The cloud population and onset of the Madden-Julian Oscillation over the Indian Ocean during DYNAMO-AMIE

Scott W. Powell1 and Robert A. Houze Jr.1

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[1] Variability of the cloud population in the central equatorial Indian Ocean was observed in the context of the Madden-Julian Oscillation (MJO) during the Dynamics of the Madden-Julian Oscillation (DYNAMO) and Atmospheric Radiation Measurement Madden-Julian Investigation Experiment (AMIE) field campaigns. Radar observations from the polarimetric S-band radar on Addu Atoll in the Maldives characterize the types of convective and stratiform radar echoes and the heights their 20 dBZ contours reach. To gain insight into the relationship between clouds and humidification of the troposphere leading up to and during an active MJO event, the work relates variability of the observed precipitation structure to that of tropospheric humidity and upper level zonal wind. The variability in stratiform precipitation areas dominates variability in the nature of precipitating convection associated with the MJO. Areal coverage of precipitating radar echo, convective echo top height, and tropospheric humidity above 850 hPa rapidly increase over ~3–7 days near MJO onset. This rate of increase is substantially faster than the 10–20 days needed for buildup of moisture prior to MJO onset as hypothesized by the “discharge-recharge” hypothesis. Convective echoes become more common during the days prior to MJO onset, and the increased convection occurs before low-tropospheric moistening. The upper troposphere rapidly moistens as the first widespread stratiform region passes over an area. Thus, clouds likely play a role in tropospheric humidification. Whether increased low-tropospheric humidity causes vertical growth of convection has not yet been determined.


1. Introduction

[2] Since the late 1960s, field projects deployed in low latitudes have yielded information about the structure and behavior of deep convection, and in the tropics, geostationary satellites have provided a global view via visible and infrared sensing. During the past 15 years, passive microwave and active sensors (radar and lidar) aboard Earth-orbiting platforms have observed tropical clouds and precipitation in greater detail. Despite these years of intensive observations of tropical convection, one dominant mode of tropical convection occurring on intraseasonal time scales (30 to 90 days) remains poorly understood.

[3] Madden and Julian [1971, 1972] first noted an intraseasonal quasiperiodic cycle in deep convection over the Indian Ocean and western Pacific tropical warm pool, which is now widely known as the Madden-Julian Oscillation (MJO). The MJO has been extensively documented (see review by Zhang [2005]). Its convectively active signature moves slowly eastward at about 5 m s⁻¹ [e.g., Knutson et al., 1986]. The moist envelope of deep convection develops anywhere between the Indian Ocean and the tropical west Pacific but most commonly over the Indian Ocean. The phenomenon occurs on an irregular interval—roughly every 30 to 90 days. The large-scale wind structure of the MJO is often described as a combined Kelvin-Rossby mode coupled to deep convection through vertical distribution of heating [e.g., Gill, 1980; Nougues-Paegle et al., 1989; Houze et al., 2000].

[4] Hypotheses explaining the mechanisms through which the onset of convection occurs over the Indian Ocean include Knutson and Weickmann’s [1987] idea that circumnavigating upper tropospheric velocity potential and zonal wind anomalies occurring as a Kelvin wave response to a previous MJO may trigger the next MJO. Seo and Kim [2003] conclude that the Kelvin wave response to one MJO circumnavigates the globe and generates a new cycle. Virts and Wallace [2010] support this idea by showing evidence that a circumnavigating Kelvin wave associated with the MJO exists and modulates the frequency of cirrus in the tropical tropopause transition layer (TTL). However, while Matthews [2000] suggests that...
mean sea level pressure anomalies associated with a Kelvin mode circumnavigate and coincide with the beginning of the next MJO cycle, Matthews [2008] conducts empirical orthogonal function (EOF) analysis on filtered outgoing longwave radiation (OLR) anomalies and shows that for only 60% of MJO cases can the negative anomalies be definitively traced back to a prior MJO event. More recently, a modeling study by Ray and Li [2013] suggests that in the long-term mean sense, extratropical influences have a greater influence on generating an MJO than tropical influences such as circumnavigating Kelvin modes. Nonetheless, such circumnavigating features are responsible for at least some individual cases of MJO onset. [5] Bladé and Hartmann [1993] pointed out that no dynamical link has been established between upper level wind anomalies and the formation of convergence in the lower troposphere that would result in enhanced convection. They offered a discharge-recharge theory of MJO onset, which proceeds as follows. After an MJO event passes, large-scale subsidence rapidly stabilizes and lowers the relative humidity through most of the troposphere. As time goes on, humidity gradually increases in the lower to middle troposphere, and a new MJO cycle commences once the atmosphere becomes sufficiently unstable and moist to support deep convection again. By compositing tropical rawinsonde data, Kemball-Cook and Weare [2001] conclude that both wave-conditional instability of the second kind [Lau and Peng, 1987; Wang and Rui, 1990; Salby et al., 1994] and discharge-recharge processes may be responsible for MJO onset and propagation. [6] One hypothesis regarding how the recharge moistening occurs is that cumulus clouds—and particularly cumulus congestus clouds [Johnson et al., 1999]—are responsible for “pre-conditioning” or “recharging” the troposphere prior to an MJO over time scales of several weeks. Benedict and Randall [2007] add detail to the proposed interaction between cumulus clouds and tropospheric moistening on the basis of compositing satellite and reanalysis data relative to a filtered maximum in rainfall during several MJO events. Their results appear to support the discharge-recharge process. They suggest that low-level heating and moistening by cumulus clouds gradually condition the lower troposphere for explosive convective development over a 10 to 15 day period prior to MJO onset. A positive feedback is suggested in which clouds grow taller as moist static energy (MSE) is transported vertically; the deeper clouds then transport MSE to even higher levels, and the process continues until the environment is moist and unstable enough to support deep convection for widespread convective events to occur. Kemball-Cook and Weare [2001] suggest a similar process occurring over the 20 days prior to MJO onset. Support for this hypothesis has been offered in numerous other composite studies of rawinsonde, satellite, model, or reanalysis data [e.g., Kiladis et al., 2005; Masunaga et al., 2006; Haertel et al., 2008; Wu and Deng, 2013]. [7] Recent studies of tropical cumulus suggest that on time scales of hours to 2 or 3 days, premoistening of the free troposphere by shallow convection and congestus cannot explain rapid transitions to deep convection. Hohenegger and Stevens [2013] find that over the tropical Atlantic, local cumulus congestus clouds transition into deep convection more quickly than the time needed for congestus clouds to moisten the troposphere. Furthermore, they show that the probability of a congestus cloud developing into a cumulonimbus does not increase for longer-lived congestus. Kumar et al. [2013] similarly show that the time scale needed for deep convection to develop near Darwin, Australia, is only a few hours. They conclude that large-scale dynamics are important in the transition from shallow to deep clouds. Masunaga [2013] shows that for congestus clouds and mesoscale convective systems throughout the tropics, free tropospheric moistening occurs as early as a day before convective clouds begin to deepen—or 2–3 days before the peak of cloud cover associated with a mesoscale system. While all three studies show that humidification of the environment is not necessary for deep cloud development locally, none exclude the possibility that premoistening could promote deep and widespread convection on longer time scales, such as that proposed to be relevant to the MJO. [8] An alternative hypothesis is that large-scale moistening may occur gradually prior to MJO onset at a given location as a result of horizontal moisture advection, particularly in the lower troposphere. The increase of lower tropospheric moisture would favor development of shallow convection, which would act as agents to transport MSE vertically. The largest term in the large-scale composite MSE budget over the Indian Ocean on an intraseasonal time scale is horizontal advection [Maloney, 2009]. Andersen and Kuang [2012] examine the effects of the advection of MSE on MJO propagation and again show that horizontal advection dominates the MSE budget. MSE buildup occurs to the east of convection, and drying occurs to the west of, or behind, the convection. The advection of MSE may be related to boundary layer moisture convergence, which has been hypothesized to influence the discharge-recharge cycle as well [Hendon and Salby, 1994; Maloney and Hartmann, 1998; Matthews, 2000; Seo and Kim, 2003]. Such studies provide a potential mechanism for why new convection is favored downstream of currently active convection. In other words, they potentially explain the eastward propagation of the MJO. They do not, however, explain how MJO onset suddenly occurs over the Indian Ocean in the absence of upstream convection. [9] Instances of the MJO were observed during the Tropical Ocean–Global Atmosphere Coupled Ocean–Atmosphere Response Experiment (TOGA COARE) [Godfrey et al., 1998] in 1992–1993 and the Mirai Indian Ocean cruise for the Study of the MJO-convection Onset (MISMO) [Yoneyama et al., 2008] in 2006. TOGA COARE provided detailed observations in segments of two MJO events over the tropical west Pacific, and MISMO documented one weak MJO event over the Indian Ocean. During TOGA COARE, rawinsondes showed humidification of the troposphere [Lin and Johnson, 1996] prior to onset of MJO-related convection over the tropical west Pacific. Katsumata et al. [2009] attributed humidification of the upper troposphere in MISMO primarily to eastward propagating mesoscale cloud systems. [10] The Dynamics of the Madden-Julian Oscillation (DYNAMO) and the ARM (Atmospheric Radiation Measurement) Madden-Julian Oscillation Experiment (AMIE) were the U.S. contributions to an international collaboration studying the cloud population of the Indian Ocean in the context of the MJO [Yoneyama et al., 2013]. This cooperative experiment was unprecedented because of the amount and duration of intensive radar and rawinsonde observations over and around the Indian Ocean during boreal winter. The long period of
intensive observations allowed for detailed observations of the variation of convection throughout several cycles of the MJO in an area where the initial onset of MJO-related convection frequently occurs. One major goal of the experiment was to explore hypothesized mechanisms for onset and propagation of an MJO in and through the Indian Ocean and tropical west Pacific.

[11] This paper examines radar and rawinsonde data sets collected during DYNAMO/AMIE to describe the structure, development, and organization of MJO-related convection over the central equatorial Indian Ocean and investigate the relationship between vertical growth of convective clouds, humidification of the troposphere, and MJO onset where MJO-related convection initiates.

2. Radar and Rawinsonde Observations

[12] The intensive observing period (IOP) for DYNAMO/AMIE spanned the period 1 October 2011 through 9 February 2012. Several scanning precipitation and vertically pointing cloud radars were deployed at locations around the central equatorial Indian Ocean during DYNAMO/AMIE. Observations by these radars provide an unprecedented view of the three-dimensional structure of the convective entities making up the population of clouds in the MJO. Figure 1 shows the locations of each radar platform and rawinsonde launch site in the IOP area. At Addu, three radar sites were located within 13 km of each other: the C-band Shared Mobile Atmospheric Research and Teaching Radar (SMART-R); the ARM mobile facility AMF2, which housed a vertically pointing cloud radar; and the dual-wavelength Doppler polarimetric S-band and Ka-band (S-PolKa) radar system described below. The extensive radar data are supplemented by soundings taken every 3 h at the primary DYNAMO/AMIE sites and at lower frequencies at other locations throughout the Indian Ocean and maritime and Indian subcontinents. The 3-hourly rawinsonde launches at Gan Island (0.69°S, 73.15°E) on Addu Atoll provided details on the vertical profiles of wind, temperature, and moisture content in the immediate vicinity of the S-PolKa radar. Detailed information about the sondes, including bias correcting and smoothing of the sonde data, is available in Long and Holdridge [2012].

[13] For this study, we concentrate on the radar and rawinsonde observations made on Addu Atoll. The primary radar system used for our analysis is the dual-wavelength, dual-polarimetric S-PolKa from the National Center for Atmospheric Research. The system consists of two radars, with wavelengths of 10 cm (S-band) and 0.8 cm (Kw-band), mounted on the same platform with matching beam widths of 0.91°. S-PolKa executed a scan strategy consisting of surveillance scans at elevation angles up to 11° for 5 min, followed by 10 min of range height indicator scans (RHIs, i.e., scanning in elevation at a fixed azimuth) spaced at intervals of 2° of azimuth in the northeast and southeast sectors of radar coverage. Each RHI sector was located over open ocean and scanned up to 41°. Spacing between elevation angles for RHIs was 0.5°. The scanned sectors were bounded by azimuths 6° and 82° and 114° and 141.9°, respectively. Then the scan sequence was repeated continuously from 28 September 2011 through 15 January 2012, uninterrupted with operator-selected scans. Technical issues caused only a few minor interruptions. Thus, the radar obtained a statistically homogeneous data set for all time periods, whether suppressed, active, or intermediate in terms of the cloud population, defined herein as the entire ensemble of clouds observable by S-PolKa. More information on the S-PolKa system, as well as the available data sets may be found at http://www.eol.ucar.edu/projects/dynamo/spol/SpolKa_DYNAMO_UsersGuide.toc.html.

3. Radar-Derived Products

[14] Before deriving products from the S-PolKa data, the reflectivity field was interpolated to a Cartesian grid with 500 m horizontal and vertical resolution using both surveillance scans and RHIs. Data within 12 km of the radar site is excluded because sampling of the tops of echoes is limited within this range by the maximum elevation angle of the radar. All data sets extend out to 150 km from the radar site. The areal coverage of the gridded data set is 70,685 km².
3.1 Convective/Stratiform Classification of Radar Echoes

Reflectivity values measured by S-PolKa were analyzed to determine if a volume of radar echo was convective or stratiform. Convective precipitation areas are characterized by strong vertical motion, heavy rain, and a low-level to mid-level maximum in diabatic heating, depending on cloud depth. Stratiform regions have less intense vertical motion, lighter precipitation, a radar bright band near the 0°C level where melting precipitation occurs, and a diabatic heating maximum (minimum) in the upper (lower) troposphere. The bright band is only a few hundred meters thick, and therefore, the radar only resolves it well near the radar, before the beam widens too much with distance from the antenna. The algorithm used to classify clouds observed as either convective or stratiform therefore emphasize the horizontal variation of radar echo intensity. The algorithm used here is a modification of the method used by Houze [1973], Churchill and Houze [1984], Steiner et al. [1995], and Yuter and Houze [1997]. The algorithm, including input parameters used, is described briefly by Houze [1997] and in greater detail in Didlake and Houze [2009, Appendix A]. For our analysis, we use radar reflectivity at 2.5 km. Background reflectivity was computed using the mean reflectivity of all pixels within 11 km of each data point. Echoes at 2.5 km exceeding 40 dBZ are automatically classified as convective, and echoes weaker than 5 dBZ are not classified. These choices have been calibrated manually, as advocated by Steiner et al. [1995], by checking algorithm output against RHI observations close to the radar, where it is possible to determine whether or not the algorithm is misclassifying echoes with bright bands as convective. The 11 km radius and 40 dBZ threshold values minimize such misclassification.

3.2 Precipitation Rates

The radar-derived precipitation rates computed by S-PolKa utilize reflectivity (Z), differential reflectivity (ZDR), and the specific differential phase (KDP) of the S-band. A “hybrid” rain rate algorithm is employed to determine the best relationship among Z, ZDR, KDP, and rain rate to use. A detailed explanation of the algorithm with references included may be found at http://www.eol.ucar.edu/projects/dynamo/spol/parameters/rain_rate/rain_rates.html.

3.3 Echo Top Heights

Echo top heights are computed only for high-elevation RHI scans because they capture the top of all echoes except those close to the radar. S-PolKa surveillance scans, which were confined to lower elevation angles, as noted above, did not always detect the tops of echoes. Echo top heights represent the maximum height at which some threshold of reflectivity—20 dBZ for applications in this paper—is observed. Echo top heights are determined by finding the highest 20 dBZ echo in a single contiguous echo volume by searching downward from the maximum height (20 km) of the radar data set. We choose a 20 dBZ threshold because high reflectivity, such as 40 dBZ, occurs less frequently and often results in a small sample size of echoes. Because our echo top detection algorithm starts from the top of the data set, too small of a threshold could erroneously classify anvil cloud that extends above a convective element as the top of the convective core. Additionally, a 20 dBZ threshold is easily comparable on the large scale with data from the Tropical Rainfall Measuring Mission (TRMM) precipitation radar, which has a minimum detectable reflectivity of about 17 dBZ.

4 Summary of Convection Observed at S-PolKa During DYNAMO

S-PolKa observed three 1 to 2 week long periods of enhanced precipitation during DYNAMO, each of which was followed by a period of reduced precipitation. These large-scale convective events (hereafter, LCEs) are referred to as LCE1, LCE2, and LCE3. Although the radar continuously sampled a relatively small domain, each convective event was part of an MJO event that propagated...
Gottschalck et al. [2013] have documented the MJO events observed during DYNAMO. Figure 2 shows plots of total radar-estimated mean precipitation within the 150 km range of S-PolKa as well as the amounts of total rainfall classified as either convective or stratiform. Satellite imagery and satellite-based precipitation estimates suggest that precipitation amounts during LCE3 were smaller than during LCE1 and LCE2 because less convection was observed near S-PolKa than elsewhere in the Indian Ocean; the strongest convection occurred east of 80°E [Gottschalck et al., 2013, Figures 4 and 13].

For purposes of discussing the convection during the IOP, we have subjectively classified radar echoes according to their areal coverage of stratiform precipitation. Figure 3 shows typical examples of what we refer herein to as isolated convection, echo clusters, squall lines, and mesoscale convective systems (MCSs). Isolated cells have little associated stratiform, and echo clusters usually have only small areas of associated stratiform. Squall lines are deep convective elements organized into a rapidly propagating convective echo line, which may or may not have an attached stratiform echo. Squall lines are often located along cold pool boundaries that advance laterally from convective downdrafts—in this case with a west-to-east component of motion. MCSs are systems consisting of a combination of widespread stratiform precipitation and deep convective cells, which in some cases are embedded in the stratiform precipitation and in other cases trail behind a squall line. The largest MCSs have stratiform regions hundreds of kilometers in dimension. Extremely large MCSs with such gigantic stratiform regions occurred in TOGA COARE [e.g., Chen et al., 1996, Figure 12] as well as in DYNAMO/AMIE, where they were seen both by the S-PolKa radar on Addu Atoll and the shipborne radar on the Revelle (www.atmos.washington.edu/~houze/DYNAMO-AMIE/).

The first LCE began in early October after anomalous westerlies above 500 hPa transporting dry air from eastern Africa and the Arabian Peninsula subsided. Isolated convective cells extending as high as ~8 km were observed through 10 October. By 12 October, individual convective elements began to form into clusters of precipitating echoes, which included convective elements and small, lightly precipitating stratiform areas surrounding them. Figure 2 indicates that at this time, the amount of stratiform precipitation began to increase. A large MCS moved within range and over S-PolKa on 16 October, and widespread stratiform precipitation was subsequently observed every other day through 30 October.

Figure 3. (5 Nov, 21 UTC) Examples of isolated shallow cells, (11 Nov, 18 UTC) small clusters of convection with limited stratiform, (31 Oct, 05 UTC) a squall line with associated stratiform area, and (23 Nov, 21 UTC) mesoscale convection with widespread stratiform coverage. The boxes in each 2 × 2 table correspond to the percentage of rainfall that is convective (top, left) and stratiform (bottom, left) and the fraction of the radar domain covered by convective (top, right) and stratiform (bottom, right) echoes for each period. Dates for each event are shown.
On days during which widespread stratiform echo was not present, the radar observed mostly isolated convective elements and several small clusters of precipitation. A similar 2 day periodicity in mesoscale organization was also observed during TOGA COARE and MISO, during which convection was probably linked to the periodicity of westward propagating inertio-gravity waves [Takayabu, 1994; Chen et al., 1996; Yamada et al., 2010]. Recent analysis of DYNAMO and AMIE data suggests that the 2 day periodicity in precipitation (Figure 2) at Addu Atoll may also be linked to inertio-gravity waves [Zuluaga and Houze, 2013]. The 2 day time scale may also represent the time required for the atmosphere to again become unstable after a large mesoscale event [Chen and Houze, 1997a, 1997b]. What is new here is that the stratiform component of the precipitation follows the 2 day variation along with the total rainfall.

[21] The second LCE occurred during November and at first exhibited isolated convective cells and small echo clusters. A few of these convective elements generated down-drafts strong enough to form cold pools that resulted in squall line formation. After a few isolated convective cells formed on 4–9 November, large clusters began to develop from 10 to 17 November. Large MCSs occurred on 18, 23, and 27–28 November. The periodicity of widespread stratiform precipitation was notably different during LCE2 than during LCE1. Widespread stratiform precipitation during LCE2 was observed only about every 4 days, similar to the 3–4 day and 6–8 day periodicity in convection seen over the Indian Ocean and attributed to mixed Rossby gravity waves and equatorial Rossby waves during MISO [Yasumaga et al., 2010]. (The S-band radar on S-PolKa was not operational from 0700 UTC, 20 November to 1100 UTC, 21 November. During that time, convection is subjectively characterized using observations from the nearby SMART-R.) On days between individual MCS events, convection mostly took the form of small clusters or squall lines and not isolated convective cells as was the case during LCE1. For the remainder of the paper, “rainy periods” refer to those during a LCE for which the hourly, radar-estimated rainfall averaged over the entire domain was at least 0.1 mm. “Dry periods” are those that fall below the same threshold. A domain-averaged hourly rainfall of 0.1 mm is typically found when part of a large stratiform region occupies a small portion along the outer edge of the radar domain or when echo clusters are observed. About two thirds of the total time during an active LCE is a “rainy period”; thus, the sample sizes of dry periods and rainy periods for a 0.1 mm threshold are high enough so that differences in humidity profiles or convection between the two categories might be statistically robust. Results in later sections have some sensitivity to the threshold used, and this sensitivity is further explored in Appendix A.

[22] Convection prior to LCE3 mostly consisted of shallow cumulus and some taller isolated convective elements and echo clusters through 7 December. An MCS passed westward over and north of S-PolKa on 8 December; a review of geosynchronous satellite imagery reveals that this MCS was an isolated, local event at the time and was not associated with an MJO LCE. Widespread dry conditions persisted before and after the 8 December MCS. With the exception of the 8 December event, precipitation amounts remained low until 15 December, when several squall lines developed in and/or passed through the radar domain. Some large echo clusters and limited stratiform precipitation were observed during the second half of December. A domain average of more than 7 mm d$^{-1}$ in precipitation was estimated each day between 19 December and 24 December. During this period, squall lines with deep convective cores frequently developed near the radar in proximity to large MCSs that mostly remained outside of the radar domain. As such, the highest precipitation amounts likely occurred outside the range of S-PolKa, and this at least partially explains why less rainfall was observed during LCE3 than during LCE1 and LCE2.

5. Radar-Derived Statistics of Precipitating Clouds

[23] The continuous time series of radar data contains high-frequency variability superimposed on the MJO dynamics. Simply filtering the data set for variability that corresponds with the traditionally accepted frequency of the MJO —20–100 days—yields a smooth, sinusoidal structure of the filtered variable and thus may fail to capture abrupt changes in the cloud population that are related to MJO development. In other words, filtering on the MJO time scale aliases the higher-frequency variability MJO time scale in ways that can be misleading to any interpretation of MJO behavior. Also, the MJO is not a true wave. It is a statistical construct made up of Kelvin and Rossby wave components [Nogués-Paegle et al., 1989; Rui and Wang, 1990; Houze et al., 2000], and the deep convection during its active period at any location interacts with and modifies the large-scale wave fields with a source of latent heating, which is tied to atmospheric and oceanic thermodynamics. Any given realization of the MJO contains these large-scale wave components and convection, and it is further modulated by higher frequency phenomena, in different combinations. To interpret any specific MJO event, its individual wave and convective components must be recognized and taken into account. Therefore, to characterize the nature of the convection relative to the MJO on various time scales, we separate the radar data into discrete intervals of time and derive the statistics of convection during each such period. Our intervals are 3 days in temporal width—enough to smooth over high-frequency variability in the cloud population of 2 days or less but short enough to capture changes that occur over as little as 3 days. Each interval lags—or is lagged by—a Day 0 that is defined relative to a maximum in precipitation determined by Fourier filtering the radar-derived precipitation for 20–60 day variability. The filtered time series yields peaks that occur during each of LCE1, LCE2, and LCE3. Each peak represents Day 0 for that MJO-related LCE; they occur on 22 October, 22 November, and 19 December for LCE1, LCE2, and LCE3, respectively. (Because of the filtering, Day 0 may not actually represent the day on which the most precipitation was observed during the LCE.) The cloud population is then studied within each of the following intervals relative to Day 0: −16 to −14 days, −13 to −11 days, −10 to −8 days, −7 to −5 days, −4 to −2 days, −1 to +1 days, +2 to +4 days, +5 to +7 days, +8 to +10 days, +11 to +13 days, and +14 to +16 days.

[24] One common method of indexing the MJO in terms of environmental parameters is that introduced by Wheeler and Hendon [2004; hereafter WH], who utilized global fields of outgoing longwave radiation and zonal wind anomalies at 200 and 850 hPa. The first two EOFs of these fields are used...
to provide an index for the MJO; the index describes the MJO as if it were a wave consisting of eight different phases. The phase is determined in real time based on the horizontal distributions of the anomalies and their projections onto different linear combinations of the first two EOFs for each field. Any statistics presented in terms of the WH MJO phase simply document the variable of interest at one location given a known structure of global OLR and zonal wind anomalies. As noted above, this approach imposes a wavelike interpretation that obscures the higher frequency variability within an MJO phase. Furthermore, one should employ caution when considering the temporal variation of a quantity in terms of MJO index because the duration of each phase may differ between MJO events and from the duration of other phases within a single MJO event. For these reasons, we choose not to composite our data by WH MJO phase for purposes of evaluating the evolution of atmospheric variables. However, for the reader’s reference, Table 1 provides the WH MJO phase for each date during DYNAMO. Each phase is also marked on the upper axis of Figure 2 and is noted for comparison with our approach in Figure 4.

[25] Figure 4 illustrates the fraction of area within the S-PolKa domain that was classified as either convective or stratiform during each WH MJO phase and during each lag interval. For the latter, results are presented for a composite of the three events and for each individual event. When compositing by WH MJO phase (Figure 4a), a peak in stratiform areal coverage occurs during phases 1 and 2, with an apparent rapid increase in stratiform radar echo between phases 8 and 1, which is consistent with Barnes and Houze’s [2013] analysis of 14 years of TRMM satellite precipitation radar data. However, since the time period of each phase differs for each LCE, such a composite yields little about the actual time required for the increase in stratiform precipitation to occur. When compositing by lag interval for stratiform echoes (Figure 4b), a peak in stratiform echo area occurs at +2 to +4 days. The areal coverage of stratiform radar echo appears to increase steadily at about the same rate for 2 weeks prior to the maximum when all three LCEs are combined into a composite.

[26] However, the duration of each LCE seen in Figure 2 is different, and MJO onset need not occur at the same time relative to Day 0 for each LCE. During LCE1, a rapid increase in stratiform radar echo occurred between −10 to −8 and −7 to −5 days, and after a decrease in stratiform areal coverage, another increase occurred between −1 to +1

<table>
<thead>
<tr>
<th>Phase</th>
<th>Dates</th>
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<tbody>
<tr>
<td>1</td>
<td>15–19 October, 17–20 November</td>
</tr>
<tr>
<td>2</td>
<td>20–29 October, 21–25 November</td>
</tr>
<tr>
<td>3</td>
<td>30 October to 1 November, 26–30 November</td>
</tr>
<tr>
<td>4</td>
<td>2–5 November, 1–5 December, 13–24 December</td>
</tr>
<tr>
<td>5</td>
<td>6–8 November, 6–12 December, 25–28 December</td>
</tr>
<tr>
<td>6</td>
<td>1–4 October, 9 November, 29 December to 7 January, 10–16 January</td>
</tr>
<tr>
<td>7</td>
<td>5–8 October, 10–12 November, 8–9 January</td>
</tr>
<tr>
<td>8</td>
<td>9–14 October, 13–16 November</td>
</tr>
</tbody>
</table>

Table 1. Dates of Occurrence of Each WH MJO Phase During DYNAMO-AMIE

<table>
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<tr>
<th>Composite (time)</th>
<th>Fraction of radar domain</th>
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<tr>
<td>−15 −12 −09 −06 −03 00 +03 +06 +09 +12 +15</td>
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</tr>
<tr>
<td>0.25 0.20 0.15 0.10 0.05 0.00</td>
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Figure 4. (a) Mean fraction of radar domain occupied by either stratiform or convective precipitating echoes during each phase of the MJO as described by WH. (b) Same as Figure 4a but composited in 3 day intervals relative to Day 0 instead of by WH MJO phase for stratiform echoes during each LCE and composited among the three LCEs. (c) Same as Figure 4b but for convective echoes.
and +2 to +4 days. The separate increases may have been associated with passages of equatorial Kelvin waves [Gottschalck et al., 2013, Figure 13]. During LCE2 and LCE3, a similar increase was observed, but it occurred between −7 to −5 days and −1 to +1 days. A maximum in stratiform echo at −13 to −11 days during LCE3 was associated with an MCS that passed over S-PolKa on 8–9 December and was not associated with an MJO (section 4). Sharp increases in stratiform areal coverage were observed during each LCE; however, compositing the three events together around a precipitation maximum effectively smoothes out the rapid increases and prevents the detection of such changes. We could, alternatively, composite our results relative to the observed rapid increase in stratiform areal coverage. We would then preserve the rapid increase in stratiform in the composite, but we would then introduce an unrealistically gradual decrease in the stratiform areal coverage at the end of a LCE, which was not observed during any LCE. Many studies, like those mentioned in section 1, composite atmospheric variables relative to a precipitation maximum or OLR minimum at some location. This procedure smoothes out sharp increases and/or decreases in variables that might change rapidly near MJO onset during any single case. Our results thus underscore the importance of studying MJO onset in terms of each individual LCE rather than a composite of several events.

[27] We also note from our results that the expansiveness of widespread stratiform rain dominated the variability in areal coverage of precipitation echoes on a roughly 30 day time scale. Such an observation is an obvious one since stratiform regions are generally much larger than their parent convective cores, but we make the point here because the variability in the stratiform component has important implications on the tropospheric diabatic heating profile, which in turn, affects the anomalous circulation associated with an MJO. Additionally, large areas of stratiform enhance cloud radiative feedback, which have been suggested, but not proven, to have an effect on MJO maintenance and propagation [e.g., Raymond, 2001; Bony and Emanuel, 2005; Ma and Kang, 2011; Kim et al., 2013; Wang et al., 2013]. Generally, extensive stratiform precipitation areas develop in association with the deepest convective cores [Houze, 2004]. Deep convective clouds were detected during active and suppressed MJO conditions; however, the areal coverage of convective cells was greatest within 2–4 days of Day 0 during each LCE (Figure 4c), and areal coverage of convective echoes does not always increase as sharply as the areal coverage of stratiform echoes. Particularly during LCE1, a gradual increase in the areal coverage in convective echo is observed.

[28] To study how the depth of convection changed in time at Addu Atoll, we examine top heights of S-PolKa radar echoes classified as convective. Figure 5 is a time series of the probability distribution function (PDF) of 20 dBZ echo top height, which is simply referred to as “echo top” through much of the remainder of the paper, for convective echoes only (section 3.1). The 20 dBZ threshold occasionally extended above 10 km, and depending on their heights, these echoes generally signify convection producing moderate surface rainfall or containing small graupel (shown by applying the polarimetric particle identification algorithm of Vivekanandan et al. [1999] to the S-PolKa data) at altitudes above about 4 km. Hourly data were smoothed by averaging within 3-hourly intervals to match the temporal resolution of rawinsonde data, and the PDF accumulated over each 3 h interval has been normalized to 1. Yellows and reds indicate the height at which an echo top is most likely to be observed. The modal distribution of echo top height peaks near 8 km during rainy periods in LCE1 and LCE2, while the modal distribution decreases to between 4 and 6 km between LCEs. The 2 day (4 to 6 day) periodicity in modal distribution of convective echo tops seen in Figure 5 corresponds to variability in precipitation during LCE1 (LCE2) seen in Figure 2.

[29] The date of Day 0 for each LCE is marked along the upper axis of Figure 5. Figure 6a composites PDFs of convective echo tops represented in Figure 5 by lag interval relative to each Day 0. Shallow boundary layer cumulus and deep cumulonimbus were present throughout the IOP, although deep convective echoes, as expected, were more common during a LCE. On average, during the LCEs and between −4 and +4 days, the PDF peaks near 7.5 km, and 33 (12%) of echo tops were higher than 7 (8) km. During inactive periods between 14 and 16 days on either side of Day 0, the PDF peaks between 4 and 5.5 km and 11 (5%) of echo tops were higher than 7 (8) km.
periods come deeper during precipitation maxima. Therefore, the 20 dBZ convective echo tops maximize. The echoes also be- and reaches a maximum, the areal coverage and number of 6a, we observe that when convective precipitation increases during other times. When combined with Figures 4c and 6, we observe that when convective precipitation increases and reaches a maximum, the areal coverage and number of 20 dBZ convective echo tops maximizes. The echoes also become deeper during precipitation maxima. Therefore, the increased amount of convective precipitation near Day 0 occurs not primarily because individual convective elements precipitate more but rather because more convective ele- ments or more widespread convective echoes are present. This fact is consistent with Zuluaga and Houze’s [2013] finding that within the ~2–4 day precipitation episodes when most rain falls during a LCE, deep and wide convective cores maximize in number. This behavior suggests some scale similarity in the behavior of convective elements between the higher-frequency episodes and the events on the spatial and temporal scale of the MJO.

[30] We have established that onset of MJO-related con- vention over the Indian Ocean coincides with large and fairly sudden increases in stratiform precipitation and modal depth of convective cores. To gain some insight on the relationship of humidification to convective cell depth, and thus poten- tially MJO development, we divide the PDF of echo top height during the active phases into rainy periods and dry periods (recall definitions in section 4). The PDFs for these categories and for WH phases 8–3 and 4–7 are seen in Figure 6c. PDFs for these combinations of WH phases are made because LCE1 and LCE2 occur during WH phases 8 through 3 (Figure 2). Because LCE3 was not as well sampled, the PDFs do not include days after 12 December. The PDF during rainy periods closely resembles that of WH phases 8–3, mainly because most of the echoes observed during an active LCE occur during rainy periods. Meanwhile, the PDF for dry periods during the active phases peaks around 5 km and looks more similar to the echo top height distribution during the inactive WH phases 4–7; it is even a little lower. The difference between the PDFs for rainy pe- riods and dry periods is statistically significant (Appendix B). Thus, on some days during a LCE, convective echo tops were distributed as if the environment was suppressed even though MJO onset had already occurred.

6. Tropospheric Humidification Observed in Rawinsonde Data

[31] While S-PolKa gives us detailed information about the three-dimensional structure of convection during the IOP, it provides little information about the local or large-scale hu- midification and drying of the troposphere associated with the MJO. In order to examine the relationship between cloud development and humidity, we examine the data obtained by rawinsondes launched from Gan Island (section 2). The 4 month length of the data set allows for documentation of the intraseasonal variability, and the 3 h frequency of the launches permits detection of high-frequency variations of wind direction, temperature, and humidity. Figure 7 shows anomalies of zonal wind (u′), meridional wind (v′), tempera- ture (T′), and anomalous specific humidity divided by its time mean (q′/s, or “fractional difference” in text). Anomalies were computed by first manually quality controlling the data set for obvious inaccuracies. After questionable or bad data were removed, missing observations were filled in by linearly interpo- lating between the nearest available measurements. The data were then smoothed vertically into 5 hPa bins. The time mean was then computed for each bin over the entire period of the data set, and anomalies were computed by subtracting the time mean at each pressure level from the interpolated data. Anomalies of specific humidity are given as fractional.
differences from the time mean in order to highlight changes in the humidity in the upper troposphere, where the absolute change in humidity was small.

Figure 7d shows that during three distinct periods—one each in October, November, and December—anomalously high values of humidity occurred throughout most of the troposphere, particularly between 700 hPa and 200 hPa. The events of moistening were separated by periods during which a deep layer of anomalously dry air was present. The dry and moist periods were similar to those seen during TOGA COARE in the western Pacific [Brown and Zhang, 1997]. Above the 500 hPa level, humidity increased slightly during the second half of January; however, very dry air dominated the entire tropospheric column over Gan during most of January. Slight positive anomalies in temperature were concurrent with positive moisture anomalies, especially above 500 hPa, consistent with previous studies [Hendon and Salby, 1994; Yanai et al., 2000; Kiladis et al., 2005]. A transition from westerly to easterly zonal wind anomalies in the TTL occurred simultaneously with positive humidity and temperature anomalies. No variability on a similar time scale is noted in the $v'$ field.

Of major interest for detection of intraseasonal variability related to the MJO are changes in zonal wind, temperature, or humidity on time scales of roughly 20–60 days. As mentioned in section 4, a continuous time series of data reveals many signals of higher frequency than that of the MJO. Figure 8a is the same as Figure 5, except that it is smoothed to a 72 h interval to eliminate very high frequency signal such as diurnal variability and waves of 2 day frequency. The modal distribution of the smoothed time series, depicted by the red line, peaks during each LCE and increases from 5–6 km to about 8 km over 3 days near the beginning times of LCE1 and LCE2. Two such sudden increases in the modal distribution are observed during the onset of LCE3. Figure 8b is the time series of $q^*$ smoothed to 72 h intervals. The Eulerian derivative of $q^*$, when positive, is plotted in gray contours to illustrate the time scale of moistening. The first striking feature is that humidification through the troposphere occurred more quickly prior to MJO onset than prescribed by the discharge-recharge hypothesis. For LCE1, moistening began between 850 and 700 hPa after a maximum dry anomaly on 8 October. The first moist anomaly between 850 and 700 hPa was seen on 11 October. By 15 October, a moist anomaly extended vertically to 200 hPa upon arrival of the first MCS associated with LCE1 near S-PolKa. Convective echo tops increased rapidly between 14 and 16 October, near the end of the moistening period. On 13 November, anomalous humidity was observed as high as 250 hPa after small echo clusters passed over Addu Atoll; however, drying occurred above 850 hPa during the subsequent 3 days. Some moistening continued between 850 hPa and 700 hPa during this time, and although humidity became anomalously low above 500 hPa on 14–16 November, a rapid rise in convective echo tops was seen at that time. Then rapid humidification of the troposphere above 600 hPa was observed.

Figure 7. Anomalies of (a) zonal wind ($u'$), (b) meridional wind ($v'$), (c) temperature ($T'$), and (d) fractional difference of specific humidity from the mean ($q^*$). All are computed with rawinsonde data from sondes launched from Gan Island. All anomalies are calculated relative to the mean value from 1 October 2011 to 9 February 2012. Data have been interpolated temporally to fill in for missing soundings or bad data, and vertical resolution of the data after smoothing is 5 hPa. The dates denoted on the top axis represent the Day 0 associated with each large-scale convective event. Note that the color scale is reversed in Figure 7d.
during the following 3 days. The 8–9 December MCS moistened the troposphere as high as 550 hPa. Drying occurred throughout the troposphere from 10 to 13 December, and an increase in convective echo tops between 13 and 16 December was concurrent with rapid moistening between 700 to 300 hPa during the same period. Thus, the increase in convective echo top heights occurred at different times relative to moistening in the low to middle troposphere. During LCE1 and LCE2, low-level moistening preceded convective echo top increases. During LCE3, the increase in low-level moisture and convective echo top heights are nearly simultaneous, though a moist anomaly was present at 850 hPa for several days prior to the increase. During the three LCEs, mid-level moisture increased before (during LCE1), after (during LCE2), and at the same time (during LCE3) as the increase in convective echo tops. During each LCE, additional moistening occurred in the upper troposphere above 500 hPa for several days after the occurrence of the first MCS observed by S-PolKa.

**Figure 8.** (a) Same as Figure 5 but smoothed to 72 h intervals. The solid red line follows the modal distribution of echo top height. (b) Time series of $q^*$ smoothed to 72 h intervals. The Eulerian derivative of $q^*$ is shown in gray contours only where positive. The solid black line is the same as the red line in Figure 5a. The dates denoted on the top axis represent the Day 0 associated with each large-scale convective event.

**Figure 9.** Composite median relative humidity vertical profiles computed from rawinsonde data at Gan Island. The solid (dashed) black line is the relative humidity profile for MCS days during MJO active (inactive) phases 8–3 (4–7). The blue and red lines show the relative humidity profiles, respectively, for rainy periods and dry periods.
Table 2. Maximum Lagged Cross-Correlation Coefficients and the Lag (in Days) at Which They Occur (in Parentheses) for Filtered Specific Humidity Anomaly Time Series at 850 hPa, 700 hPa, 500 hPa, and 300 hPa, as well as for the Filtered Time Series of Stratiform and Convective Areal Coverage

<table>
<thead>
<tr>
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<th>Convective</th>
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<th>q_700′</th>
<th>q_500′</th>
<th>q_300′</th>
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<td>q_850′</td>
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<td>q_500′</td>
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<td>0.90 (−0.5)</td>
<td>0.76 (−1.5)</td>
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<td></td>
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<tr>
<td>q_300′</td>
<td>0.85 (+4)</td>
<td>0.94 (−2.125)</td>
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</tr>
<tr>
<td>Stratiform</td>
<td>0.92 (+2.125)</td>
<td></td>
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</tr>
</tbody>
</table>

*Positive lags indicate that the quantity listed in that column occurs first. Because of the small sample size, none of the correlations are statistically significant.

We may also use the rawinsonde data set to gain more insight into the relationship between tropospheric humidity and echo top PDFs of the S-PolKa observations. Figure 9 contains median relative humidity (RH) profiles for the rawinsonde data that correspond to the time periods composited in Figure 6c. All profiles are remarkably consistent below 925 hPa; this is indicative of the persistently warm, moist marine boundary layer. The remainder of this discussion will refer to portions of the RH profiles above 925 hPa. Not surprisingly, the RH profile during rainy periods closely parallels the RH profile composited over phases 3–8 and is about 2 to 5% (absolute change in RH) greater below 400 hPa. The RH profile during dry periods is close to, and even 1–3% less than that during phases 4–7 up to 800 hPa. Above 800 hPa, the RH profile during dry periods is between the profiles for phases 4–7 and phases 8–3, and it parallels the profile for phases 8–3 while remaining 5–10% lower. Thus, RH during dry periods at levels between 850 hPa and 400 hPa is typically 10–15% lower than during rainy periods. Because of the small sample sizes involved and the temporal autocorrelation of the RH time series, none of the profiles are statistically different at any level using a Mann-Whitney U test. Nonetheless, in Figure 6c, we saw that convective echo top heights were significantly lower during dry periods than during rainy periods. Here we see that the humidity profile for dry periods is also lower, though it is moister than the profile during MJO inactive phases through much of the troposphere. That RH in the lower troposphere during dry periods is close to that during phases 4–7 may have a physically meaningful explanation. Prior studies [e.g., Muller et al., 2009; Wang and Sobel, 2012] suggest that low-level moisture may have some control on precipitation. Decreased moisture in the lower troposphere during inactive MJO conditions, or during dry periods within active MJO conditions, could restrict the amount and depth of convection that forms. At the same time, the humidity profiles during these periods may simply be lower because fewer clouds are present. Thus, we are motivated to further investigate the temporal relationship between convection and environmental humidity.

7. Lag-Correlation Analysis of Precipitation Echo and Tropospheric Humidity

7.1. The 20–60 Day Filtered Time Series

A slew of studies referenced herein [Hendon and Salby, 1994; Maloney and Hartmann, 1998; Kemball-Cook and Weare, 2001; Kiladis et al., 2005; Benedict and Randall, 2007] and many others have used a band-pass filtered time series of atmospheric variables, such as OLR or humidity, to determine the relationship relevant on the time scale of the MJO between those and other variables. Such methodology is appropriate if the variables of interest are known to evolve on the time scale for which they are filtered. These studies generally show a gradual buildup of moisture prior to onset of convection. Regardless of the time scale of moisture buildup, the low-level humidity increases prior to an increase in convection in a time series filtered for MJO-variability; thus, prior observational, reanalysis, and modeling studies have concluded that the low-level moisture increase is critical for MJO onset. While moisture buildup may be necessary in the case of onset of a LCE downstream from the region of initial MJO convective onset, our results show that the time scale of moistening and convective buildup prior to MJO onset is less than the traditional 10–100 day frequency used in band-pass filtering. Thus, we have no reason to expect that a band-pass filtered time series will accurately describe the

Table 3. Maximum Lagged Cross-Correlation Coefficients (With Lag in Hours in Parentheses) Between Convective/Stratiform Areal Coverage and Unfiltered, Unsmoothed Specific Humidity Anomalies for 1 October to 15 January Using Various Smoothing Periods

<table>
<thead>
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<tr>
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</tr>
<tr>
<td>Conv. 500′</td>
<td>0.49 (+6)</td>
</tr>
<tr>
<td>Conv. 600′</td>
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<tr>
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</tr>
<tr>
<td>Strat. 400′</td>
<td>0.45 (−3)</td>
</tr>
<tr>
<td>Strat. 500′</td>
<td>0.55 (+3)</td>
</tr>
<tr>
<td>Strat. 600′</td>
<td>0.52 (+3)</td>
</tr>
<tr>
<td>Conv. Strat.</td>
<td>0.81 (+3)</td>
</tr>
</tbody>
</table>

*All correlation values that are in bold are statistically significant at the 95% level. Variables correlated are shown in columns 1 and 2. Positive lags indicate that Variable 1 comes first. (Conv = Convective areaal coverage; Strat = Stratiform areaal coverage).
relationship between humidity and convection prior to and during MJO onset. Instead, we are specifically interested in the variability that projects onto frequencies of less than 10 days, which is the signal that is lost by many prior studies through filtering.

[36] To demonstrate the effect of filtering DYNAMO data, we first examine what happens when we filter the time series of humidity anomalies and convective/stratiform areal coverage with a simple 20–60 day band-pass Fourier filter. Table 2 shows the lag-correlation analysis for the filtered time series. The correlation coefficient is shown for the lag during which two time series are most correlated. A few results are notable, though because of the small sample size of MJO events sampled, none of the results are statistically significant. First, in the lower troposphere below 700 hPa, moistening begins prior to an increase in stratiform (convective) echo area by up to 4 (2) days. Second, the humidification at 300 hPa lags the humidification at 850 hPa by about 5 days. Finally, we note that the maxima in 500 and 300 hPa humidity occur after the peak of stratiform areal coverage. The lag correlations in Table 2 suggest that at the beginning of an MJO LCE, the number of convective echoes begins to increase after a couple of days of low-level moistening. The convection then widens and becomes characterized by large stratiform regions that moisten the upper troposphere.

7.2. Unfiltered and Smoothed Time Series

[37] Table 3 contains maximum lagged correlations of the same variables using their unfiltered time series. We have smoothed the time series with various intervals so that we remove very high-frequency variability but preserve the variability that appears to be important for moistening and convective buildup prior to MJO onset, just as we did in section 6 for Figure 8. As we increase the smoothing, we remove additional high-frequency variability (i.e., diurnal frequency) or apparently random variability at the site that is not representative of the large-scale humidity field. Such a procedure generally yields higher correlation values at the expense of the time series length and thus the statistical significance of the correlations. Note that the lags included in Table 3 can best be thought of as intervals. For example, when using smoothing over 24 h, the width of a single unit of time is 24 h. Thus, a lag of 0 only implies that the two variables lag each other by something between −12 h to +12 h.

[38] Several statistically robust results are shown in Table 3. First, most variables correlate well with each other at lag periods of less than a day regardless of the smoothing used. Second, the areal coverage of convective and stratiform elements are highly correlated with specific humidity anomalies, and convective elements, as expected, lead stratiform elements by a few hours. Third, upper tropospheric moistening occurs at near the same time or slightly after stratiform areal coverage increases. Moistening above 500 hPa occurs about 3 h after stratiform areal coverage increases. Finally, the areal coverage of convective echoes increases prior to humidity anomalies throughout the troposphere. When no smoothing is used, convective areal coverage precedes humidification at 700 (300) hPa by about 3 (9) h. The latter two results suggest that convective elements are at least partially responsible for moistening the lower troposphere, and moistening of the upper troposphere is also likely due to the presence of increased cloudiness, particularly stratiform regions and subsequently occurring anvil clouds. Riley et al. [2011] and Del Genio et al. [2012] also found that high-altitude ice anvil clouds were most prevalent after a widespread stratiform event, and Masunaga [2013] shows that the peak humidity in the upper troposphere occurs after the maximum in cloud cover associated with an MCS. The persistence of an upper tropospheric moist anomaly after an active LCE ends (Figures 7 and 8) further suggests that clouds play a vital role in moistening the upper troposphere. Moist anomalies above 500 (300) hPa persisted for a few days after low-level moist anomalies gave way to drier conditions during LCE1 and LCE2 (LCE3).

[39] Combined with other findings from this study, the lag correlations just described make a case that convection with low-level to mid-level tops contributes significantly to the moistening observed at Addu Atoll prior to onset of a LCE. A review of Figure 8b reveals that moistening, at least during LCE1 and LCE2, occurred at levels near 850 hPa before convective echo tops increased. Relative to the respective Day 0 for the filtered precipitation time series for LCE1 and LCE2, the moistening occurred primarily during the lag intervals of −13 to −11 and −10 to −8 days, or about 2–7 days prior to the buildup in convective echo top height. We see in Figure 6b that the number of 20 dBZ convective echo tops at 5 km or lower increases during these lag intervals. Furthermore, Figure 4c shows that the areal coverage (or number) of convective echoes at 2.5 km increases at −13 to −11 days for both LCEs, and it continues to increase for LCE1 after that interval. All of this evidence strongly supports the notion that prior to LCE1 and LCE2, the number of convective echoes below 5 km increased prior to an increase in the modal depth of convection and prior to an increase in stratiform areal coverage. Very likely, more numerous and widespread convective elements can detrain more water into the lower troposphere, and this can explain why humidity throughout the troposphere lags behind convective area coverage.

8. Conclusions

[40] This study is unique because never before has a powerful S-band radar system such as S-PolKa been located and operated continuously for several months where MJO-related convection is known to first appear before propagating eastward. The DYNAMO/AMIE radar operation in the central equatorial Indian Ocean documented the cloud population over a 3.5 month period with the S-PolKa radar system nearly colocated with 3-hourly rawinsondes.

[41] Three distinct 1 to 2 week long large-scale convective events (LCES) occurred at Addu during the intensive observing period. Although clouds of all depths were present during active and suppressed periods, the variability in convection associated with the MJO was dominated by the variability in the areal coverage of cloud systems exhibiting deep precipitating stratiform areas, as has been noted previously by Zuluaga and Houze [2013] and Barnes and Houze [2013]. Large echo clusters and MCs containing large areas of stratiform precipitation contributed over one third of the total precipitation of the cloud populations occurring during LCES, while suppressed periods featured isolated convective cells and some echo clusters with small amounts of stratiform precipitation. Since the stratiform components of the cloud
systems originate as convective cores, the increased stratiform rain during LCEs indicates a temporary maximum in the upward mass flux of water within deep convection during active MJO periods. Whether this transport was due simply to the greater number of convective cores present or occurred because stronger updrafts transport more water vertically remains to be determined but is not important when considering the net latent heat release, which depends only on the net vertical transport. More efficient detrainment from convective elements can also more easily allow for the formation of widespread stratiform regions. The greater proportion of stratiform rain during LCEs thus implies [Houze, 1982, 1989] that the heating profile is more top heavy (i.e., has a maximum in the upper troposphere) during rainy portions of LCEs.

[42] Several observations indicate that the cloud population and humidity field change rapidly near the beginning of a LCE. The areal coverage of stratiform precipitation increases rapidly over about 3–7 days as the primary type of convection making up the cloud population shifts from isolated convective clouds and medium-sized precipitating clusters to large MCSs. A rapid increase in the modal distribution of convective echo top height from 5–6 km to 8 km closely corresponds in time to humidity increases in the portion of the troposphere lying above 850 hPa that occur a week or less prior to the increase in modal echo top height, if prior at all. Little variation in humidity occurs below 850 hPa; a warm, moist marine boundary layer is observed at all times. These results contradict the proposed time scale for the discharge-recharge mechanism for MJO onset, which describes the vertical buildup of moist static energy as occurring gradually over 10 to 20 days. We have shown that results consistent with the time scale for the discharge-recharge hypothesis can be obtained by compositing our results over all three LCEs because sharp changes in humidity and the convective cloud population are smoothed out. Results supporting the 10–20 day time scale of discharge-recharge do not arise when examining individual realizations. No gradual “recharging” of humidity or convective depth is actually observed. This paper highlights the importance of investigating MJO onset on a case study basis.

[43] We also have documented the relationship between the buildup of humidity through the depth of the troposphere and the associated buildup of convection. When filtered for 20–60 day variability, humidity appears to increase a couple of days before convective areal coverage increases. Stratiform echoes are observed a couple more days later, followed by moistening of the upper troposphere. However, the time scale for changes in humidity and convection prior to MJO onset is shown by this study to be less than 10 days. The filtered time series lose information about high-frequency events and sharp changes in either field that are critical in describing how MJO onset occurs during each individual LCE. When we do not filter for the 20–60 day variability, several results from this study suggest that clouds moisten the lower troposphere prior to a buildup in convective echo top associated with MJO onset:

[44] 1. Humidity anomalies above 850 hPa lag convective areal coverage by less than 1 day and perhaps even only a few hours. Anomalies in convective areal coverage do not lag behind humidity anomalies.

[45] 2. The number of convective echoes in the lower troposphere and the number of 20 dBZ echo tops observed below 500 hPa both increase during the period that moistening occurs in the lower troposphere prior to MJO onset.

[46] 3. The composite relative humidity profile for dry periods during an MJO active phase is very similar to the relative humidity profile during MJO inactive phases below 800 hPa. The probability distribution functions of 20 dBZ echo top heights during these two times are also very similar.

[47] Additionally, we show evidence that stratiform and anvil elements contribute to moistening in the upper troposphere because humidity anomalies in the upper troposphere lag stratiform areal coverage. Upper tropospheric humidity anomalies caused by presence of anvil cloud persist for a few days after a large-scale convective event ends.

[48] We cannot draw any conclusions about the effects of low-level and mid-level moisture on the eventual increase in convective echo top and the onset of a LCE associated with development of an MJO. During one LCE, increases in lower tropospheric humidity occurred at the same time as convective depth increased. Also, a quick look at the smoothed, unfiltered time series of humidity (Figure 8b) shows that mid-level moistening occurred before, after, and concurrently with, the observed rapid buildup of convective echo tops associated with LCE1, LCE2, and LCE3, respectively. Additionally, we have not explored potential effects of large-scale convergence or advection of moisture into the region prior to convective buildup [e.g., Maloney, 2009], which could also contribute to the increase in low-level moisture, though we have shown that humidity anomalies did not appear to precede anomalies in convective areal coverage in the cases considered here. We have not proven or disproven that an increase in low-level moisture is necessary for convection to grow deeper; the time scale for such an increase need not be 10–20 days but rather only a few days. The convective buildup observed is probably related to the “building block” hypothesis proposed by Mapes et al. [2006], which essentially proposes that similar convective lifecycles are observed in association with the dynamics of waves or disturbances of different periodicities. The lifecycle is stretched temporally based on the periodicity of the disturbance affecting the convection. In the MJO cases seen in our paper, a transition from shallow to deep and widespread convection is observed over a few days. Such a transition is similar to the evolution of the cloud population during events with even shorter time scales, such as mesoscale convective systems. Zuluaga and Houze [2013] further detail and support such a claim. We have also briefly shown that changes in the upper level zonal wind and temperature anomalies occur on the same frequency as the MJO events occur. This concurrence indicates that upper level dynamics may also have an impact on widespread, organized convection—a topic that needs to be explored in a future study.

[49] Our current description of convection related to the MJO is not intended to fully explain the mechanisms responsible for onset and propagation of convection but rather provide some detail on the relationship between convection and tropospheric humidity leading up to MJO onset. While the DYNAMO/AMIE data analyzed herein do not support the discharge-recharge hypothesis, we have also not yet explained definitively why clouds generally grow taller in the 3 to 7 days prior to MJO onset. Also, the current analysis only examines humidity and convection within a small sample...
domain located within a much larger area in which MJO onset occurs. The vast DYNAMO/AMIE data set includes not only instrumentation used in this paper but also an array of precipitation and cloud radars over the Indian Ocean and tropical west Pacific that provide information on variability of the cloud population in other regions. Use of a broader set of deployed instruments, reanalysis, and satellite data should provide more understanding of the threedimensional processes, potentially including upper tropospheric dynamics, involved in MJO onset. Future numerical simulations of cloud systems can also be anchored to the observational data set and provide insight on the relative roles of various processes that control the intraseasonal variability in tropospheric moisture and convection. Such experiments will yield detailed water budgets that describe the transfer of water between convective clouds and their anvils, which act as agents for moistening the troposphere, and between clouds and the surrounding environment.

Appendix A: Sensitivity of Quantities Derived for “Rainy” and “Dry” Periods to Precipitation Threshold

The selection of a rainfall threshold to separate “rainy” and “dry” periods during an active LCE is a somewhat subjective process. If the threshold is too low (i.e., 0 mm), then an insufficient amount of the time series will be classified as a dry period to make any meaningful statistical comparison of echo top PDFs during rainy and dry periods. If the threshold is too high, then dry periods are biased wet by times when a significant amount of rainfall is falling over some portion of the radar domain. An hourly domain-averaged precipitation estimate of 0.25 mm (or 6 mm d\(^{-1}\)) typically occurs during the beginning or end of a short-term (1 to 2 day long) precipitation event (Figure 2), and the precipitation echo during such times is classified either by many isolated convective echo clusters or stratiform in some part of the radar domain. Thus, any threshold above 0.25 mm is likely too high. Figure A1 shows two panels: They are duplicates of Figures 6c and 9, except using a rainfall threshold of 0.25 mm. The peak of the PDF for rainy periods in Figure A1a is at the same location as that in Figure 6c, though the probability is about 0.01 higher. The PDF for dry periods in Figure A1a is shifted upward by 0.5 km or less, which is less than the resolution of the interpolated radar data set. Figure A1b reveals RH profiles for rainy and dry periods using a threshold of 0.25 mm. Both profiles are shifted toward moister conditions than in Figure 9, but the difference in RH between 850 hPa and 500 hPa remains about 10–15%. Additionally, the dry period RH profile maintains its shape and remains between the profiles for phases 8–3 and phases 4–7. The RH profiles below 800 hPa for dry periods and phases 4–7 remain nearly identical. Thus, we determine that our main conclusions are not unduly influenced by the choice of rainfall threshold given in separating times during an active LCE into rainy and dry periods.

Appendix B: Testing Statistical Significance for Echo Top PDFs

Statistical significance among differences in echo top PDFs described in section 5 and RH profiles shown in section 6 is determined using a Mann-Whitney \(U\) test with a 95% confidence level. For echo top PDFs, each separate contiguous echo object observed by S-PolKa theoretically represents a separate degree of freedom (DOF). Determining the exact DOF would thus require a complicated radar echo tracking algorithm or determining spatial and temporal autocorrelation of radar data at each data point for 3.5 months. Instead we make an unrealistically conservative assumption that very few individual echo objects exist. For a rainfall threshold of 0.1 mm, the number of convective echoes detected during rainy periods is greater than that observed during dry periods by a factor of about 12. Suppose we assume that only one new contiguous convective echo is observed each hour and we assume that 12 times that amount is observed during wet conditions. The DOF for dry periods in Figure 6c is equal to the number of hours classified as falling within a dry period, or 133; and the DOF for rainy periods is 12*(number of hours classified as rainy period), or 12*190 = 2280.

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