Deep Convective Systems Observed by A-Train in the Tropical Indo-Pacific Region Affected by the MJO

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ABSTRACT

In the Indo-Pacific region, mesoscale convective systems (MCSs) occur in a pattern consistent with the eastward propagation of the large-scale convective envelope of the Madden–Julian oscillation (MJO). MCSs are major contributors to the total precipitation. Over the open ocean they tend to be merged or connected systems, while over the Maritime Continent area they tend to be separated or discrete. Over all regions affected by the MJO, connected systems increase in frequency during the active phase of the MJO. Characteristics of each type of MCS (separated or connected) do not vary much over MJO-affected regions. However, separated and connected MCSs differ in structure from each other. Connected MCSs have a larger size and produce less but colder-topped anvil cloud. For both connected and separated MCSs, larger systems tend to have colder cloud tops and less warmer-topped anvil cloud. The maximum height of MCS precipitating cores varies only slightly, and the variation is related to sea surface temperature. Enhanced large-scale convection, greater frequency of occurrence of connected MCSs, and increased midtroposphere moisture coincide, regardless of the region, season, or large-scale conditions (such as the concurrent phase of the MJO), suggesting that the coexistence of these phenomena is likely the nature of deep convection in this region. The increase of midtroposphere moisture observed in all convective regimes during large-scale convectively active phases suggests that the source of midtroposphere moisture is not local or instantaneous and that the accumulation of midtroposphere moisture over MJO-affected regions needs to be better understood.

1. Introduction

Deep convective systems are a prominent feature of tropical convection, precipitation, and upper-level clouds. These cloud systems play important roles in modulating diabatic heating and affecting the general circulation (Houze 1982; Hartmann et al. 1984; Schumacher et al. 2004; etc.). Deep convective systems are also an important source of upper-level ice clouds and moisture. The upper-level water vapor and cirriform clouds are crucial factors for better understanding climate feedback. The deepness of convection and the efficiency of deep convection to produce upper-level clouds need to be better understood for their potential impacts on cloud feedback (Hartmann et al. 2001; Lindzen et al. 2001). Deep convective cloud systems are thus key factors in understanding global weather and climate. The objective of this study is to document specifically the population, structure, and organization of deep convective systems associated with the Madden–Julian oscillation (MJO; Madden and Julian 1971) to help understand the fundamental nature of tropical deep convection, to determine how deep convection varies with tropical environmental conditions, and to assess its role in the MJO.

The MJO dominates the variability of tropical convection in the Indo-Pacific region at the intraseasonal time scale (20–100 days). Since the 1970s, much effort has been expended to explore MJO characteristics, structure, initiation, propagation, and its important effects on weather and climate, including tropical cyclone activity, monsoons, El Niño–Southern Oscillation, and extratropical weather [see reviews by Madden and Julian (1994), Lau and Waliser (2005), and Zhang (2005)]. While its effects on a broad spectrum of weather events and climate variability have been recognized and extensively documented, most current general circulation models (GCMs) have difficulty reproducing even the gross features of the MJO (Lin et al. 2006; Hartmann and
Hendon 2007). For GCMs capable of reproducing some salient features of the MJO, the precipitation is generally too weak and the observed coherence between precipitation and low-level zonal wind associated with the MJO in general is not reproduced (Zhang et al. 2006).

A variety of factors are candidates for improving the MJO simulation in GCMs, cumulus parameterization, model mean background state, air–sea interaction, and model resolution [see review by Zhang (2005)]. Among these factors, deficiencies in the cumulus parameterization in GCMs have long been considered a primary limiting factor to simulating the MJO. GCMs that reproduce some (but not all) observed MJO features have shown great sensitivity to treatments affecting cumulus convection, which include the entrainment rate (Tokioka et al. 1988), types of closure (Slingo et al. 1996), treatments of convection that directly affect the lower- and midtroposphere moisture conditions (Wang and Schlesinger 1999; Maloney and Hartmann 2001; Zhang and Mu 2005; etc.), the model vertical resolution that affects the representation of moistening effect of midlevel clouds (Inness et al. 2001), vertical heating profiles (Park et al. 1990; Zhang and Mu 2005; Li et al. 2009), and the use of explicit cloud-resolving schemes capable of simulating mesoscale structure of convection (e.g., Grabowski 2001; Miura et al. 2007; Khairoutdinov et al. 2008; Benedict and Randall 2009). The role of cumulus parameterization in simulating the MJO in climate models points to the need for better understanding of the nature and variability of deep convective systems in the MJO.

Gross convective features and large-scale thermodynamic and dynamic structures of the MJO have been documented since the 1970s via field observations, satellite data, and reanalysis fields (Madden and Julian 1994; Zhang 2005). Owing to their large spatiotemporal coverage, satellite observations of radiation, clouds, and precipitation have become an important avenue to explore this intraseasonal large-scale phenomenon. In the past few decades, the use of domain-mean values or patterns of tropical precipitation and cloudiness has been the main way to represent large-scale tropical convection in both observations and model diagnosis. However, studies based on infrared satellite images have shown that deep cloud clusters exhibit rich multiscale temporal and spatial variability within the MJO (e.g., Nakazawa 1988; Lau et al. 1991; Sui and Lau 1992; Hendon and Liebmann 1994; Chen et al. 1996).

The use of mean cloud patterns in satellite-based studies obscures the fact that convection occurs in cloud clusters and exhibits different forms of organization, characterized by horizontal size, deepness, and locations relative to other systems. Such structural differences in deep convective systems have been shown to be associated with different physical characteristics and environmental conditions of convective systems; in particular, clustering of MCSs is favored by oceanic conditions and tends to produce more thick anvils and thus more “stratiform” dynamic and microphysical processes (Yuan and Houze 2010, hereafter YH10; Yuan et al. 2011, hereafter YHH11). Larger fractions of stratiform rain are found in precipitation features with larger horizontal extent in both South Asian premonsoon and monsoon (Romatschke and Houze 2011a,b). The size of precipitation features might be related to some “critical” values of atmosphere column water vapor path as suggested by Peters et al. (2009). These physical characteristics are crucial for understanding precipitation processes and determine the diabatic heating structure, which affects structures of large-scale tropical circulation (Houze 1982, 1989; Hartmann et al. 1984; Raymond 1994, 1995; Schumacher et al. 2004; etc.). A recent study also suggested that domain-mean properties of cloud and precipitation could not effectively reflect variability in physical processes of convection during the MJO (Tromeur and Ros sow 2010).

Tropical convective systems with wide ranges of sizes have long been observed based on precipitation and cloud measurements from radars, microwave imagers, and infrared and visible sensors (Lopez 1977; Houze and Cheng 1977; Cheng and Houze 1979; Williams and Houze 1987; Mapes and Houze 1993; Mohr and Zipser 1996; Cheng et al. 1996; YH10; and others). Early theoretical and modeling studies also suggested that tropical deep convection has the tendency to self-aggregate, with new deep convection favored in the vicinity of past convection (Held et al. 1993; Mapes 1993; Tompkins 2001; Bretherton et al. 2005). A “moisture–stratiform instability,” which relates the midtroposphere moisture deficit to the second baroclinic mode of convective heating (i.e., stratiform precipitation or congestus clouds) has been suggested as being important for determining the large-scale wave growth at wavelengths close to MJO scale (Kuang 2008). Haertel et al. (2008) demonstrated the sensitivity of the MJO to the relative amounts of deep, medium, and shallow convection in the cloud population. Despite all of these studies, mechanisms determining the organization and upscaling of deep convection remain poorly understood and are considered to be a fundamental factor limiting our ability to reproduce MJO features in GCMs (Zhang 2005). Hence, better understanding of the variability of the population and structure of deep convective systems should provide insight into the processes that lead to large-scale intraseasonal variability of precipitation and clouds and fuel future research for better understanding the moist convection. That is the goal of this study.
Infrared images from geostationary satellites provide excellent temporal resolution for cloud-cluster tracking but with limited quantitative information on physical properties of precipitation and clouds because the infrared channel used (usually 11 \micro\textmu m) quickly saturates when deep clouds are present. On the other hand, present-day lower-orbit satellites carrying multiple passive and active remote sensing instruments provide collocated comprehensive and more physically based estimates of precipitation and cloud properties, which construct a suitable set of data to comprehensively study deep convective systems (Mohr and Zipser 1996; Nesbitt et al. 2000; Zipser et al. 2006; Romatschke and Houze 2010; YH10; YHH11). The trade-off is the relative lack of spatiotemporal coverage of the cloud systems, making the life-cycle tracking of each individual system impossible. In YH10 deep cloud systems were objectively identified by combining CloudSat Cloud Profiling Radar (CPR) data with co-orbiting satellites of the A-Train constellation (Stephens et al. 2002; L’Ecuyer and Jiang 2010). Specifically, YH10 used data from the Advanced Microwave Scanning Radiometer for Earth Observing System (EOS) (AMSR-E) and the Moderate Resolution Imaging Spectroradiometer (MODIS; King et al. 1996) on board the National Aeronautics and Space Administration (NASA)’s \textit{Aqua} satellite to identify precipitating cores and nonraining anvil clouds of deep cloud systems so that anvil clouds could be separated from their parent MCSs for study with CloudSat. This A-Train-based database of MCSs allows analysis of detailed properties/structures of deep convective systems when combined with other A-Train measurements toward better understanding of processes related to deep convection. In this study, we use cloud systems identified in YH10 to investigate properties and organization of deep convective systems in relation to the MJO.

Data and methodology used in this study are introduced in section 2. Occurrence of deep convective systems in Indo-Pacific region during MJO is examined in section 3. The variability of the characteristics of mesoscale convective systems and whether these characteristics are influenced by the MJO are the focus of section 4. Moisture field associated with mesoscale convective systems is analyzed in section 5, with conclusions and summary in section 6.

2. Data and methodology

a. Data sources and technique for identifying MCSs

This study is based on four data sources: MODIS cloud product (MYD006_L2; Platnick et al. 2003), AMSR-E instantaneous precipitation retrievals (AE_Rain), AMSR-E-retrieved daily sea surface temperature (SST), and the European Centre for Medium-Range Weather Forecasts (ECMWF) Interim Re-Analysis (ERA-Interim). We use the same method to define categories of deep cloud systems as in YH10. Our technique combines information from MODIS and AMSR-E since neither sensor alone provides a satisfactory description of an MCS, which is defined in terms of the size and coldness of the cloud top together with the size, intensity, and substructure of the rainfall of a convective entity. We use the brightness temperature of MODIS channel 31 of 10.8 \micro\textmu m (Tb11) to determine the cloud-top properties, and the instantaneous precipitation retrieval from the AE_Rain product (Kummerow et al. 2001; Wilheit et al. 2003; Kummerow and Ferraro 2007) to determine the rainfall properties. Tb11 is used to identify the cloud tops exceeding certain degrees of coldness and breadth. Criteria for determining whether the cloud top and rainfall meet criteria associated with an MCS are discussed by YH10 and are based on the statistical distributions of the satellite-observed quantities and descriptions of MCSs in the literature. The features of high-cloud systems identified by these criteria are listed in Table 1, which is adapted from YH10. To identify an observed cloud system as an MCS, we first use Tb11 to identify specific features of cloud-top temperature and rainfall separately. Using the terminology of the table, high-cloud complex (HCC) is first identified as a region of cold cloud, which is then subdivided into one or more high-cloud systems (HCSs), which are cold centers within the HCC. Then we use AMSR-E to locate raining areas [precipitation features (PFs) in Table 1]. If a PF coincides with an HCS, it constitutes the raining core (RC) of the HCS. Within the RC we identify each heavy-rain area or HRA. From these features, we determine that an HCS is an MCS if it satisfies the four criteria listed in Table 1. These criteria assure that the cloud top is cold enough over a broad-enough area, reaches a sufficiently low temperature (high top), has a big-enough rain area and that the rain is intense enough to be consistent with the generally recognized criteria of MCSs. The MCSs so identified are further categorized according to whether they are either separated MCSs (SMCSs) or connected MCSs (CMCSs). The latter are MCSs connected by a common PF with at least two other MCSs and probably are a result of mergers of MCSs (Williams and Houze 1987; Mapes and Houze 1993).

The separation of SMCSs and CMCSs is based solely on their spatial structure as determined by multiple sensors. The definition is not based on any a priori assumption of physical or dynamic differences. However, previous results based on this classification have consistently
shown that differences between CMCSs and SMCSs likely occur for physical reasons. As shown in YH10 (Fig. 9), and as will be presented later in this paper, these two categories of active MCSs favor different large-scale environments. Both small and large separated MCSs occur ubiquitously in continental convective regions while large separated MCSs are preferred in the maritime continental environment. The connected MCSs occur most frequently over warm oceans, possibly for the reasons discussed in YH10 (e.g., sustainability of the moisture supply). The vertical structures observed by the CloudSat CPR (Figs. 12–13 in YH10) show that continental MCSs have lesser amounts of very thick anvil clouds and more thin anvil clouds, consistent with different microphysical processes occurring over land and ocean. YH10 also showed that CMCSs have lesser amounts of thinner anvil clouds and suggest that their spatial organization (tendency to cluster close together) is likely one reason for this difference. YHH11 further showed that thick anvil clouds close to their precipitating cores in the CMCSs have less spread in the values of radar reflectivity factor in the upper portions of clouds, which might indicate that more cloud particles are from stratiform-type microphysics processes. These previous results are, moreover, generally consistent with previous findings from ground-based cloud-radar measurements (Cetrone and Houze 2009).

Table 1 details more specifically the criteria used to determine each of these features identified in MODIS and AMSR-E data. The HCCs are defined as spatially contiguous pixels with Tb11 < 260 K. YH10 found that the 260-K contour includes almost all anvils with tops above 10 km and thicker than 6 km (verified by CloudSat Cloud Profiling Radar). The information about the extent and intensity of the rain within precipitating portions of each HCS obtained from the AMSR-E rain-rate product showed that active mesoscale convective systems (MCSs) defined in this way are responsible for 56% (YH10) of total tropical rainfall and thus represent the primary latent heat source of different tropical zones.

To understand the dynamic/thermodynamic fields associated with the identified MCSs in this study we use pressure-level daily fields of vertical and horizontal velocities, temperature, and humidity from the ERA-Interim (Dee et al. 2011). In addition, the daily-mean SST, precipitation, column-integrated water vapor path (WVP) derived from AMSR-E is obtained from Remote Sensing Systems, which is a microwave-based retrieval (Wentz and Spencer 1998) providing optimally interpolated (OI; Reynolds and Smith 1994) SST at \( \frac{1}{4} \)° (~25 km) resolution. In this product, the impact of daytime solar heating on SST is estimated and removed using a simple empirical model of diurnal warming (Gentemann et al. 2003) so that both good daytime and nighttime retrievals provide nearly complete global coverage each day. Additional quality control processes remove the contamination of SSTs by rain. Comparisons with in situ buoy measurements show that microwave-retrieved SST has a 0.08°C mean bias and 0.57°C standard deviation (Gentemann et al. 2004). As shown in Wentz and Spencer (1998), based on the same retrieving algorithm, the mean difference between WVP retrieved from the earlier version of microwave imager and quality-controlled radiosonde data is negligible for all weather conditions. Under nonraining conditions the variance of the difference is

<table>
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<tr>
<th>Abbreviation</th>
<th>Name</th>
<th>Definition</th>
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<tr>
<td>HCC</td>
<td>High-cloud complex</td>
<td>Region of MODIS Tb11 contained within a single 260-K isotherm</td>
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<tr>
<td>HCS</td>
<td>High-cloud system</td>
<td>Portion of HCC associated with a particular minimum value of Tb11</td>
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<tr>
<td>PF</td>
<td>Precipitation feature</td>
<td>Region of AMSR-E AE_Rain parameter surrounded by 1 mm h(^{-1}) contour</td>
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<td>RC</td>
<td>Raining core</td>
<td>Portion of any PF overlapping and/or located within an HCS</td>
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<tr>
<td>HRA</td>
<td>Heavy-rain area</td>
<td>Portion of PF greater than 6 mm h(^{-1})</td>
</tr>
<tr>
<td>MCS</td>
<td>Mesoscale convective system</td>
<td>Any HCS whose largest RC satisfies the following criteria:</td>
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<td>1) Exceeds 2000 km(^2) in total area</td>
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<td>2) Accounts for more than 70% of the total area with rain rate</td>
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<td>greater than 1 mm h(^{-1}) inside the HCS</td>
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<td>3) Minimum cloud-top temperature above the RC (indicated by ( T_{b11RC1min} )) is less than 220 K</td>
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<td>4) More than 10% of RC is occupied by HRAs</td>
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<tr>
<td>Separated MCS</td>
<td>The largest RC of the MCS is part of a PF that contains less than three dominant RCs of any MCS.</td>
<td></td>
</tr>
<tr>
<td>Connected MCS</td>
<td>The largest RC of the MCS is part of a PF that contains dominant RCs of at least three MCSs.</td>
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Table 1. Features identified in YH10 by automated algorithms applied to the MODIS Tb11 and AMSR-E AE_Rain data products. The cloud-top minimum temperature \( T_{b11RC1min} \) is defined as the mean Tb11 of the coldest decile of the largest rain core (RC1).
nearly zero while under heavy-rain conditions the root-mean-square of their difference is about 2.5 mm.

b. Gridding of data

As described in YH10, the MODIS Tb11 and the AE_Rain precipitation at their footprint resolutions and the precipitation data were preprocessed to match each MODIS pixel. SST, precipitation, and WVP data from Remote Sensing Systems (http://www.ssmi.com) are provided at 25 km × 25 km resolution; they are re-gridded to 1.5° to match the ERA-Interim data for analysis in section 5. Dynamic and thermodynamic parameters from ERA-Interim are kept at their original resolution as provided (1.5° × 1.5°).

c. MJO phases and the phase composite

The MJO phases are defined using the MJO index of Wheeler and Hendon (2004, hereafter WH04). This index is based on the first two empirical orthogonal functions (EOFs) of the combined fields of near-equatorially-averaged zonal winds at 850 and 200 hPa and satellite-observed outgoing longwave radiation (OLR). Further details are online at http://cawcr.gov.au/staff/mwheeler/maproom/RMM/ as well as in WH04. This index captures variability on the intraseasonal time scale of the MJO and thus can identify the MJO without time filtering. To stratify variability associated with the MJO we further limit our analysis to strong MJOs in boreal winters only (i.e., daily MJO indices with amplitude greater than one during October–April). Our analysis contains about 13–15 MJOs over 4 yr.

The WH04 daily index identifies eight phases of the MJO related to different geographical locations of the large-scale convective centers of the MJO. Our phase composites are calculated by

\[ V_j = \frac{1}{N_j} \sum_{i=1}^{N_j} V_{ij}, \]

where \( V \) is the variable under investigation (SST, precipitation, etc.), \( N_j \) is the total number of days of phase \( j \), and \( V_{ij} \) is the daily mean of \( V \) in the \( i \)th day of phase \( j \).

In a previous study, Roundy et al. (2009) showed that the daily real-time multivariate MJO (RMM) (WH04) index is very noisy and does not always correspond to the MJO, even when its amplitude is greater than one. Hence we also perform a test based on the 5-day running mean of the RMMs. We found our composites based on two methods show almost the same results. Thus we conclude that the day-to-day noise would not significantly change our conclusions based on the large sample composite. Additionally, A-Train satellites fly over the equator only at 0130 and 1330 local time so that our sample cannot resolve the diurnal cycle of deep convection. However, according to this study, we found similar results either using daytime or nighttime data alone. Therefore, we show results based on all measurements combined.

3. Occurrence of deep cloud systems during the MJO over the Indo-Pacific warm-pool region

a. Large-scale features

We start by constructing composites of the major large-scale variables used in this study. The composite large-scale features associated with the thermodynamic conditions (SST), adiabatic processes (500-hPa-level vertical velocity \( \omega_{500} \)), and convection (represented by AMSR-E rainfall) of the eight WH04 MJO phases are shown in Fig. 1. The results show a well-defined MJO life cycle composite over the Indo-Pacific region. Figure 1a shows that large-scale precipitation (convection) starts and becomes organized in phases 2 and 3, as indicated by widespread intense precipitation over the eastern Indian Ocean (EIO). Then the large-scale convective center propagates eastward across the Maritime Continent (MA; phase 4, 5), on to the western Pacific Ocean (WP; phases 6, 7), and finally to the South Pacific convergence zone (SPCZ; phases 8, 1). The variance of precipitation across the eight MJO phases over the MA region is less than that of surrounding oceans. Possible factors contributing to this characteristic are likely: the stronger diurnal cycle in convection in MA region (Houze et al. 1981; Churchill and Houze 1984; Williams and Houze 1987) competes with the MJO for moisture and energy; topography interferes with the low-level moisture convergence believed to be crucial to the MJO (Salby and Hendon 1994; Wang and Li 1994; Zhang and Hendon 1997); insufficient surface evaporation from the thinner effective surface mixed layer over the MA region inhibits organized convection (Maloney and Sobel 2004; Sobel et al. 2008). The large-scale midtroposphere upward motion (Fig. 1b) shows coherent patterns with precipitation in all phases, which is not surprising since the latent heat release and the adiabatic cooling are major terms determining the atmospheric energy balance over deep convective regions. The SST varies systematically in relation to the convection, especially over the EIO, where anomalies reach 2°C. The anomalies over the EIO peak in the restoring and preonset phases (8, 1), decrease when convective activity is maximum (phases 2, 3), reach their minimum in the following convective weakening stage (phases 4, 5), and begin to recover
FIG. 1. Composites of daily mean of (a) AMSR-E sea surface temperature, (b) ERA-Interim vertical pressure velocity at 500 hPa ($\omega_{500}$), and (c) AMSR-E instantaneous rainfall rate in eight phases of the MJO. The composites are based on 4 yr (2007–10) of data.
upward in the suppressed stage (phases 6, 7). The phase difference between SST and precipitation (i.e., the SST anomaly leads the precipitation anomaly) is consistent with previous MJO studies (Stephens et al. 2004; Lau and Wu 2010).

b. Contributions to the clouds and precipitation of deep convection

Since the composite MJO life cycle (Fig. 1) captures the large-scale signals of large-scale deep convection, adiabatic processes, and sea surface conditions over the Indo-Pacific region, we have confidence to further explore the variability and organization of deep convective systems under the same framework. For this purpose, we make use of the categories of MCSs defined by YH10 (Table 1). In subsequent discussions, we will refer to the smallest (largest) 25% of the SMCSs as small (large) SMCSs. Figure 2 shows the spatial distribution of the number concentration of CMCSs, large SMCSs, and small SMCSs. The frequency of each type strongly depends on whether they occur over ocean or over the islands of the MA. CMCSs form mostly over the open ocean, while large SMCSs occur most frequently over the MA. Small SMCSs slightly favor the island region, and their occurrence has less phase dependence over the oceans. These different island versus sea behaviors were found in YH10 to be a characteristic of deep convective systems over the whole tropics. One possible reason is that the “sustainability” of MCSs over warm oceans favors the formation of stratiform precipitation region (Yuter and Houze 1998).

The occurrence of MCSs shown in Fig. 2 is generally consistent with the precipitation patterns shown in Fig. 1. Increased numbers of MCSs accompany enhanced large-scale precipitation over all regions, except that the convective systems take different forms: CMCSs dominate over oceans, while large SMCSs dominate over the MA region. Although the variation in number of small SMCSs is similar to other types of MCSs, they only produce about \( \frac{1}{10} \) as much precipitation as large SMCSs (YH10). That is, higher occurrences of the smallest SMCSs are not connected with increased precipitation.

Figures 1 and 2 both suggest that differences within each pair of the four biphase groups—(8, 1), (2, 3), (4, 5), and (6, 7)—are much smaller than differences between different pairs. So to reduce the complexity of our results we construct composites of these four biphases, and we further limit our regional analysis to be close to 10°S–10°N since in boreal winter most of the MJO convective signal is in this tropical belt (Figs. 1 and 2). We further divide our analysis into four subregions to fully explore the spatial and temporal discrepancies/consistencies of the behavior of deep convective systems in the MJO (Fig. 3). To demonstrate the dependence of MCSs on underlying surface types, the MA subregion is defined to consist mainly of island areas. Figure 4 shows the contributions to cloud cover and total precipitation of deep convective systems in the four subregions. To test the robustness of our results we performed \( t \) tests on the mean of daily-averaged value of each variable for each subregion. The effective numbers of degrees of freedom were estimated following Bretherton et al. (1999) based on temporal autocorrelation coefficients computed using daily-mean time series of each variable.

The fractional coverage (percentage of total area covered) of CMCSs, SMCSs, and HCSs excluding active MCSs (HCSXs) show the same phase variation for all subregions (Fig. 4). They all peak at the local active phases of the MJO and reach their minima in suppressed phases, indicating the overall increase and decrease of the total convective activity, respectively. The total fractional coverage of these three types of high-cloud systems varies greatly (up to 1.5–2.5 times; Fig. 4a4), which suggests that the domain-mean changes in convective impacts on the MJO in each subregion might result from significant changes in the spatial coverage and/or intensity of the largest horizontal scales of convection. Among the four subregions, the EIO (WP) exhibits the largest (smallest) variation [2.5 (1.5) times]. This difference in magnitude of variation is consistent with the SST variation shown in Fig. 1, which also shows that largest/smallest variations in both the pattern and magnitude of SST are observed in EIO/ WP.

Despite the similar temporal variation, there are two notable differences in the nature of the fractional coverage among different regions and different types of deep convective systems:

1) In the MA, CMCSs have the least fractional coverage overall compared to the other three oceanic subregions, consistent with Fig. 2.
2) The relative change of the fractional coverage of CMCSs appears to be the largest among the three types of systems in all subregions. This variation is most pronounced in the EIO, where the fractional coverage of CMCSs varies up to 5 times compared to 2.5 times for SMCSs. A similar behavior can be distinguished in the relative contribution to precipitation (percentage of total precipitation produced). Figures 4b1–b3 show that in terms of their contribution to rainfall, only CMCSs share a marked variation from phase to phase in the relative contribution to precipitation. Comparison of Figs. 4a4 and b4 shows that although these high-cloud systems only cover 20%–40% of the area, they produce 60%–90% of total latent heat in any phase of the MJO and in
FIG. 2. Number concentration of MCSs observed during the eight phases of the MJO. Active MCSs are divided into (a) the smallest 25% of SMCSs, (b) the largest 25% of SMCSs, and (c) CMCSs. The number concentration is defined as number of systems instantaneously observed in a 1000 km $\times$ 1000 km box.
any geographical subregion. This fact indicates the primary role of mesoscale cloud systems in determining the latent heat structure of the MJO-affected regions of the tropics. The major change in the cloud population of the MJO is that the latent heat release could be largely influenced by CMCSs in the active phase of the MJO, especially in the EIO region. This fact has implications for the vertical distribution of latent heating since the CMCSs are likely to have the largest stratiform components and thus more top-heavy heating profiles (Houze 1982, 1989).

4. General characteristics of deep convective systems over the Indo-Pacific region

Section 3b shows that the prevalence of deep convective systems varies systematically in relation to phases of the MJO. Now we explore whether detailed characteristics of the systems themselves also vary from one MJO phase to the next. More specifically, we investigate the change in anvil cloud production and overall deepness of cloud systems from convectively active to suppressed conditions.

a. Anvil cloud production

As defined in YH10, the active MCSs identified by our methodology are considered to be in a relatively mature stage, meaning that they must contain one dominant large raining core (Table 1). The systems not identified by the methodology as active MCSs could be growing or decaying MCSs or systems never becoming MCSs. These other life cycle stages cannot be determined from the twice-daily A-Train data. Our methodology thus allows us to focus specifically on the active, mature MCSs, and to quantify anvil production for those systems. In doing so, we consider two parameters: the size of the system and the ratio $R$ of the area of its largest raining core to its overall area. This ratio is related to the efficiency of a system to produce anvil clouds (the part of a HCS lying outside of its raining cores).

As described in section 3b, we divided our sample into four biphases and four subregions, yielding a total of 16 ensembles of active MCSs. Examination of these ensembles shows that for each type of active MCS (CMCS or SMCS), the size distribution does not have distinct relationships with either the phase of MJO or the sub-regions. However, all ensembles robustly produce a difference in size distributions between different types of MCSs. Figure 5a shows that the CMCSs have relatively more large systems than SMCSs. Similar information is found for $R$; differences between two types of MCSs are much greater than variability within each type of MCS. Figure 5b illustrates that CMCSs consistently generate less anvil cloud for the same size of raining area for all sizes; that is, CMCSs are systematically more efficient at producing precipitation than are SMCSs. The closely located curves based on the largest raining core (RC1) and the total raining area show less than a 10% difference for all sizes, which suggest that the RC1s in fact represent the dominant cores of MCSs. In YH10 an area of 2000 km$^2$ is used as the threshold of size of RC1 to identify MCSs, which sets the lower limit of $R$. As shown in Fig. 5b, it is apparent that for very small systems the magnitude of $R$ is constrained by this threshold limit while for larger systems the much lower value of the threshold limit unlikely regulates the $R$. Because the areal coverage is determined primarily by large systems (one can roughly derive that based on Fig. 5a), the overall areal-weighted $R$ is approximately 0.32 and 0.22 for CMCSs and SMCSs, respectively. Therefore, SMCSs produce on average greater than 67% more anvil cloud for the same size of raining area. One factor resulting in the lower efficiency of CMCSs to produce horizontal extensive anvil clouds is likely the relative spatial locations of this type of MCSs. As defined in YH10, CMCSs share a common precipitation feature (i.e., spatially contiguous precipitating area observed by AMSR-E) among multiple MCSs, which means this type of MCS is clustered together, leaving less space for outer, thinner anvil. CloudSat radar data lead to a similar conclusion (YH10).

b. Deepness of MCSs indicated by the brightness temperature distribution for all Indian Ocean and western Pacific MCSs

The vertical structure of convective clouds is important for understanding the precipitation processes, the moistening effect on the troposphere, and the cloud radiative heating profile. Bulk and internal properties of MCS clouds, including likely microphysical mechanisms, were obtained by YH10 and YHH11 using the merged MCS dataset of the present study along with CloudSat CPR measurements. However, because of the
small sample size of CloudSat CPR the analysis could only be applied to all systems (i.e., not broken down into subsets). The passive cloud remote sensing retrievals only capture the column-mean properties from the tops of clouds. However, the larger sample size and the physically based relationship between clouds and the Tb11 provide valuable information on the variability of the vertical structures of MCSs.

Figure 6 shows the variability of the probability density function (PDF) of the Tb11 patterns of three types...
of high-cloud systems. To compute the PDF we first discretized the Tb11 and the size of MCS into \( Y_T \) and \( X_S \) bins, respectively. Then we sorted one type of system into \( X_S \) groups according to their size. For each group the PDF \( P_{ik} \) of Tb11 of each system’s raining portion or anvil portion is computed as

\[
P_{ik} = \frac{A_{ik}}{S_k},
\]

with \( i = 1, 2, \ldots Y_T \) and \( k = 1, 2, \ldots N_j \).

In (2) \( S_k \) is the area of the raining/anvil portion of the \( k \)th system, \( A_{ik} \) is the area of its raining/anvil portion falling in the \( r \)th bin of Tb11, and \( N_j \) is the total number of systems in the \( j \)th size bin. The mean PDF \( P_{ij} \) of each group can be computed either as

\[
P_{ij} = \frac{\sum_{k=1}^{N} P_{ik} S_k}{\sum_{k=1}^{N} S_k},
\]

or

\[
P_{ij} = \frac{\sum_{k=1}^{N} P_{ik}}{N_j},
\]

with \( i = 1, 2, \ldots Y_T \) and \( j = 1, 2, \ldots X_S \).

Practically, we found that (3a) and (3b) produce very similar results and here we only present results computed by using (3a). Values of \( P_{ij} \) were computed for CMCSs, SMCSs, and HCSXs separately. Because no significant/robust difference was found among different subregions and different MJO phases, we use data from the whole Indo-Pacific region from all phases.

The 10.8-\( \mu \)m channel is an atmospheric window channel, within which gases and aerosols in the atmosphere only lightly absorb radiation without the presence of clouds. Hence, for deep clouds (optically thick with high cloud tops), Tb11 is usually a good index of cloud-top height, especially for deep precipitating clouds. On the other hand, the interpretation of Tb11 of nonprecipitating anvil clouds could easily result from cloud properties other than the cloud-top height (YHH11). Therefore, we separate the cloud tops located over the raining cores of cloud from those of the nonprecipitating anvil portions (left and right columns of Fig. 6).

Figures 6a and 6c shows that the raining cores of active MCSs tend to have a single dominant mode, with a value less than 220 K (equivalent to the emission level of 12–13 km). We take this value to be the most likely cloud-top (emission level) temperature over the raining cores of the type of deep convective system under consideration. Precipitation areas within HCSXs have warmer cloud tops, especially for the smallest systems, whose cloud tops shift to a warmer mode between 255 and 260 K (\(-6.5 \) km). This warmer mode is likely related to precipitating congestus clouds or growing systems not having yet reached maximum height. All Tb11 distributions have strong size dependencies, for which raining
cores of larger active MCSs have colder and narrower Tb11 distributions. This behavior suggests that larger precipitating cores usually penetrate deeper and have less horizontal heterogeneity, which may partly be due to relatively less entrainment of dry air in the case of larger precipitating cores (Lucas et al. 1994a,b) and likely larger stratiform fractions within large systems (Romatschke and Houze 2011a,b). Rain cores of CMCSs appear to have slightly narrower/colder distributions, which might be because CMCSs in general have
larger raining cores, they are clustered together, and they generally have larger stratiform regions.

The anvil clouds in large active MCSs have warmer and broader Tb11 distributions than over the raining cores. According to YH10 and YHH11, we should expect this result because compared to the raining cores anvils have 1) slightly lower cloud tops, 2) overall lower cloud-hydrometeor contents, and/or 3) larger variability in the amount and properties of hydrometeor contents. Factors 1 and 2 result in lower emission levels and factor 3 induces broader distributions. The fewer occurrences of anvil clouds with warmer cloud tops (>240 K) for large CMCSs compared to that of SCMCs is also consistent with the fact that CMCSs have a lesser amount of thinner anvil cloud (section 4a above and YH10). A warmer mode starts to present itself in the Tb11 distribution of anvil clouds as MCS size decreases, and it becomes dominant in smaller systems, suggesting that the latter tend to have a greater proportion of thin anvil cloud. The tendency of larger systems to produce relatively more thick anvil cloud would be consistent with the larger fraction of stratiform rain production in large raining features (Romatschke and Houze 2011a,b). Tb11 distributions of nonprecipitating clouds in HCSXs consistently show that warmer effective cloud tops dominate and little similarity is found in the distribution patterns compared to Fig. 6e. This is likely because HCSXs have precipitating areas occupying less than 7% of their total area and most statistics of their nonprecipitating clouds are from clouds not relevant to these simultaneously existing raining cores.

In addition to information about cloud structures, Fig. 6 provides some useful hints about the link between Tb11 and precipitation. Figure 6 shows that there is almost no specific Tb11 value that could be used to separate precipitating and anvil clouds except for the coldest tops (≤190 K) that capture the deepest part of very large systems. Actually, cloud clusters defined by any single threshold of Tb11 warmer than that will most likely represent a mixture of cold anvils and precipitating areas together, and the relative mixture of different types of clouds depends on the organization of convection (i.e., type and size of systems). This finding indicates that a purely infrared-based precipitation estimate using a universal relationship (such as the global precipitation index; Arkin and Meisner 1987) might produce biases even in a large spatial domain across different large-scale convective scenes (such as the MJO active phase region). Moreover, the relative amounts of different types of clouds (deep raining, thick and thin anvils, etc.) in a cloud cluster depend on the organization and the physical processes of systems associated with each cluster. So representations of deep convective activity by statistics of single sensor data report at grid points (no coherent cloud object organization recognized by comparison with surrounding pixels) or Tb11 cloud clusters alone (no physically based precipitation information) would not be able to depict accurately the behaviors of specific forms of deep convection, such as those shown by Fig. 2.

c. Does deepness of MCSs vary with phase of the MJO?

As shown in Fig. 6, larger active MCSs tend to have colder cloud tops. This result was obtained for all subregions in our analysis and for all phases of the MJO. We wish to know if these temperature properties vary from one phase of the MJO to the next. To answer this question, we first determine if there is a relationship between cloud height and the profile of equivalent potential temperature:

\[ \theta_e = \theta \exp \left( \frac{Lq}{C_p T} \right), \]  

where \[ \theta = T \left( \frac{p_e}{p} \right)^{R/C_p} \]  

where \( \theta, L, q, C_p, T, R, p_e, \) and \( p \) are potential temperature, latent heat of vaporization at 0°C (2.5 × 10^8 J kg⁻¹), specific humidity, specific heat at constant pressure (1004 J kg⁻¹), absolute temperature, dry gas constant (287 J K⁻¹ kg⁻¹), surface pressure, and pressure, respectively.

For this investigation, we look for empirical relations between the Tb_min of MCSs and the intercept temperature \( T_{\text{int}} \) defined as the temperature of the level at which the \( \theta_e \) is equal to the \( \theta_s \) near the surface \( \theta_{\text{sfc}} \). We assume that Tb_min is indicative of the maximum height that undiluted convective parcels could attain (as in the “hot tower” concept of Riehl and Malkus 1958). For MCSs with very large precipitating cores this approximation is probably reasonable. The value of \( T_{\text{int}} \) is computed using the SST and temperature profiles from the ERA-Interim. To estimate the \( \theta_{\text{sfc}} \), we assume that the near-surface relative humidity is 80% universally and then find the level in the upper troposphere where \( \theta_e \) can most closely match the \( \theta_{\text{sfc}} \). Figure 7a shows scatterplots of mean \( T_{\text{int}} \) against the mean Tb_min of large MCSs (size > 10^4.5 km²). Since we use SST data, only the three oceanic subregions are presented in the plot. Of these three regions, only the EIO exhibits large variations of SST (Fig. 1a), and no significant relationship is
found between \( T_{\text{int}} \) and \( T_{\text{bmin}} \) in the northwestern Pacific (NWP) and southwestern Pacific (SWP) where the variation of \( \theta_{\text{e}} \) is relatively small. Larger variations of both the \( T_{\text{int}} \) (\( \sim 7 \) K) and the \( T_{\text{bmin}} \) (\( \sim 5 \) K) occur in the EIO, and they are well correlated with a slope of 1.25. We compared results for near-surface relative humidity of 75% and 85%, and they primarily show similar linear relations except for a systematic decrease in the \( T_{\text{int}} \) with increased humidity. If the relative humidity actually varies significantly temporally and spatially, then the relationship of \( T_{\text{int}} \) and \( T_{\text{bmin}} \) would be noisier; therefore, the results in Fig. 7a must be accepted with caution. Also, a perfectly undiluted parcel probably rarely occurs in real convection. The actual height that a convective parcel can reach usually is largely affected by the entrainment of dry air as well as the environmental lapse rate at upper levels. The use of the \( \theta_{\text{e}} \) difference as a measure of buoyancy also ignores the additional latent heat released when ice particles are produced (Lucas et al. 1994a; Zipser 2003). With these caveats in mind, we note that if the slope of 1.25 (or some similar slope) is real, then Fig. 7a suggests that the intercept height increases more than does the cloud-top height, as the cloud-top height increases (\( T_{\text{bmin}} \) decreases).

Since the maximum height of the MCSs varies systematically with SST, we might expect that if the effective cloud-top height varies with phase of the MJO it would be most apparent in the EIO region, where the SST varies substantially from colder in the active phases to warmer in the suppressed phases in that region (Fig. 1). Figure 7b shows the anomaly of mean \( T_{\text{bmin}} \) of MCSs in EIO. In general the MCS tops are about 5°C colder in the active phase than the suppressed phase. This difference corresponds to about 0.5 km in cloud-top height—only a slight difference. In other words, the deepness of MCSs is not strongly associated with phase of the MJO.

5. Moisture field and mesoscale convection in active and suppressed conditions

In sections 3 and 4, we showed that the occurrence of the all types of deep convective systems increases from suppressed to active phases of the MJO at all subregions of the tropical Indo-Pacific region. We also showed that the most pronounced variability in deep convective systems in the MJO is the shift in the fractional coverage of different types of systems (comparative numbers of CMCSs, SMCSs, HCSXs, etc.), while relatively little variability is found in the statistical structural properties of each type of MCS regardless of the phase of the MJO. In this section we will show that the upscaling of deep convection in the form of MCSs shifting toward the form
of CMCSs is robustly associated with anomalously high midlevel moisture.

a. The general variation of moisture in the MJO

Composites of the anomaly of averaged specific humidity across the Indo-Pacific region for the four biphases of the MJO (Fig. 8a) show the well-known equatorial westward-tilt structures in the MJO (Kiladis et al. 2005; Tian et al. 2006; Haertel et al. 2008; etc.). The most significant change in the specific humidity profile is in the middle troposphere, where the vapor content always peaks where the local convection is enhanced. Moreover, the enhanced convection in the MJO is usually associated with the horizontal convergence in a deep layer from the surface to around 400 hPa (which can be seen in the zonal wind component in Fig. 8b). In general, the westward-tilted moisture field would be consistent with the moisture convergence anomaly in the lower troposphere leading the enhanced deep convection and anomalous upward transport of water vapor. Since the corresponding variation in temperature in midtroposphere is small (−0.2 to +0.2°C; not shown),

![Figure 8](image-url)
the variation in midtroposphere relative humidity is large (−15% to +15%; not shown). The increased water vapor in the midtroposphere can come from deep convection (especially, the evaporation of precipitating and nonprecipitating anvil clouds and/or mesoscale secondary circulations), shallow convection (detraining and/or evaporating congestus clouds), and widespread upward motion associated with large-scale convergence (e.g., associated with equatorial waves or the MJO). Figure 8b shows the composite of absolute upper-level vertical shear. The wind shear is dominated by the zonal wind. The magnitude of shear is greater in the Indian Ocean than in the western Pacific. The strongest shear mostly follows enhanced midlevel moistening and deep convection, which is consistent with previous findings (e.g., Kiladis et al. 2005). Theoretical and modeling results suggest that the vertical shear should significantly alter the structure of equatorial waves (e.g., Holton 1970, 1971; Lindzen 1971; Boyd 1978; Zhang and Webster 1989; Wang and Xie 1996; Xie and Wang 1996). These potential effects on the MJO need to be investigated more thoroughly and are beyond the scope of this study. Here we want to mention that a weaker upper-level vertical shear tends to keep upper-level clouds and moisture aligned with low-level active convection that generates them. Thus, with other conditions held unchanged, weaker shears favor development of larger MCSs, possibly favoring CMCSs. Looking at Fig. 8, the causality between the variation of shear and the intensity of deep convection is unclear, except that over EIO and MA the strongest shear is present in transition phases of the MJO in which deep convection is subsiding (cf. Figs. 2 and 4).

Previous studies point out that moisture conditions play crucial roles in both the development and self-organization of deep convection. Held et al. (1993) suggested that deep convection can more easily develop in places where the midtroposphere is moist and also tends to maintain moisture in middle and upper troposphere. The moist midtroposphere, combined with other factors (weaker wind shear, radiative heating, and surface fluxes), likely builds a positive feedback between the deep convection and the moisture field to help convection to self-aggregate (Tompkins 2001; Bretherton et al. 2005). Such self-aggregation tends to result in a spatially inhomogeneous precipitation and moisture field in contrast to the general thinking of a statistical equilibrium state. It also may favor CMCS development. A recent study also suggested that a very dry midtropospheric profile can suppress deep convection in favor of a shallow convection regime (Derbyshire et al. 2004). Yuter and Houze (1998) hypothesized that the sustainability of water vapor in the boundary layer is a favorable condition for developing large and long-lasting MCSs, which favors their occurrence over warm oceans with widespread moist boundary layers undergoing only slight diurnal modulation. Increased midtroposphere moisture coincides with both increased occurrence of all deep convective systems and relatively more frequent occurrence of CMCSs, which likely suggests that a moist midtroposphere is favorable for the development of mesoscale convective systems and it seems reasonable to further expect that extreme conditions favorable for MCSs might result in more CMCSs. However, whether the increased moisture is a consequence of or causative factor for enhanced deep and/or mesoscale-organized convection is not clear from these data or any other previous analyses of which we are aware. Observations have shown that the grid-mean (2.5° × 2.5°) column relative humidity (column WVP divided by the column saturated WVP) has very strong correlations with the magnitude of total precipitation rate on daily scales and almost exactly in phase (Bretherton et al. 2004). The exact reason for such a relationship is still an open question, and the possibility that the WVP is primarily the consequence of deep convection (simultaneously coexisting) cannot be ruled out. In that case, the WVP is strictly (statistically) related to local precipitation intensity and the increased midlevel moisture may be primarily a result of the expanding of convective regimes (i.e., the total occurrence of deep convective systems). According to that idea, the WVP would be solely a function of precipitation rather than related to the form of MCSs. Furthermore, whether the coherent changes in the form of deep convective system organization and the associated moisture field are specifically related to the MJO or rooted in the fundamental nature of moist convection remains a separate interesting question. Next we will attempt to address these questions.

b. Vertical motion as a proxy for active and suppressed conditions

As shown in section 3b the variability of the form of deep convective systems is strongly reflected in their relative contribution to precipitation (Fig. 4b1). CMCSs produce a significantly greater fraction of total precipitation in active phases of the MJO in all subregions. Hence, we investigate the variability of the relative contribution to precipitation of CMCSs and the column-integrated WVP by constructing 16 composites by subdividing the data into

1) four regions,
2) two seasons (October–March or “winter” and April–September or “summer”), and
3) strong/weak (or no) MJO (the amplitude of WH04 RMM indices greater than/less than or equal to 1).

The domain-mean $\omega_{500}$ is used as a proxy of the overall strength of domain-mean convection on a large scale (thousands of kilometers). This proxy is based on the fact that deep convection dominates ascending regimes and stronger large-scale upward motions are usually related to enhanced deep convection while subsiding regimes are primarily associated with shallow clouds (Yuan and Hartmann 2008).

Because convection patterns vary with regions and seasons, we define four quartiles based on the cumulative frequency of occurrence of $\omega_{500}$ for each subregion in each season, with quartile 1 being the most convectively active days and quartile 4 representing the most suppressed days for each ensemble (Figs. 9a,b). The increase of the relative contribution to precipitation from CMCSs with increasing overall strength of the large-scale convection (domain mean of the $\omega_{500}$ proxy) is clearly seen in Fig. 9c. The difference between quartile 1 and quartile 4 is about 13% as measured by the difference between median lines, which indicates up to 3 times variation in the relative contribution to precipitation of CMCSs since the mean of all 64 ensembles ($\overline{F}$ in Fig. 9c) is about 13.3%. This shows that under such separation according to large-scale convective scenario, the shifting of form of MCSs is well stratified regardless whether it is restricted to the strong winter season MJO. Next we will investigate how WVP within a more local, fixed convective regime varies in different large-scale convective scenarios.

The associated anomalies in WVP are shown in Fig. 10 based on 1.5° gridded daily-mean data from ERA-Interim and AMSR-E retrievals. To illustrate the spatial variability of moisture anomalies we divided the composites into seven convective regimes corresponding to different gridpoint values of AMSR-E daily-mean precipitation. Composites for column relative humidity (derived based on ERA-Interim temperature and moisture; not shown) show results similar to those of WVP, given that atmospheric temperature anomalies are small across both convective regimes and large-scale convective scenarios (not shown). We then focus only on results and discussions regarding the WVP. As shown in Fig. 10, the middle point of each precipitation bin ranges from 0 to 39.8 mm day$^{-1}$. Here we consider the daily-mean grid-mean precipitation to be a good indicator of the overall amount of deep convection for a day at the 1.5° grid scale. The data are grouped by quartiles of the domain-mean $\omega_{500}$. As noted in the preceding paragraph, the quartiles are proxies for active (quartile 1) to suppressed (quartile 4) large-scale mean conditions. Such
widespread conditions could be determined by phase of the MJO or some other large-scale dynamical condition affecting the whole domain. It is well known that deep convection can occur locally under widespread suppressed conditions. Thus, the plots show how locally active convection might behave in an otherwise large-scale active or suppressed condition. Figures 10a1–a3 show that grid-mean column WVP increases with increasing local precipitation, consistent with Bretherton et al. (2004). In addition, the ERA-Interim data show that the increase of WVP toward stronger convective regimes is largely due to an increase of WVP (>8 mm) in midlevels.
The low-level WVP is only about 2 mm. We have found that the composites of ERA-Interim and AMSR-E WVP in general agree with each other quantitatively (cf. Figs. 10a1 and 10a2 with Fig. 10a3). Not surprisingly, Fig. 10b4 shows that the variance of precipitation composites is negligible because the composite is based on fixed bins of the same data. Although the precipitation (or the mean strength of convection) within each convective regime does not vary perceptibly, the WVP of the same convective regime varies considerably according to the strength of the large-scale upward motion and net rain amount. In the right-hand column (Figs. 10b1 and 10b3), the midlevel WVP and column WVP anomalies are seen to progressively decrease from quartiles 1 to 4, across all convective regimes ranging from regimes with no precipitation to very intense precipitation (>35 mm day$^{-1}$). Figure 10b4 further confirms that relative variations in precipitation within each convective regime are negligible. These plots show that the 16 ensembles of midtroposphere conditions (850–400 hPa) over all convective regimes consistently have an (10%–25%) increase in WVP from quartile 4 to quartile 1. These results suggest that the amount of local convection is not the primary explanation for the varying midlevel moisture contents of the four regimes. And the more humid midtroposphere is associated with the phenomenon that MCSs tend to cluster and form spatially larger active systems (CMCSs). To the extent that the upward motion and net precipitation regimes are proxies for the MJO phases (quartile 1—active, quartile 4—suppressed), one could conclude that the intensity of local convection is not the proximate cause of the increased mass of water vapor at midlevels on the daily scale in active phases of the MJO. Other explanations could be sustained water vapor from previous convection, the spread of moisture from another convective area, or large-scale moisture convergence (e.g., due to wave-forced uplift).

Figures 10a2 and 10b2 show that the WVP below 850 hPa does not vary much among different convective regimes or different quartiles, which indicates that a robust moist layer is present at all times, whether conditions are locally convectively active or convectively suppressed over a wide domain. This result shows that all the phases of the MJO are well supplied with low-level moisture, and that the variability from phase to phase is more likely a free-atmospheric variation above the moist layer.

We also used the grid-mean $\omega_{500}$ instead of precipitation to divide our data into different dynamical regimes associated with different strengths of $\omega_{500}$ and obtained similar results (not shown). Gradual increases of midlevel WVP from quartile 4 to quartile 1 were found over all dynamical regimes, including regimes with strong subsiding motions and very dry troposphere (~43-mm ERA-Interim column WVP).

6. Conclusions

This study has examined the characteristics of deep convective systems in association with the boreal-winter MJO, as defined by WH04 strong MJO indices. As found to be the case by many previous studies [see review by Zhang (2005)], our composites show well-defined MJO features in SST, $\omega_{500}$, precipitation, water vapor, and zonal wind field in the Indo-Pacific region of the tropics. We have further found that the MCS occurrence field propagates eastward in a pattern resembling precipitation in the MJO.

In addition to general features associated with MJO, we have brought to light new information associated with the organization of deep convective systems during phases of the MJO. High-cloud systems, which we have categorized as CMCSs, SMCSs, and HCSXs, contribute 20%–40% of domain-mean cloud cover and 60%–90% of the total precipitation. CMCSs have the largest increase in their relative occurrence over all subregions of the Indo-Pacific zone in the local active phase of MJO. The mesoscale organization of deep convection within the MJO, moreover, has clear geographical dependence. CMCSs favor oceanic conditions, and the largest SMCSs occur more frequently over the MA area. In the MJO, deep convection thus not only changes in its overall frequency of occurrence but also varies in its horizontal scale and form of organization.

Despite the overall shift in the MCS organization (preferred type of MCS) from one phase of the MJO to another, the statistical characteristics of a given type of deep cloud system do not vary much. Large differences though are found in the structures of different types of systems. CMCSs overall have larger horizontal size, and they generally produce horizontally less extensive but relatively colder (lower Tb11) anvil clouds. The spatial clustering that characterizes the larger MCSs is one possible reason for the smaller anvil fraction of CMCSs. The structural differences among different systems are not well reflected in their Tb11 patterns alone. Representing cloud processes (such as precipitation) through methods that consider pixel values without consideration of their mesoscale cloud context of that grid point, or through purely infrared-based cloud-cluster methods, is thus largely limited. Larger MCSs usually penetrate deeper (i.e., have colder tops) than smaller MCSs, and they systematically but only slightly change their maximum height (in association with SST differences) with the MJO phases. The main difference in mesoscale convective system behavior from phase to phase is in the
number of large SMCSs and CMCS, which are more prevalent in the active phases.

By examining gridpoint values of precipitation in several subregions of the Indo-Pacific region, we find that the midlevel water vapor content is always highest when the domain-averaged conditions correspond to convectively active conditions (more domainwide precipitation) and lowest when the domain-averaged conditions correspond to widespread convectively suppressed conditions (less domainwide precipitation), regardless of whether or not deep convection is occurring locally (on the ~150-km grid scale). This enhancement in domain-scale convection is coherently associated with MCSs shifting toward their most exaggerated mesoscale form (i.e., CMCSs). Such information cannot be extracted either from gridpoint data or cloud-cluster occurrence if the clusters have not been categorized according to the spatial organizations of MCSs (i.e., SMCSs versus CMCSs). This coherence suggests that the increased relative frequency of occurrence of CMCSs in particular is likely associated with increased midlevel moisture or some other physical factor. Moreover, the increase in midlevel moisture spreading over all convective regimes further suggests that the increase of moisture in the midtroposphere is due to either nonlocal processes or the accumulation of moisture from previous convection. The moisture content of the atmosphere below the 850-hPa level shows much smaller variability with respect to either local or domain-averaged convective strength, indicating that the variability in convection associated with the MJO is connected with changing conditions in the free atmosphere above the persistent low-level moist layer.

Deep convective systems alter their preferred modes of organization when large-scale convection is enhanced along with the moistening of the middle troposphere. This result has broader implications than the crucial role of convection in the MJO. In a general way, it suggests a certain type of cloud feedback. When SST increases in a warm climate the deep convection is enhanced overall. However, such enhancement of deep convection may not only correspond to an increase of the general intensity of all convection as proposed in the “iris” hypothesis (Lindzen et al. 2001). Instead, the convective systems might shift their organization toward having larger precipitating areas. As suggested by previous work (YHH11; Romatschke and Houze 2011a,b), the relative amount of stratiform precipitation alters with the size and type of deep convective systems. It is well known that the variation of the fraction of stratiform rain usually corresponds to variation of the structure of diabatic heating, which might result in different mesoscale/large-scale circulation that can further affect properties and occurrence of upper-level clouds (Hartmann et al. 1984; Schumacher et al. 2004). Hence, mechanisms controlling upper-level ice clouds associated with deep convection in response to the increase of underlying SST might be much more complicated than that proposed in the iris hypothesis and awaits further study. This inference again points out the need for further investigation of mechanisms controlling the mesoscale structure and organization of deep convection and associated humidification processes.

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