable to the sum of all anomalous zonal advection terms: advection of the mean temperature by anomalous winds $-u' \partial T'/\partial x$, the anomalous nonlinear zonal eddy advection of temperature, $-(u' \partial T'/\partial x)'$, and the zonal advection of the anomalous temperature by the mean wind $(-\bar{u} \partial T'/\partial x)$. This implies that the $\partial T'/\partial t$ cannot be interpreted in the context of the weak temperature gradient approximation (Sobel et al. 2001), otherwise vertical advection of temperature would balance diabatic heating. It can be concluded that $\partial T'/\partial t$ associated to the MJO at Manus is caused by a combination of subgrid heat sources, and vertical and zonal advection by the large-scale circulation.
Figure 4.18: same as in Fig. 4.11 except for (a, b) anomalous EC-ARM physics temperature tendencies, and (c-h) different terms of Eq. (4.5) using ERAI.
4.5 Results Part III: Evolution of Global MJO Signals

In this section we present the global three-dimensional structure of the MJO signals diagnosed from the previous section using ERAI. Non-MJO events have no propagating features, their evolution is briefly discussed in sub-section 4.5.4. The comparisons of sounding and ERAI composites in the previous two sections of results showed that this reanalysis product captures most of the prominent local structure and evolution of anomalies in winds, pressure/geopotential height and temperature of both MJO and non-MJO events at Manus. While soundings from Manus are available for the data assimilation system that produced ERAI, this is not the main reason for their good match at this location. Ling et al. (2014) demonstrated that the ECMWF forecast system is not very sensitive to data sources from any specific locations or even regions. It is the global data input, including those from satellite observations, that is responsible for the fidelity of the forecast system. Therefore, it is safe to assume that ERAI captures the structure and evolution of the MJO equally well at other tropical locations.

We first present the spatial structure and evolution of the rainfall anomalies (Fig. 4.19) as a background for large-scale signals of the MJO to be discussed in this section. Robust positive rainfall anomalies that propagate eastward start over the Indian Ocean on day -15 and end over the central Pacific on day 20. This composite bears many similarities to that solely based on the Wheeler and Hendon (2004) MJO index (Zhang et al. 2012) with MJO phases 1 – 8 corresponding to days -15 to 20 here. Positive rainfall anomalies after day 10 are associated to the south convergence Pacific zone (Matthews et al. 1996) that have no discernable eastward propagation. Therefore, periods other than that from day -15 to day 10 can be considered as without MJO convection. As it will be
shown latter, the absence of precipitation anomalies is very important for the results of this section. We further explored the statistical significance of the absence of signals in precipitation anomalies form days other than -15 to 10 by regridding them into a coarser grid, we choose a 5° x 5°. As shown in Fig. 4.20, our previous diagnosis still holds.

Figure 4.19: Composites of TRMM 3B42 rainfall before and after the rainfall peaks of MJO events at Manus. Only results significant at the 95% confidence level are shown.
Figure 4.20 as Fig. 4.21 but regridded into 5° x 5° mesh.

4.5.1 Zonal-Vertical Structure

The zonal-vertical structure in temperature shows the known boomerang shape of positive anomalies (Kiladis et al. 2005) over the Indian Ocean starting from approximately day -15, when the MJO is initiated there (Fig. 4.21d). These temperature anomalies, $T'$, along with strong mid-upper tropospheric positive anomalies in geopotential height, $Z'$, propagate eastward at approximately 5 m s$^{-1}$ in the following 15 days from the Indian Ocean to the western Pacific. The most interesting feature is the continuous slow eastward propagation ($\sim$ 3 m s$^{-1}$, red dashed line) of positive $T'$ and $Z'$ in
the mid-upper troposphere from days 20 to 35 (Fig. 4.21 k-n). These signals have no boomerang shape and no associated rainfall (Fig. 4.21). The amplitude and strength of these signals decrease gradually during their propagation. Their e-folding decay rates are about 15 - 20 days. Decay rates due to linear mechanical damping and cooling in idealized models typically vary from as 2.5 days (Wu et al 2000) to 20 days (Hayashi and Miyahara 1987). It is also notable that the signals in upper-level $T'$ are in quadrature with $Z'$: positive $T'$ below and negative $T'$ above the local maximum in $Z'$. This result suggests that the cold anomalies close to the tropopause are consequence of the positive mid-tropospheric $Z'$ anomalies (Eq. 4.6). Previous studies (Sherwood et al. 2003) have attributed similar cold temperature anomalies during convective events to adiabatic lofting.

Fast eastward propagating signals in $T'$ and $Z'$ during days -10 to 5 over the Pacific and Atlantic are quickly separated from the main MJO structure. These signals are different from the slowly eastward propagating signals in $T'$ and $Z'$. Such fast propagating signals were previously documented (Milliff and Madden 1996; Matthews 2000; Yanai et al. 2000) and are presumably related to fast dry equatorial Kelvin waves (Sobel and Kim 2012).

There are two other interesting signals in $T'$ and $Z'$. First, negative mid-upper tropospheric $T'$ and $Z'$ during days -30 to -15 (Figs. 4.21 a-d) moving eastward from approximately 120 to 160 E at about 3.4 ms\(^{-1}\), passing Manus before the MJO rainfall peaks there. Similar temperature signals have been documented over the Indian Ocean (Mathews 2008; Ling et al. 2013). Second, positive mid-upper tropospheric $T'$ and $Z'$ emerge over the central Indian Ocean on day -25 and subsequently move eastward at
about 5 m s\(^{-1}\) from the Indian Ocean to the western Pacific (Fig. 4.21 b-g). They are always larger to the east of their local rainfall peak.

Figure 4.21: Longitude-height composites of \(T'\) (shaded) and \(Z'\) (contour interval 5 m, negative in green) averaged between 10.5 S and 10.5 N before and after the rainfall (purple) peak of MJO events at Manus. Only results significant at the 95% confidence level in temperature are shown; those in \(Z'\) are hatched. Vertical lines mark the longitude of Manus (147° E). The red dashed line approximately follows peaks in \(Z'\), with its averaged speed of 2.8 m s\(^{-1}\).
The structural evolution of the zonal-vertical circulation of the MJO also shows interesting features (Fig. 4.22). On day -30, there is a mid-upper tropospheric anomalous descent over the eastern Indian Ocean (Fig. 4.22a). This signal seems to propagate eastward slowly and, with its strength and statistical significance varying in time, becomes the downward branch of the “mini-Walker” circulation associated with MJO convection established on day -15 (Fig. 4.22d). The upward branch of this mini-Walker circulation manifests the onset of MJO convection over the Indian Ocean with a strong ascent and negative anomalies in low-level pressure (Figs. 4.22 d-g). Afterwards, strong downward westerly momentum transport is seen as another major mechanism for the typical westerly wind bursts associated to the MJO (Houze et al. 2000) from days 5 to 20 (Fig. 4.22 h-k).

The structures of the zonal-vertical circulation and $Z'$ associated to the MJO are consistent with the observed evolution of these fields at Manus (Fig. 4.14). Positive mid-upper tropospheric $Z'$ tend to have upward (downward) anomalous wind to the west (east) of its peak. Also, the largest anomalous upward motions tend to occur to the west of the largest lower tropospheric negative $Z'$. 
Figure 4.22: Longitude-height composites of \( Z' \) (shaded) and standardized anomalous wind vectors. \( Z' \) significant at the 95% confidence level are marked by red crosses. Vectors significant at the 95% confidence level are black. The red line in the left panel approximately follows peaks in \( Z' \) over the Indian Ocean, with its averaged speed of 4.9 m s\(^{-1}\). Yellow lines are rainfall (mm day\(^{-1}\)) anomalies. Other conventions are the same as in Fig. 4.17.
4.5.2 Horizontal Structure

The distinct anomalous signals in mid-upper tropospheric $T'$ and $Z'$ discussed in the previous section undergo a horizontal structural evolution. Fig. 4.23 shows their averages through a layer of 400 – 175 hPa (~8 – 13 km), where their signals are the strongest (Figs. 4.21 and 4.22). As expected from Fig. 7 and the hydrostatic balance (Eq. 4.6), $T'$ and $Z'$ tend to be collocated with each other. Their strongest signals in this layer start on day -10 in a “swallowtail” pattern of positive anomalies (Fig. 4.23e). This structure of the MJO has also been observed in rainfall (Zhang and Ling 2012). Starting from day -5, there are equatorial signals in $T'$ moving eastward very fast away from the ‘head’ of the swallowtail, at a gravity wave speeds (>20 m s$^{-1}$). These signals reach Africa by day 5 (Fig. 4.23h). Meanwhile, the ‘tails’ propagate at the known propagation speed of the MJO of about 5 m s$^{-1}$. From days 10 to 35 (Figs. 4.23 e-n), the tails continue their propagation over the central Pacific Ocean but at a slower speed (~ 3 m s$^{-1}$). They gradually merge toward the equator during this period, and disappear after they reach South America (after day 35).
Figure 4.23: Composites of anomalies in $T'$ (shaded), $Z'$ (contour interval 5 m, negative in green), and standardized anomalous wind vectors, averaged over the 400 - 175 hPa layer before and after the rainfall peak of MJO events at Manus (day 0). Only results significant at the 95% confidence level in $T'$ and wind are shown; those in $Z'$ are hatched.

Another interesting feature of the slowly propagating signals in $T'$ and $Z'$ is the equatorially symmetric pair of negative anomalies that start during day 10 over the Indian Ocean (Fig. 4.23i). This pair of negative anomalies appears to move in concert with the
positive anomalies over the central and eastern Pacific approximately from days 10 to 30 (Figs. 4.23 i-m). During this period they flank equatorial positive anomalies over the Indian Ocean that can be traced back to the detached head of the swallowtail pattern.

Prior to any rainfall signal in the composite (Fig. 4.20), there are negative $T'$ emerging on day -25 in the equatorial Indian Ocean and Maritime Continent, which move eastward slowly in a broken swallowtail pattern, and vanish as positive $T'$ are established on day -10 (Figs. 4.23 b-e). The most prominent wind anomalies during this period are located over the tropical western and central Pacific, but they do not seem to be directly related to the onset of MJO convection over the Indian Ocean (Figs. 4.23 a-d).

4.5.3 Meridional-Vertical Structure

On a meridional-vertical cross-section at the longitude of Manus (147° E), negative signals in $T'$ appear first near the equator in the upper troposphere on day -20 (Fig. 4.24a). They split into a north and south part in the following 5 days (Fig. 4.24b). Such splitting repeats for the upper-level positive $T'$ and $Z'$ (Figs. 4.24 d – g) as the convection center of the MJO propagates eastward (Fig. 4.23 e-g). Negative $T'$ in the mid troposphere, centered at about 8 km, appear again after the rainfall peak at Manus disappears on days 10 - 15 (Figs. 4.24 g and h). The meridional-vertical structure of the slow propagating $T'$ and $Z'$ over the central and eastern Pacific shows its unique equatorially symmetric structure (Fig. 4.25). During days 15 to 35 (Fig. 4.25 c-g), the structure of $Z'$ maintains its equatorial symmetry during its eastward propagation while their two centers tend merge toward the equator as seen in Fig. 4.25. There are no
statistically significant anomalies connecting this structure to the mid-latitudes or the lower troposphere during this period.

Figure 4.24: Latitude-height composites of anomalies in $T'$ (shaded), $Z'$ (contour interval 5 m, negative in green), and standardized anomalous wind vectors at the approximated longitude of Manus (147.5° E) before and after the rainfall (purple) peak of MJO events (day 0). Only results significant at the 95% confidence level are shown for $T'$ and wind; those in $Z'$ are hatched.
Figure 4.25: same as in Fig. 10 except following the red dashed line in Fig. 7 and without wind vectors.
4.5.4 Non-MJO Events

The comparison between MJO and non-MJO events helps to distinguish signals unique to the MJO from those that are universally related to large-scale convection in the tropics (Gill 1980). Almost everything we have discussed is unique to the MJO with one exception, the pairs of positively anomalies in upper-tropospheric geopotential height and mid-tropospheric temperature straddling the equator, resembling a swallowtail, near the rainfall peak (Figs. 4.26 a-d), and its meridional dissipation (Fig. 4.26 e-f). These anomalies during non-MJO events, however, do not move eastward, are much weaker and equatorially constrained than those during MJO events.

![Figure 4.26](image)

Figure 4.26: same as Fig. 4.19 for a-c, Fig. 4.17 for d, and Fig. 4.20 for e and f but for non-MJO events. The contour interval is 5 m in all panels.
4.5.5 ERAI Tendencies

This subsection highlights processes key to the mid-upper tropospheric signals in $T'$ and $Z'$ (e.g. Fig. 4.25) by analyzing the anomalous subgrid and advective temperature tendencies (left hand side terms of Eq. 4.5). The local tendencies of $T'$ and $Z'$ over the central and eastern Pacific propagate slowly (~3 m s$^{-1}$) and they weaken as they propagate. Consequently, the $\partial T'/\partial t$ and $\partial Z'/\partial t$ signals of $T'$ and $Z'$ over these regions are nearly undistinguishable (Figs. 4.27 h-n). Propagation signals in $\partial T'/\partial t$ and $\partial Z'/\partial t$ are much evident when $T'$ and $Z'$ developed in the Indian Ocean and strengthen in their propagation towards the tropical western Pacific (Fig. 4.27 a-g). Interestingly, signals in $\partial T'/\partial t$ and $\partial Z'/\partial t$ over the Indian Ocean can be detected as early as day -30 (Fig. 4.27a), when there are hardly any positive rainfall anomalies there (Fig. 4.20a, and purple lines in Fig. 4.27a). Figure 4.28 shows the vertical advection of mean temperature by anomalous vertical motions, $-(p/p_s)^{R/C_p} \omega' \partial \Theta / \partial p$. The strongest signals in this term are related to rainfall anomalies over the Indian and western Pacific Oceans and are almost canceled by $Q_1/c_p$ as demonstrated in section 4.4.5 (Figs. 4.18 c-f). But there are positive anomalies in $-(p/p_s)^{R/C_p} \omega' \partial \Theta / \partial p$ that propagate eastward from days -30 to 30. The propagation speed of these signals is in average close to 4.9 m s$^{-1}$ from the Indian Ocean to the western Pacific (red line across Figs. 4.28 a-g), and 2.8 m s$^{-1}$ from the western to the eastern Pacific (red line across Figs. 4.28 h-n). These positive anomalies sometimes merge with or are overwhelmed by the dominant signals from convection over the eastern Indian and western Pacific Oceans. But they persistently move eastward. From the Indian to the western Pacific Ocean, they lead the onset of MJO convection, and from the central and eastern Pacific they remain without MJO convection. Although these propagating
signals are indeed week, no other term of the right hand side of Eq. (4.5) showed discernable eastward propagation throughout this period. These signals of $-(p/p_s)^{R/Cp} \omega' \partial \bar{T}/\partial p$ imply anomalous downward advection of temperature. Eastward propagating downward vertical velocity anomalies, $\omega'$, can be discerned intermittently (Fig. 4.22).

Figure 4.27: Longitude-height composites of $\partial T'/\partial t$ (shaded) and $\partial Z'/\partial t$ (contour interval 0.5 m day$^{-1}$, negative in green) before and after the rainfall (purple lines) peaks of MJO events at Manus. All signals in $\partial T'/\partial t$ are shown. Statistical significance at the 95% confidence level in $\partial Z'/\partial t$ is hatched. Red dashed lines in the left and right panels approximately follow the peaks in $\partial Z'/\partial t$, their speeds are 4.9 and 2.8 m s$^{-1}$ respectively. Other conventions are the same as in Fig. 4.17.
Figure 4.28: Same as Fig. 4.17 except for $-(p/p)_R \omega \hat{\Theta}/\partial p$. Statistical significance at 95% confidence level is contoured. Red lines are from Fig. 4.22.
4.6 Summary and Discussion

We explored the idea that shallow cloud moistening and diabatic heating is crucial to the MJO (Kemball-Cook and Weare 2001; Benedict and Randall 2007) using more than 10 years of observations from the ARM Manus site. We compared the evolution of shallow clouds, their moisture and heating tendencies in MJO and non-MJO large-scale convective events. Such comparisons are part of our hypothesis testing because mechanisms common to both types of large-scale convective events cannot explain the existence and observed behavior of the MJO. The results derived from the test of this hypothesis are:

1. Moistening by both precipitating and non-precipitating shallow clouds is substantial, but it appears to exist all the time and fluctuate randomly before and after rainfall peaks of both MJO and non-MJO events (e.g. Fig. 4.7). From this result we propose that shallow cumulus clouds provide background moistening to the lower troposphere.

2. The commonly observed low-level moisture increases leading to the rainfall peaks of the MJO seem to be caused by anomalous nonlinear zonal advection. Background large-scale moistening and drying due to subgrid processes, vertical advection, and nonlinear meridional advection largely cancel each other during both MJO and non-MJO events.

3. Most of the features we present are common for both MJO and non-MJO events over Manus. One main difference between MJO and non-MJO events is severe drying near rainfall peaks that exists only in MJO events.

4. Shallow cloud diabatic heating could not explain low-level convergence prior the rainfall peak of MJO events (Fig. 4.14a).
Based on these results, we fail to reject the null hypothesis that shallow-cloud moistening and diabatic heating is not crucial to lead to MJO events rainfall peaks. We thus conclude that shallow clouds are a persistent source of background moistening to the lower troposphere; they can be a dominant source of low-level moistening during convectively suppressed periods (Johnson et al. 2015), but the observed increase in low-level moisture leading to the rainfall peak of the MJO cannot be attributed to them. This low-level moisture increases leading to the rainfall peak of the MJO is mainly a consequence of the circulation.

The results from this study confirm ubiquitous shallow clouds found in the TRMM observations (Barnes and Houze 2013) and in other satellite observations (Riley et al. 2011; DelGenio et al. 2012). Our results are also consistent with cloud-permitting model simulations of Janiga and Zhang (2015), which show moistening by shallow clouds exists evidently only in the suppressed phase of the MJO. The perpetual moistening by shallow clouds needs to be considered in models that represent multi-cloud evolution (Khouider and Majda 2006; Majda et al. 2007; Thual et al. 2014). We repeated the moisture budget analysis (section 4.5) after applying intraseasonal filtering to the daily anomalies. The result indicates that removing high-frequency variability may lead to an overestimated role of moistening by subgrid processes. This implies the importance of synoptic perturbations during MJO events as suggested previously (Majda and Stechmann 2009). However, our results show no evident difference in their roles between MJO and non-MJO events. To further explore these results, we repeated our ERAI moisture budget analysis by applying the 5-day running mean after all tendencies were
calculated. The results we obtained do not change any aspect of the diagnosis we have presented.

Some of our results contrast with some previous studies. Meridional moisture advection has been previously suggested as essential to the increasing moisture in the boundary layer (Wang 1988) and lower troposphere (Maloney 2009) leading to MJO rainfall peaks. In our diagnostics, the main effect of meridional advection is drying. We also estimated moistening of congestus clouds (cumulus clouds with tops above the freezing level but lower than 9 km) using the method described in section 3.2. Rainfall from these clouds is larger than their LWC available for evaporation, and consequently their effect is a net moisture sink without other processes, such as the circulation in response to their diabatic heating. This result is consistent with the drying tendencies from subgrid processes we found in the EC-ARM data, and with recent numerical simulations (Mechem and Oberthatler 2013), but contrary to the popular believe that congestus clouds precondition the environment through detrainment in daily time scales (e.g. Benedict and Randall 2007). Our results suggest that such preconditioning might be overwhelmed drying through precipitation. The crudeness of our moistening estimates and attenuation of cloud radar echoes due to heavy rain may, however, undermine the results on congestus clouds.

Our analyses lead us to propose a different view of the evolution of the MJO. We presented signals of the MJO that cannot that cannot be explained by local anomalies in convection. These signals can be separated in two sets:
1. Signals over Manus: baroclinic geopotential height anomalies (Fig. 4.11, left column) and positive mid-upper tropospheric temperature anomalies (Fig. 4.12 left column). These signals are robust over a 15 – 30 days period before and after MJO rainfall peaks. With similar anomalies in rainfall and diabatic heating, non-MJO events (no eastward propagation) produce signals in temperature and geopotential height with much smaller amplitudes and shorter durations.

2. Signals over the central and eastern Pacific: positive mid-upper tropospheric temperature anomalies (e.g. Fig. 4.23) with a slow eastward propagation (~3 m s\(^{-1}\)) and decay rate (~15 days) in the absence of MJO convection.

Both sets of signals are consistent with each other, and no single term of temperature equation anomalies (Eq. 4.3) we analyzed can explain either set of signals. Diabatic heating, mean and anomalous vertical and zonal flux convergence, including the nonlinear eddy components, are all needed to account for their tendencies.

Fig. 4.29 presents the new perspective of the evolution of the MJO derived from this study. Particularly, this figure synthesizes the MJO signals in \(T'\) and \(Z'\) from Figs. 4.15, 4.21, 4.22, 4.23, and 4.27. In this diagram, mid-upper tropospheric \(T'\) and \(Z'\) signals are in the hydrostatic balance. They first appear over the equatorial Indian Ocean before convective initiation of the MJO. There, \(T'\) and \(Z'\) signals and their associated vertical motions are relative weak (Fig. 4.22a). They are amplified by convection as they propagate eastward at a slow speed (~3 m s\(^{-1}\)). During the mature stage of the MJO, the \(T'\) and \(Z'\) signals and its associated circulation evolve into the canonical MJO pattern (Fig. 4.23b). This well documented structure propagates eastward across the Maritime
Continent into the western Pacific at the known MJO speed of about 5 m s\(^{-1}\). After MJO convection ceases, the \(T'\) and \(Z'\) signals continue their eastward propagation over the central and eastern Pacific with again a slower speed close to 3 m s\(^{-1}\) (Fig. 4.29 c). The signals over the Indian Ocean prior to convective initiation of the MJO we presented in this study are consistent with previous studies (Matthews 2007; Straub 2013; Ling et al. 2013). Those over the central and eastern Pacific need verification using different data and methods. If confirmed, they might bear intriguing implications.

Figure 4.29: Schematic evolution of MJO anomalies in wind (arrows), temperature, geopotential height (colors: positive or \(H\) in green and negative or \(L\) in blue), and deep convection (red ellipses) based on Figs. 4.11, 4.12, 4.17, 4.18 and 4.22. The long lasting positive geopotential height anomalies are highlighted in yellow.
The stark contrast of the temperature and geopotential height anomalies between MJO and non-MJO signals in Manus soundings suggests that they are not simply atmospheric responses to local convective heating. They must be part of an MJO structure that is independent of local convection. Our diagnostics of the global reanalysis data confirms this. The slowly propagating signals over the central and eastern Pacific may be interpreted in at least two different ways: They are remnants of the previously existing MJO event, or they represent an intrinsic structure of the MJO that continues its eastward journey with or without MJO convection.

If the slowly propagating $T'$ and $Z'$ signals over the central and eastern Pacific (e.g. Fig. 4.23 k) are simply remnants of previous MJO events, there should be evidence of similar signals in model simulations forced by MJO-like heating sources. Hayashi and Miyahara (1987), using a three dimensional linear model, found a Rossby-Kelvin type of response (Gill 1980) propagating eastward only when their model had an intraseasonal forcing at all tropical longitudes (their Fig.3a). Their model did not produce eastward propagating signals when the forcing was constrained to certain regions (their Fig. 3b). They used a damping rate of 20 days which is consistent with the e-folding rates of the $T'$ and $Z'$ signals we discuss, ~15 to 20 days. Other studies (Hayashi and Golder 1987, 1988) reported a quick dissipation of the dynamical signals after the cessation of MJO-like heating perturbations in a primitive-equations global climate model. Hendon and Salby (1996), using a linear primitive model and assuming damping rates of 10 days, showed that only the ‘Kelvin’ type of response continues propagating eastward after a MJO heating source was ceased. In accordance with Hendon and Salby (1996), Yamagata and Hayashi (1984) reported only a Kelvin type of response eastward propagating in a
two dimensional linear model days forced by an intraseasonal heating source using
damping rates of 2.2 days. Salby and Garcia (1987) analyzed the response to a
longitudinally localized random heating source on a linear primitive model using a more
realistic damping rate profile (an exponential function of height), with values close to 10
days in the troposphere. Consistent with previous results, they found a dominantly
eastward propagating response with Kelvin wave characteristics.

Based on the reports of the studies discussed above, we have no reason to expect
the $T'$ and $Z'$ signals presented in this study are remnants of previous MJO events. Our
results suggest that there is a dry structure intrinsic to the MJO. A possible test to
disprove such conclusion would be seeking for signals similar to the $T'$ and $Z'$ we present
in a dry, nonlinear, three-dimensional, global model forced by observed diabatic heating
of the MJO (Zhang et al. 2010; Jiang et al. 2011; Ling and Zhang 2011; Zhang and Ling
2012) using different damping rates. It would also remain to be seen if there are slow and
dry MJO signals in global models capable of reproducing the MJO (Zhang and Mu 2005;
Miura et al. 2007; Thayer-Calder and Randall 2009; Maloney 2009; Zhang and Song

A mathematical expression for a dry MJO wave is unknown, but if it exists, it
might be a solution to a nonlinear and three-dimensional system. This is why it is not
included in the solution to a linear shallow water system (Matsuno 1966). Previous
attempts of deriving a mathematical expression of the MJO without convective heating
have reached only limited success (Van Tuyl 1987; Zou and Cho 2000; Wedi and
Smolarkiewicz 2010). Consistently with these studies, the spatial characteristics of the
slowly propagating $T'$ and $Z'$ signals (e.g. Fig. 4.23 k) suggest the hypothesized dry MJO
structure has nonlinear Rossby wave dynamics at its core (Boyd 1980; Wedi and Smolarkiewicz 2010).

Analogous to convectively coupled equatorial waves (Kiladis et al. 2009), the typical MJO signals can be a consequence of interaction between convective heating and a dry dynamical structure. If there is a dry dynamical structure of the MJO as suggested by our diagnosis, it should exist in all global primitive equation models, and it should be detectable in global models that are capable of reproducing the MJO (Zhang and Mu 2005; Miura et al. 2007; Thayer-Calder and Randall 2009; Maloney 2009; Zhang and Song 2009; DelGenio et al. 2012). Most numerical models fail to produce the MJO no more mysteriously than they fail to reproduce equatorial Rossby, Kelvin and other types of waves (Huang et al. 2013). Their deficient parameterization schemes are fully responsible for such failure (Wang and Schlesinger, 1999; Maloney and Hartmann 2001). Recent attempts of using a regional model to simulate the MJO have shown that deficient parameterization schemes not only fail to nourish the MJO signals, but they may act to destroy them (Ray et al. 2009; Ulate et al. 2014 and 2015). If convective heating serves as an effective energy source to a pre-existing dry dynamical structure of the MJO, it may also serve as an effective mechanism to drain energy from it, depending on the relative spatial structures and timing of the two.

Our results by no means imply convection and moisture is unimportant to the MJO. The sensitivity of tropical convection to its ambient moisture (Tokioka et al. 1988; Kiladis et al. 2005; Benedict and Randal 2009; Sobel and Maloney 2012) is no doubt a crucial piece of the MJO puzzle. Our results add an alternative possibility to most existing theoretical explorations of the MJO that vitally depend on convective heating. If
such a dry dynamical structure of the MJO indeed exists, then many seemingly competing ideas of MJO mechanisms would all complement each other. It would no longer need to be debated whether the MJO is driven by convection-circulation interaction (Lau and Peng 1987; Hendon 1988), cloud-radiation interaction (Hu and Randall 1994, 1995; Raymond 2001), air-sea coupling (Neelin et al. 1997; Zhang and McPhaden 2000; Hendon 2000), lateral influences from the extratropics (Lau and Peng 1987; Hsu et al. 1990; Knudson and Weickmann 1987; Ray et al. 2009; Ray and Zhang 2010), upstream influences from circumnavigating remnants of previous MJO events (Knudson et al. 1986; Wang and Li 1994; Ray and Li 2013), scale interaction (Slingo et al. 2003; Moncrieff 2004; Majda and Biello 2004; Peatman et al. 2014), stochastic processes (Neelin and Yu 1994; Yu and Neelin 1994; Thual et al. 2014), or other possible mechanisms. They all serve as sources of energy to amplify and sustain the dynamic structure of the MJO, but with different efficiencies. The key question would then be why these virtually perpetual processes only energize the MJO very occasionally. This question, unfortunately, cannot be fully addressed before we know precisely the structure of the dry the MJO as suggested in our results.
CHAPTER 5: On the Role of Shallow Convection and Surface Zonal Circulation

5.1 Background

Previous studies have shown that sufficient shallow convective heating can produce a robust steady state response of surface and low-level winds in idealized models (Wu 2003; Zhang and Hagos 2009). Surface equatorial winds over the tropical Atlantic in models are notorious for their westerly biases and their association to Amazonian rainfall during boreal spring. This study investigates the hypothesis that insufficient shallow convective heating over the Amazonia is a cause for surface weak equatorial surface easterlies or westerly biases over the equatorial Atlantic in AGCMs. This idea attempts to refine previous findings suggesting that westerly biases are originated in the misrepresentation of moist processes over the Amazonia (Biasuti et al. 2006; Chang et al. 2007; Richter and Xie 2008; Richter et al. 2011).

The tropical Atlantic atmosphere and ocean host a large variety of important and intriguing weather and climate phenomena. They include the inter-tropical convergence zone (ITCZ), trade winds, easterly waves, hurricanes, the equatorial branches of the meridional overturning circulations in both the ocean and atmosphere, an equatorial cold tongue, and zonal and meridional modes of the tropical Atlantic variability (Xie and Carton 2004). This region is influenced by many remote factors. They include two major centers of atmospheric deep convection over the Amazonia (Wang and Fu 2007) and West Africa (Ruiz-Barradas et al. 2003), the North Atlantic decadal variability (Xie and Tanimoto 1998), and teleconnections with the Pacific Ocean (Aceituno 1988; Chiang et al. 2002; Klein et al. 1999; Enfield and Mayer 1997). Given this complexity, it is of little
surprise that most state-of-the-art global climate models (GCMs) suffer from misrepresentations of variability mechanisms (Breugem et al. 2006) and persistent systematic biases in this region (Davey et al. 2002; Deser et al. 2006; Richter and Xie 2008).

Besides the westerly bias (DeWitt et al. 2005), other notable AGCM biases over the equatorial Atlantic during boreal spring include a weak east-to-west equatorial gradient in sea-level pressure (SLP) (Chang et al. 2007; Richter and Xie 2008), and an underestimation of low-level clouds (also known as stratocumulus cloud decks) off the African coast and over the southeast Atlantic (Huang et al. 2007; Hu et al. 2011), a southward shift of the ITZC (Deser et al. 2006), and deficient (excessive) rainfall over the Amazonia (West Africa) (Richter and Xie 2008). These problems have shown an enigmatic persistency through decades of considerable model improvements in other aspects.

Westerly biases in most atmospheric-oceanic GCMs of Coupled Model Intercomparison Project phase 3 (CMIP3, Meehl et al. 2005) (Fig. 5.1b) are more severe in all seasons than those in their atmospheric-only counterparts of Atmospheric Model Intercomparison Project phase 2 (AMIP2, Glecker et al. 1996) (Fig. 5.1a). In both coupled and uncoupled GCMs, maximum westerly biases occur during March, April, and May (MAM). In the latest generation of AGCMs from CMIP5 (Figs. 5.1c), these westerly biases are only modestly reduced from those in CMIP3 (Fig. 5.1a). In coupled GCMs warm oceanic biases in the eastern equatorial Atlantic (Seo et al. 2006; Breugem et al. 2008; Patricola et al. 2012) can be a source for SLP biases, but not in AGCMs. The
westerly biases in AGCMs are generally attributed to the SLP biases (Biasuti et al. 2006; Chang et al. 2007; Richter and Xie 2008; Richter and Xie 2011).

Figure 5.1: Monthly mean biases in surface zonal wind ($u'$) over the equatorial Atlantic (2S to 2N, 40W to 10W) in (a) AMIP2 models, (b) CMIP3 models and (c) AMIP models of CMIP5. The ICOADS data were used to estimate the biases.

An erroneous zonal gradient in SLP along the Atlantic equator in AGCMs can be introduced by deficient rainfall over the Amazonia (Chang et al. 2008; Wahl et al. 2010) or excessive rainfall over West Africa (Richter et al. 2012) during boreal spring, and by the influence of the monsoon in the Gulf of Guinea during the spring-summer transition (Okumura and Xie 2004). Chang et al. (2008) showed that a reduction in Amazonian rainfall deficit in a model can lead to a better simulation of the zonal gradient in SLP and
thus a reduction in the westerly bias. A more accurate simulation of the Amazonian rainfall may come from better representations of convection (Betts and Jakob 2002), land processes and land-atmosphere interactions (Richter et al. 2012), or a better representation of the remote forcing from the Pacific (Tozuka et al. 2011). We hypothesize that systematically insufficient shallow convection over the Amazonia would result in the biases SLP biases along the Atlantic equator.

The method of this study is presented in section 5.2. The results, and a summary and discussion are presented in sections 5.3 and 5.4 respectively.

5.2 Methods

The methods of this chapter are divided in two subsections. First, the connection between sea level pressure and westerly biases is diagnosed using a simple boundary layer model (Stevens et al. 2002) (subsection 5.2.1). Second, the connection between biases in equatorial SLP, zonal wind, and diabatic heating over the Amazonia from AGCM outputs was diagnosed by means of a tercile approach from a multi model ensemble since reanalysis products were found unreliable to evaluate model diabatic heating errors over the Amazonia (subsection 5.2.2).

5.2.1 Idealized Zonal Surface Winds

Stevens et al. (2002) demonstrated that mean surface winds over the tropical oceans can be studied using a simple well-mixed boundary-layer model. The zonal momentum equation for this mixed-layer model at the equator is

\[-\alpha \frac{\partial p}{\partial x} - U \frac{|V| C_d}{h} + (U_T - U) \frac{\omega_E}{h} = 0 \tag{5.1}\]
where $U$ is the zonal component of the mixed-layer bulk wind vector, $V$, which represents surface wind, $U_T$ the zonal component of the wind vector immediately atop the mixed layer, $\omega_E$ the entrainment velocity across the top of the mixed-layer, $C_d$ the surface drag coefficient, $h$ the depth of the mixed layer and $\alpha$ the specific volume. This equation describes the momentum balance of equatorial boundary-layer winds by the pressure gradient force (first term), surface drag or friction (second), and entrainment (third).

For simplicity, ignoring the meridional component and assuming a mean easterly wind we obtain from Eq. (5.1)

$$U = \frac{1}{2C_d} \left\{ \omega_E \pm [\omega_E^2 + 4C_d(-\alpha \frac{\partial p}{\partial x} h - U_T \omega_E)]^{1/2} \right\}$$

(5.2)

A scale analysis using $\omega_E = 1.2 \times 10^{-2} \text{ m s}^{-2}$ (McGauley et al. 2004), $U_T = 10 \text{ m s}^{-1}$, $h = 1000 \text{ m}$, $-a \frac{\partial p}{\partial x} = 2 \times 10^{-5} \text{ m s}^{-2}$ (see section 4.2), and $C_d = 1/900$ (Garrat 1992) indicates that the terms with $\omega_E$ in (5.2) cannot be neglected in comparison to the pressure gradient term on the right hand side of (5.2). The surface westerly bias can therefore come from errors in either the pressure gradient force (Chang et al. 2008; Richter and Xie 2008), or the momentum entrainment, or both.

Two factors determine the SLP gradient force over the equatorial oceans. The first is an isolated diabatic-heating source in the troposphere (Gill 1980), or more precisely, its vertical gradient (DeMaria 1982; Hartmann et al. 1984; Wu et al. 2000; Wu et al. 2003; Schumacher et al. 2004). The second is the SST gradient (Lindzen and Nigam 1987). Biases in the SLP gradient force can be introduced via biases in SST in coupled simulations, but not in AMIP simulations; Observed monthly SST is prescribed in AMIP
simulations. We therefore focus on two possible root causes for the westerly bias in AGCMs: erroneous vertical profiles of diabatic heating over the Amazonia (sections 5.3.1 and 5.3.2) and erroneous entrainment of momentum over the equatorial Atlantic Ocean (section 5.3.3).

5.2.2 Heating Profiles and the Multi-model Tercile Approach

The apparent heating source, $Q_1$ (Yanai et al. 1973) was estimated to represent the vertical structure of diabatic heating in the AMIP2 models and ERA-Interim (ERAI, Dee et al. 2011) reanalysis. $Q_1$ was also estimated from radiosonde data of the Large-Scale Biosphere Atmosphere experiment (LBA, Silva Diaz et al. 2002). Direct outputs of diabatic heating profiles from the Climate Forecast System Reanalysis (CFSR, Saha et al. 2010) and the Modern Era Retrospective-Analysis for Research and Applications (MERRA, Rienecker et al. 2011) were also used. $Q_1$ was computed from (5.3) using centered and forward difference schemes for horizontal and vertical derivatives, respectively,

$$\frac{Q_1}{c_p} = \frac{\partial T}{\partial t} + V \cdot \nabla T + \omega \left( \frac{\partial T}{\partial p} + \frac{\alpha}{c_p} \right)$$  (5.3)

where $T$ is mean temperature, $\omega$ vertical velocity, $V$ the horizontal velocity vector, $c_p$ heat capacity at constant pressure and $\alpha$ specific volume. All data were interpolated to a horizontal grid of 1°. There are 14 vertical levels (10 m and 1000, 925, 850, 700, 600, 500, 400, 300, 250, 200, 150, 100 and 70 hPa) in all model simulations. Table 1 lists the temporal resolutions and periods covered by the used data.
Two main domains were analyzed. One is over the equatorial Atlantic Ocean (as in Richter and Xie 2008), and the other over the Amazonia (Fig. 5.2). Surface zonal winds and momentum entrainment were averaged over the equatorial Atlantic domain. Due to missing data in observations (ICOADS), the zonal SLP gradient was estimated as the slope of a least squared line fitting all data points along the equatorial Atlantic domain. Diabatic heating profiles were averaged over the Amazonian domain. The total number of MAM samples (540) in the AMIP2 simulations was used as the degree of freedom in correlation significance tests.

Figure 5.2: Analysis domains for $Q_1$ profiles (over the Amazonia) and zonal surface wind (over the equatorial Atlantic Ocean), the location of LBA (star), mean AMIP2 surface wind biases against the ICOADS data (arrows, magnitude $> 0.5$ m s$^{-1}$), and mean continental precipitation ($> 6$ mm day$^{-1}$) from observations (shades) and the AMIP2 simulations (contours) during MAM.
The first issue to address is whether profiles of diabatic heating from the three reanalysis products, either as their direct output of temperature tendency terms (MERRA, CFSR) or estimated as $Q_1$ (ERAI), could be used as surrogates for unavailable observations to estimate biases in $Q_1$ in the AGCM simulations. To evaluate the reliability of reanalysis heating profiles, we compared them to the mean LBA heating profile (Fig. 5.3). CFSR correctly captures the maximum amplitude (only slightly lower than the observed) of the LBA profile, but it overestimates its strength at low levels (800-500 hPa). The ERAI profile has the smallest root-mean-square error, but its heating peak is weaker than the observed by 20% and it also overestimates its amplitude at low levels. The MERRA profile differs the most from the LBA profile; its amplitude is the smallest and it does not have any clear peak. In addition, none of the reanalyses reproduced the spatial pattern of the observed rainfall, and all of them underestimated the total rainfall amount at the LBA location and over the Amazonia (Fig. 5.4). This simple comparison suggests that errors in the three reanalyses not only come from incorrect vertical structure of diabatic heating over the LBA region, they also come from incorrect spatial distribution of rainfall. It might be safer to assume that their heating profiles over the Amazonia during boreal spring are not reliable either. Therefore, reanalysis heating profiles were not used as surrogates of observations.
Figure 5.3: Mean profiles of diabatic heating from LBA observations and reanalyses at the LBA location, averaged over the LBA period of November 1998 – February 1999. The root-mean-squared errors (RMSE) of the reanalyses are given.

Figure 5.4: Continental precipitation from CMAP observations and reanalyses averaged over the LBA period of November 1998 – February 1999. The root-mean-squared errors (RMSE) of the reanalyses are given.

Without a reliable reference of heating profiles, we took a different approach to diagnose connections between diabatic heating in the Amazonia and the westerly bias. We calculated the errors in surface zonal wind ($u'$, hereafter primes denote biases) with respect to ICOADS for each month of simulation in each of the eight AMIP2 models.
(Table 2.3). We treated each model simulation as a member of an ensemble (hereafter AMIP2 ensemble), which includes the simulations from the eight models. The probability distribution of $u'$ (Fig. 5.5a) demonstrates that westerly biases exist in more than 75% of all months in the AMIP2 ensemble during MAM. All samples of $u'$ were grouped into three equal parts (terciles) based on their amplitude. The top tercile (to the right of the right vertical dashed line in Fig. 5.5a) represents the portion of the AMIP2 ensemble with large $u'$ or westerly biases, and the bottom tercile (to the left of the left vertical dashed line) the portion with small or no westerly biases or even easterly biases. The corresponding distribution of monthly errors in the zonal SLP gradient force ($-a\partial p'/\partial x$, hereafter $P'$) is shown in Fig. 5.5b. Positive $P'$ represents eastward SLP gradient force corresponding to the surface westerly bias. Possible connections of biases in surface zonal wind $u'$ and zonal SLP gradient force $P'$ with diabatic heating profiles in the Amazonia can be diagnosed by comparing the top and bottom terciles of $u'$ and $P'$.

Figure 5.5: Probability distributions of biases in (a) surface zonal wind ($u'$) and (b) zonal SLP gradient force ($P'$) over the equatorial Atlantic in the AMIP2 ensemble during MAM. Dashed lines separate the total population into terciles.
5.3 Results

5.3.1 Multi-Model Ensemble

The first result from this ensemble approach is given in Fig. 5.6. The panels show the difference between the top and bottom terciles of $u'$ (left column) and $P'$ (right) in surface zonal wind, surface wind vectors, precipitation, and SLP. The tercile difference (hereafter denoted by $\delta$) is the average over the top tercile subtracted by the average over the bottom tercile. Larger westerlies in the top terciles (positive $\delta u'$ in Figs. 5.6a and e) are expected by our difference definition. Positive $\delta P'$ with higher pressure over the Amazonia and lower pressure over the eastern Atlantic (Figs. 5.6c and g) is consistent with positive $\delta u'$. The difference in the zonal gradient in SLP is about 80 Pa from 10 to 40º W (Figs. 6d and e), which is consistent with its analogous using observations in Richter and Xie (2008). The corresponding force is $2 \times 10^{-5}$ m s$^{-2}$. If the zonal gradient in SLP bias is related to insufficient rainfall over the Amazonia as previously suggested (Richter and Xie 2008), one would expect a precipitation deficit there in the top tercile. There is no obvious deficit in precipitation over the Amazonia (Figs. 5.6b and f). Instead, excessive precipitation is found over the Brazilian Nordeste and tropical Atlantic Ocean, which is possibly related to a characteristic southward drift or expansion bias of the Atlantic ITCZ in models (Biasutti et al. 2006; Deser et al. 2006). Tercile differences in SST, consistent with the notion that SST is not responsible for the westerly bias in AGCMs, are very small and do not show any identifiable pattern (not shown). The similar patterns in the tercile differences in $u'$ and $P'$ further confirms that surface westerlies and the zonal SLP gradient force biases are closely connected.
We now present an explanation for $u'$ and $P'$ in terms of $Q_1$ over the Amazonia. Fig. 5.7 shows the vertical profiles of $Q_1$ over the Amazonia (Fig. 5.2) averaged in the top and bottom $u'$ terciles. In each month of March, April, and May, and during MAM, the amplitude of lower-tropospheric $Q_1$ in the top tercile (larger westerly biases) is smaller than that in the bottom tercile. This result holds even if the profiles in each tercile are randomly selected (Fig. 5.8). The probability distributions of $Q_1$ at 850 - 700 hPa levels in the top and bottom terciles of $u'$ substantially differ (at the 95% confidence level according to the Kolmogorov-Smirnov test), with more months of weak low-level $Q_1$ in the top tercile (larger westerly biases) than in the bottom tercile (Fig. 5.9a). These differences are not all due to a deficit in the total rainfall amount in the bottom tercile.
(Figs. 5.6b and f); the ratios of the vertically integrated $Q_1$ (proportional to the total rainfall amount) between the two terciles are 0.85 – 0.96.

Figure 5.7: Mean $Q_1$ profiles over the Amazonia in the top and bottom terciles of surface zonal wind biases ($u'$) from the AMIP2 ensemble. The vertically integrated $Q_1$ ratio (QR) between both profiles (top over bottom) is given.

Figure 5.8: As in Fig. 5.7 except that randomly selected profiles are shown in gray.
The linear correlation coefficients between $u'$ and $Q_1$ reach a negative peak (-0.36) at 850 hPa (Fig. 5.10a) and a positive peak (0.12) at 300 hPa. This is consistent with Fig. 5.7 in that larger westerly biases are associated with weaker low-level $Q_1$. If larger westerly biases were caused by an underestimation of precipitation alone regardless of the vertical structure of $Q_1$, months of larger westerly bias (larger $u'$) would be drier months. No correlation between $u'$ over the equatorial Atlantic and rainfall over the Amazonia was found\(^2\). Similar analyses showed no evident connection between $u'$, and $Q_1$ over equatorial West Africa.

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\(^2\)Richter and Xie (2008) did not calculate the correlation between Amazonian rainfall and westerly biases in the models.
Figure 5.10: Correlation coefficients (black line) and their 95% confidence level limits (shaded) between standardized $Q_1$ over the Amazonia and biases in (a) surface zonal wind ($u'$) and (b) zonal SLP gradient force ($P'$) from the AMIP2 ensemble during MAM.

The plausible physical connections between $u'$ and $Q_1$ in the Amazonia are further supported by their connection to $P'$. Consistent with Fig. 5.9a, the probability distributions of $Q_1$ at 850 - 700 hPa levels substantially differ (at the 95% confidence level) between the top and bottom $P'$ terciles, with more samples of weak low-level $Q_1$ in the top tercile (larger biases of positive $P'$) than in the bottom tercile (Fig. 5.9b). The correlation between $P'$ and $Q_1$ (Fig. 5.10b) also reaches a negative peak at 850 hPa, suggesting that positive $P'$ is related to insufficient $Q_1$ at this level. Figs. 5.10a and b indicate that biases in surface westerly and zonal SLP gradient force are both related to insufficient lower-tropospheric $Q_1$. The correlation between $u'$ and $P'$ is 0.41 (significant at the 95% confidence level).

To provide a broader perspective of the connection between the westerly bias over the equatorial Atlantic and $Q_1$ over the Amazonia we examined the $u'$ tercile differences in tropospheric zonal wind ($\delta u$) and those in $Q_1$ ($\delta Q_1$) through the entire year (Fig. 5.11). Excessive low-level westerlies in the top tercile compared to the bottom tercile ($\delta u > 0$)
exist through the entire year, but become stronger and deeper during boreal spring. Maximum westerlies are located at 850 hPa during MAM. During June and July they extend up to the 200 hPa level but with a smaller amplitude than during MAM. After July they gradually become weaker and shallower. $\delta Q_1$ and $\delta u$ roughly correspond to each other. In the lower troposphere, excessive westerlies ($\delta u > 0$) in the top tercile are generally accompanied by deficit in $Q_1$ ($\delta Q_1 < 0$), and excessive easterlies in the upper troposphere ($\delta u < 0$) by excessive $Q_1$ ($\delta Q_1 > 0$). The $Q_1$ deficit in low-levels begins after December and continues through the early months of the year. During MAM, negative $\delta Q_1$ at 850 hPa roughly matches the maximum in westerly $\delta u$ at that level. In May, the deficit in $\delta Q_1$ reaches its maximum near 600 hPa and becomes shallower and weaker afterwards, as correspondingly do the excessive westerlies. In the upper troposphere, there are excessive easterlies with much larger amplitude and two peaks: one from February to May, and the other from August to October, both at 150 hPa. Upper-level excessive easterlies appear to be associated with excessive upper-level $Q_1$. Their peaks occur roughly in the same seasons. The upper-level excessive $Q_1$ alone, however large, is obviously not the reason for the excessive surface westerlies. Excessive low-level westerlies and upper-level easterlies indicate a weakening of the Atlantic zonal circulation in the top tercile. This is possibly a combined result of insufficient $Q_1$ in the lower troposphere and excessive $Q_1$ in the upper troposphere. The correspondence between $\delta u$ and $\delta Q_1$ is, however, not perfect. For example, there is an apparent lag between the peaks of low-level $\delta Q_1$ and lower-tropospheric $\delta u$. Other factors for the westerly bias must also exist.
5.3.2 Individual Models

Fig. 5.12 shows the difference in $Q_1$ over the Amazonia between the top and bottom $u^*$ terciles in each individual AMIP2 model during MAM. Although $\delta Q_1$ shows different vertical structures, all models suffer from weaker 850 – 700 hPa $Q_1$ in the top tercile compared to the bottom tercile. Deficient $Q_1$ exists at all levels in some models (Fig. 5.12b), while only at low levels in others (Fig. 5.12b). We now explore if the mean vertical $Q_1$ can explain the degree of the westerly biases among the models. Fig. 5.13a shows a scatter diagram of mean low-level $Q_1$ and mean westerly biases during MAM. The symbol size is proportional to the vertically integrated heating over the Amazonia. The severity of the westerly biases among these models does not depend on the vertically integrated $Q_1$, and apparently neither on the mean low-level $Q_1$. The later relationship, however, can be better analyzed by comparing models with similar vertically integrated $Q_1$ (Fig. 5.14). The two models with the largest vertically integrated $Q_1$ (1.8 - 1.9°C day$^{-1}$, Fig. 5.14a) share similar vertical structures, however the one with slightly stronger low-
level $Q_1$ (open star) suffers from slightly less severe westerly biases (Fig. 5.13a). The two models with intermediate vertically integrated $Q_1$ (1.2 - 1.3º C day$^{-1}$, Fig. 5.14b) best exemplify the possible effect of low-level $Q_1$ on the westerly biases. The one with weaker lower low-level $Q_1$ (solid diamond) suffers from more severe westerly biases (Fig. 5.13a) than the one with stronger low-level $Q_1$ (solid triangle). There is no relationship between low-level $Q_1$ and the westerly biases among the four models with relatively weak vertically integrated $Q_1$ (0.9 - 1.1º C day$^{-1}$, Fig. 5.14c). Other factors must be in play. One of them is explained in the next section.

![Figure 5.12](image_url)

**Figure 5.12:** Differences in $Q_1$ over the Amazonia between the top and bottom terciles in surface zonal wind biases ($u'$) from individual AMIP2 models with deficient $Q_1$ (a) at all levels and (b) only at low levels during MAM.
Figure 5.13: Scatter diagram between (a) low-level (850-700 hPa) $Q_1$ over the Amazonia and (b) entrainment over the equatorial Atlantic ($E$) against mean surface zonal wind biases ($u'$) over the equatorial Atlantic from each AMIP2 model during MAM. In a, the symbol size is proportional to the vertically (1000 to 100 hPa) averaged $Q_1$ for each model. In b, biases in zonal SLP gradient force ($P'$, ms$^{-2} \times 10^5$) are contoured.

Figure 5.14: Mean $Q_1$ profile over the Amazonia for each AMIP2 model with its vertically averaged $Q_1$ in the range of (a) 1.8 - 1.9º C day$^{-1}$, (b) 1.2 - 1.3º C day$^{-1}$, and (c) 0.9 - 1.1º C day$^{-1}$ during MAM.
5.3.3 Momentum Entrainment

As discussed in section 5.2.1, entrainment is an essential component of the momentum balance in the boundary layer over the tropical oceans. Its mistreatment in AGCMs may potentially lead to biases in surface wind. In this section, we examine the possible relation between erroneous entrainment and the westerly bias over the equatorial Atlantic in addition to biases in the zonal SLP gradient force in the AMIP2 ensemble.

Calculating momentum entrainment in GCM simulations requires the entrainment velocity ($\omega_E$), wind throughout the boundary layer, and the boundary-layer depth, $h$. They are not available from standard model output. We estimated the entrainment term ($E$) in the AMIP2 models and observations over the tropical Atlantic domain (Fig. 5.2) as a residual of the zonal momentum Eq. (5.1):

$$E = D - P$$  \hspace{1cm} (5.4)

where $D=U(|V|C_d/h)$ is surface friction and $P=-a\partial p/\partial x$. We assumed the mean boundary layer wind ($U$) as $u_s$, and a boundary layer height ($h$) of 1000m. Mean $E$ in observations is negative (Fig. 5.15a), indicating easterly momentum is mixed downward from the lower troposphere into the mixed layer and helps to maintain or enhance the surface easterlies. During MAM when the westerly bias is the largest, mean $E$ in the AMIP2 ensemble is negative but weaker than in observations. The amplitude of $E$ during these months is larger than that in the zonal SLP gradient force $P$ (Fig. 5.15b), suggesting a dominant role of $E$. During the other months, mean $E$ in the AMIP2 ensemble is positive, but its amplitude is much smaller than that of $P$, suggesting that the role of $E$ is not
dominant. In observations, however, the amplitude of $E$ is larger than that of $P$ all the time.

![Figure 5.15: Monthly (a) entrainment ($E$) and (b) the ratio of $E$ and zonal SLP gradient force ($P$) over the equatorial Atlantic from the AMIP2 ensemble and ICOADS data.](image)

Larger $u'$ is mainly due to larger, positive $P'$ (Fig. 5.13b). The possible effect of entrainment on the westerly bias $u'$, however, must be evaluated with respect to a constant pressure gradient force bias $P'$. Among models with similar $P'$ (e. g., between 0.8-1.2 x10$^5$ ms$^{-2}$ in Fig. 5.13b), models with weaker negative entrainment $E$ suffer from larger westerly biases $u'$ than those with stronger negative $E$. This partially explains the independence of $u'$ from low-level $Q_1$ in Fig. 5.13a, (e. g., open square and shaded dot). This result suggests that insufficient easterly momentum could lead to westerly biases.

The connection between monthly westerly biases $u'$ and errors in entrainment ($E' = D' – P'$) is examined through their correlation in bins of similar $P'$ from the AMIP2 ensemble (Fig. 5.16). The correlations between $u'$ and $E'$ are positive in all $P'$ bins. This positive correlation indicates that, at fixed $P'$, westerly biases tend to become more
severe \( (u' > 0) \) when models suffer from westerly entrainment biases \( (E' > 0) \), namely, insufficient easterly entrainment.

Figure 5.16: Correlation coefficients (dots) and their limits of the 95% confidence level (vertical bars) between surface zonal wind biases \( (u') \) and entrainment biases \( (E') \) in a given bin of zonal SLP gradient force \( (P') \) from the AMIP2 ensemble during MAM. The sample size is given for each bin as its percentage of the total (540 months), their sum accounts for 95% of the total, bins with sample sizes less than 2% of the total (on the tails of the \( P' \) distribution) were excluded.

5.4 Summary and Conclusions

We have analyzed the connection between boreal spring Amazonian diabatic hating structures and surface westerly biases over the equatorial Atlantic from eight AGCM outputs. We used a simple a model for a well-mixed boundary layer over the tropical oceans to diagnose their connection. We found that the connection between low-level diabatic heating and surface westerly biases critically depends on the boundary layer momentum entrainment: the westerly biases tend to be large when the lower-tropospheric (850-700 hPa) diabatic heating over Amazonia and boundary layer easterly momentum over the equatorial Atlantic is weak. Our results suggest that insufficient shallow convective heating over the Amazonia and vertical momentum mixing across the top of
the boundary layer over the equatorial Atlantic are two possible root causes of the equatorial Atlantic westerly bias in AGCMs.

Our results along with previous studies suggest that a westerly bias in an AGCM would occur when: 1) Amazonian rainfall is underestimated, 2) low-level diabatic heating in the Amazonia is too weak, even if the total rainfall amount is well reproduced, and 3) the vertical mixing of zonal momentum across the top of the equatorial Atlantic boundary layer is too weak (in the presence of low-tropospheric easterlies), even if the total rainfall amount and vertical structure of diabatic heating in the Amazonia are well reproduced. There can certainly be other causes. One of those could be the bias in diabatic heating over the tropical Atlantic related to the erroneous southward shift of the ITCZ. We tried to identify a relation between the ITCZ bias and the westerly bias in the AMIP2 models we diagnosed but found none.

Further diagnoses and modeling work, as well as reliable observations of diabatic heating profiles over the Amazonia and boundary-layer entrainment over the equatorial Atlantic Ocean are needed to confirm our diagnosis. We have conducted numerical simulations using a regional mesoscale atmospheric model covering the tropical Atlantic, South America and Africa. When the diabatic heating profiles over the Amazonia were modified to be weaker in the lower troposphere, the model produced more severe westerly bias over the equatorial Atlantic. This result is expected. It is consistent to previous studies that examined the sensitivity of the vertical structure of the atmospheric large-scale circulation to vertical heating profiles (Hartmann et al. 1984; Wu et al. 2000; Wu et al. 2003; Schumacher et al. 2004; Li et al. 2008; Zhang and Hagos 2009). Our experiments showed insensitivity of westerly bias to the diabatic heating profiles and
precipitation amounts over equatorial West Africa, consistent to Chang et al. (2008). It would be interesting to investigate the sensitivity of the westerly bias to boundary layer parameterization schemes with different efficiencies of vertical momentum mixing. It would also be interesting to estimate momentum entrainment bias in other regions, if this is a common model bias, it should not be constrained to the equatorial Atlantic region.

The implication of our results is that there might be no simple or single remedy for the westerly bias in GCMs. This may be why this problem has been so stubborn and persistent up to the new generation of CMIP5 models. Efforts of advancing models in a holistic way must continue before this problem is completely solved.
CHAPTER 6: Concluding Remarks

This study investigated the role of shallow clouds in the variability of the deep tropics with the aim to better understand the importance of their moistening and heating effects with respect to that of other processes in these regions. This work addressed three problems of tropical meteorology: the environmental controls and feedbacks of shallow clouds in tropical regimes of deep convection (Chapter 3), the role of these clouds in the MJO (Chapter 4, see Zermeño-Díaz et al. 2015, and Zermeño-Díaz and Zhang 2015), and the possible connection of equatorial Atlantic surface circulation and Amazonian shallow convection (Chapter 5, see Zermeño-Díaz and Zhang 2013).

Our results show that the variability of shallow clouds in the deep tropics shares one important characteristic with subtropical shallow clouds. Low-frequency (~monthly) anomalies in occurrence and depth of shallow clouds in the deep tropics are related to lower tropospheric stability parameters (Klein and Hartman 1993), but only during certain conditions (regimes of negative anomalies of LTS). These clouds interact with their environment through their heating and moistening effects. The moistening and heating effects of shallow clouds are found to be not negligible but they lack coherent timings with those in low-level humidity, temperature, or wind. They appear to have a background role, at least in relatively slow phenomena. These results are particularly revealing in the context of the MJO. Low-level moisture increases during the MJO at Manus cannot be attributed to shallow cloud moistening. Although shallow cloud moistening is an important source of low tropospheric moisture, our results suggest indicate it is not critical to the MJO. Even more, this study presents evidence of the
existence of a MJO structure that propagates without any evident connection to moist processes. Such structure propagates eastward at an ultra slow speed (3 m s\(^{-1}\)). Its shape, resembling a Rossby soliton (Boyd 1980), suggests nonlinear dynamics might be essential to it. If this structure is at the heart of the MJO, the results from this study suggest that moist processes, including shallow convection, are important sources of energy to the shape the typical structure of the MJO (e.g. Kiladis et al. 2005; Zhang and Ling 2012), but they might not be crucial for its existence and propagation. The importance of shallow convection in the tropical variability and circulation seems to be case-dependent and subject to the interaction with other factors, as exemplified by the westerly biases over the equatorial Atlantic. Finally, the results from this study highlight the importance of clouds and boundary layer processes not only in the equatorial Pacific Ocean and Atlantic regions, but in general in the variability of the tropics.
CHAPTER 7: Significance

The results from this study are important in several contexts. From a methodological point of view, this study presents a new method to estimate bulk shallow cloud moistening by condensation. This method has certainly several weaknesses, but it has the advantage of estimating moistening from field observations. This study is also the first to apply a simplified boundary layer model to diagnose wind biases in GCMs. From a scientific point of view, the results from this study indicate that shallow clouds provide a non-negligible background moistening to the lower troposphere. This result is important to understanding the role of these clouds in relatively slow phenomena, such as the MJO. Also in the context of the MJO, this study reveals new and structurally different aspects of the MJO after more than 40 years of research in this topic. These results defy current ideas of the nature of the MJO as a mode that essentially depend on moisture anomalies. In the context of the tropical Atlantic variability, results from this imply that the adequate representation of rainfall over the Amazonia in standard GCM simulations needs to be associated to an adequate representation of the vertical structure of diabatic heating over this region in order to improve the westerly biases over the equatorial Atlantic. This study is the first to show evidence of a possible misconnection between the free troposphere and the boundary layer processes in GCMs associated to erroneous momentum entrainment.
Appendix A: Attenuation Threshold

Cloud profiling radars are suitable for detecting the bases and tops of hydrometeor layers in the atmosphere (Clothiaux et al. 2000). However, heavy rain can cause severe attenuation of radar signals and thus result in underestimation of true echo top heights. Here, echo top heights of the Ka-band (35 GHz) Zenith pointing Radar (KAZR), deployed on Addu Atoll (0° 42’S, 73° 9’ E) during the DYNAMO (Dynamics of the Madden-Julian Oscillation) field campaign (October 2011- January 2012, Yoneyama et al 2013) are compared to a merged reflectivity data set (Feng et al. 2014) based on simultaneous observations of the KAZR and two precipitation radars, the National Center for Atmospheric Research S-band dual polarimetric (S-Pol) radar, and Shared Mobile Atmospheric Research and Teaching Radar (SMART-R) of the Texas A&M University (Table 1). This merged data will hereafter be referred to as K-SPOL-SR (Table 1). The KAZR is more accurate than the precipitation radars to detect cloud echo tops under non-precipitating conditions due to its higher sensitivity to small cloud particles. Under precipitating conditions, attenuated reflectivity from the KAZR is replaced by those from the S-POL and SMART-R. The merged data are treated here as a benchmark without rainfall attenuation. Figure A.1a shows median differences between echo top heights from K-SPOL-SR and the KAZR alone as a function of the surface rain rate and the KAZR echo top height. Positive (negative) differences correspond to underestimated (overestimated) echo top heights by the KAZR in comparison to those by the K-SPOL-SR. The area to the left of the dashed line marks the regime where shallow KAZR echo tops are not likely to be attenuated by rain. A relatively large fraction of KAZR observations falls into this regime (Fig. A.1b). KAZR echo top heights lower than 4 km
height with rainfall rates larger than 25 mm hr\(^{-1}\) are likely to be underestimated by more than 1 km due to attenuation. When rain rates are less than 25 mm hr\(^{-1}\), KAZR echo tops lower than the freezing level (~4.5 km) are unlikely to suffer from attenuation even though echo tops at higher level may. A 25 mm hr\(^{-1}\) threshold was used to flag KAZR echo tops lower than the freezing level as attenuated. Using this threshold to distinguish the real and the spurious shallow echoes in KAZR had 75% accuracy (75% of 1-minute profiles were correctly identified). The 25% error is mainly caused by real shallow echoes that were classified as spurious ones. Higher thresholds (e.g., 30 mm hr\(^{-1}\)) showed slightly higher accuracy (0.78) but produced misclassification of spurious shallow echoes as true shallow echoes. The MMCR reflectivity data suffer from the same attenuation problem. At the end, a conservative approach is used and all MMCR shallow echoes (top height < 4 km) at Manus Island with surface rain rates exceeding the 25 mm hr\(^{-1}\) threshold were excluded.
Figure A.1: (a) Median differences in echo top heights between the K-SPOL-SR and KAZR at Gan during DYNAMO and (b) KAZR sample size as functions of the surface rain rate. Dashed lines in (a) at 25 mm hr$^{-1}$ and 4 km height are discussed in the text.
Appendix B: Estimation of Free-Tropospheric LWC of Shallow Clouds

The cloud liquid water content (LWC) above the boundary layer was estimated by assuming a increase linearly with height, \( z \) (Han et al. 1994). This approach was applied to both non-precipitating and precipitating shallow clouds. Following Wood and Taylor (2001):

\[
\frac{\Delta \text{LWC}}{\Delta z} = az
\]  

(A.1)

where \( a \) is a constant. The vertical integral of LWC from the surface to the cloud-top height, \( H \), is roughly the liquid water path, LWP:

\[
LWP \approx \int_0^H \text{LWC}(z)dz = 0.5aH^2
\]  

(A.2)

Solving (A.1) for \( a \) and substituting it into (A.2) yields:

\[
\frac{\Delta \text{LWC}}{\Delta z} = (2LWP/H^2)z
\]  

(A.3)

The vertically-integrated free tropospheric LWC (\(<\text{LWC}>\)) of a cloud can be obtained by integrating (A.3) from the top of the boundary layer \((h_b)\) to the top of the cloud \((H)\):

\[
<\text{LWC}> = \frac{LWP}{H^2}(H-h)^2
\]  

(A.4)
Appendix C: Event Identification Method

MJO and non-MJO large-scale convective events were identified through two steps.

Step 1: Identification of large-scale events over Manus Island. This was done using anomalies of daily rainfall from TRMM. After applying pentad and 10° longitudinal running means to the rainfall data, an area-average time series was generated over the Manus domain (142.375°E- 152.375°E, 7.125°S - 3.125°N) (Fig. C.1). A convective event was identified when the time series exceeded one standard deviation for at least 3 consecutive days. The day of the rainfall peak was defined as “day 0”. Positive rainfall anomalies before or after day 0 were considered as part of this convective event. A total of 37 events were detected during the extended boreal winter (October-April) of 2001- 2013 (Table 4.1).

Step 2: classification of MJO and non-MJO events. This was done using two criteria, one is based on a rainfall tracking method (Ling et al. 2013), the other on the Wheeler and Hendon (2004) RMM index. The tracking method was used to determine the propagation characteristics of the selected convective events. The smoothed daily TRMM rainfall anomalies from step 1 were further averaged over 12 ° S and 8° N (Fig. C.1) and used to generate a longitude-time diagram for each event. On a given diagram, possible directions and speeds of zonal propagation (either eastward or westward) are represented by a set of straight lines (hereafter tracking lines) passing through the longitude of Manus at day 0. The range of eastward propagation speeds is 1 - 14 m s⁻¹, with an interval of 0.1 m s⁻¹. This broad range is to cover all possible MJO speeds. Westward tracked speeds cover the same range but include 0 m s⁻¹ for stationary events.
Rainfall anomalies were integrated along each of the tracking lines starting from day 0, at the longitude of Manus: backward in time towards the west along eastward tracking lines for possible eastward propagation, and towards the east along westward tracking lines for possible westward propagation. Integrations along the tracking lines stopped when rainfall anomalies become lower than one standard deviation. There is one exception: if rainfall anomalies are lower than one standard deviation over the Maritime Continent (between 90° and 145° E, Fig. C.1) and become greater than one standard deviation...
deviation again over the Indian Ocean (west of 90° E), the integration would continue and include both positive and negative anomalies. This exception was applied to include MJO events that weaken over the Maritime Continent before they reach the Western Pacific (Matthews 2008). Finally, the integrated rainfall anomaly of each eastward tracking line was averaged using the number of points where rainfall anomalies are larger than one standard deviation (all points were used for westward trajectories). The direction and mean speed of a given event were assigned by the tracking line with the largest averaged rainfall anomalies.

A convective event was classified as an MJO if it satisfies one of the following two conditions; otherwise, it was classified as a non-MJO event.

**Condition 1 (C1, or tracking condition):** The western end point of the eastward tracking line is west of 90˚E and rainfall anomalies greater than one standard deviation occupy at least half of the tracking line.

**Condition 2 (C2, or RMM condition).** The phase of the RMM index at day 0 is 4, 5, 6, or 7, and its amplitude is larger than 1 in three consecutive preceding phases.

Figure C.2 presents three examples. In the first example (Figs C.2 a and d) both tracking and RMM conditions are satisfied. Even though its positive rainfall anomalies over MC are moderate, the speed of this event is 6.4 m s⁻¹. This case is evidently a strong MJO event. The second example (Figs C.2 b and e) only satisfies the tracking condition (C1). There are negative rainfall anomalies over MC along the tracking line, but there is robust eastward propagation of 5.3 m s⁻¹ of positive rainfall anomalies that starts from the IO and continues to the east of Manus. The amplitude of the RMM index is greater than 1
in phase 5 only, between the west end point (blue dot) and day 0 (red dot) on the track. The RMM condition (C2) is not satisfied. This is a weak MJO event. The third example (Figs C.2c and f) is a non-MJO event. Its strong positive rainfall anomalies are associated with a westward propagating perturbation. Day 0 occurs when the RMM amplitude is less than 1 in phase 1. Its large amplitude in phases 4 - 7 represents a previous MJO event.

Figure C.2: Upper panels: longitude-time diagrams of daily rainfall anomalies. Lower panels: the RMM phase diagrams of three large-scale convective events over Manus Island. Contours in the longitude-time diagrams mark positive and negative one standard deviations. Tracking lines are orange. The RMM diagrams cover days -25 to 10. Red dots correspond to day 0 at Manus and blue dots the end points of the tracks.
Table A.1 presents the frequency of occurrence of the events we identified with respect of each other. Most (86%) identified MJO events are primary as defined by Matthews (2004), meaning they are not immediately (within 40 days) preceded by another MJO event, and are not involved with successive MJO events, meaning they are not immediately followed by another MJO event. There is a small chance (5%) that they are immediately preceded by non-MJO events, and moderate chance (22%) that they are immediately followed by non-MJO events. Overall, 18% of MJO events are immediately preceded by another large-scale convective event, MJO or non-MJO, and 36% are immediately followed by it. The identified non-MJO events are more likely to be immediately preceded by either MJO or non-MJO events (47%) than followed by them (27%). In combination, there is an equal chance (roughly 30%) for a large-scale convective event over Manus, MJO or non-MJO, to be preceded or followed by another large-scale convective event. This statistics suggest that there is a tendency for large-scale convective events to follow each other, but MJO events rarely emerge from such consecutive events.

Table C.1 Number of MJO and non-MJO events (from Table 2) preceded (followed) by other MJO and non-MJO events within 40 days before (after) the rainfall peak at Manus Island. Note that some of the preceding and following MJO and non-MJO events are outside the analysis period. So the number of non-MJO events following MJO events may not be the same as the number of MJO events preceding non-MJO event, for example.

<table>
<thead>
<tr>
<th></th>
<th>MJO (22)</th>
<th>Non-MJO (15)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Preceded by MJO</td>
<td>3</td>
<td>4</td>
</tr>
<tr>
<td>Preceded by non-MJO</td>
<td>1</td>
<td>3</td>
</tr>
<tr>
<td>Preceded by no event</td>
<td>18</td>
<td>8</td>
</tr>
<tr>
<td>Followed by MJO</td>
<td>3</td>
<td>2</td>
</tr>
<tr>
<td>Followed by non-MJO</td>
<td>5</td>
<td>2</td>
</tr>
<tr>
<td>Followed by no event</td>
<td>14</td>
<td>11</td>
</tr>
</tbody>
</table>
MJO events selected by our procedure, in general, start over the Indian Ocean approximately 15 to 20 days before its peak over Manus. Rainfall peaks over Manus mostly occur in RMM phase 5, but also in phases 4, 6 and 7. A total of 22 MJO and 15 non-MJO events were identified (Table 2). From the 22 MJO events, 12 events satisfy both conditions. These are generally stronger than the other 10. The RMM condition detects strong MJO events but misses the relatively weak ones. The MJO events that satisfied the RMM (tracking) condition also satisfy the tracking (RMM) condition, but none event satisfied the RMM condition alone. In other words, the precipitation condition identified weak MJO events that the RMM condition missed.

Gottschalck et al. (2013) also reported a case of MJO event that the RMM index fails to identify. The advantage of the tracking method is not only identifying both strong and weak MJO events, but providing quantitative measures of the strength and propagation speed of convective signals for all identified MJO events. We analyzed separately strong and weak MJO events but found no qualitative differences, for these reason, all MJO events were treated equal. We also used a slightly different threshold that requires positive rainfall anomalies to last at least 5, instead of 3, consecutive days. As a result, 18 MJO events and 5 non-MJO events were selected. This indicates that the local convective periods at Manus are about 20 days or longer for most MJO events, but shorter for most non-MJO events. To maintain a sufficient sample size for non-MJO statistics, we decided to use the threshold of 3 consecutive days. We inspected each individual non-MJO event to see if it was associated with a tropical cyclone. We found none. Of the 22 MJO events, 8 might be “successive” (Matthews 2008) with a preceding MJO event within 60 days. This information bears no significance to local structures and
evolution of MJO events at Manus, but it may be relevant to large-scale structures and evolution of the MJO over the Indian Ocean.

In general, the RMM phase corresponds linearly to the local rainfall evolution (Fig C.3). This relation has also been shown using DYNAMO radar rainfall estimates (Xu and Rutledge 2014). Figure C.3a shows that MJO rainfall start over the IO (phases 1 and 2), approximately 15 to 20 days before its peak over Manus. Rainfall peaks over Manus mostly occur in phase 5, but also in phases 4, 6 and 7. Similar results are obtained using other indexes solely based on outgoing longwave radiation (OLR) anomalies (Fig. C.3b), such as the OLR MJO index (Kiladis et al. 2014). The inability of using the RMM or OMI indexes alone to identify times or locations of MJO rainfall peaks is another reason for using the tracking method in this study.

Figure C.3: Joint probability density distributions of phases of the Real Multivariate MJO (RMM) index and the OLR MJO index (OMI) (repeated once) against the lag to the rainfall peak of MJO MJO events at Manus. Only phases with amplitude larger than 1 were considered.
Appendix D: MMCR Modes and ARSCL ZC Correction

The Millimeter Cloud Radar (MMCR, Kollias et al. 2007) and the Ka Band Zenith pointing Radar (KAZR) are both 35-GHz vertically pointing radars. MMCR radars were replaced by the newer KAZR in the ARM sites. The dynamic range of both radars is about 7 orders of magnitude, from -50 to 20 dBZ (Kollias et al. 2007). The sensitivity of each mode is different so it targets different groups of clouds (Moran et al. 1998). MMCR operated in 4 modes:

Mode 1 or boundary layer mode: it has phase coding and has a 90 m range resolution.

Mode 2 or cirrus mode: it has phase coding and has a 45 m range resolution.

Mode 3 or general mode: it has no phase coding and has a 45 m range resolution.

Mode 4 or robust mode: it has no phase coding and has a 90 m range resolution.

KAZR operates in 2 modes (Bharadwaj and Chandrasekar 2012), approximately comparable with MMCR modes 2 and 3. The Active Remotely-Sensed Cloud Locations (ARSCL, Clotheaux et al. 2000) is value added product that merges the profiling modes of each radar (4 in MMCR, 2 in KAZR), filters significant atmospheric echoes (cloud boundaries), de-aliases radar Doppler velocities, and removes insect noise and brag scattering using a combination of multiple active remote-sensing instruments (cloud radar, lidar) of the ARM facility. ARSCL products are available for Manus during the period of operation of the MMCR (~1998-2011).
Figure E.1a shows the equivalent reflectivity (Ze) frequency distribution by height (also known as CFAD) for the MMCR observational period at Manus (Table 2.1) from ARSCL. Ze from mode 4 is manifested as a low tropospheric (below the 4 km height) local maximum at 40 dBZ, but its values can be larger than 50 dBZ (red box in Fig. E.1a). Mode-4 extends the dynamic range of the MMCR to 10 orders of magnitude producing a discontinuity in Ze profiles (Fig E.2a). To avoid this discontinuity, we excluded mode-4 Ze from the ARSCL reflectivity. Mode-4 Ze was included in the ARSCL based on this procedure: “if the mode 4 velocity estimate does exceed the mode-3 Nyquist velocity and the mode 4 data signal-to-noise ratio is greater than 10 dB, we replace the mode 3 data at the grid point by the mode 4 data” (Clothiaux et al. 2000). We applied the opposite of this procedure to insert back Ze from mode 3 in place of mode 4. We modified the conditions as follows. Mode-3 Ze was used instead of that of mode 4 when $2 \text{ m s}^{-1}$ was exceeded and mode-3 Ze was greater than 0 dBZ. The range of these conditions is wider compared to the original to avoid discontinuities at the boundaries were mode-4 was originally. Fig E.1b shows the resulting frequency distribution by height of Ze. Despite these efforts, remnants of mode-4 are still present but they are much less than before (Fig E.2). This procedure does not affect the spatial distribution of the atmospheric echoes detected originally by ARSCL.
Figure D.1: Frequency distributions by height of Ze from (a) ARSCL and (b) ARSCL-123 (see text for details) over Manus Island.

Figure D.2: Example of reflectivity from (a) ARSC and (b) ARSCL-123. See text for details.
Appendix E: Sensitivity to Detect Precipitation in Clouds

In this appendix we explore the sensitivity of the classification of shallow clouds as either precipitating or non-precipitating to the instruments available, and the conditions we used to detect rainfall (Mather et al. 2007). A given shallow cloud was classified as precipitating if one of the following three conditions was met in any minute (see subsection 3.2.2): i) reflectivity greater than 0 dBZ below 4 km, ii) gauge rain rate greater than 0.1 mm hr\(^{-1}\) (the lowest reliable value of the instrument), iii) wet MWR window. Otherwise the shallow cloud was classified as non-precipitating. The occurrence of precipitating and non-precipitating clouds is virtually the same if rainfall is detected using only the first the condition (Fig. E.1), and they are qualitatively the same if the threshold of 0 dBZ is varied (Fig. E.2). On the other hand, the occurrence can vary substantially when rainfall is detected using only the optical rain gauge (Fig. E.3), and when different minimum thresholds are used. The sensitivity to the gauge threshold was found particularly large from 2001 to 2005 (not shown). This discrepancy suggests different instrument sensitivities during the analyzed period. However, the rainfall detection from the gauge and microwave radiometer is not dominant factor (Fig. E.1).
Figure E.1 Time series of occurrence (min) in (a) non-precipitating (NP) and (b) precipitating (P) shallow clouds (SC). Precipitation was detected using the (black) standard condition and only a reflectivity threshold of 0 dBZ below 4 km. See text for more details.

Figure E.2 same as Fig. E.1 but using a reflectivity threshold of -15 dBZ.
Figure E.3 same as Fig. E.1 but precipitation was detected using a threshold in gauge rainfall of 0.1 mm hr$^{-1}$. 
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