The Role of Shallow Cloud Moistening in MJO and Non-MJO Convective Events over the ARM Manus Site

DAVID M. ZERMEÑO-DÍAZ AND CHIDONG ZHANG

Rosenstiel School of Marine and Atmospheric Science, University of Miami, Miami, Florida

PAVLOS KOLLIAS AND HEIKE KALESSE*

Department of Atmospheric and Oceanic Sciences, McGill University, Montreal, Quebec, Canada

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ABSTRACT

Observations from the Atmospheric Radiation Measurement Program (ARM) site at Manus Island in the western Pacific and (re)analysis products are used to investigate moistening by shallow cumulus clouds and by the circulation in large-scale convective events. Large-scale convective events are defined as rainfall anomalies larger than one standard deviation for a minimum of three consecutive days over a 10° × 10° domain centered at Manus. These events are categorized into two groups: Madden–Julian oscillation (MJO) events, with eastward propagation, and non-MJO events, without propagation. Shallow cumulus clouds are identified as continuous time–height echoes from 1-min cloud radar observations with their tops below the freezing level and their bases within the boundary layer. Daily moistening tendencies of shallow clouds, estimated from differences between their mean liquid water content and precipitation over their presumed life spans, and those of physical processes and advection from (re)analysis products are compared with local moistening tendencies from soundings. Increases in low-level moisture before rainfall peaks of MJO and non-MJO events are evident in both observations and reanalyses. Before and after the rainfall peaks of these events, precipitating and nonprecipitating shallow clouds exist all the time, but their occurrence fluctuates randomly. Their contributions to moisture tendencies through evaporation of condensed water are evident. These clouds provide perpetual background moistening to the lower troposphere but do not cause the observed increase in low-level moisture leading to rainfall peaks. Such moisture increase is mainly caused by anomalous nonlinear zonal advection.

1. Introduction

In the tropics, shallow cumulus clouds (herein called shallow clouds) are the most populous cloud type (Lau and Wu 2003; Masunaga and Kummerow 2006) and produce about 20% of the total rainfall (Short and Nakamura 2000). They are embryos for tropical deep convective disturbances in different time scales (Mapes et al. 2006). Diabatic heating and moistening effects from shallow clouds have been suggested to be particularly important to the Madden–Julian oscillation (MJO; Madden and Julian 1971, 1972). 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). Yanai et al. (1973) first explicitly pointed out the importance of moistening of the lower troposphere by shallow clouds in the development of deep convection. Such a moistening role of shallow clouds has subsequently been supported by observations (Nitta and Esbensen 1974; Johnson and Lin 1997). Moistening by shallow clouds has been assumed to be 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). However, arguments for the moistening role of shallow clouds in the MJO have mostly been based on the simultaneous increases in both low-level moisture and abundance of shallow clouds. No quantitative estimate of shallow cloud
moistening in the MJO has yet been available directly from observations.

Large-scale cloud distributions during the life cycle of the MJO have been studied largely using satellite observations (Lau and Wu 2003; Masunaga et al. 2008; Riley et al. 2011; Del Genio et al. 2012; Barnes and Houze 2013). These observations have shown a relatively large population of shallow clouds during suppressed phases of the MJO. The detection of shallow clouds from space is challenging because of partial beam-filling conditions (cloud sizes smaller than beam widths), the proximity of these clouds to the surface, and weak reflectivity from nonprecipitating or weakly precipitating shallow cumulus clouds (Sassen and Wang 2008; Zuidema and Mapes 2008). 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 U.S. Department of Energy’s Atmospheric Radiation Measurement Program (ARM) site at Manus Island in the western Pacific (Long et al. 2013) provides unique long-term (1996–2014) observations from a suite of instruments, including vertically pointing cloud radars, radiosondes, rain gauges, and others. Data from Manus have been used to evaluate satellite observations (Hollars et al. 2004) and model simulations (Chen and Del Genio 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 seasonal cycle (Mather 2005) and the MJO cycle (Deng et al. 2013). The long record of shallow clouds in a tropical deep convective regime observed by the cloud radar at Manus is a particularly valuable asset to the study of the MJO.

In this study, we evaluated the possible moistening role of shallow clouds in the MJO in comparison to that of the circulation. We proposed a null hypothesis that shallow clouds do not contribute to the commonly observed increase in low-level moisture 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 effects of shallow clouds on observed moisture tendencies and compare them to moistening effects by the circulation derived from reanalysis products. Shallow clouds observed by the vertically pointing radar at Manus Island were defined as time–height continuous radar echoes with their bases below 2 km and tops below the freezing level (approximately 4.5-km height). This definition of shallow clouds 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 layered (thin) clouds in the mid- and upper troposphere. Our shallow clouds should also be distinguished from congestus clouds, with tops above the freezing level and containing ice particles (Johnson et al. 1999). We attempted to develop a method to estimate moistening by shallow clouds 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 that will be discussed (section 5). The main reason to include the moistening estimates, however crude they might be, is to compare to other estimates (e.g., Bellenger et al. 2015) and to motivate quantitative assessments of shallow cloud moistening in future research, particularly using field observations.

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; 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 critical for the MJO.

The data and methods used are further described in sections 2 and 3, respectively. Results are presented in section 4. A summary and discussions are given in section 5.

2. Data

Table 1 lists the data used in this study. They include ground and satellite observations and data assimilation products. Ground observations are from the ARM site at Manus Island (2°3′S, 147°25′E), which have been collected since 1996 (Ackerman et al. 1999; Ackerman and Stokes 2003; Long et al. 2013). Observations from a vertically pointing 35-GHz (Ka band) millimeter cloud radar (MMCR; Kollia et al. 2007a) at this site were the primary sources of shallow cloud observations. These data were processed by the Active Remote Sensing of Clouds algorithm (ARSL; Clothiaux et al. 2000). Data from Manus also included observations from a microwave radiometer (MWR), upper-air soundings, a micropulse lidar (MPL), a ceilometer, and optical rain gauges. Other data used are rainfall estimates from the Tropical Rainfall Measuring Mission (TRMM 3B42v7; 0.25° × 0.25°; Kummerow et al. 2000); rainfall, specific humidity, and its physical tendency term from the operational
analysis (0.56° × 0.56°) of the European Centre for Medium Range Weather Forecasts (ECMWF) prepared for the Manus site (EC-ARM; ECMWF 2011); rainfall, specific humidity, and wind from the ECMWF interim reanalysis (ERAI; 1.5° × 1.5°; Dee et al. 2011); and the Wheeler and Hendon (2004) real-time multivariate MJO (RMM) index. The analysis period is 3 April 2001–7 March 2011, during which the Manus ARSCL and sounding data overlapped. Data gaps, however, still exist (Fig. 2 of Kalesse and Kollias 2013).

Soundings at Manus are regularly available at 1100 and 2300 UTC (2100 and 0900 local time). Their averages were used to represent daily means. Soundings with melting-level heights outside the typical range of 3.5–5.5 km (Geerts and Dawei 2004) were considered unreliable and excluded from our study. Sounding estimates of boundary layer heights from four methods (C. Sivaraman et al. 2012, meeting presentation) were averaged to provide a daily mean best estimate.

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 observations over Manus, however, have a temporal resolution of 1 min (Table 1). To homogenize these data and for computational efficiency, ARSCL reflectivity was averaged into 1-min profiles in a vertical range from 100 m to 22 km, with a 100-m resolution. Before the averaging, missing-value flags were placed at levels where no hydrometeors were detected. Daily means of soundings and EC-ARM data 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).

### 3. Methodology

The general strategy of this study is to compare the local moisture tendencies at Manus with the evolution of shallow clouds (section 3a), their moistening through evaporation of liquid water, and moisture tendencies due to the circulation (section 3b) during MJO and non-MJO large-scale convective events (section 3c).

Daily anomalies were calculated by removing the daily seasonal cycle. A 5-day running mean was applied to all data after daily anomalies (including moisture budget terms; see section 3b) were calculated. With a low-level wind speed of 5 m s⁻¹, the 5-day running mean made the point observations at Manus representative...
of a horizontal scale of about 20 degrees. A centered difference scheme was used to compute horizontal derivatives, and a forward scheme for vertical derivatives. The 5-day smoothed climatological daily mean was used to fill missing observations of boundary layer or melting-level heights in the soundings. A two-tailed Student’s t test was applied to composites of anomalies to mark results significant at the 95% confidence level.

a. Shallow cloud identification

The shallow cloud identification involved two steps: (i) removing radar profiles that are attenuated by rain and (ii) selecting echoes that represent shallow clouds.

1) Exclusion of strongly attenuated radar profiles

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

2) Shallow cloud classification

The vertically pointing MMCR can detect clouds passing over or standing above. We treated its temporarily adjacent hydrometeor echoes as objects that represent shallow clouds when their bases (first echo levels from the ground) were below 2 km (close to the hourly maximum height of the boundary layer; Fig. 1c) and their tops were below the freezing level. This classification is consistent with previous studies (Kollia et al. 2007b; Riley et al. 2011; Deng et al. 2013; Feng et al. 2014). In a joint probability distribution of echo-top height and echo depths (Fig. 1a), there are four distinct groups of clouds. One is along the diagonal line in Fig. 1a. 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.\(^1\) Clouds with echo-top heights below 5 km are the primary targets of this study. Figure 1b shows the total echo-top population based on 1-min observations and their fractions that were classified as shallow. The fraction of shallow clouds with rain rates greater than 25 mm h\(^{-1}\) and excluded from this study [section 3a(1)] is small. Varying the rainfall threshold slightly does not affect the subsequent results.

A given shallow cloud object was classified as precipitating if any 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 h\(^{-1}\), and water in the MWR window. Otherwise, the shallow cloud was classified as nonprecipitating.

\(^{1}\) The three clusters in Fig. 1a must not be mistaken as observational support of trimodal tropical convection discussed by Johnson et al. (1999).
Ground precipitation was detected in 46% of the shallow clouds, of which 7% had rain rates larger than 25 mm h$^{-1}$, our threshold for intolerable radar echo attenuation. About 78% of the shallow nonprecipitating clouds exhibited virga. Virga was identified when the cloud base detected by the MPL or ceilometer was higher than the lowest MMCR echo level (separated by at least one vertical grid point; 0.1 km). In this study, shallow echoes with virga are considered nonprecipitating because they do not remove moisture from the atmosphere. The vertical distributions of precipitating and nonprecipitating shallow cloud fractions are shown in Fig. 1b.

b. Moisture tendencies

The continuity equation of water vapor $q$ averaged over a given area is as follows (e.g., Johnson 1976):

$$\frac{\partial q}{\partial t} = -\mathbf{V} \cdot (\mathbf{V} q) - \frac{Q_2}{L},$$

(1)

where $\mathbf{V}$ is the three-dimensional wind vector, $\mathbf{V}$ is the three-dimensional wind vector, $L$ is latent heat of condensation, and $Q_2$ is the moisture sink (Yanai et al. 1973). The moisture sink represents the combined effects of evaporation $e$, condensation $c$, and vertical turbulent fluxes $\partial (\omega^* q^*)/\partial p$:

$$Q_2/L = c - e + \partial (\omega^* q^*)/\partial p,$$

(2)

where $\omega$ is the vertical wind velocity, $p$ is pressure, the braces denote areal mean, and an asterisk denotes deviation from the mean. In Eqs. (1) and (2), $q$, $\mathbf{V}$, $c$, and $e$ are all areal means, but the braces are not used for simplicity. For gridded products (e.g., reanalysis and model simulations), the area for the mean 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). A connection between point measurements and an areal mean might partially come from temporal smoothing (see section 3). Although it is not clear what the area for the mean should be for point measurements, it will be shown that the moisture variability observed by Manus soundings is comparable to that from EC-ARM interpolated to Manus and from ERAI averaged over a domain of $10^\circ \times 10^\circ$ centered at Manus (herein called the Manus domain; see Fig. C1 in appendix C). This suggests that Eqs. (1) and (2) can be applied to temporally smoothed observations at Manus to represent its domain.

Three processes determine the local moisture tendency $\partial q/\partial t$: moisture flux divergence by the circulation $-\mathbf{V} \cdot (\mathbf{V} q)$, liquid-to-vapor conversion $c - e$, and vertical turbulent moisture fluxes $\partial (\omega^* q^*)/\partial p$. The first two processes can, to a certain degree, be estimated using the observations and reanalysis products described in section 2. A method of estimating the role of shallow clouds in $\partial q/\partial t$ at Manus through $c - e$ is discussed in section 3b(1). Methods of diagnosing roles of the circulation and $Q_2$ in $\partial q/\partial t$ over the Manus domain using the reanalysis products are discussed in section 3b(2).

We have no means to estimate $\partial (\omega^* q^*)/\partial p$. This term will be ignored except when we discuss the limitations of our approach and the differences between our estimates of shallow cloud moistening and $Q_2$.

1) MOISTURE TENDENCIES BY SHALLOW CLOUDS

Our method aims to estimate bulk (daily mean, averaged over the lower troposphere) moistening by shallow clouds from observations. All water vapor that enters a cloud through its base may condense or not. Over a life-span of a cloud, its total condensed water is removed through either precipitation or evaporation. Condensed water can, to a certain extent, be measured. Our approach of estimating the moistening effect of shallow clouds includes only evaporation of condensed water. Moistening by detrainment of uncondensed vapor (Langhans et al. 2015) and the circulation in response to cloud diabatic heating (Chikira 2014; Janiga and Zhang 2015, manuscript submitted to J. Atmos. Sci.) and other factors (e.g., updraft strength, detrainment levels) cannot be estimated from the Manus observations and is not included in our approach.

Without turbulent fluxes and flux divergence by the circulation [to be considered in section 3b(2)], the water vapor tendency is determined solely by the liquid-to-vapor conversion:

$$\frac{\partial q}{\partial t} = e - c.$$

(3)

We explore the bulk moistening effect of shallow clouds on the lower troposphere:

$$\left\langle \frac{\partial q}{\partial t} \right\rangle = \left\langle e - c \right\rangle,$$

(4)

where the angle brackets indicate the integral $\int_{\Delta t} \frac{\partial q}{\partial t}(\cdot) \, dp$, and $h$ and $H$ are the levels of the cloud base and top, respectively. Over the 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

$$\left\langle \frac{\partial q}{\partial t} \right\rangle_{\Delta t} \approx \frac{\Delta q}{\Delta t} \approx \frac{\text{LWC}}{\Delta t} - P,$$

(5)
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 subperiod \( t \) during which a shallow cloud was observed. Notice that the anvil part of the cloud of Fig. 2 would not be counted as part of the 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:

\[
M = \sum_{i=1}^{n} \left( \frac{\langle \text{LWC} \rangle}{\Delta t} - P \right) \Delta t, \tag{6}
\]

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. 1994; Wood and Taylor 2001) in shallow clouds, it can be shown (appendix B) that \( \langle \text{LWC} \rangle \) is related to the liquid water path as

\[
\langle \text{LWC} \rangle = \langle \text{LWP} \rangle (H - h_b)^2, \tag{7}
\]

where \( h_b \) is the height of the boundary layer top. Thus, from Eq. (6),

\[
M = \sum_{i=1}^{n} \left[ \frac{\langle \text{LWP} \rangle (H - h_b)^2}{\Delta t} - P \right] \Delta t. \tag{8}
\]

At Manus, LWP is available from MWR measurements (Table 1); \( H \) was obtained from the MMCR (section 3a), \( h_b \) was obtained from the soundings, and \( P \) was obtained from the optical rain gauges. Numerical studies (e.g., Zhao and Austin 2005) have suggested that typical life spans of shallow nonprecipitating and precipitating clouds are about 18 and 25 min, respectively. These studies 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. We choose a range of \( \Delta t \) between 30 and 90 min. This range defines the uncertainty of our moistening estimates. The value of \( M \) was estimated separately for shallow precipitating and non-precipitating clouds.

We also estimated \( \langle \text{LWC} \rangle \) using two LWC–reflectivity relations for stratocumulus clouds (Liao and Sassen 1994; Frisch et al. 1995). However, because the reflectivity increases as a power of 6 with the diameter of the cloud particles, these estimates suffered from errors with orders of magnitude larger than those based on LWP, especially for precipitating clouds. These results are not reported here.

We will discuss in more detail the limitations of our moistening estimates in section 5.

2) LARGE-SCALE MOISTURE TENDENCIES

Using ERAI, we diagnosed the moisture budget for both MJO and non-MJO convective events over the Manus domain. For consistency with previous studies (e.g., Johnson et al. 2015) that diagnosed \( Q_z \), we used the uncompressed version of Eq. (1):

\[
\frac{\partial q}{\partial t} = -\mathbf{V} \cdot \nabla q - \frac{Q_z}{L}. \tag{8}
\]

From Eq. (8), \( -Q_z/L \) was computed as the residual of the other terms. We decomposed variables in Eq. (8) into
their daily seasonal cycles (denoted by overbars) and their daily anomalies from the seasonal cycles (primes), which include both synoptic and intraseasonal perturbations. The anomalous moisture tendency equation is

\[
\frac{\partial q'}{\partial t} = -\mathbf{V}' \cdot \nabla q - \mathbf{V} \cdot \nabla q' - (\mathbf{V}' \cdot \nabla q') - \frac{Q'}{L} = -\left( \frac{\partial q'}{\partial x} + \frac{\partial q'}{\partial y} + \frac{\partial q'}{\partial p} \right) - \left( \frac{\partial q}{\partial x} + \frac{\partial q}{\partial y} + \frac{\partial q}{\partial p} \right) - \left( \omega \frac{\partial q}{\partial p} - \omega \frac{\partial q'}{\partial p} \right) - \frac{Q'}{L},
\]

where \( u \) and \( v \) are the zonal and meridional wind components, respectively. The first group of terms on the right-hand side of Eq. (9), \(-\mathbf{V}' \cdot \nabla q\), represents advection of mean humidity by anomalous wind; and the second group, \(-\mathbf{V} \cdot \nabla q'\), represents advection of anomalous humidity by mean wind. The third group, \(-(\mathbf{V}' \cdot \nabla q' - \mathbf{V} \cdot \nabla q')\), the difference between nonlinear advection of humidity and its seasonal mean, is referred to as anomalous nonlinear advection of humidity and is expressed herein as \(-(\mathbf{V}' \cdot \nabla q')\). The terms of Eq. (9) were averaged over the Manus domain after the 5-day running mean was applied.

c. Event identification

Large-scale convective events over Manus were identified using the TRMM 3B42 precipitation data during extended boreal winter (October–April). Events that showed unambiguous eastward propagation were classified as MJO events; the others were classified as non-MJO events. Details of this procedure are described in appendix C. A total of 22 MJO events and 15 non-MJO events were identified during the analysis period, of which 13 MJO events and 11 non-MJO events were covered by MMCR and sounding observations at Manus (Table 2). We used the latter events in our analysis. Each event was covered, on average, 91% of the time by the MMCR and 80% by the soundings from days 0 to 8 km (Fig. 4c). These ten- }
same for MJO events and, to a lesser degree, for non-MJO events (Figs. 5a,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. 5c,d). Low-level (below the freezing level; $\sim$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 with the increase in the depth of the moisture layer (Fig. 4a). The main task of this study is to explore the degree to which these IWV tendencies can be explained in terms of moistening by shallow clouds (sections 4b–4d) using the observations at Manus and by the circulation (section 4e) using the reanalysis products.

b. Variability of shallow clouds

Previous studies 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). This is also evident in our analysis. For both MJO and non-MJO events, the daily frequency of occurrence is higher for non-precipitating than for precipitating shallow clouds (Figs. 6a–d). Their mean size and occurrence undergo random fluctuations, but there is no qualitative difference between them in either type of event (Figs. 6c,d,g,h). There are no clear increases in their occurrence or depth (echo-top height) in total (Figs. 6a–d) or anomalies (Figs. 6e–h) leading to the rainfall peaks. Anomalies in daily frequency of nonprecipitating shallow clouds during both types of events are qualitatively related to those in boundary layer moisture (cf. blue and red lines of Figs. 7a–d): positive (negative) anomalies in shallow nonprecipitating clouds tend to occur more frequently with negative (positive) boundary layer moisture anomalies. A dry (moist) boundary layer is detrimental (nourishing) for cumulus clouds to grow deep, so there tend to be more (less) shallow nonprecipitating clouds when the boundary layer moisture is relatively low (high). This is particularly evident during the suppressed phase of MJO events, from days $-25$ to $-20$. While this result makes physical sense, there is no statistically significant correlation between shallow clouds and boundary layer moisture. We found not evident relation between shallow cloud occurrence and other environmental parameters, such as the boundary layer height and lower-tropospheric stability (Figs. 7e,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.

c. Variability of shallow cloud moistening

Composites of estimated daily moistening tendency by shallow clouds $M$ are shown in Figs. 8a–d, where the largest and smallest values correspond to $\Delta t = 30$ and $90$ min in Eq. (7), respectively, and the red line corresponds 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 large-scale convective 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$^{-1}$). On the other hand, moistening per cloud occurrence of precipitating clouds is larger ($1.4$ mm day$^{-1}$ h$^{-1}$) than that of nonprecipitating shallow clouds ($1$ mm day$^{-1}$ h$^{-1}$). The combined moistening by total shallow clouds ($1.5$ mm day$^{-1}$) is similar during MJO and non-MJO events. The magnitude of $M$ is comparable to the sounding-based low-level IWV tendencies (Fig. 8i,j) prior to rainfall peaks of both types of events (up to $1.2$ mm day$^{-1}$). However, neither the total nor the anomalies of $M$ increases toward the rainfall peaks; they seem to fluctuate randomly in both types of convective events (Figs. 8 a–h), as does the cloud occurrence (Figs. 6a–d). This result is not an artifact of our shallow
FIG. 3. Composites of longitude–time diagrams of daily anomalies in rainfall (colors) and 850-hPa zonal wind (contour interval of 1 m s$^{-1}$; solid contours for westerlies and dashed for easterlies; thick solid for zero) for (a) MJO and (b) non-MJO events over Manus Island, and (c),(d) their corresponding RMM phase diagrams (thick black lines for the composites; purple for individual events) from days −35 to 25 and (e),(f) spatial rainfall anomalies (colors) and 850-hPa zonal wind vectors at day 0. Tracking lines are orange; red dots correspond to day 0 at Manus; and blue dots correspond to the end point of the track (see appendix C). Only results significant at the 95% confidence level are shown.
cloud moistening estimation method. The daily mean LWC of both types of clouds is nearly invariable during both types of convective events (Figs. 8g,h). The temporal fluctuations in $M$ mainly come from the cloud occurrence.

In summary, the estimated moistening tendencies of precipitating and nonprecipitating shallow clouds do not vary in concert with the observed lower-tropospheric moisture tendency in either type of convective event. Shallow clouds are a nonnegligible 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.

d. 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 (Figs. 9a,b). They are dominated by condensation that leads to precipitation. There are two semi-perpetual peaks below and above the freezing level,
respectively, which have been observed at many tropical locations (Yanai et al. 1973; Johnson et al. 1996). During MJO events, the anomalous moisture tendencies (\(\frac{\partial q}{\partial t}\)physics) show a deepening of drying from the low to the midtroposphere at the same time that the rain rate increases toward its peak (Fig. 9c). This implies a growth in the depth of precipitating clouds. There is an apparent deepening in positive (\(\frac{\partial q}{\partial t}\)physics), starting at the low levels (2–4 km) from days -20 to -25 and reaching 8–10 km on day -5 (Fig. 9c). But this repeats again immediately after the rainfall peak with greater strength. So it cannot be related to the development of deep convection of the MJO. Anomalous drying and moistening are less significant during non-MJO events (Fig. 9d). These results imply that shallow cloud moistening in EC-ARM, if any, is overwhelmed by the drying from precipitation processes; consequently, it cannot be detected from the total physical tendency term.

Figure 10 clearly demonstrates that neither \(\frac{\partial q}{\partial t}\) nor \(\frac{\partial q}{\partial t}\)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.

e. Budget of moisture anomalies in ERAI

Anomalies in the local moisture tendency are similar in the sounding observations (Figs. 4c,d) and EC-ARM (Figs. 4g,h) at Manus and in ERAI (Figs. 11a) over the Manus domain for both MJO and non-MJO events. The apparent moisture sink \(-Q^c/L\) (Figs. 11c,f) from ERAI and the subgrid moisture tendency \(\frac{\partial q}{\partial t}\)physics in EC-ARM (Fig. 9c,d) are also alike. These comparisons lend confidence to use ERAI to diagnose the moisture budget of Eq. (9).
The largest terms of Eq. (9) are $-Q_0^2/L$ (Figs. 11e,f) and $-\omega \partial q/\partial p$ (Figs. 11g,h). They vary in tandem but with the opposite signs. Their combination results in a small but persistent positive tendency (moistening) in the lower troposphere with relatively small fluctuations before and after the rainfall peaks of both MJO and non-MJO events (Figs. 11i,j). Among the terms on the right-hand side of Eq. (9), only the anomalous nonlinear zonal moisture advection $-(u' \cdot \partial q/\partial x)$ shares similarities with anomalies in the low-level positive moisture tendency within 10 days prior to the rainfall peaks for both types of convective events (Fig. 11a–c).

These results are more evident and easier to quantify by vertically integrating the terms of Eq. (9) over the lower troposphere, from 1000 to 600 hPa (approximately from the surface to 4.5 km) and by diagnosing them in groups (Figs. 12). In Figs. 12a and 12b the combination of $-Q_0^2/L$ and $-\mathbf{V}' \cdot \mathbf{v}\tilde{q}$ is positive and remains approximately constant (0.3–0.8 mm day$^{-1}$) during both types of convective events. It can be interpreted as a source of background moistening, similar to the estimated moisture tendency of shallow clouds at Manus (section 4c). The vertical component of $-\mathbf{V}' \cdot \mathbf{v}\tilde{q}$, as shown in Figs. 11g and 11h, is much larger than the horizontal ones. During

![Fig. 6. Time–height composites of daily occurrence frequency of (a),(b) nonprecipitating (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. (i),(j) Total daily gauge rain rates are also shown. Contours mark results significant at the 95% confidence level.](image-url)
both types of convective events, the combination of drying by the anomalous nonlinear advection $-\mathbf{V}' \cdot \mathbf{v}_d$ (dark blue in Fig. 12c) and the background moistening by $-\mathbf{V}' \cdot \mathbf{v}_b - \frac{Q'_0}{L}$ (magenta in Fig. 12c) almost matches the local moisture tendency (black). But $-\mathbf{V}' \cdot \mathbf{v}_d - \frac{Q'_0}{L}$ (green in Figs. 12c and 12d) and drying by the meridional component of the anomalous nonlinear advection $-(u' \partial q'/\partial y)'$ (red in Figs. 12e and 12f) almost cancel each other. In consequence, the zonal component of the anomalous nonlinear advection alone $-(u' \partial q'/\partial x)'$ (orange in Figs. 12e and 12f) can well explain the positive anomalies in the moisture tendency during both MJO and non-MJO events.

Advection of moisture anomalies by the seasonal mean circulation $-\nabla \cdot \mathbf{v}_d$ does not play any role in the increase of the low-level moisture tendency prior to day $-5$. But this term is needed to explain the sharp drying after day $-5$. Although the balance between these terms in MJO and non-MJO events (Figs. 11 and 12) is qualitatively similar, there is one major difference. Negative anomalies in the low-level tendency near the rainfall peaks are stronger in MJO events than in non-MJO events (black in Figs. 12c–f). This difference is consistent with the observed boundary layer drying in MJO events (Figs. 4c,d).

Another way to examine the column moisture budget is to integrate Eq. (2) from the surface to the top of the atmosphere (denoted here by square brackets), which yields, for anomalies,

$$[\frac{Q'_0}{L}] + E' = [e' - c'] = P',$$

where $E$ is surface evaporation and $P$ is the surface rain rate. Negative $P'$ indicates net moistening because of subgrid evaporation $e'$ dominating subgrid condensation $c'$. There is negative $P'$ approximately from days $-10$ to $-20$ and from days $-10$ to $-15$ in MJO and non-MJO events, respectively, based on all three rainfall datasets (Figs. 11k,l or 12g,h). With the existence of other types of clouds in this period, the implied moistening from this result can be partially attributed to shallow clouds. This is consistent with our result.

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**Fig. 7.** Composites of anomalies in daily (a),(b) occurrence of NP and P shallow clouds; (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.
that shallow clouds provide a perpetual background moistening effect, which might be particularly notable during convectively suppressed periods.

5. Summary and discussion

In this study, we explored the idea that moistening by shallow clouds is essential to the MJO (Johnson et al. 1999; Kemball-Cook and Weare 2001; Benedict and Randall 2007) using more than 10 years of observations from the ARM Manus site. We examined the evolution of observed shallow cloud occurrence, estimated moisture tendencies because of precipitating and nonprecipitating shallow clouds, and compared them 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.

We have presented four main results: 1) shallow clouds, both precipitating and nonprecipitating, are ubiquitous with random fluctuations; 2) their moistening is almost perpetual without any apparent increase toward rainfall peaks; 3) the observed increase in low-level moisture leading to rainfall peaks is mostly due to anomalous nonlinear zonal advection; and 4) these features 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.

Based on these results, we fail to reject the null hypothesis that shallow cloud moistening does not contribute to the observed low-level moisture increase leading to rainfall peaks of MJO events. We thus conclude that shallow clouds are a persistent source of

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**FIG. 8.** Composites of daily moistening from (a),(b) NP and (c),(d) P shallow clouds; (e),(f) tendencies of water vapor mixing ratio from the soundings integrated (denoted by square brackets) over the lower troposphere (from 0.1 to 4.5 km); (g),(h) daily mean LWC of SC; and (i),(j) gauge rain rates for (left) MJO and (right) non-MJO events at Manus Island. In (a)–(h) red lines (shaded areas) mark \( \Delta t = 60 \) (30–90) min.
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 they do not explain the observed increase in low-level moisture leading to the rainfall peak of the MJO. This low-level moisture increase leading to the rainfall peak of the MJO is mainly a consequence of the anomalous 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; Del Genio et al. 2012). Our results are also consistent with cloud-permitting model simulations of Janiga and Zhang (2015, manuscript submitted to J. Atmos. Sci.), which show that 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 4e) 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.

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 3b. Rainfall from these clouds is larger than their LWC available for evaporation; 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 that we found in the EC-ARM data and with recent numerical simulations (Mechem and Oberthaler 2013). But it suggests that moistening by congestus clouds through detrainment (Johnson et al. 1999; Waite and Khouider 2010) may be overwhelmed by their drying through precipitation, at least on the daily time scale. The crudeness of our moistening estimates and attenuation of cloud radar echoes due to heavy rain may, however, undermine our estimates of moistening by congestus cloud.

Bellenger et al. (2015), using shipborne lidar and sounding data over the central equatorial Indian Ocean,
estimated the magnitude of moistening by shallow clouds between 1.5 and 4 km to be 10–20 g kg\(^{-1}\) day\(^{-1}\) (25–50 mm day\(^{-1}\)) on the scale of a few tens of minutes 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 as 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, manuscript submitted to J. Atmos. Sci.). 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 accurately estimating moistening by shallow clouds using field observations and uncertainties in the current knowledge on this subject.

Our study suffers from several limitations. The small sample size of the diagnosed events (13 MJO events and 11 non-MJO events) and the point measurements at Manus Island are constraints from the available data. Composites for all large-scale convective events presented in Table 2 (22 MJO and 15 non-MJO events) using ERAI have led to virtually the same results. It is desirable to confirm our results using a larger sample size and perhaps at different locations. The nature of the point measurements and the instrumentation availability at Manus forced us to make several assumptions in our moistening estimates. These assumptions are as follows:

(i) 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. Such data are not yet available from tropical deep convective regions.

(ii) MMCR echoes are reliable when the rain rate is lower than 25 mm h\(^{-1}\). This is a conservative approach that may have led to an underestimate of the total amount of shallow clouds. This choice of the threshold is based on a relatively small number of observations (appendix A). The issue
of attenuation by rain in cloud radar echoes needs more rigorous treatment (Feng et al. 2009).

(iii) Precipitation from shallow clouds is adequately measured by collocated rain gauges. This assumption neglects evaporated shallow cloud rain drops 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.

(iv) 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 h\(^{-1}\) or wet MWR window). This is adapted from previous studies (Mather et al. 2007) but needs to be verified.

(v) 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.
(vi) 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 with 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, manuscript submitted to *J. Atmos. Sci.*).

(vii) LWC increases linearly with cloud depth (adiabaticity). This was adapted from previous studies (Han et al. 1994; Wood and Taylor 2001). More realistic estimates of LWC in shallow clouds are needed. We 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 et al. 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. Our conclusions—namely, shallow clouds provide background moistening to the lower troposphere but are not responsible for the observed increase in low-level moisture leading to rainfall peaks of the MJO—were drawn from the main results at Manus summarized earlier in this section. It does not critically depend on our estimates of shallow cloud moistening and thus is not affected by the limitations of our estimates of shallow cloud moistening.

The method that we present to estimate shallow cloud moistening is nevertheless primitive. 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.

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**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 field campaign (October 2011–January 2012; Yoneyama et al. 2013) are compared to a merged reflectivity dataset (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 the Shared Mobile Atmospheric Research and Teaching Radar (SMART-R) of Texas A&M University (Table 1). These merged data will herein be referred to as K-SPOL-SR (Table 1). The KAZR is more accurate than the precipitation radars to detect cloud echo tops under nonprecipitating conditions because of 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 A1a 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. A1b). KAZR echo-top heights lower than 4-km height with rainfall rates larger than 25 mm h$^{-1}$ are likely to be underestimated by more than 1 km because of attenuation. When rain rates are less than 25 mm h$^{-1}$, KAZR echo tops lower than the freezing level ($\sim$4.5 km) are unlikely to suffer from attenuation, even though echo tops at higher levels may. A 25 mm h$^{-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-min 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 h$^{-1}$) 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 below 4 km) at Manus Island with surface rain rates exceeding the 25 mm h\(^{-1}\) threshold were excluded.

APPENDIX B

Liquid Water Content of Shallow Clouds

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

\[
\frac{\Delta \text{LWC}}{\Delta z} = a z, \quad (B1)
\]

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):

\[
\text{LWP} \int_0^H \text{LWC}(z) \, dz = 0.5 a H^2. \quad (B2)
\]

Solving Eq. (B1) for \(a\) and substituting it into Eq. (B2) yields

\[
\frac{\Delta \text{LWC}}{\Delta z} = (2\text{LWP}/H^2)z. \quad (B3)
\]

The vertically integrated free-tropospheric LWC (\(\langle \text{LWC} \rangle\)) of a cloud can be obtained by integrating Eq. (B3) from the top of the boundary layer \(h_b\) to \(H\):

\[
\langle \text{LWC} \rangle = \left(\text{LWP}/H^2\right)(H - h_b)^2. \quad (B4)
\]

APPENDIX C

Event Identification Method

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

The first step is identification of large-scale events over Manus Island. Anomalies of daily rainfall from TRMM were used for this. After applying pentad and 10° longitudinal running means to the rainfall data, an areally averaged time series was generated for the Manus domain (Fig. C1). A convective event was identified when the time series exceeded one standard deviation for at least three 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) during 2001–13 (Table 2).

The second step is 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), and the other is based 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–8°N 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 (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 toward the west along eastward tracking lines for possible eastward propagation and toward the east along westward tracking lines for possible westward propagation. Integrations along the tracking lines stopped when rainfall anomalies became lower than one standard deviation. There is one exception: if rainfall anomalies are lower than one standard deviation over the Maritime Continent (between 90°E and 145°E; Fig. C1) and become greater than one standard 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 tracking line with the largest averaged rainfall anomalies determined the direction and mean speed of a given event.

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.

---

**Fig. C1.** Composites of daily anomalous TRMM rain rates (mm day\(^{-1}\)) for each RMM phase with the RMM amplitude greater than 1 during October–April. Red stars mark the location of Manus Island. Blue boxes define the Manus domain. Only results significant at the 95% confidence level are shown.
In general, MJO events selected by our procedure start over the Indian Ocean approximately 15–20 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 events. MJO events that satisfied the RMM condition also satisfy the tracking condition, but not vice versa. The tracking condition identified weak MJO events that the RMM condition missed.

Table C1 presents the frequency of occurrence of the events that we identified with respect to each other. Most (86%) identified MJO events are primary as defined by Matthews (2008), 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 a 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. These 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.

REFERENCES


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