Convective Momentum Transport by Rainbands within a Madden–Julian Oscillation in a Global Nonhydrostatic Model with Explicit Deep Convective Processes. Part I: Methodology and General Results

TOMOKI MIYAKAWA* AND YUKARI N. TAKAYABU
Atmosphere and Ocean Research Institute, Kashiwa, Chiba, Japan

TOMOE NASUNO
Research Institute for Global Change, JAMSTEC, Yokohama, Kanagawa, Japan

HIROAKI MIURA AND MASAKI SATOH
Atmosphere and Ocean Research Institute, Kashiwa, Chiba, and Research Institute for Global Change, JAMSTEC, Yokohama, Kanagawa, Japan

MITCHELL W. MONCRIEFF
National Center for Atmospheric Research,* Boulder, Colorado

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ABSTRACT

The convective momentum transport (CMT) properties of 13,215 rainbands within a Madden–Julian oscillation (MJO) event simulated by a global nonhydrostatic model are examined. CMT vectors, which represent horizontal accelerations to the mean winds due to momentum flux convergences of deviation winds, are derived for each rainband. The CMT vectors are composited according to their locations relative to the MJO center.

While a similar number of rainbands are detected in the eastern and western halves of the MJO convective envelope, CMT vectors with large zonal components are most plentiful between 0° and 20° to the west of the MJO center. The zonal components of the CMT vectors exhibit a coherent directionality and have a well-organized three-layer structure: positive near the surface, negative in the low to mid-troposphere, and positive in the upper troposphere. In the low to midtroposphere, where the longitudinal difference in the mean zonal wind across the MJO is 10 m s⁻¹ on average, the net acceleration due to CMT contributes about 216 m s⁻².

Possible roles of the CMT are proposed. First, the CMT delays the eastward progress of the low- to mid-troposphere westerly wind, hence delaying the eastward migration of the convectively favorable region and reducing the propagation speed of the entire MJO. Second, the CMT tilts the MJO flow structure westward with height. Furthermore, the CMT counteracts the momentum transport due to large-scale flows that result from the tilted structure.

1. Introduction

The Madden–Julian oscillation (MJO), first identified by Madden and Julian (1971, 1972), is one of the most significant intraseasonal atmospheric variabilities in the tropics. In the Eastern Hemisphere, which this paper focuses on, the MJO signal appears as a convectively active region that propagates eastward at a phase speed of about
5 m s\(^{-1}\) (Madden and Julian 1994; Hendon and Salby 1994; Zhang 2005; Waliser 2006). This convective envelope has a zonal scale of around 5000 km and consists of cloud clusters of various scales. Multiscale structures exist within the envelope; satellite observations indicate that cloud clusters (CCs) of order 10–100 km comprise eastward-propagating super cloud clusters (SCCs) with scales of several hundred to 1000 km (Nakazawa 1988; Takayabu 1994a,b). Convective organizations (i.e., CCs, SCCs, and MJOs) exhibit similar archetypal structures repeatedly over various scales (Moncrieff 1992, 2004; Mapes et al. 2006). Many papers refer to the scale of CCs as “mesoscale” and SCC as “synoptic scale” (e.g., Nakazawa 1988; Biello et al. 2007), whereas some others (e.g., Houze et al. 2000) refer to the former as “smaller mesoscale” and the latter as “larger mesoscale.” For brevity, in the present study, the term mesoscale will refer to the CC scale and the SCC scale combined. Although recognizing the importance of distinguishing the CC-scale and SCC-scale effects, we limit our scope to the interaction between the mesoscale (i.e., CC scale and SCC scale combined) convection with the MJO-scale ambient flows (defined later as horizontal mean values over 5°-radius circles). Close examination of the CC and SCC scale are left for future studies.

One way to study the multiscale interactions in the MJO is to explicitly represent the cumulus convection with extremely high resolution globally. The required resolution to resolve the cumulus convection is thought to be higher than 1 km (e.g., Khairoutdinov et al. 2009), but it is impossible to conduct a global simulation with such a fine mesh, even with state-of-the-art computational resources. Therefore, alternative approaches are taken, one being the cloud-resolving convective parameterization (CRCP; Grabowski 2001) or “superparameterization” approach described by Randall et al. (2003), in which 2D cloud-resolving models are embedded in each grid cell of a general circulation model (GCM). Unlike most GCMs that implement conventional cumulus parameterizations, Benedict and Randall (2009) produced robust atmospheric variability in intraseasonal space and time (30–60-day) scales by adopting the superparameterization in their GCM. Another approach is to explicitly compute convection globally using available high resolution, currently 3.5 km (7 km in this study), such as in the Nonhydrostatic Icosahedral Atmospheric Model (NICAM; Tomita and Satoh 2004; Satoh et al. 2008). These approaches attempt to resolve the mesoscale organization of convection, which is considered to play essential roles in the tropics (e.g., Houze et al. 2000; Moncrieff 2010). However, these two methods have their respective limitations: two-dimensionality or insufficient resolution to represent the embedded cumulus cells may affect the mesoscale features.

Among various effects of the mesoscale systems, we focus on horizontal accelerations due to momentum flux convergence of wind deviations associated with organized mesoscale convection [traditionally called convective momentum transport (CMT)]. Moncrieff and Klinker (1997) showed in a global model that organized systems transport momentum in a fundamentally different way from cumulus schemes. Mesoscale cloud clusters [or mesoscale convective systems (MCSs)] often consist of linear-shaped, highly convective regions visualized as squall lines with a broad stratiform region (Houze 2004). The width of the convective elements within a typical MCS is on the order of 10 km, which is subgrid-scale for most global models (e.g., GCMs). The width is only about 100–200 km even in an SCC (Fig. 1). Most GCMs either neglect the subgrid momentum transport entirely or parameterize it in a simplified fashion (Zhang and Wu 2003; Wu et al. 2007) as a downgradient momentum mixing process that weakens the vertical shear. However, it has long been known from both observational and model studies that squall lines do not always operate with downgradient momentum transports. LeMone (1983) showed, from aircraft observations, that intense squall lines often produce or reinforce the vertical shear (upgradient) of horizontal momentum in the direction normal to their alignment because of the horizontal gradient of mesoscale pressure anomalies produced by the convective system. Moncrieff (1981, 1992) took a theoretical approach and constructed a 2D nonlinear model based on Lagrangian conservation properties, which quantified the upgradient effects due to momentum flux divergence associated with squall lines. LeMone and Moncrieff (1994) confirmed that CMT represented by the 2D archetypal model of Moncrieff (1992) compared favorably with observational data. Studies using 2D/3D cloud-resolving models also show upgradient effects (e.g., Yang and Houze 1996; Mechem et al. 2006).

Although qualitatively successful attempts have been made to represent the effects in cumulus parameterization (Zhang and Cho 1991a,b; Wu and Yanai 1994; Kershaw and Gregory 1997; Gregory et al. 1997), the role of CMT in the resolved-scale circulation remains obscure. It is difficult to quantitatively evaluate the CMT schemes by comparing with observational data because the momentum equations of the resolved-scale flow consist of resolved-scale terms of large magnitude.

\footnote{In this paper, the term “acceleration” refers to both positive acceleration and negative acceleration (deceleration).}
and in a close balance. Slight errors in the calculation of the resolved-scale terms can result in large errors in the budget residual that are assumed to be due to CMT. Because the momentum tendency terms of the resolved-scale flow are often of comparable magnitude to the CMT effects, uncertainty in the CMT parameterization scheme can lead directly to uncertainty in the momentum tendency.

In regard to MJOs, Tung and Yanai (2002a,b) performed a momentum budget residual analysis using observational data from the Tropical Ocean Global Atmosphere Coupled Ocean–Atmosphere Response Experiment (TOGA COARE). They showed that zonal components of CMT associated with rainbands near the leading edge of the lower-tropospheric westerly winds have a net upgradient effect, whereas the downgradient effects are dominant elsewhere. Lin et al. (2005) performed a similar momentum budget residual analysis on reanalysis data from the National Centers for Environmental Prediction (NCEP) and the European Centre for Medium-range Weather Forecasts (ECMWF). The zonal momentum budget residuals of their analysis indicated that the downgradient effects dominate within the MJOs. This is partly consistent with Tung and Yanai
(2002a,b). However, the upgradient elements near the leading edges were not captured. Houze et al. (2000) showed from airborne and shipborne Doppler radar data of TOGA COARE that downdrafts associated with stratiform regions of large mesoscale convective cloud systems transport momentum from the midtroposphere down to near the surface. In the “westerly onset regions,” where the midlevel wind is easterly, the momentum transport decreases the westerly near the surface. In the regions where the westerly is deeper and the midlevel wind is also a strong westerly, the momentum transport increases the westerly near the surface. They further suggest that the momentum transport by downdrafts associated with stratiform regions may affect the formation of mesoscale organization of convective systems, depending on the ambient midlevel wind.

In theoretical respects, some recent studies investigate the roles of acceleration due to the momentum flux convergence associated with CC- and/or SCC-scale flows within the MJO, using idealized models that consider interactions between different scales. Moncrieff (2004) proposed that the MJO itself can be considered as a convective archetype on the large scale. By taking into account the meso-β-scale (20–200 km) organization of the large-scale convective organization highlighted in Grabowski and Moncrieff (2002)—the westerly wind burst, the vertical and meridional transport of horizontal momentum, and the equatorial superrotation—were explained. Based on a systematic multiscale theory introduced in Majda and Klein (2003), a series of studies (Majda and Biello 2004; Biello and Majda 2005, 2006) constructed a family of theoretical MJO models that incorporate the momentum and temperature flux convergence associated with CC- or SCC-scale convective organizations into the planetary-scale flows. Biello et al. (2007) point out that the equatorial superrotation associated with the MJO-scale flow is due to the vertical upscale momentum flux from SCC scales, which reinforces the horizontally convergent flow due to the MJO-scale mean heating. Majda and Stechmann (2009) reproduced the westerly wind burst phase of the MJO with a simplified dynamical model that considers the momentum flux convergence associated with convectively coupled waves within the MJO.

Despite valuable insights from various previous studies, the spatial distributions and ensemble effects of the CMT associated with the rainbands within the MJOs have yet to be clarified. Detailed examination of the CMT within the MJOs requires three-dimensional datasets with resolved mesoscale flows and broad spatial coverage. Such datasets are currently available only from global nonhydrostatic models that explicitly represent deep convective circulations. Herein, taking advantage of the explicit 3D mesoscale convection and broad spatial coverage of NICAM, we explore the effects of CMT within the convective envelope of an MJO. We analyze an MJO event that was simulated successfully by NICAM (Miura et al. 2007a; Liu et al. 2009).

The objectives of this study are (i) to calculate the acceleration due to CMT directly for each rainband and document their spatial distribution relative to the MJO and (ii) to quantitatively evaluate the net effect of CMT and discuss its role in the MJO.

The paper is structured as follows. The datasets and analysis methods we used are described in section 2, along with an introduction to the concept of CMT. The results of the analyses are shown in section 3. Section 4 discusses the possible roles of the CMT, and key issues are identified. Summary and conclusions are given in section 5.

2. Data and methodology

a. Datasets

NICAM adopts an icosahedral horizontal grid system that almost homogeneously covers the globe. The governing equations are nonhydrostatic and fully compressible. The model has 40 vertical layers with a terrain-following grid system. The model top height is 38 km, and the layer thicknesses increase gradually with altitude. The cloud microphysics scheme of Grabowski (1998) is applied with no cumulus parameterization. The radiation scheme of Sekiguchi and Nakajima (2008) is applied. The subgrid turbulence is represented by the Mellor–Yamada level-2 scheme (Mellor and Yamada 1982), with modifications that include the effect of moisture (Noda et al. 2010). Further details can be found in Tomita and Satoh (2004), Miura et al. (2007b), and Satoh et al. (2008).

We use the 7-km mesh NICAM output data from the MJO experiment performed by Miura et al. (2007a), which successfully reproduced the MJO propagation (Liu et al. 2009; Nasuno et al. 2009). The 3.5-km mesh in Miura et al. (2007a) produced a realistic MJO in terms of cloud and precipitation properties (Inoue et al. 2008; Sato et al. 2009). However, only a week-long integration was performed at this resolution, and the 3D datasets required for the calculation of CMT were archived only once per day. The 7-km mesh experiment is 32 days long, starting 15 December 2006. Full 3D datasets were archived every 6 h, and 2D datasets (e.g., surface rain) are available every 1.5 h. The initial atmospheric conditions are obtained from the 1.0° mesh NCEP final analysis. Spatial and temporal variations in sea surface temperature
(SST) are interpolated from the weekly Reynolds SST (Reynolds and Smith 1994).

The effective resolution, roughly 7 times the mesh size, is about 50 km (Skamarock 2004). The MJO contains convective structures on the CC, SCC, and MJO scales (Nasuno et al. 2009). Whereas the relatively small CCs are only marginally resolved, large CCs and SCCs are effectively resolved. The finescale cumulus cells are not represented. The distribution and structures of the rainbands compare favorably to observations, as shown later in section 3. The 7-km mesh model permits mesoscale pressure gradients, which are essential for upgradient CMT (Moncrieff 1992). Although the quantitative values of CMT may be revised in the future by higher-resolution experiments that resolve the mesoscale convective structures more in detail, analyses of the 7-km mesh dataset provide useful insights. In summary, this is the first 3D dataset that allows explicit evaluation of CMT in an MJO event.

More information on this particular MJO experiment can be found in the following papers. Masunaga et al. (2008) compared the precipitation features between NICAM and satellite observation data from the Tropical Rainfall Measurement Mission (TRMM). Nasuno et al. (2009) analyzed the eastward-propagating squall lines within the convective region. Liu et al. (2009) analyzed the evolution of the structure, amplitude, and phase of this MJO.

The NICAM is usually limited to short-term simulations because of the extremely large amount of computation required. Therefore, its capability of reproducing multiple MJO cycles in boreal winter in long runs has not been comprehensively tested. However, a 3-month-long run for boreal summer (July–August 2004) reproduced a reasonable MJO cycle (Oouchi et al. 2009).

The NICAM’s simulations of rainbands and the MJO structure were evaluated using the surface rain products retrieved from TRMM and zonal winds from the reanalysis data from the Japan Meteorological Agency (JMA) Climate Data Assimilation System (JCDAS). Reanalysis data from NCEP and outgoing longwave radiation (OLR) data from the National Oceanic and Atmospheric Administration (NOAA) are used to obtain climatological data to locate the MJO convection center, following Wheeler and Hendon (2004, hereafter WH04).

b. Methodology

The design of the CMT analysis of this study is based on the budget residual analysis by Tung and Yanai (2002a,b); that is, we use the Reynolds-averaging technique. The small-scale terms (traditionally denoted by a prime) are deviations from horizontally averaged values (denoted by an overbar). However, in the analysis corresponding to objective (i), stated in section 1, we identify each rainband and treat the associated CMT distinctively. In the quantitative analysis corresponding to objective (ii), we integrate the CMT effect over an MJO composite. Such differences require modifications to the analysis method of Tung and Yanai (2002a,b). Location of the MJO center and identification of rainbands are also required. The following subsections define the CMT terms starting from the momentum equation, describe the procedures to locate the MJO center and to identify the rainbands, and introduce two analysis methods designed to address objectives (i) and (ii), respectively.

1) ACCELERATION DUE TO CMT

The tendency equation for the mean zonal wind is given by

\[
\frac{\partial u}{\partial t} = \frac{1}{\rho} \frac{\partial \rho uu}{\partial x} - \frac{1}{\rho} \frac{\partial \rho vu}{\partial y} - \frac{1}{\rho} \frac{\partial \rho w}{\partial z} - \frac{1}{\rho} \frac{\partial \rho u'}{\partial x} + \frac{f \sigma}{\rho} + \overline{F},
\]

where \( u, v, \) and \( w \) are the zonal, meridional, and vertical winds, respectively. Also, \( \rho, \rho, f, \) and \( F \) are the density, pressure, Coriolis parameter, and friction term, respectively; \( F \) includes the forcing due to sub-grid-scale turbulence and the drag due to cloud water and cloud ice. In this study, we determine the overbar terms by their horizontal mean values within a 5°-radius circular area. Primes denote the deviations from this horizontal mean. A radius of 5° was chosen so that most MCSs, including their broad stratiform regions, fit inside the circle. The area of the circle is comparable to the area of the intensive flux array region described in Tung and Yanai (2002a,b), as well as to the grid cell size of GCMs with low to moderate resolutions.

For later convenience, the sum of the right-hand side of Eq. (1) except for \( \overline{F} \) is denoted by \( A \), and zonal acceleration due to zonal, meridional, and vertical

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For later convenience, the sum of the right-hand side of Eq. (1) except for \( \overline{F} \) is denoted by \( A \), and zonal acceleration due to zonal, meridional, and vertical
momentum flux convergence associated with deviation winds [terms in the second line of Eq.(1)] are denoted by \( X_x, Y_x, \) and \( Z_x, \) respectively:

\[
X_x = -\frac{1}{\rho} \frac{\partial \mu u}{\partial x}; \quad Y_x = -\frac{1}{\rho} \frac{\partial \mu u}{\partial y}; \quad Z_x = -\frac{1}{\rho} \frac{\partial \mu v}{\partial z}. \tag{2}
\]

Positive/negative \( X_x, Y_x, \) and \( Z_x, \) describe the positive/negative acceleration on \( \vec{\mu}. \) As will be shown in the following section, \( X_x \) and \( Y_x \) are small compared to the other advection terms and \( Z_x. \) We focus mainly on \( Z_x, \) and call it the zonal acceleration due to CMT. Similarly, we consider the tendency equation for the mean meridional wind and define

\[
X_y = -\frac{1}{\rho} \frac{\partial \mu u}{\partial x}; \quad Y_y = -\frac{1}{\rho} \frac{\partial \mu v}{\partial y}; \quad Z_y = -\frac{1}{\rho} \frac{\partial \mu v}{\partial z}. \tag{3}
\]

Here, \( Z_y \) is the primary meridional acceleration term due to CMT. Thus, the acceleration due to CMT on the \( 5^\circ \)-radius circle scale horizontal wind \([\vec{\mu}, \vec{\nu}]\) is given by

\[
[Z_x, Z_y]. \tag{4}
\]

2) Determination of the MJO Center

We use the seasonally independent MJO monitoring indices, real-time multivariate MJO series 1 (RMM1) and 2 (RMM2), introduced by WH04. WH04 applies a set of empirical orthogonal function (EOF) analyses on a combined daily equatorial observational dataset of OLR, 850-hPa zonal winds, and 200-hPa zonal winds, meridionally averaged over \( 15^\circ N - 15^\circ S. \) The seasonal cycles and interannual variability are removed from the dataset prior to the EOF analyses. Hereafter these data are referred to as “anomaly data.” RMM1 and RMM2 for a certain day are obtained by projecting anomaly data of the day onto the two leading EOFs derived by performing the above analysis on the observed anomaly dataset of 1979–2001. The leading observation-based EOF pair, provided freely from the Centre for Australian Weather and Climate Research (online at http://www.cawcr.gov.au/staff/mwheeler/maproom/RMM/eof1and2.htm), describes the spatial structure of MJOs (Fig. 1 of WH04). We derive the anomaly data from the NICAM dataset, project them on to the leading observation-based EOF pair to calculate RMM1 and RMM2 for each simulated day, and reconstruct the NICAM MJO structure of the anomalous equatorial OLR, 850-hPa zonal winds, and 200-hPa zonal winds from the RMM indices and the two leading observation-based EOFs. Detailed comparison of RMM plots from NICAM and from observations can be found in Liu et al. (2009). In this study, we focus mainly on the “transition area” of the MJO, where the dominant type of convection changes from cumulus congestus to cumulonimbus. Many previous studies (e.g., Kemball-Cook and Weare 2001; Kikuchi and Takayabu 2004) have shown that the transition occurs where the lower-tropospheric westerly winds of the MJO intrude and gradually deepen. Therefore, we define the MJO center as the converging zero point of the reconstructed 850-hPa zonal wind anomaly. Meridional shifts are not considered in the determination of the MJO center. The MJO center determined by the above procedure moved eastward from \( 53^\circ E \) to \( 175^\circ W \) during the month-long experiment. The average propagation speed of the MJO across \( 100^\circ E \) and \( 170^\circ E, \) the primary region examined later in this study, was \( 4.8 \) m s\(^{-1}. \) In the process described above, the validity of projecting the NICAM anomalies onto observed EOF structures may be questioned. However, the essence is that the definition of the MJO center is designed to represent the converging zero point of the 850-hPa zonal wind anomaly. The reconstructing process is done mostly to decrease noise. The MJO center simply defined by a converging zero point of the 5-day running mean 850-hPa zonal wind does not significantly alter the general results of this paper.

3) Detection of Rainbands

Individual rainbands are detected by the surface rain data from NICAM and the TRMM Microwave Imager (TMI). Prior to the detection process, horizontal averages are taken on both datasets to equalize their resolutions and reduce noise. The grid intervals of the averaged data are about 23 km. Grid cells with more than \( 0.3 \) mm h\(^{-1}\) of rain are labeled as raining grid cells, and continuous regions of raining grid cells having an area larger than 2000 km\(^2\) are labeled as rainbands. For each of the detected rainbands, a straight line that connects the two farthest raining grid cells is recorded as the longer axis of the rainband. The defined longer axes do not always represent linear structures in the rainbands.

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\(^3\) The term \( Z_x, \) also includes momentum flux divergence due to gravity waves.
but the substantial results of the analysis in this paper are not highly sensitive to differences in the axis determination technique. Dark shades in Figs. 1a and 1b show an example of combined TMI surface rain of eight consecutive orbits and a snapshot of surface rain by NICAM, respectively. Vectors show the 850-hPa-level horizontal winds by JCDAS and 1.6-km-level horizontal winds by NICAM, respectively. The light shade in Fig. 1b shows the region where the OLR is below 210 W m\(^{-2}\), representing clouds. Figure 1c shows samples of detected rainband axes.

Rainband detection was performed over the equatorial Indian Ocean and the western to central Pacific Ocean covering 20\(^\circ\)N–20\(^\circ\)S, 40\(^\circ\)E–140\(^\circ\)W. While the path of TMI covers about \(\frac{2}{3}\) of the region twice per day, the NICAM surface rain data are archived 16 times per day and cover the entire region. Therefore, NICAM has roughly a 10–15 times better chance than TMI of detecting rainbands.

4) METHOD I

Method I, depicted schematically in Fig. 2, is designed for surveying the spatial distribution of acceleration due to CMT of individual rainbands. The 5\(^\circ\)-radius circles are determined for each detected rainband by setting the center of the circles at the center of the rainband axis. Then the horizontal mean and deviations are calculated within the circle. Considering that multiple rainbands might exist in a single circle, the area assumed to be strongly affected by the rainband in question is determined as a rectangular area with side lengths proportional to the longer axis length. For a longer axis length of \(l\), a slightly longer length 1.2\(l\) is chosen for the axis-parallel side of the rectangle, to include the perturbation effects that probably extend a little further than the ends of the axis. The lengths of the axis-normal sides are also proportional to the longer axis length \(l\). In the present study, they are set to 1.2\(l\), so that most of the stratiform regions are included.\(^5\) If the rectangular area (always a square area in this study) does not entirely fit into the 5\(^\circ\) radius circle, the protruding part is excluded. This area will be called the “near-rainband area” in this paper. We define the “CMT vector” \([Z^*_x, Z^*_y]\) as the contribution of the near-rainband area to \([Z_x, Z_y]\); that is,

\[
[Z^*_x, Z^*_y] = \left[ \sum_{k=1}^{n} \frac{1}{\bar{p}} \frac{\partial \mu_k w_k}{\partial z}, \sum_{k=1}^{n} \frac{1}{\bar{p}} \frac{\partial \nu_k w_k}{\partial z} \right] \frac{1}{N},
\]

where \(k, n,\) and \(N\) are labels for each grid cell within the near-rainband area, the number of grid cells within the near-rainband area, and the number of grid cells within the 5\(^\circ\)-radius circle, respectively. The CMT vector is calculated for each individual rainband at each model layer. They are then composited according to their location relative to the MJO center.

The contribution to \([Z^*_x, Z^*_y]\) from the area outside of the near-rainband area but inside the 5\(^\circ\)-radius circle can be similarly defined as

\[
[Z^{**}_x, Z^{**}_y] = \frac{1}{N} \sum_{j=1}^{N} \left[ \frac{1}{\bar{p}} \frac{\partial \mu_j w_j}{\partial z}, \frac{1}{\bar{p}} \frac{\partial \nu_j w_j}{\partial z} \right],
\]

where \(j\) is the label for each grid cell outside of the near-rainband area but inside the 5\(^\circ\)-radius circle. Naturally,

\[
[Z_x, Z_y] = [Z^*_x, Z^*_y] + [Z^{**}_x, Z^{**}_y].
\]

The separation between \([Z^*_x, Z^*_y]\) and \([Z^{**}_x, Z^{**}_y]\) is necessary in order to distinctively extract the CMT effect of a single rainband. An intense rainband is often accompanied by smaller and weaker rainbands. If we define the CMT vector of individual rainbands by \([Z_x, Z_y]\) instead of \([Z^*_x, Z^*_y]\), the CMT vectors calculated for the rainbands nearby an intense rainband will also include the effect of the intense rainband, and be large in

\[^{5}\text{The length of the axis-normal sides is shortened if we intend to focus more specifically on the highly convective parts.}\]
Since the objective of method I is to survey the spatial distribution of the acceleration due to CMT of individual rainbands, we want to avoid such multiple counting to the extent possible; otherwise, the result (Fig. 9) may be misleading (not shown). The denominator \( N \) is fixed and therefore not a function of rainband size. A circular shape is applied for the calculation of the ambient flow in order to treat the rainbands equally, regardless of their alignment.

While \([Z^*, Z^*]\) represents the CMT effect of individual rainbands on the ambient flow, \([Z, Z]\) represents the total CMT effect on the ambient flow. Note that \([Z, Z]\) will be used in method II, where we will quantify the net CMT effect on the ambient flow.

An example of a rainband in the convectively active area of the NICAM MJO is shown in Fig. 3. Note that this extremely significant case leads the deep convective phase of the MJO. The colors in Fig. 3a are the surface rain, visualizing a linear structure of the convective region. The purple line indicates the automatically determined longer axis of this rainband, and the black circle is the 5\(^{\circ}\)-radius circular area used to calculate the horizontal averages denoted by overbars (e.g., \(\bar{u}, \bar{w}\)).

The center of the circle is equivalent to the center of the purple line. This rainband travels eastward at about 7 m s\(^{-1}\). The horizontal winds at a height of 1.6 km (vectors) show significant convergence along the leading edge. Strong vertical winds are consistently concentrated...
along the leading edge of this rainband (Fig. 3b). The vertical profile of $\tilde{\eta}$ (Fig. 3c) shows that the environmental wind shear was strongly east-southeasterly. Figure 3d shows vertical profiles of $\tilde{\eta}$ (blue), $\tilde{\omega}$ (purple), and $Z^u_x$ (red). Figure 3e shows the zonal–height cross section of $u'\omega'$ and the tilted nature of the airflow along the red arrows in Figs. 3a and 3b.

Organized vertically tilted rainbands lead to a vertical convergence of momentum flux that contributes to $Z^u_x$ (Moncrieff 1992, 1997). For example, the westward-tilted rainband in Fig. 3e produces negative $u'\omega'$, which provides positive $Z^u_x$ in the lower troposphere and negative $Z^u_x$ in the midtroposphere (see the red line in Fig. 3d). The CMT reinforces the shear (upgradient transport) where the environmental shear is easterly, a typical condition west of the MJO center. This rainband is a linear-shaped squall line, associated with a convective system categorized as an SCC in standard classifications. Although many rainbands found in CCs and SCCs have this structure, there also exist rainbands with more complicated structures and/or propagation features. Numerous rainbands have multiple squall-line axes traveling in different directions. Note that we adopt an Eulerian view here following Tung and Yanai (2002a,b). Some previous studies use the moving (Lagrangian) frame of reference (e.g., Moncrieff 1992) to examine the change in momentum across a rainband. The Eulerian approach is more readily determined from GCM output.

5) METHOD II

Although method I gives a useful description of the spatial distribution of the CMT vectors, it is not suitable for evaluating the net effect of unevenly distributed CMT vectors on the large-scale flow (e.g., $\tilde{\eta}$). Method II, on the other hand, is designed to quantitatively evaluate the effect of a CMT ensemble on $\tilde{\eta}$. For method II, the $5^\circ$-radius circles are distributed evenly over the equator, as schematically shown in Fig. 4. The circles are overlapped to reduce noise produced by the rainbands propagating in and out of the circle during the 6-h interval of the archived data. The zonal intervals of the circles are $1^\circ$.

For each circle, overbars and primes are as in method I, and the vertical profiles of the terms associated with $\delta \tilde{\eta}/\delta t$ (e.g., $X_x$, $Y_x$, $Z_x$) are calculated for each circle. Note that in method II, the total $Z_x$ is the zonal CMT effect, instead of $Z^u_x$, which represents the CMT of individual rainbands. The term $Z_x$ represents the total CMT effect of all rainbands within the $5^\circ$ circle. It includes extra effects from regions that are not classified as a near-rainband area, such as momentum transport by convectively excited gravity waves that propagate out of the near-rainband area, and CMT associated with convection below the threshold of rainband detection. However, these effects are small within the convective envelope of this MJO and do not affect the results significantly (not shown). The zonal–height structures of the terms for each time step are computed by aligning the vertical profiles derived from the circles. These zonal–height structures are composited according to the longitude of the MJO center. Thus, we construct the averaged MJO-relative zonal–height structures of the terms associated with $\delta \tilde{\eta}/\delta t$. The individual contribution of each term to the longitudinal change of $\tilde{\eta}$ across the MJO can then be quantified.

3. Results

a. Overall characteristics and comparison to observations

Figure 5 shows the frequency distribution of the rainbands detected from TMI and NICAM. Statistics are analyzed for rainbands detected between 15 December 2006 and 15 January 2007: a total of 16 546 rainbands from TMI and 226 462 from NICAM. Note that the rainbands are counted individually in each orbit or 6-hourly snapshot, so long-lived rainbands will be counted several times. The ratio of the numbers detected by TMI and NICAM (1:13.7) is consistent with the ratio (1:12.9) expected from the distinct data sampling. The smoothness of the frequency distribution map appears to be quite different, mainly because of the different sample size. When the rainband detection is performed on the NICAM surface rain data masked by individual passes of the TMI data (nearly 1.5-hourly) in order to equalize the sample shape and size, the number of detected rainbands reduces to 17 669. The smoothness of the frequency distribution also reduces (not shown) to be more similar to the TMI distribution in Fig. 5a. While we do not expect NICAM to realistically

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6 The overlapping of circles does not cause multiple countings of CMT effects. It is almost identical to taking zonal running means.

7 TMI covers about $\frac{2}{3}$ of the analysis region per day on its northward and southward paths respectively, whereas the NICAM precipitation data covers the entire region every 1.5 h.
reproduce each observed rainband throughout a month-long experiment, the spatial distributions agree reasonably well. However, the number of rainbands of NICAM seems overestimated in the central to eastern Pacific intertropical convergence zone and in the western Indian Ocean. We chose to further analyze the 12°N–12°S, 100°–170°E region, where the MJO is active, the rainband cases are plentiful, and their distributions are in fairly good agreement. Of the 72 033 rainbands identified within this main analysis region (MAR), 18 009 had a complete 3D dataset for CMT calculation.

Figure 6 is similar to Fig. 5 but shows the frequency distribution relative to the MJO center (zero longitude). Only the rainbands detected within the MAR are included. Total numbers of the rainbands within ±50° from the MJO center are 4021 in TMI and 52 921 in NICAM; 13 215 NICAM rainbands had a complete 3D dataset. Again, the distributions agree well except that NICAM is smoother than TMI because of the larger sample size. Rainbands are concentrated within ±30° from the MJO center.

Figure 7 compares the size distributions of rainbands in TMI and NICAM. While Inoue et al. (2008) shows that 7-km mesh NICAM tends to produce larger cloud clusters compared to those observed by the Japanese Geostationary Meteorological Satellite-5 (GMS-5), the rainbands in NICAM tend to have smaller horizontal coverage than those in TMI.

We also examined the alignment of the rainbands. In TMI, 37% (63%) of the identified rainband axes were aligned closer to the north–south (east–west) direction than to the east–west (north–south). NICAM had 23% (77%) closer to the north–south (east–west). Although the axis alignments of TMI and NICAM qualitatively agree in that they tend to be in the east–west direction, NICAM tends to produce more east–west rainbands than observed in TMI. Previous studies (e.g., LeMone 1983) suggest that such a bias would underestimate the zonal CMT effects. However, this conclusion may be premature because the simple axis definition in the present study cannot distinctively handle multiple linear structures that often appear when low-level shear and midlevel shear are approximately parallel (e.g., Dudhia and Moncrieff 1987; LeMone et al. 1998). Many east–west-aligned rainbands in NICAM featured a short north–south (shear normal) linear structure associated with a large upgradient zonal CMT, and a long east–west (shear parallel) linear structure associated with small CMT (not shown). Therefore, the bias is arguably not sensitively related to CMT.

Note that the frequency distributions, size distributions, and axis directions also depend on the choice of thresholds and to the degree of smoothing performed.
prior to the rainband detection. The results from TMI depend on the sensitivity of the instruments. The substantive analyses of CMT, on the other hand, are designed to be relatively less sensitive to these issues.

Figure 8 shows the longitude–height cross section of the zonal winds composited relative to the MJO center within the MAR. Although the wind speed is greater in NICAM than JCDAS, the overall structures, such as the depth and the level of maximum of the low- to midlevel westerly winds west of the MJO center, are in good agreement. Note that the stronger MJO-scale zonal wind anomalies in NICAM may cause bias (e.g., overestimation of the CMT effects described in this paper). Along with the results shown in previous papers on this MJO simulation (e.g., Miura et al. 2007a; Nasuno et al. 2009; Liu et al. 2009), Figs. 5–8 indicate that the distribution of rainbands and the vertical profile of zonal winds are simulated reasonably well by NICAM.

b. Results of the method I analysis

Figure 9 shows the plan views of the CMT vectors \([Z_x^*, Z_y^*]\) that correspond to the individual rainbands. Vectors with positive/negative zonal components, indicated by red/blue, are located relative to the MJO center. CMT vectors are drawn only when the absolute value exceeds \(2.31 \times 10^{-5} \text{ m s}^{-2}\) (i.e., 2.0 m s\(^{-1}\) day\(^{-1}\)). CMT vectors with large absolute values are concentrated to the west of the MJO center, especially in the \(-20^\circ-0^\circ\) locale. Note that rainbands also exist to the east of the MJO center as evident in Fig. 6. CMT vectors associated with those rainbands are sparse simply because they have small absolute values.

A striking feature in Fig. 9 is that the CMT vectors are highly organized, evincing strong directivity, rather than random as anticipated from the various structures of
the rainbands. Near the surface (0.63 km), CMT vectors with positive zonal components dominate: the westerly winds near the surface are accelerated. In the low to midtroposphere (2.49–6.24 km), the zonal components of the CMT vectors change sign and decelerate the westerly winds. In the upper troposphere (11.99 km, 13.08 km), the CMT vectors with positive zonal components again dominate. At 8.34 and 10.03 km, the CMT vectors with both positive and negative zonal components coexist. In this study, the above highly directed structure of CMT vectors between 2° and 0° from the MJO center will hereafter be called the “three-layer structure,” regardless of the 8.34–10.03-km layer.

Although the MAR contains many islands that tend to trigger convection, the CMT vectors that form this three-layer structure do not show a strong dependence on the land–ocean distribution. The three-layer structure occurs throughout the MAR as the western part of the MJO passes (not shown).

The three-layer structure is consistent with the momentum budget analysis of the NICAM MJO simulation briefly reported by T. Nasuno et al. (2011, unpublished manuscript). Using a similar Reynolds-average analysis, they average over smaller horizontal areas in order to imitate the grid cell sizes of high-resolution GCMs.

The time resolution of the 3D data used to derive the CMT vectors is rather coarse (6 h) compared to the rapid life cycle of convective systems. Therefore, the temporal representativeness of the momentum budget calculated from 6-hourly snapshots was evaluated by comparing the change in $\ddot{u}$ between two time steps to the change in a 6-h integration of a linearly interpolated tendency equation. They were in good agreement, indicating that the tendency Eq. (1) is not significantly affected by high-frequency effects, so the CMT vectors have sufficient temporal representativeness. This is confirmed later in the analysis using method II.

Figure 10 shows a longitude–height cross section relative to the MJO center of $u$ composite (contours) and the zonal component of the composite CMT vectors (those within ±3° from the equator and having zonal components with absolute values larger than 2.0 m s$^{-1}$ day$^{-1}$ as indicated by the arrowheads) over the equator. As in Fig. 9, a three-layer structure occurs west of the MJO center. CMT vectors with positive zonal components are dominant up to 1.6 km; negative zonal components are dominant between 2 and 6.5 km; positive and negative zonal components coexist between 7 and 10 km; and positive zonal components dominate above 11 km.
The CMT vectors that belong to numerous rainbands with various structures have clear directivity. This important issue will be further described in T. Miyakawa et al. (2012, unpublished manuscript, hereafter Part II), along with the detailed analysis of typical rainbands in NICAM. Owing to their strong directivity, the CMT vectors, as a group, likely have significant impact on the longitudinal change of the zonal wind across the MJO. A quantitative description and evaluation of this group impact is the main objective of the method II analysis.

c. Results of the method II analysis

Figure 11 is a snapshot of the longitude–height cross section of the vertical profiles of $u$ and $Z_x$ calculated in each circle over the equator in Fig. 4. The MJO center is located near 146°E and profiles exist only where the longitude is within the range of the MAR. Typical of MJOs, the lower-tropospheric westerly penetrates below the upper-level easterly. This structure resembles the composite in Fig. 10 but is not as smooth. The structure of $Z_x$ resembles the three-layer composite structure in the analysis using method I. For instance, between 0° and −20°, the acceleration is positive near the surface, negative at 2–6 km, and positive above 11 km.

Figures 12a–d show the MJO center-relative composites of $u$, $\Delta u/\Delta t$, $A$, and $Z_x$, respectively. Note that $\Delta u$ is the difference of $u$ between successive (6-hourly) time steps; also, $A$ is defined on the right-hand side of Eq. (1), with the small forcing term $F$ neglected (i.e., $\Delta u/\Delta t$ and $A$ approximate $\partial u/\partial t$). Figure 12a is almost identical to Fig. 8b except that it is smoother. Figure 12b shows that zonal wind changes associated with the MJO occurs mainly between 120° ($x_f$) and 230° ($x_r$) relative to the MJO center. Figure 12c resembles Fig. 12b, indicating that 6-hourly snapshots of $A$ approximate $\Delta u/\Delta t$, the 6-h mean values of $\partial u/\partial t$. Figure 12d shows the zonal-height dependence of $Z_x$. This is consistent with the three-layer structure in Figs. 9 and 10 west of the MJO center. Note that the magnitude of $Z_x$ is large compared to $\Delta u/\Delta t$ and $A$. The structure of the CMT is consistent with previous studies: mostly downgradient, as in Tung and Yanai.

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Subscripts $r$ and $f$ stand for “rear” and “front,” respectively.
Upgradient CMT effects occur in the low to midtroposphere between 0° and 20° west of the MJO center where the westerlies are shallow, as pointed out by Tung and Yanai (2002b), although details are different. In particular, the upper component of the three-layer structure in the shallow westerly region is not robust in Tung and Yanai (2002b).

To quantitatively evaluate the effect of $Z_x$ on $u$, we focus on longitudinal differences in the mean zonal wind across the MJO, $\overline{u}_x - \overline{u}_f$, where $\overline{u}_x$ and $\overline{u}_f$ are the $u$ values at $x = x_f$ and $x = x_r$ respectively. Figure 13a shows the respective vertical profiles of $\overline{u}_x - \overline{u}_f$, $(1/c)\int_{x_r}^{x_f} (\Delta \overline{u}/\Delta t) \, dx$, and $(1/c)\int_{x_r}^{x_f} A \, dx$, where $c = 4.8$ m s$^{-1}$ is the mean eastward phase speed of the MJO throughout the MAR. Assuming the MJO phase speed is constant, integration in the $x$ direction divided by $c$ is equivalent to time integration. A stationary observer initially at the MJO-relative position $x_f$ at time $t = t_f$ experiences a change in $\overline{u}$ of $\overline{u}_x - \overline{u}_f$, and since $dx/dt = -c$, it follows that

$$\overline{u}_x - \overline{u}_f = \int_{t_f}^{t_r} (\partial \overline{u}/\partial t) \, dt = -(1/c) \int_{x_f}^{x_r} (\partial \overline{u}/\partial t) \, dx$$

The good agreement between the profiles of $\overline{u}_x - \overline{u}_f$ and $(1/c)\int_{x_r}^{x_f} (\Delta \overline{u}/\Delta t) \, dx$ in Fig. 13a supports the assumption of a constant eastward phase speed. That the profile

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**FIG. 10.** Composite zonal-height cross section relative to the MJO center of equatorial zonal wind (contours) and acceleration due to CMT (arrowheads; red/blue indicate positive/negative values) with absolute values exceeding 2.0 m s$^{-1}$ day$^{-1}$ and located within 3° of the equator.

**FIG. 11.** Snapshot of a zonal–height cross section derived by connecting the vertical profiles calculated at each 5°-radius circle distributed along the equator (0° on the horizontal axis indicates the center of the MJO) for (a) $\overline{u}$ and (b) $Z_x$, the zonal CMT term. Negative values are shaded.
of \((1/c)\int_{x'}^{x'} A \, dx\) agrees with the profiles of \(\pi_r - \pi_f\) and \((1/c)\int_{x'}^{x'} (\Delta \pi/\Delta t) \, dx\) confirms that the 6-hourly snapshots of \(A\) approximate the 6-h mean zonal tendency. Subtle differences are likely due to the neglect of \(\bar{F}\) and the variance not captured by the 6-hourly sampling.

Since the terms on the right-hand side of Eq. (1) are included in \(A\) (except for \(\bar{F}\)) and can be treated separately, their contributions to \(\pi_r - \pi_f\) can be quantified. Figure 13b shows the vertical profiles of \((1/c)\int_{x'}^{x'} X_r \, dx\), \((1/c)\int_{x'}^{x'} Y_r \, dx\), and \((1/c)\int_{x'}^{x'} Z_r \, dx\), the contributions to \(\pi_r - \pi_f\) of the terms associated with the correlations of the deviation winds. As pointed out in previous studies, the contributions of \(X_r\) and \(Y_r\) are much smaller than that of \(Z_r\). Although not negligible, they almost cancel each other. Note that \(X_r\), \(Y_r\), and especially \(Z_r\) represent effects not resolved in standard GCMs. Between 2 and 6.5 km in height, the sum of these perturbations, \((X_r + Y_r + Z_r)\), strongly affects the westerly zonal wind tendency: the mean value of \(\bar{u}_r\) is 10 m s\(^{-1}\) while the mean value of \((1/c)\int_{x'}^{x'} (X_r + Y_r + Z_r) \, dx\) is \(-16\) m s\(^{-1}\).

Figure 13c shows the contributions of the large-scale terms\(^9\) \((1/c)\int_{x'}^{x'} M_{FCX} \, dx\), \((1/c)\int_{x'}^{x'} M_{FCY} \, dx\), \((1/c)\int_{x'}^{x'} M_{FCZ} \, dx\), and \((1/c)\int_{x'}^{x'} M_{PGF} \, dx\), where

\(^9\) Here \(M\) stands for “mean scale,” and the subscript FC stands for flux convergence in the zonal \((x)\) direction. Also, FCY and FCZ are similar to FCX but for the meridional \((y)\) and vertical \((z)\) directions, respectively. PGF stands for pressure gradient force. The contribution of \(\bar{f}\pi\) is not shown because it is very small over the equator.
While the contributions of the mean-scale terms are much larger than \( Z_x \) in the upper troposphere, they are comparable at lower altitudes. In the 2–6.5-km layer, \( Z_x \), \( M_{FCZ} \), and \( M_{PGF} \) dominate, and \( Z_x \) and \( M_{FCZ} \) almost equal
cancel. However, the two terms may not remain in balance when \( Z_x \) alters. Possible consequences of excluding \( Z_x \) in GCMs are discussed in the next section. The significance of \( M_{FCZ} \) and \( M_{PGF} \) will be examined in subsequent papers.

4. Discussion

A striking feature pointed out in the previous section is that although the rainbands have complicated structures, the CMT vectors in Fig. 9 are well organized and work in concert upon the MJO. The directivity of the CMT vectors contributes to the tendency of \( \pi \) (Figs. 12 and 13). The reasons will be discussed in Part II. We suggest two possible roles for CMT. As shown in Figs. 10 and 12, while CMT contributes positively to the westerly tendency near the surface, it contributes negatively to the westerly tendency between 2 and 6.5 km, and to the easterly tendency above 11 km. In other words, while the CMT cooperatively affects the eastward progress of the westerly winds near the surface, it delays the westerlies in the low to midtroposphere. In the upper troposphere, CMT delays the eastward progress of the easterlies. Note that these CMT effects are not resolved by traditional climate models (mesh size \( \sim 100 \) km), although they may be crudely resolved by high-resolution global numerical weather prediction models (e.g., Moncrieff and Klinker 1997).

Figure 14a describes a possible role of CMT in the low to midtroposphere. For deep convection over tropical oceans, horizontal convergence in the low to midtroposphere is more important than the horizontal convergence near the surface (e.g., Sherwood 1999). The surface boundary layer is already sufficiently moist, therefore horizontal convergence in the low to midtroposphere controls deep convection by providing moisture to counter the entrainment of environmental air. The CMT delays the eastward progress of the low- to midtropospheric westerly winds and the longitudinally convergent areas favorable for deep convection, delaying the eastward shift of atmospheric heating due to latent heat release. This will result in a slower eastward shift of the Matsuno–Webster–Gill response (Matsuno 1966; Webster 1972; Gill 1980) to atmospheric heating involving the entire MJO.10 This view is supported by the composited distribution pattern of relative humidity within the MJO in Fig. 15a: CMT prevents or suppresses the eastward advection of moisture-rich air.

Moistening of the low to midtroposphere to the east of the MJO center can also occur because of shallow (congestus) convection or external disturbances (e.g., Kikuchi and Takayabu 2004). The ratio of the contribution of the eastward moisture advection, the shallow convection, and the external disturbances to the moisture budget may be an important factor for the MJO phase speed.

The CMT effect near the surface will also enhance the latent heat flux to the west of the MJO center. The maximum positive near-surface CMT is in phase with the strong convection, and ahead of the low-level westerly wind maximum (Fig. 15). The lag between the maxima of convection and low-level westerlies suggests that the nonlinear WISHE theory described in Sobel et al. (2010) would retard MJO propagation. The CMT effect appears to reduce the degree to which the nonlinear WISHE retards propagation, and it may strengthen MJO amplification by enhancing the overall latent heat flux.

As shown in Fig. 14b, the negative CMT at the 2–6.5-km layer tilts the westerly flow structure westward with height. Note that \( M_{FCZ} = -\partial \rho \bar{w} / \partial z \) is related to the westward tilt (i.e., to \( \partial \bar{w} / \partial z \) in the “transition area”). Should the

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10 Whether or not the eastward propagation is essentially controlled by the eastward migration of convective heating remains a matter of debate.
westward tilt of the westerly wind with height become smaller when the CMT is excluded, $M_{FCZ}$ may be reduced, canceling the absence of negative CMT. If so, the delaying effect of CMT suggested above may be insignificant.

The CMT delays the eastward propagation of the MJO and/or tilts the flow structure of the MJO westward with height. Representing the eastward phase speed and the westward-tilted structure are difficulties that many current GCMs face with the MJO (Lin et al. 2006, 2008; Zhang et al. 2006; Kim et al. 2009).

Importantly, most GCMs (especially climate models) entirely neglect subgrid CMT effects, although some recent GCMs include parameterization of CMT as a momentum mixing process. Although a well-designed numerical experiment is necessary to clarify the CMT effects, such an experiment may not immediately improve the MJO in GCMs that implement a CMT parameterization. Rather, we need to understand the similarities and differences between the parameterized CMT and explicit CMT, which in our paper includes organized mesoscale momentum transport by squall lines. The exact role of CMT in the MJO remains an open question, but it is likely that mesoscale organization of convection and the associated mesoscale pressure gradients are important.

Part II will show that the three-layer structure arises from the combined effects of downgradient and upgradient CMT. The latter cannot be adequately represented by a mixing-type CMT parameterization. On the other hand, some recent GCMs that implement cumulus schemes with high sensitivity to relative humidity show significant improvement in the eastward propagation of MJOs (Zhang and Mu 2005a,b; Bechtold et al. 2008; Chikira and Sugiyama 2010; Chikira 2010; Kim et al. 2009).
Comparisons of the CMT of NICAM and parameterized CMT in such GCMs will be a next step. Along with CMT effects, MJO flow structures (e.g., westward vertical tilt) and moisture budgets in the MJOs produced by NICAM and by GCMs should be compared. These studies are ongoing in collaboration with several users of GCMs. The need for cumulus schemes capable of producing upgradient CMTs and the evaluation of preceding schemes of this type will be discussed in future studies. In addition, the separate roles of CC- and SCC-scale processes (e.g., Moncrieff 2004; Majda and Stechmann 2009), which were not investigated in this study, also deserve closer examination.

It is essential to verify the CMT distributions and quantify the magnitude of the CMT effects associated with MJOs. Incompleteness of our study are that (i) it is a single case study, (ii) the 7-km mesh is too coarse for a realistic explicit representation of detailed structures of mesoscale convection, and (iii) the availability of observational data on momentum is very limited. Higher resolution (e.g., 3.5-km mesh) will soon enable multiple case studies to be simulated, so the first two items are likely to be addressed in the near future. A comparison of $u'w'$ in squall lines produced by regional nonhydrostatic models of various resolutions conducted by Weisman et al. (1997) shows that the magnitudes of $u'w'$ tend to be larger at finer resolution. They suggest that 4-km mesh represents most of the features produced with a 1-km mesh.

The third item, the lack of observational data, is more troublesome. The existing observations most suitable for comparison with this study are the budget residuals of two MJO events that occurred during December 1992–February 1993 in TOGA COARE, shown in Figs. 8 and 9 of Tung and Yanai (2002b). However, it is difficult to say whether or not these figures agree with our results. The budget residual representing the zonal CMT acceleration near the leading edge of the second MJO event (their Fig. 9) has a similar structure to the three-layer structure described in this paper. However, the first MJO event (their Fig. 8) has less robust negative (positive) acceleration in the low to midtroposphere (upper troposphere). The significant difference between these two MJO events implies that a large sample of MJO events is required for a reasonable comparison, from both observations and models. Note that the budget residual analysis of Tung and Yanai (2002b) was conducted at a single location, whereas the three-layer structure shown in Figs. 9, 10, and 12 is a composite view over the entire MAR. The sensitivity of the budget residual technique to small observational errors also strengthens the need for a larger sample number. More observational studies along the lines of Tung and Yanai (2002b) will be helpful.

5. Summary and conclusions

The analysis of a global nonhydrostatic model NICAM, which reproduced an MJO event with multiscale convective structure, quantitatively evaluated the structure of CMT vectors and the net contribution of CMT effects on the longitudinal difference of the zonal wind across the MJO. Even though the model resolution is too coarse to fully resolve the mesoscale circulations, the NICAM MJO case is the first 3D dataset in which the CMT can be explicitly evaluated for a realistic multiscale MJO event.

The distribution of rainbands in NICAM agrees with satellite observations reasonably well. Of the 72 033 rainbands identified within the MAR, 18 009 had a complete 3D dataset necessary for calculating CMT. The flow structures of the rainband samples produced in NICAM also reasonably resembled those documented in previous observational and model studies, including CMTs with up- and/or downgradient features.

While numerous rainbands occur to the east and to the west of the MJO center, the result of an analysis with method I shows that strong acceleration due to CMT is most prevalent between 0° and 20° to the west. An unexpected feature was revealed. Although the rainbands exhibit various and complicated structures, the zonal components of the CMT vectors have a well-organized three-layer structure: positive near the surface (below 1.6 km), negative at the low to midtroposphere (2–6.5 km), and positive again at the upper troposphere (above 11 km). The directions of the CMT vectors between 7 and 10 km in height were not significant.

The quantitative analysis with method II shows that the net acceleration due to CMT contributes largely to the longitudinal differences in $u$ across the MJO. Between 2 and 6.5 km in height, where the longitudinal difference in the mean zonal wind across the MJO is 10 m s$^{-1}$ on average, the net acceleration due to CMT contributes about $-16$ m s$^{-1}$.

It was suggested that the CMT delays the eastward MJO propagation, described as follows. CMT delays the eastward progress of westerly winds and zonally convergent regions in the low to midtroposphere. Because the low- to midtropospheric convergence is viewed as a controlling factor for deep convection in the tropics, the CMT also delays the development of deep convection to the east of the MJO center. This retards the eastward migration of the entire MJO pattern, which is principally a Matsuno–Webster–Gill response to the convective heating.

Another suggested role of the CMT is to tilt the flow structure of the MJO westward with height. If so, the negative contribution of CMT in the tendency equation of...
\( \bar{\sigma} \) is balanced by the positive contribution from \(-\frac{1}{\rho} \frac{\partial \rho \mathbf{v}}{\partial z} \), which is conjectured to increase (decrease) when the MJO flow is more (less) tilted westward.

To the extent verifiable by the 7-km resolution NICAM, the zonal components of CMT associated with mesoscale convection work in concert, contributing negatively and significantly to the zonal wind tendency in the low to midtroposphere near the MJO center. The contribution of CMT to the zonal wind tendency has been quantified. The CMT effects likely have essential roles associated with the eastward propagation of the MJO and/or with its vertically westward-tilted flow structure. However, it is necessary to verify our results with a larger sample of simulated MJO events and with more observational data on CMT. These data are not yet available but are expected in the near future. It is also necessary to examine how the CMT contributes to the momentum budget when model resolution is improved.

Convective elements that form the three-layer structure will be further examined in Part II, with respect to their up- and downgradient features and environmental wind shear. Comparisons between the CMT effects in NICAM with those in recent GCMs with parameterizations are in progress. The roles of the SCC-scale in the multiscale process of the MJO certainly deserve closer investigation.

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