Convective Momentum Transport Observed during the TOGA COARE IOP. Part II: Case Studies

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ABSTRACT

Convective momentum transport (CMT) associated with the Madden–Julian oscillation (MJO), tropical waves, squall and nonsquall mesoscale convective systems (MCSs), and the diurnal cycle is studied by examining the momentum budget residual \( X = (X, Y) \) deduced from the objectively analyzed in situ observations during the Tropical Ocean Global Atmosphere Coupled Ocean–Atmosphere Response Experiment (TOGA COARE) intensive observing period (IOP; November 1992–February 1993). Using wavelet transform, time evolution of signals of these disturbances in the time series of \( |X| \) and \( I_{BB} \) (an index for deep convection), averaged over the intensive flux array (IFA), is analyzed. Signals of disturbances with periods \( \approx 1 \text{ day} \) in \( |X| \) generally evolve in phase with those in \( I_{BB} \). During the convective phase of MJO, signals in both \( |X| \) and \( I_{BB} \) with shorter periods are also enhanced. Frequency distribution of IFA-mean \( E = -\nabla \cdot X \) in the troposphere is examined. The mean \( E \) is positive, that is, kinetic energy (K) transfer is downscale, about 60%–65% of time in the lower troposphere below 500 hPa, and between 200 hPa and the tropopause. However, in the upper troposphere, between 350–200 hPa, upscale and downscale K transfers occur with nearly equal frequency. Different frequency distributions near the surface, the middle troposphere, and near the tropopause suggest the existence of different regimes of K transfer associated with various convective and boundary layer processes. Furthermore, the frequency of direction of CMT on mesoscale convective organizations documented in many previous observations is found to be detectable at the 2.5° × 2.5° objective analysis. Couplets of vorticity and vorticity budget residual Z appear in the upper levels with nonsquall MCSs. Upscale K transfer is found in the line-normal direction of a squall line. During the westerly wind phase of the MJO, convection appears to play dual roles. First, as the westerlies are initiated in the lower troposphere, CMT is typically upgradient and may help maintain middle-level easterly shear. Thus the upscale K transfer may help trigger the westerly wind burst (WWB). Second, at the later stage with strong lower- to middle-level westerlies, CMT is mostly downgradient and reduces the middle-level zonal wind shear.

1. Introduction

This paper is the second part (hereafter Part II) of a study of convective momentum transport (CMT) observed during the intensive observing period (IOP; November 1992–February 1993) of the Tropical Ocean Global Atmosphere (TOGA) Coupled Ocean–Atmosphere Response Experiment (COARE). In the first part (Tung and Yanai 2002, hereafter Part I), we examined the general features of the large-scale momentum budget residual \( X = (X, Y) \) obtained in the vicinity of the Intensive Flux Array (IFA). The time series of \( X \) and \( Y \) exhibit multiscale temporal behavior, showing modulations by the Madden–Julian oscillation (MJO; Madden and Julian 1971, 1972) and other disturbances. The power spectra of \( X, Y \), and \( I_{BB} \) (an index of deep convective activity) showed remarkable similarity, suggesting a dynamical link between deep cumulus convection and the large-scale horizontal motion, via CMT which is being modulated by various atmospheric disturbances. The IOP-mean \( X \) acts to decelerate the mean flow. The averaged vertical momentum transport is downgradient, and kinetic energy is transferred from the large-scale motion to subgrid-scale eddies. However, many previous studies suggested that the directions of instantaneous CMT are dependent on the mode of convective organization. In Part II, we focus on cases of the CMT in various convective events during the IOP.

The influence of CMT on the environmental flow in the immediate vicinity of vigorous convection has long been documented by many early investigators. Through aircraft observations, it has been recognized that the flow surrounding a large cumulonimbus resembles that around a solid cylinder in fluid (e.g., Newton and Newton 1959; Fujita and Grandoso 1968; Fankhauser 1971; Ramond 1978). Furthermore, satellite observations have shown ship wave patterns trailing overshooting cumulonimbus tops (Levizzani and Setvák 1996; Wang 2001).
Detailed studies by modeling (Rotunno and Klemp 1982) and observations (e.g., LeMone et al. 1988a) showed, however, that the apparent resemblance of the flow around cumulus convection to that past a solid obstacle arises from the difference between ambient momentum and in-cloud momentum carried upward from the cloud base.

A more complicated picture holds for the dynamical interaction between an ensemble of cumuli and its environment. One of the most important achievements of the Global Atmospheric Research Program (GARP) Atlantic Tropical Experiment (GATE) in 1974 is the advancement of our knowledge of mesoscale convective systems (MCSs) in the Tropics. The MCSs are further categorized into nonsquall and squall systems. Because of a lack of reliable surface pressure analysis during GATE, only Shapiro and Stevens (1980) attempted to use X to diagnose CMT. Instead, several authors (e.g., Tollerud and Esbensen 1983; Sui and Yanai 1986; Sui et al. 1989) analyzed vorticity budget residual Z to infer CMT.

After Ooyama (1980, cited in Houze and Betts 1981) discovered a “vorticity couplet” in the upper troposphere of a nonsquall MCS, Tollerud and Esbensen (1983) not only confirmed this finding but also showed that a similar vorticity couplet appeared on a composite of nonsquall MCS cases during phase III of GATE. In addition, a remarkable couplet of Z was found collocating with the large-scale vorticity couplet at the developing and mature stages of an MCS. Sui and Yanai (1986) found similar Z couplets with several nonsquall MCSs observed during GATE phase III. It was concluded that momentum redistribution by convection is the major mechanism for the development and maintenance of the vorticity couplet. Sui et al. (1989) further examined Z and quantitatively verified the roles of CMT in modifying the large-scale vorticity field with a parameterization scheme that maintains a consistency between the effects of cumulus convection on the large-scale momentum and vorticity fields (Esbensen et al. 1987).

Moreover, by using multiaircraft observation during GATE, LeMone (1983) found that the vertical transport of horizontal momentum normal to a squall line is upgradient; that is, it is against the vertical momentum gradient. On the other hand, the transport of momentum parallel to the squall line is downgradient. LeMone et al. (1984) confirmed and elaborated these findings with systematic evaluations of nine squall lines during GATE. It has been further found that the “selective” CMT in a squall line is mainly due to the convection-induced perturbation pressure gradient resulting from the interaction between convective updraft and strong lower-tropospheric vertical shear of horizontal wind (LeMone et al. 1988a). From observations, it is well known that strong environmental wind shear is crucial in the formation of convective bands (e.g., Barnes and Sieckman 1984; Alexander and Young 1992). The convective-scale pressure perturbation is weaker in a low-shear environment, as observed by LeMone et al. (1988b). Therefore, it is less likely to find upgradient CMT associated with a nonsquall MCS because such a system tends to form in a low-shear environment.

Numerous MCSs have been observed during the TOGA COARE IOP (e.g., Hildebrand 1998; Lewis et al. 1998; Roux 1998; Chong and Bousquet 1999; Halverson et al. 1999; Kingsmill and Houze 1999; Bousquet and Chong 2000; Protat and Lemaître 2001a,b). More complicated structure and evolution of MCSs have been revealed. Among the most extensively studied is a squall line around 9.5°S, 159°E, on 22 February 1993, either through observations (Jorgensen et al. 1997; LeMone et al. 1998) or by numerical modeling (Trier et al. 1996, 1997, 1998; Montmerle et al. 2000). LeMone et al. (1998) confirmed the importance of environmental wind shear in organizing MCSs. They also noted that, in events that squall lines initially developed normal- to low-level wind shears, additional midlevel wind shears may cause the MCSs to evolve into nonsquall organization at later stages. Such departure from the original linear organization, as indicated by the simulation of Trier et al. (1998), may have direct impact on CMT and its effects on large-scale flow.

The main purpose of this paper is to present the CMT observed in active convective events under various large-scale settings during TOGA COARE IOP. Data and methods are discussed in section 2. In section 3, we show the link between deep convection and the momentum budget residual X at various timescales using the wavelet transforms. This section also shows the frequency distribution of downscale and upscale kinetic energy transfer at various levels. In section 4, examples of vorticity and Z couplets, and deceleration/acceleration of the large-scale motion associated with different types of convective organizations are presented. The directions of CMT associated with two events of the MJO are discussed in section 5. Finally, a summary and conclusions are given in section 6.

2. Data and methods
2a. TOGA COARE IOP datasets

The major dataset for this study is a 6-hourly objective analysis of the merged radio soundings and wind profiler data archived in the University Corporation for Atmospheric Research (UCAR) (Loehrer et al. 1996) during the IOP (1 November 1992–28 February 1993). The objective analysis (OBAUKLA; see Part I for details) covers 20°S–20°N, 130°E–180°, centering on the IFA with a horizontal grid size of 2.5° × 2.5°. There are 43 vertical levels, including the surface, 37 levels from 1000 to 100 hPa, at 25-hPa intervals, and additional 5 levels (80, 70, 50, 30, and 10 hPa). In addition, as in Part I, the box covering 5°S–0°, 152.5°–157.5°E is referred to the “IFA region” for brevity.
The momentum budget residual \( \mathbf{X} = (X, Y) \) is obtained using \( \text{OBA}_{\text{UCLA}} \) and is interpreted as the acceleration of the grid-scale (large scale) horizontal wind due to the convergence of subgrid-scale convective and/or turbulent momentum fluxes:

\[
X = \left( \frac{\partial \mathbf{v}}{\partial t} + \mathbf{v} \cdot \nabla \mathbf{v} + \frac{\partial \mathbf{v}}{\partial p} + \nabla \mathbf{\phi} + \lambda \mathbf{k} \times \mathbf{v} \right)
\]

\[
\approx - \frac{\partial \mathbf{v} \cdot \mathbf{\omega}}{\partial p},
\]

(1)

in which \( \mathbf{v} \) is the horizontal velocity, \( \omega = dp/dt \) the vertical \( p \) velocity, \( \mathbf{\phi} \) the geopotential, \( \lambda \) the Coriolis parameter, \( \nabla \) the isobaric del operator. The overbar represents the running horizontal average over a large-scale area, and the prime denotes the deviation from the average. As in Part I, the horizontal convergence of eddy momentum flux is ignored. The approximation then leads to the expression for the vorticity budget residual (e.g., Esbensen et al. 1987; Sui et al. 1989):

\[
\begin{align*}
Z &= \frac{\partial \mathbf{v}}{\partial t} + \mathbf{v} \cdot \nabla \mathbf{v} + \mathbf{v} \cdot \nabla \mathbf{\omega} + \frac{\partial \mathbf{v}}{\partial p} \\
&\quad + \mathbf{k} \cdot \nabla \mathbf{\omega} \times \frac{\partial \mathbf{v}}{\partial p} \\
&= \mathbf{k} \cdot \nabla \times \left( - \frac{\partial \mathbf{v} \cdot \mathbf{\omega}}{\partial p} \right).
\end{align*}
\]

(2)

where \( \mathbf{k} = \mathbf{k} \nabla \times \mathbf{v} \) is the relative vorticity and \( \mathbf{\eta} = \lambda + \mathbf{k} \mathbf{\omega} \) the absolute vorticity. From (1) and (2), we see that \( Z \) is a result of nonuniform horizontal distribution of vertical convergence of eddy vertical flux of momentum.

The vertical flux of horizontal momentum associated with convection at an arbitrary level \( p \) is obtained by integrating \( \mathbf{X} \) from the surface pressure \( P_s \) to \( p \):

\[
F = (F_x, F_y) = \left( \frac{\rho \mathbf{v} \cdot \mathbf{w}}{\rho} \right) \bigg|_{P_s} \\
= \left( \frac{\rho \mathbf{v} \cdot \mathbf{w}}{\rho} \right) \bigg|_{P_s} - \frac{1}{g} \int_{P_s}^{P} \mathbf{X} \, dp.
\]

(3)

with \( w = -\omega \rho g \) the vertical density, \( \rho \) the density of air, and \( \frac{\partial \mathbf{v} \cdot \mathbf{w}}{\partial p} \bigg|_{P_s} = -\mathbf{\tau} \) the momentum flux at surface. The surface stress \( \mathbf{\tau} \) is estimated by the bulk parameterization method (Fairall et al. 1996).

The use of raw (unadjusted) estimates of \( \mathbf{X} \) in (3), however, often yields unrealistically large values of \( \mathbf{F} \) at upper levels. To reduce systematic errors in \( \mathbf{X} \) (and thus in \( \mathbf{F} \)), we impose an additional constraint that \( \mathbf{F} \) vanishes in the lower stratosphere. In practice, we set \( \rho \mathbf{v} \cdot \mathbf{w} \bigg|_{P_{so}} = 0 \), where \( P_{so} = 10 \) hPa, the highest data level in \( \text{OBA}_{\text{UCLA}} \). Then the estimates of \( \mathbf{X} \) are “adjusted” to satisfy this additional condition. Readers are referred to Part I for details about the adjustment. In Part II, we continue the practice in Part I in which unadjusted values of \( \mathbf{X} \) are used for analyses that require

continuity in time (i.e., the wavelet transforms); adjusted values are emphasized when physical constraint in vertical is required.

Furthermore, we use several indicators of convective activities in this paper. A subset of the 0.1° × 0.1° \( I_{\text{rain}} \) is used to measure deep convective activity and define organizations of MCSs. Here, \( I_{\text{rain}} \) is deduced from values of equivalent blackbody temperature \( (T_{\text{BB}}) \) that are no larger than 225 K as indicated by the \( \text{Geostationary Meteorological Satellite (GMS) IR radiance data (Nakazawa 1995).} \) The 5-day-mean gauge- and satellite-based rainfall rate from the Climate Prediction Center (CPC) Merged Analysis of Precipitation (CMAP; Xie and Arkin 1997) is also utilized. In addition, following Yana et al. (1973), the apparent heating \( Q_a \) is calculated and interpreted by

\[
Q_a = \frac{c_p}{\rho} \left( \frac{\partial \mathbf{\theta}}{\partial t} + \mathbf{v} \cdot \nabla \mathbf{\theta} + \frac{\partial \mathbf{\theta}}{\partial p} \right)
\]

\[
\approx - \frac{\partial \mathbf{\nabla} \cdot \mathbf{\theta}}{\partial p} + Q_e + L(\mathbf{\tau} - \mathbf{\tau}_o),
\]

(4)

where \( c_p \) is the specific heat of air at constant pressure, \( R \) the gas constant of dry air, \( p \), the 1000-hPa reference pressure, and \( \theta \) the potential temperature. The quantity, \( \mathbf{\tau} \) is the dry static energy; \( s = c_p T + gz \), in which \( T \) is temperature; \( Q_e \) is radiative heating rate; \( L \) the latent heat of vaporization; and \( e \) and \( e \) are condensation and evaporation per unit mass of air respectively. Equation (4) shows that \( Q_a \) is the total effect of convergence of convective sensible heat flux, radiative heating, and net condensation heating. The detailed computation scheme for \( Q_a \) has been presented in Part I.

b. Wavelet transform and multiresolution decomposition

Power spectral analysis showed that time series of \( X, Y \), and \( I_{\text{rain}} \) bear signals of the MJO, various tropical waves, and diurnal cycle (Part I). We employ wavelet transform (e.g., Lau and Weng 1995; Torrence and Compo 1998; Mallat 1999) in order to further visualize the time evolution of CMT at a certain time and frequency. The wavelet transform is a tool to analyze time series that contain nonstationary power at many different frequencies. In meteorology, the wavelet transform has been used by many authors (e.g., Sato and Yamada 1994; Weng and Lau 1994; see also a review by Torrence and Compo 1998).

A mother wavelet \( \psi \) is a basis function that has zero mean and is localized in both time and frequency domain. It is scaled with parameter \( s' \) and translated by parameter \( \tau' \) to give a (daughter) wavelet

\[
\psi_{s',\tau'}(t) = \frac{1}{\sqrt{s'}} \psi \left( \frac{t - \tau'}{s'} \right).
\]

(5)

The continuous wavelet transform (CWT; Grossmann
and Morlet (1984) of a time series $A(t)$ at the timescale $s$ and position $\tau$ is computed by correlating $A(t)$ with a wavelet

$$WA(\tau, s) = \int_{-\infty}^{+\infty} A(t) \frac{1}{\sqrt{s}} \psi^*_s(\frac{t - \tau}{s}) \, dt,$$

(6)

in which $\psi^*_s(t)$ is the complex conjugate of $\psi_s(t)$. This definition shows that the CWT, accomplished by translating wavelets of various scales over the length of $A(t)$, is a measure of similarity between the basis functions (wavelets) and the signal $[A(t)]$. For each timescale (period) chosen, the calculated CWT coefficients express the closeness of the signal to a wavelet at the designated scale. Furthermore, the CWT coefficients lie in a scale–translation domain, which is presented as the period–time domain in our study.

The commonly known CWT algorithm has the advantage of viewing events of interest at desired period and time. However, it provides highly redundant information for the reconstruction of the original signal. The discrete wavelet transform (DWT) algorithm is concise and mathematically rigorous for the reconstruction. It has been used by Sato and Yamada (1994), Yano et al. (2001a,b) and others. In DWT, the scale parameter ($s$) is discretized onto a logarithmic grid with 2 as the base, and the translation parameter ($\tau$) is adjusted with respect to the timescales; therefore, there is a different sampling rate at each scale. In addition, by applying an inverse DWT to the wavelet coefficients at certain scale, partial information of the original time series is reconstructed at the scale. Following such a procedure, the original time series can be decomposed into those of various scales, which is known as multiresolution decomposition. In section 3, both CWT and DWT are utilized. A Morlet wavelet (Fig. 1a) is adopted into the CWT algorithm, and a Meyer wavelet (Fig. 1b) is used for the DWT algorithm developed by Kolaczyk (1994) in order to produce the multiresolution decomposition.

3. Characteristics of CMT events in the IFA during IOP

a. Time evolution of $X$ with convection

In Part I, we found that the IOP time series of $X$, $Y$, and $I_{\text{mb}}$ in the IFA display similar multiscale behavior with spectral peaks corresponding to the MJO, several known tropical waves, and the diurnal cycle. The coupling between convection and the large-scale tropical motions has been studied by many investigators (e.g., Nitta 1970; Salby and Hendon 1994; Takayabu 1994a,b; Chen and Houze 1997; Tung et al. 1999; Wheeler et al. 2000; Yanai et al. 2000). The similarity between the power spectra of $X$, $Y$ at 500 hPa, and $I_{\text{mb}}$ in the frequency domain has suggested a dynamical coupling between cumulus convection and the large-scale motions through CMT. Here we further examine the correspondence between the time series of $|X| = \sqrt{X^2 + Y^2}$ at 500 hPa and $I_{\text{mb}}$ using wavelet transforms.

Figures 2a and 2b, respectively, show the wavelet coefficients of $|X|$ at 500 hPa and $I_{\text{mb}}$ in the IFA region obtained from the CWTs of their time series. The wavelet coefficients are plotted in the period–time domain. To simplify the representation, the coefficients are normalized at each sampled period, and only positive values are plotted. If $|X|$ is induced by deep convection, then we expect to see simultaneous occurrence of large wavelet coefficients of $|X|$ and $I_{\text{mb}}$. Figure 2 generally confirms this expectation. The most dominant MJO signals around periods 30–60 days in late December and early February appear both in Figs. 2a and 2b. Similarity is also found between 8–16 days, in the range of equatorial Rossby wave (e.g., Kiladis and Wheeler 1995; Pires et al. 1997) and the lifetime of super cloud clusters (Nakazawa 1988, 1995). Furthermore, it is noted that, in both Figs. 2a and 2b, wavelet coefficients at periods around 2–16 days are enhanced during the MJO events, suggesting nonlinear interactions between oscillations at various scales. Such phase-locking relationship between MJO and synoptic disturbances was previously reported in the tropical convection (with $I_{\text{mb}}$, in Lau et al. 1991) and temperature fields (with $T_{\text{mb}}$, in Weng and Lau 1994).

Results shown in Figs. 3a and 3b are multiresolution decompositions of the $|X|$ and $I_{\text{mb}}$ time series via inverse Meyer DWTs. The original time series can be reconstructed by adding the individual time series together. In this form, relationships between $|X|$ and $I_{\text{mb}}$ at smaller timescales are more clearly recognized. We find general correspondence between the features in Figs. 3a and 3b. However, $|X|$ varies more widely than $I_{\text{mb}}$ particularly at the shortest timescales. As seen in Fig. 3, $|X|$ has relatively strong signals even during the “quiet times” of $I_{\text{mb}}$ series (e.g., late November) at the 12-h period. Aside from possible errors in estimating $X$, such strong variation could be attributed to the following reasons: 1) the semidiurnal cycle is only marginally resolved by the 6-hourly observations, 2) the wavelet transform naturally generates coarser-frequency resolution at higher-frequency scales, and 3) less significant, short-lived convective events may not be well captured by the IFA-averaged $I_{\text{mb}}$. 

![Fig. 1. Mother wavelets: (a) the Morlet wavelet for CWT and (b) the Meyer wavelet for DWT.](image-url)
b. Frequency distribution of kinetic energy transfer

In Part I, we have shown that as far as the IOP mean is concerned, $E = -\mathbf{v} \cdot \mathbf{X}$, where $\mathbf{v} \cdot \mathbf{X}$ is the rate of work done by the subgrid-scale frictional force, is positive in the IFA. That means, kinetic energy (thereafter $K$) is, on average, being transferred from the large-scale motion to the subgrid-scale convection and turbulence. However, the large standard deviation of $E$ has suggested that upscale $K$ transfer also occurs frequently.

Figure 4a shows the relative frequency (in percentage) of downscale $K$ transfer ($E > 0$) in the troposphere without regard to its magnitude, with $E$ estimated using the unadjusted and adjusted $\mathbf{X}$ (see section 2a). We remark that the computation of 4-month $E$ in the IFA
yields no zero value. Downscale $\mathcal{K}$ transfer occurs around 60%–65% of the 478 cases at each level between surface and 500 hPa. From 500 hPa and above, the relative frequency of downscale transfer drops gradually; around 350–200 hPa, up- and downscale $\mathcal{K}$ transfers tend to occur at nearly equal frequency. Above 200 hPa, and below the tropopause (~125 hPa), the frequency for downscale $\mathcal{K}$ transfer to occur increases to around 60% again.

The frequency distribution of $\mathcal{E}$ obtained with adjusted $\mathcal{X}$ as a function of pressure from the surface to 100 hPa is shown in Fig. 4b. To produce the frequency distribution plot, the abscissa is equally divided into intervals of $10^{-4}$ m$^2$ s$^{-3}$. The frequency distribution is in general positively skewed, as downscale $\mathcal{K}$ transfer occurs more frequently and with more events of larger amplitudes. Moreover, the frequency distribution at low values of $\mathcal{E}$ has two maxima, one near the surface and the other around 600–400 hPa. The lower maximum frequency, possibly representing the effects of boundary layer process, is more concentrated in small positive $\mathcal{E}$. The upper maximum that seems to be collocated with the deep convective mass flux is more evenly centered on both small positive and negative values of $\mathcal{E}$. It is noted that the frequency distribution above 200 hPa spreads out significantly. Aside from possible measurement errors, this spread may indicate very different characteristics of the $\mathcal{K}$ transfer near the tropopause where detrainment from the top of deep convection may take place. Furthermore, the narrowing of the frequency distribution around 500–400 hPa and the flattening above 200 hPa in Fig. 4b are in part contributed by the weaker wind in the middle atmospheric levels and the jet at higher levels throughout the IOP, since fewer changes at these levels are found in the frequency distributions of $\nabla \cdot \mathcal{X}/|\nabla|$ (not shown). Figure 4c highlights...
the frequency distributions at 950, 500, and 150 hPa. This figure clearly shows that the frequency distribution of \( E \) at 950 hPa (near the lower maximum in Fig. 4b) is more skewed than that at 500 hPa (near the upper maximum in Fig. 4b), while the latter is more concentrated in smaller values. Finally, the frequency distribution at 150 hPa is depicted with heavier tails than those at lower levels.

4. CMT and organizations of convection

This section presents features of CMT associated with several active convective events during the IOP. The main interest here is to examine the dependence of CMT on the mode of convective organization. The objective analysis (OBAUCLA) used here has a 2.5\(^\circ\) × 2.5\(^\circ\) horizontal grid mesh at 6-h time intervals. These are too coarse to resolve the detailed structure and time evolution of individual MCSs, but it is still capable of showing the environmental flows affected by CMT. With the aid of \( I_{\text{BB}} \), as well as previous documentations based on radar observations by several investigators, the locations and dimensions of MCSs can be defined.

a. Vorticity couplet associated with a nonsquall MCS

Figure 5a shows the horizontal map of \( I_{\text{BB}} \) and vorticity at 250 hPa at 1800 UTC 24 December 1992. The vorticity budget residual \( Z \) and vertical \( p \) velocity \( \omega \) at 250 hPa, at the same time, are shown in Fig. 5b. A nonsquall MCS is located in the southern half of the Outer Soundings Array (OSA), as represented by \( I_{\text{BB}} \) in Fig. 5a. The MCS is centered around 5\(^\circ\)S and 155\(^\circ\)E, as indicated by the maximum values of \( I_{\text{BB}} \) (Fig. 5a) and \( \omega \) (Fig. 5b). In Fig. 5a, there is a clearly defined vorticity couplet located over the MCS, with positive vorticity on the north side and negative vorticity on the south side of the convective core. A couplet of \( Z \) is seen in Fig. 5b, with a northwest–southeast arrangement. At this time, the large-scale tropospheric wind is westerly in lower levels and easterly in the upper levels (not shown). The results here confirm the findings by previous studies: the vorticity couplet is due to the deceleration of upper-level easterlies by mixing with lower-level westerly momentum due to CMT, and the \( Z \) couplet is evidence of the large-scale vorticity redistribution due to convective-scale circulations. Such couplets, particularly the \( Z \) couplets, are frequently observed with nonsquall MCSs during the COARE IOP.

b. Deceleration observed in a nonsquall MCS

More direct evidence of the large-scale momentum mixing by CMT in a nonsquall MCS is shown in Fig. 6. In Fig. 6a, a nonsquall MCS in the center of IFA at 0000 UTC 11 November 1992 is shown with the streamlines and \( X \) vectors at 700 hPa. It is noted that the direction of \( X \) is generally against the westerly wind passing through the MCS in IFA; that is, the large-scale westerly flow is being decelerated. The vertical profiles of tropospheric wind and \( X \) averaged over the IFA region are shown in Figs. 6b and 6c. In the figures, both adjusted and unadjusted \( X \) are shown to ensure the credibility of results. The zonal wind speed is shown to be decelerated by both adjusted and unadjusted \( X \) in the troposphere except for around 400–300 hPa (Fig. 6b). The meridional wind speed is mostly decelerated throughout the troposphere except for around 500–400 hPa where acceleration is very small (Fig. 6c). In gen-
eral, the CMT in this MCS is acting for large-scale vertical momentum mixing and deceleration of the horizontal flow speed. Not surprisingly, couplets of vorticity and $Z$ are also found with this system (not shown).

c. Selective CMT in a squall line

Figure 7a illustrates a squall line at 1800 UTC 20 February 1993, extending from 3° to 12°S. This squall line has been documented with radar observations by Lewis et al. (1998) and its environmental setting has been discussed in LeMone et al. (1998). The vertical profile of zonal wind averaged over 7.5°–2.5°S and 162.5°–165°E in Fig. 7b shows moderate to strong westerly and easterly shears, below and above 800 hPa, respectively. Around this time, the squall line has started to show some departure from the typical linear organization, probably resulted from the direction of the midlevel shear that favors the development of east–west aligned convection band extending to the rear of the old north–south squall line (LeMone et al. 1998). Nevertheless, the MCS still maintains certain typical CMT characteristics similar to those of previously studied squall lines. As seen in the 700-hPa map (Fig. 7a), $X$ is roughly along the streamlines and the westerlies are being accelerated. However, $X$ shows large deflections to the right. The averaged wind and $X$ profiles in the zonal direction (normal to the squall line, Fig. 7b) shows that the westerly from 1000 to 550 hPa is being accelerated. In the meridional direction (parallel to the squall line, Fig. 7c), the wind speed below 400 hPa is evidently decelerated. It is noted that the squall line is accelerating zonal flow and decelerating meridional flow with similar magnitudes.
5. CMT and Madden–Julian oscillation

There are two major Madden–Julian oscillation events observed in the IFA region during the IOP. The westerly wind phase of the first case (MJO1) lasted from late December to early January, and that of the second case (MJO2) from early to mid-February (see Fig. 3 of Part I). Large-scale features of the two MJOs have been well-documented (e.g., Nakazawa 1995; Lin and Johnson 1996; Yanai et al. 2000). In brief, the eastward-propagating large-scale circulations associated with the MJO are characterized by a baroclinic structure and are coupled with deep convection over the western Pacific warm pool. Furthermore, strong low-level westerly episodes called the westerly wind bursts (WWB) accompany the westerly phase of the MJO events (e.g., Sui and Lau 1992; Kiladis et al. 1994; Hartten 1996). The so-called Gill-type model (Matsumo 1966; Gill 1980) has been often used to characterize the structure of MJO over the warm pool (e.g., Hendon and Salby 1996). In such a model, convective heating centered on the equator sets off low-level easterlies on its east side through Kelvin wave response and westerly inflow on its west side via Rossby wave response. However, the coupling between the circulation and convection associated with MJO is in reality more complicated. Large-scale motions and convection propagate with different phase speeds; convection leads the characteristic westerly phase but with a slower phase speed than the latter over most of the warm pool region until reaching the dateline, where the westerlies catch up with the convection (Yanai et al. 2000). WWBs as well as convection grow stronger as the coupled system approaches the dateline. In this section, we discuss possible roles played by CMT in the MJO, which are not included in the cumulus parameterization schemes of most GCMs.

Figure 8 shows the time–height sections of (a) zonal wind $\bar{u}$, (b) zonal momentum budget residual $X$, (c) apparent heating $Q_1$, and (d) the deep convection index $I_{BB}$ and the 5-day-mean CMAP data in the IFA region during MJO1. We note that the IFA wind observations in the UCAR merged soundings and wind-profiler data archive are marked unreliable below 700 hPa around 10–15 December. Therefore, the portions of Figs. 8a and 8b in this period requires very cautious interpretations. In Fig. 8a, the westerly wind phase begins in middle December from the lower troposphere and gradually extends to the middle and upper troposphere. Intermittent strong westerlies around 800–600 hPa are seen with maxima in the middle to late December, with the most intense WWB on 1 January. After 6 January, the strong westerlies in the troposphere subside abruptly. In addition to an easterly maximum near 550 hPa around 18–20 December, strong easterlies are also found above 200 hPa throughout MJO1. Furthermore, in Fig. 8a, a zone of strong westerly shear is seen below the 800-hPa westerly maxima during 21 December–6 January, with its maximum magnitude on 1 January. There is also an easterly shear zone above the westerly maxima, with its height and strength increasing as the westerlies develop and extend into the upper troposphere.

From middle December to 1 January, $X$ and the local time change of $u$ have the same order of magnitude of $3–6 \text{ m s}^{-1} \text{ day}^{-1}$ (Figs. 8a and 8b). Because $X$ reflects the vertical convergence of convective momentum flux in the presence of convective events, the magnitude of $X$ is generally large when the indicators of convection in Figs. 8c and 8d are large. These indicators, $Q_1$, $I_{BB}$, and CMAP rainfall, agree well among themselves. Maxima of $Q_1$ and $I_{BB}$ indicate that intermittent deep convection, reflecting the passage of super cloud clusters through the IFA (Nakazawa 1995), is most active before the peak intensity of WWB in the IFA. By examining Figs. 8a–d, the convective events during 13–21 December are found to occur with significant positive $X$ within the westerly regime from 1000 to 700 hPa, while negative $X$ appears with the easterlies above 400–600 hPa. Thus, $X$ generally intensifies $|\bar{u}|$, and upscale transfer of zonal $K$ is taking place. However, after 21 December, $X$ acts to reduce the magnitudes of the westerlies between the surface and 500 hPa and the easterlies above 200 hPa, and downscale $K$ transfer occurs in the zonal direction during most of the following period until 6
Figure 8. Time-height sections of (a) $\pi$ (m s$^{-1}$), (b) $X$ (m s$^{-1}$ day$^{-1}$), (c) $Q_1/c_p$ (K day$^{-1}$), and (d) $I_{\text{BB}}$ and 5-day-mean CMAP rainfall (mm day$^{-1}$) for MJO1. All quantities are averaged in the IFA region, and 2-day running means have been applied for clearer presentations. In (a)–(c), negative values are shaded. Hatched boxes in (a) and (b) indicate unreliable data records. In (d), $I_{\text{BB}}$ missing period is blocked out.

January. Nevertheless, there are occasional moments of upscale $K$ transfer in the lower to middle troposphere on around 23–24 and 26–30 December, as well as between 400–200 hPa on 1–5 January around 400–200 hPa. It is noted that the maximum downscale $K$ transfer is associated with deep convection, for example, around 25 December. Also, deep convection after 21 December evidently weakens the easterly shear zone above 800 hPa. The peak WWB on 1 January occurs with suppressed convection, but as soon as deep convection resumes, deceleration prevails in a deep tropospheric layer.

Similarly, $\pi$, $X$, $Q_1$, $I_{\text{BB}}$, and CMAP rainfall in the IFA region during MJO2 are plotted in Figs. 9a–d. MJO2 propagates with a faster phase speed than MJO1 and the associated convection in the IFA region is relatively weaker (Yanai et al. 2000). The magnitudes of the WWB and $X$ associated with MJO2 (Figs. 9a and 9b) are smaller than those associated with MJO1 (Figs. 8a and 8b). As shown in Fig. 9a, the westerly wind
phase begins on 26 January; however, it never extends as high as that in MJO1. The peak WWB on 4 February is weaker than that on 1 January (Fig. 8a); thus the westerly shear below and easterly shear above the westerly maxima are less pronounced. Besides, MJO2 is concurrent with the reintensification of the Australian monsoon (McBride et al. 1995), and the lower to middle tropospheric westerlies persist toward the end of February.

Despite the differences, MJO2 still has many features similar to those found in MJO1. For example, in Fig. 9a, intermittent strong westerlies are also found around 800–600 hPa. As shown in Figs. 9a and 9b, $X$ and the local time change of $\pi$ are on the same order of magnitude as the westerlies intensify from late January to early February. Moreover, at the initial phase of the WWB during 26–30 January, convection occurs with upscale $K$ transfer as a positive correlation exists between $X$ and $\pi$ in a deep tropospheric layer. Similar to what was observed during MJO1, the correlation turns negative as the first burst of westerly wind takes place on 1 February. From 1–16 February, $X$ mostly acts to reduce the strength of both the lower tropospheric westerlies and the upper tropospheric easterlies. However, there are occasional incidents of lower-level westerly acceleration found around 2, 3–5, 6–7, and 10–12 February. Finally, as observed in MJO1, deep convection events between 1–16 February tend to happen with
downscale $K$ transfer and weaken the easterly shear above the westerly maxima. The WWB on 4 February is followed by westerly deceleration on 5–6 February.

According to these case studies, convection appears to play dual roles during the westerly phase of MJO. On the one hand, it transfers $K$ into large-scale flow and may help maintain vertical shear, as observed in the initial phase of the two WWB events and in the occasional incidents. On the other hand, it decelerates the large-scale flow and reduces wind shear by vertical mixing of momentum, as observed during most of the westerly wind phase of MJO, particularly after the bursts of westerlies. The accelerating role might be played by less penetrative convection such as cumulus congestus, as reported in Johnson et al. (1999) and Tung et al. (1999) or by smaller-scale and short-lived squall lines as reported in LeMone et al. (1998). Most of the squall lines are less likely to be resolved by the datasets used in this study. The decelerating role seems to be played by very deep convection (e.g., on 24–26 December and 11–13 February). It is likely that such deep convection is even responsible for bringing the westerlies upward by vertical mixing, which may explain why the westerlies extend higher in the troposphere in MJO1 than in MJO2.

6. Summary, discussion, and conclusions

a. Summary and discussion

In Part II, we examined the convective momentum transport (CMT) associated with various convective events over the western Pacific warm pool during the TOGA COARE IOP. The effects of CMT are deduced from the residual of the large-scale momentum budget with the objective analysis OBA_{UCLA}. The convective events include those related to the MJO, tropical waves, squall and nonsquall MCSs. The findings are summarized and discussed as follows.

1) Multiscale temporal behavior of CMT

In Part I, power spectral peaks of $X$, $Y$, and $I_{nm}$ were found to correspond to various disturbances. In Part II, using the wavelet transform and multiresolution decomposition, signals of these disturbances with periods $\approx$ 1 day in $|X|$ and $I_{nm}$ are found to generally evolve in phase. Furthermore, during the convective phase of MJO, signals of both $|X|$ and $I_{nm}$ with shorter periods are also excited. The synchronous enhancement of $|X|$ signals with various periods substantiates the result in Part I that $X$ and $Y$ time series have fractal characteristics. These findings not only confirm the role of CMT in modulating the large-scale motions but also further suggest its role in modulating the multiscale interaction among large-scale waves of various periods.

2) Relative frequency of downscale/upscale $K$ transfer

The frequency distribution of $E = -v \cdot X$ in the troposphere is studied. Positive/negative $E$ indicates that $K$ is transferred downscale/upscale between large- and convective-scale motions. Even though $K$ is transferred downslope on the average, upscale $K$ transfer does occur during the IOP. Regardless of magnitude, downslope $K$ transfer occurs around 60%–65% of the time from the surface to 500 hPa and between 200 hPa and the tropopause. In general, the frequency distribution of $E$ in the troposphere is positively skewed. It is noted that, however, up- and downslope $K$ transfers occur with nearly equal chance in the 350–200-hPa layer. Detailed frequency distribution also suggests the importance of downslope $K$ transfer near the surface by boundary layer processes. The distribution near the tropopause is heavier-tailed than that in the middle and lower troposphere, suggesting that $K$ transfer associated with the detrainment of deep convection is very different from that pertaining to convection in the middle troposphere.

3) CMT and MCS organizations

The dependence of the direction of CMT and associated acceleration or deceleration on the structure of significantly intense convective systems is detected even though the space and time resolutions of OBA_{UCLA} are not sufficient to describe the detailed structure and time evolution of MCSs. In the cases of nonsquall MCSs, overall deceleration of large-scale flow speed is shown. In addition, the vertical mixing of momentum by CMT is indicated with couplets in the vorticity field and $Z$ at around 250 hPa. In the case of a squall line, acceleration/deceleration of the lower-to middle-level wind speed in the line-normal/line-parallel directions are observed. These results confirm many previous mesoscale studies, including the seminal work by LeMone (1983).

4) CMT during MJO

During the westerly wind phase of the two major MJO events during IOP, $X$ and the local time change of $\pi$ are often on the same order of magnitude. It is further found that convection appears to play dual roles during the westerly phase of MJO. At the initial stage, CMT transfers $K$ into large-scale zonal flow and may help maintain middle-level easterly shear. At the later stage, however, it decelerates the large-scale zonal flow and reduces zonal wind shear by vertical mixing of momentum. The upscale $K$ transfer at the initial stage may be associated with less penetrative cumulus convection with squall lines. The occurrence of such convection during the IOP has been reported in previous studies (e.g., LeMone et al. 1998; Johnson et al. 1999; Tung et al. 1999). Deep convection is responsible for the downslope $K$ transfer, which brings westerlies upward by vertical mixing.
More frequent occurrence of deep convection in MJO1 may explain why the westerlies extend higher in the troposphere in MJO1 than in MJO2.

From the observations presented above, we conjecture the following scenario: The convection with squall lines at the leading edge of a super cloud cluster accelerates the large-scale zonal flow through upslope kinetic energy transfer and maintains vertical wind shear through upgradient CMT. On the other hand, the following fully developed, more three-dimensionally structured deep convection functions to damp the large-scale zonal flow due to downslope kinetic energy transfer, and causes vertical momentum mixing. Thus the net CMT effects might provide a zonal acceleration at the front of the eastward-propagating envelope of westerly wind of the MJO and results in WWB. Needless to say, more observation and modeling efforts are still needed to determine the effects of CMT on the MJO.

b. Implications for parameterization

This study shows that kinetic energy may be transferred from convection to large scales in events such as squall lines and WWBs in the tropical western Pacific, even though on the 4-month IOP average, large-scale K is shown to transfer downslope at the order of $O(10^{-4})$ m$^2$ s$^{-1}$ in the troposphere. Given the result that the upslope energy transfer is frequently observed at a scale comparable to the size of a GCM grid box, its occurrence should not be overlooked in large-scale modeling.

The mechanisms for upslope K transfer requires further investigations. Several papers emphasized on convection-induced pressure perturbation generated by the interaction between convective updraft and vertical wind shear (e.g., Rotunno and Klemp 1982; LeMone et al. 1988a; Jorgensen et al. 1991; Liu and Moncrieff 2001). However, as pointed out by LeMone 1983 and LeMone et al. 1984, it is also necessary to investigate the pressure perturbation attributed to convective-scale buoyancy distribution. Attempts have been made to include these effects in CMT parameterization by several authors (Zhang and Cho 1991a,b; Wu and Yanai 1994; Kershaw and Gregory 1997; Gregory et al. 1997; Grubišić and Moncrieff 2000). There have been reports of improvements of mean circulations by including the CMT parameterization in the GCMs (Zhang and McFarlane 1995; Gregory et al. 1997; Inness and Gregory 1997). However, Inness and Gregory (1997) reports that signals of the MJO have been degraded, which is probably because the parameterization scheme failed to produce upgradient CMT during the westerly wind phase of MJO. The simulation by Moncrieff and Klinker (1997) suggests that upgradient CMT may have occurred between the mid- and upper troposphere during the WWB. So far, the geometry of MCSs has not been considered in most of these parameterization schemes. Wu and Yanai (1994) considered anisotropy of mesoscale convective organization and were able to produce

upgradient as well as downgradient CMT provided that the mode of convective organization is given. Success of future CMT parameterization for GCMs may depend on the prediction of shear- and/or buoyancy-induced pressure perturbation with the knowledge of dominant convective modes.

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