Convective Momentum Transport Observed during the TOGA COARE IOP. Part I: General Features

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Abstract

The momentum budget residual, $\mathbf{X} = (X, Y)$, is estimated with objectively analyzed soundings taken during the TOGA-COARE Intensive Observing Period (November 1992-February 1993) to study the effects of convective momentum transport (CMT) over the western Pacific warm pool. The time series of X and Y exhibit multi-scale temporal behavior, showing modulations by the Madden-Julian oscillation (MJO) and other disturbances. The power spectra of X, Y, and I_{TBB} (an index of convective activity) are remarkably similar, showing peaks near 10, 4-5 and 2 days, and at the diurnal period, suggesting a link between deep cumulus convection and the acceleration/deceleration of the large-scale horizontal motion, via CMT which is being modulated by various atmospheric disturbances. The temporal behavior of X and Y can be described as fractals from 1/4 to ~ 20 and from 1/4 to ~ 16 days, respectively. Their fractal characteristics are reflected in the very large standard deviations around the small IOP means. From the analyses of the quantities $\bar{u}X/|\bar{u}|$, $\bar{v}Y/|\bar{v}|$, and $\bar{\mathbf{v}} \cdot \mathbf{X}$, the IOP-mean vertical distributions of the frictional force due to subgrid-scale eddies and the rate of kinetic energy transfer ($E = -\bar{\mathbf{v}} \cdot \mathbf{X}$) are determined. The frictional deceleration and downscale energy transfer take place in a deep tropospheric layer from the surface to 300 hPa. In addition, a concentration of large friction and energy transfer exists in a layer just below the tropopause, suggesting the contribution of momentum detrainment from the top of deep cumuli. The IOP-mean frictional deceleration and downscale energy transfer in the lower troposphere are ~ 0.5–1.0 m s^{-1} day^{-1} and ~ 1.0×10^{-4} m^2 s^{-3}, respectively. The product of eddy momentum flux with the large-scale vertical wind shear shows that the momentum transport is, on the average, downgradient, i. e., kinetic energy is converted from the large-scale motion to convection and turbulence.

1. Introduction

Organized cumulus convection interacts with its environment through vertical transports of heat, moisture, and momentum. The influences of organized cumulus convection on the large-scale thermodynamic field have been well documented [see Yanai and Johnson (1993) for a review]. An ensemble of cumulus convection modifies the temperature and moisture of the environment by the part of environmental vertical motion which compensates the convective mass flux and by the effects of heat and moisture detrained from cumulus clouds and evaporation of detrained cloud water (Ooyama 1971; Yanai et al. 1973; Arakawa and Schubert 1974). The effects of a cumulus ensemble on the large-scale temperature and moisture fields are respectively expressed (in the pressure coordinate system) as

$$Q_{1c} = -M_c \frac{\partial \bar{s}}{\partial p} + \delta(s_D - \bar{s} - L l_D) \tag{1}$$

and

$$Q_{2c} = L M_c \frac{\partial \bar{q}}{\partial p} - L\delta(q_D - \bar{q} + l_D), \tag{2}$$

where Q_{1c} and Q_{2c} are the contributions from the cumulus ensemble to the *apparent heat source* and the *apparent moisture sink*. M_c is the convective mass flux, δ the mass detrainment, $s = c_p T + \phi$ the dry static energy per unit mass of air, c_p the specific heat of air at constant pressure, T the temperature, ϕ the geopotential, q and l are respectively the mixing ratios of water vapor and liquid water, and L the latent heat of vaporization. The subscript D denotes the quantities detrained from the cumulus ensemble. The overbar denotes the running horizontal average. (1) and (2) have been derived by assuming the conservation of mass, water substance $(q + l)$ and moist static energy $(c_p T + \phi + Lq)$ of convective elements.

Assuming that the horizontal momentum of the convective element is also conserved, Ooyama (1971) and Arakawa and Mintz (1974) obtained expressions equivalent to (1) and (2) for the effects of a cumulus ensemble on the momentum field of the environment. Then cumulus clouds act to decelerate the environmental motion as *cumulus friction* (Schneider and Lindzen

1976). As Ooyama (1971) noted, however, a buoyant convective element rising though the environment with a strong vertical shear of horizontal winds may generate a horizontal pressure gradient and affects the horizontal motion of the environment.

Shapiro and Stevens (1980) included the "convection-induced" horizontal pressure gradient force in the expression for the cumulus ensemble effect on the environmental momentum in the form of

$$\mathbf{X}_c = -M_c \frac{\partial \bar{\mathbf{v}}}{\partial p} + \delta(\mathbf{v}_D - \bar{\mathbf{v}}) + \sigma\left(\frac{1}{\rho}\nabla p^*\right), \tag{3}$$

where $\mathbf{X}_c$ is the cumulus-induced acceleration of the large-scale motion (or the *cumulus friction*), $\mathbf{v}$ is the horizontal wind vector, ∇ the horizontal gradient operator, p^* the convection-induced pressure perturbation, and σ the fractional area covered by cumulus clouds. Analogous to (1) and (2), a cumulus ensemble modifies the environmental momentum field through *the part of environmental vertical motion compensating convective mass flux and the detrainment of excess momentum from clouds*. In addition, the last term in (3) shows that *the cumulus ensemble interacts with the environment through the convection-induced pressure gradient force*. Several authors have included this term in the parameterization of convective momentum transport (CMT) (Zhang and Cho 1991a, b; Wu and Yanai 1994; Kershaw and Gregory 1997; Gregory et al. 1997).

As shown by (3), the convection-induced acceleration $\boldsymbol{X}_c$ depends on not only the vertical advection of $\bar{\mathbf{v}}$ due to the compensating mass flux $(-M_c)$ and the detrainment of excess momentum, but also the perturbation pressure gradient force whose magnitude and direction would crucially depend on the geometry of convective organization. Observations during the Global Atmospheric Research Program (GARP) Atlantic Tropical Experiment (GATE) and subsequent field experiments have demonstrated that the vertical momentum transport associated with organized cumulus convection (cloud clusters) can be *downgradient* (of opposite sign to the vertical shear) or *upgradient* (of the same sign as the vertical shear), depending on the type of convective organization.

LeMone (1983) found that the vertical transport of horizontal momentum normal to a line of cumulonimbus is *upgradient*, while the transport of momentum parallel to the line is *downgradient*. The transport of the momentum associated with squall lines has been studied further by LeMone et al. (1984) and others [see LeMone and Moncrieff (1994) for a review]. For a squall line observed over Oklahoma-Kansas during AVE-SESAME (Atmospheric Variability Experiment-Severe Environmental Storms and Mesoscale Experiment), Wu and Yanai (1994) showed that the momentum budget residual acts to increase the vertical shear of line- normal component of environmental wind in the upper troposphere. For the momentum transport by convective bands, there has been an approach of parameterization of CMT independent of (3), that explicitly models the squall lines (e.g., Moncrieff 1981, 1992).

On the other hand, for a non-squall cloud cluster observed during GATE, Ooyama [unpublished manuscript 1980; cited by Houze and Betts (1981)] found a fascinating vorticity couplet in the upper troposphere, which was subsequently confirmed by Tollerud and Esbensen (1983) and by Sui and Yanai (1986). The vorticity couplet was shown to be a result of mixing of the upper-tropospheric easterly momentum with the lower-tropospheric westerly momentum carried upward by convection near the center of the cluster, providing a clear example of downgradient transport of momentum. Wu and Yanai (1994) also showed evidence of the upper tropospheric wind deceleration observed in a mesoscale convective complex (MCC) during SESAME.

There is still a need for observationally assessing the significance of the CMT effects in the large-scale atmospheric circulation. Although there has been some progress made using numerical general circulation models (GCMs) and cloud-resolving models (CRMs) (e.g., Zhang and McFarlane 1995; Inness and Gregory 1997; Mapes and Wu 2001), our knowledge of the impact of CMT on the large-scale circulation is still meager. We have not used a sufficiently long record of high-quality soundings to systematically investigate the effects of CMT on the large-scale motion.

There are basically two methods to diagnose the CMT effects with observations. One is via direct measurements of momentum transport with airborne Doppler radars (e.g., Lewis et al. 1998; Roux 1998; Bousquet and Chong 2000). The other is through the momentum budget approach using the large-scale network of upper-air soundings (e.g., Gallus and Johnson 1992; Wu and Yanai 1994; Carr and Bretherton 2001). While the former approach is more suited for detailed study of CMT associated with particular convective systems, it can cover only a small area in each measurement. In the present work, we take the second approach using the in-situ soundings taken during the Intensive Observing Period (IOP) of the Tropical Ocean Global Atmosphere (TOGA) Coupled Ocean Atmosphere Response Experiment (COARE) which provided a broad view of the convection-coupled atmosphere over the western Pacific warm pool over an extended period from 1 November 1992 through 28 February 1993.

The main purpose of Part I of this paper is to discuss general features of the momentum budget residual, $\mathbf{X}$, obtained in the vicinity of the Intensive Flux Array (IFA) of the TOGA COARE during the 4-month period. In section 2 the procedures to deduce the effects of convective transports of heat and momentum as well as the objective analysis of the upper-air soundings are described. Section 3 presents an overview of the convective activities and other significant meteorological events during the IOP. Then this section discusses the highly variant and multi-scale behaviors of the time series of the convective activity, mean flow $\bar{\mathbf{v}}$ and residual $\mathbf{X}$ using the power spectral analysis and the structure function analysis technique. In section 4, the IOP means and standard deviations of $\mathbf{X}$ and its components are presented. Then this section examines the IOP mean characteristics of the processes associated with $\mathbf{X}$ such as convection-induced deceleration and kinetic energy conversion between the mean flow and convection. Section 5 presents a summary and discussions.

2. Data and methods

a. The TOGA-COARE datasets

The general goals of the TOGA COARE were to provide a better understanding of the role of the warm-pool regions in the mean and transient states of the tropical ocean-atmosphere system (Webster and Lukas 1992). The COARE IOP lasted from 1 November 1992 to 28 February 1993 in an amplification phase of an El Niño episode (e.g. Gutzler et al. 1994). During the IOP, the observation arrays were deployed over the equatorial western Pacific warm pool, as shown in Fig. 1. Syntheses of the large-scale flow features during the TOGA COARE may be found in Velden and Young (1994) and Lin and Johnson (1996a).

Our primary data are the upper-air soundings merged with wind profiler data provided by the University Corporation for Atmospheric Research (UCAR) (Loehrer et al. 1996). The wind, temperature, humidity, time, and position of the sounding instruments were recorded at the surface and at 5-hPa intervals. Our domain of analysis covered the COARE Outer Soundings Array (OSA) including the Intensive Flux Array (IFA), with twelve 6-hourly sounding stations in total (Fig. 1). The data from nine 12-hourly sounding stations surrounding the OSA and from the buoy of the Improved Meteorological Instruments for Ships and Buoys (IMET) at the center of the IFA were auxiliary for the data quality check and filling process. Bad data were eliminated first according to the quality check flags supplied by UCAR. The remaining data formed a set of ensemble statistics for each variable of wind, temperature and humidity at each pressure level. All data deviating more than 3.5 times of the standard deviation from the ensemble mean were rejected. At every observation time, the surface pressure at each station in the OSA was adjusted to the sea level by integrating the hydrostatic equation downward. By doing so, we suppressed the potential error in the momentum budget calculation caused by the time-varying sounding release heights at ship stations, as reported by Carr and Bretherton (2001). For each station, missing data were filled with one of the following three methods: (1) via linear interpolations in vertical if the missing intervals were smaller than 50 hPa, (2) via linear regressions between the targeted station and several available stations that had shown high correlation with it, and (3) via linear interpolations in time if the missing periods were shorter than 12 hours. The final sounding composite over the twelve stations in the OSA contains no missing data from the surface to 70 hPa in the whole IOP.

In the COARE domain, we used a version of the successive correction method (e.g., Daley 1991, Chapter 3) to objectively interpolate the station soundings onto the background fields formed by the $2.5° \times 2.5°$, 6-hourly European Centre for Medium-range Weather Forecasts (ECMWF) Re-Analysis (ERA) level III-B global surface and upper-level datasets (Gibson et al. 1997). The surface objective analysis was done with nine sounding stations with launching sites no higher than 30 meters from the sea level. From 1000 to 70 hPa, all sounding stations in the OSA were used. Above 70 hPa, because of insufficient numbers of soundings, the ERA data at 50, 30, and 10 hPa were adopted. The resultant objective analysis (OBA_{UCLA} hereafter) comprises 6-hourly surface pressure (P_S), zonal wind ($\bar{u}$), meridional wind ($\bar{v}$), temperature ($\bar{T}$), water vapor mixing ratio ($\bar{q}$), and geopotential ($\bar{\phi}$) on a $2.5° \times 2.5°$ horizontal grid covering the area of $130°$-$180°$E, $20°$S-$20°$N, at 0000, 0600, 1200, and 1800 UTC. The geopotential height was derived by integrating the hydrostatic equation from sea level. There are 43 vertical levels, including the surface, 37 levels at 25-hPa intervals from 1000 to 100 hPa, and 80, 70, 50, 30, and 10 hPa levels (Fig. 2). Since the calculation of momentum budget residual requires a dataset of the best available quality, we mainly analyze the $152.5°$–$157.5°$ E, $5°$ S– Equator box covering the IFA (see Fig. 1). In the following, this box is referred to the "IFA region" for brevity.

In addition, the 6-hourly subset of deep convection index I_{TBB} in the COARE domain (Nakazawa 1995) is taken to identify deep convective centers. I_{TBB} derived from the $0.1° \times 0.1°$, hourly Geostationary Meteorological Satellite (GMS) Infrared data, is defined as

$$I_{\mathrm{TBB}} = (225 - T_{\mathrm{BB}})/5, \text{ if } T_{\mathrm{BB}} \le 225\mathrm{K},$$
$$I_{\mathrm{TBB}} = 0, \text{ if } T_{\mathrm{BB}} \ge 225\mathrm{K}. \tag{4}$$

In (4), T_{BB} is the equivalent black body temperature. Finally, the 5-day gauge- and satellite-based rainfall rate from the Climate Prediction Center (CPC) Merged Analysis of Precipitation (CMAP) (Xie and Arkin 1997) is used to indicate overall convective activity during the COARE IOP.

b. Computations of heat and momentum budget residuals

Following Yanai et al. (1973), the *apparent heat source*, Q_1, is computed and interpreted by

$$Q_1 \equiv c_p\left(\frac{p}{p_0}\right)^{\kappa}\left(\frac{\partial\bar{\theta}}{\partial t} + \bar{\mathbf{v}}\cdot\nabla\bar{\theta} + \bar{\omega}\frac{\partial\bar{\theta}}{\partial p}\right) \tag{5a}$$

$$= Q_{\mathrm{R}} + L(\bar{c} - \bar{e}) - \nabla\cdot\overline{s'\mathbf{v}'} - \frac{\partial}{\partial p}\overline{s'\omega'}, \tag{5b}$$

where θ is the potential temperature, ω the vertical p-velocity, p_0 = 1000 hPa, $\kappa = R/c_p$ with R the gas constant of dry air, Q_{R} the radiative heating rate, and c and e are the rates of condensation and evaporation (of cloud water) per unit mass of air, respectively. Similarly, the momentum budget residual, **X**, is obtained and interpreted by

$$\mathbf{X} = (X, Y) \equiv \frac{\partial\bar{\mathbf{v}}}{\partial t} + \bar{\mathbf{v}}\cdot\nabla\bar{\mathbf{v}} + \bar{\omega}\frac{\partial\bar{\mathbf{v}}}{\partial p} + \nabla\phi + \lambda\mathbf{k}\times\bar{\mathbf{v}} \tag{6a}$$

$$= -\nabla\cdot\overline{\mathbf{v}'\mathbf{v}'} - \frac{\partial}{\partial p}\overline{\mathbf{v}'\omega'}, \tag{6b}$$

where λ is the Coriolis parameter.

In deriving (5) and (6) we have assumed that the Reynolds conditions and their consequences hold with sufficient accuracy [see Yanai and Johnson (1993) for a detailed discussion]. The overbar denotes the running horizontal average with respect to a large-scale area and the prime denotes the deviation from the average. The variables resolved at the grid points of the $\mathrm{OBA}_{\mathrm{UCLA}}$ used in this study are regarded as "large-scale" and are marked with overbars when mentioned in the paper. The apparent heat source, Q_1, computed with the large-scale variables by (5a) is interpreted by (5b) to represent the total effect of radiative heating, latent heat released by net condensation, and the convergence of fluxes of sensible heat due to subgrid-scale eddies such as cumulus convection and turbulence. Likewise, the momentum budget residual, $\mathbf{X}$, obtained from (6a) is interpreted by (6b) as the acceleration of the large-scale horizontal wind through the convergence of momentum flux due to convection and other unresolved eddies. Traditionally, the eddy horizontal transport terms $-\nabla \cdot \overline{s'\mathbf{v}'}$ and $-\nabla \cdot \overline{\mathbf{v}'\mathbf{v}'}$ have been ignored in the discussion of convection-large-scale interaction. This approximation has been discussed by Arakawa and Schubert (1974) and Wu (1994).

In the actual computations, we use finite difference forms of (5a) and (6a). In Fig. 2, a vertically staggered grid for the budget calculations is shown. Q_1 and $\mathbf{X}$ are computed between two isobaric levels where the original data are given. The finite difference schemes to calculate the budgets are written to conserve certain vector identities so that computations using the advective forms agree with those from the flux forms. Accurate evaluations of Q_1 and $\mathbf{X}$ from (5a) and (6a) using observational data critically depend on the accuracy of estimates of the large-scale vertical p-velocity, $\overline{\omega}$. Near the tropopause where the static stability is large, even small errors in $\overline{\omega}$ introduce large erroneous values of Q_1. Therefore, a thermodynamical constraint $Q_1 \sim Q_\mathrm{R}$ is imposed at the tropopause level (P_T) (Nitta 1977), while P_T is determined by a stratification test at every grid point and every observation time as in Tung et al. (1999). In practice, Q_1 is approximated to zero at P_T based on Liou (1992). An adjustment scheme is applied to the vertical distribution of divergence. $\overline{\omega}$ is then obtained by integrating the continuity equation with the adjusted divergence.

If the vertical convergence term, $-\frac{\partial}{\partial p}\overline{\mathbf{v}'\omega'}$, dominates in (6b), we can estimate the vertical flux of horizontal momentum caused by convection and turbulence by integrating $\mathbf{X}$ from the surface pressure P_S to a pressure level p:

$$\mathbf{F} = (F_x, F_y) = \overline{\rho\mathbf{v}'w'}\big|_{P_S} - \frac{1}{g}\int_p^{P_\mathrm{S}} \mathbf{X}\,dp = \overline{\rho\mathbf{v}'w'}\big|_p \tag{7}$$

with the vertical velocity $w \approx -\omega/(\rho g)$, ρ the density of air. $\overline{\rho\mathbf{v}'w'}\big|_{P_S} = -\tau_\mathrm{s}$ is the momentum flux at the surface and τ_s is the surface wind stress estimated with the bulk parameterization algorithm by Fairall et al. (1996).

c. Constraint on the momentum budget residual

Because a large number of terms are involved in (6a), obtaining a reliable estimate of $\mathbf{X}$ from large-scale sounding data has been a challenging task. Especially, accurate estimates of the horizontal pressure gradient force constitutes a major problem in momentum budget computations (e.g., Holland and Rasmussen 1973). The errors in $\mathbf{X}$ contaminate $\mathbf{F}$ through (7). In addition to all the possible errors, the eddy horizontal transport term in (6b), $-\nabla \cdot \overline{\mathbf{v}'\mathbf{v}'}$, may not always be negligible. Even with very small errors in $\mathbf{X}$, the $\mathbf{F}$ at 10 hPa, i.e., $\mathbf{F}(P_{10})$, calculated from (7) can be one order of magnitude larger than the surface stress during COARE IOP. To reduce systematic errors as best as we can, a constraint on the column integrated values of $\mathbf{X}$ is imposed.

We assume the raw estimate $\mathbf{X}_o$ consists of the "true" value $\mathbf{X}_1$ and error $\mathbf{R}$, i. e.,

$$\mathbf{X}_o = \mathbf{X}_1 + \mathbf{R}\text{, and} \tag{8}$$

$$\mathbf{X}_1 = -\frac{\partial}{\partial p}\overline{\mathbf{v}'\omega'}\,. \tag{9}$$

$\mathbf{R}$ may also contain the effect of $-\nabla \cdot \overline{\mathbf{v}'\mathbf{v}'}$. Integrating (8) with (9) from P_{10} to P_S, we obtain

$$\begin{aligned} \langle \mathbf{X}_o \rangle &= \langle \mathbf{X}_1 \rangle + \langle \mathbf{R} \rangle \\ &= \overline{\rho \mathbf{v}'w'}\big|_{P_S} - \overline{\rho \mathbf{v}'w'}\big|_{P_{10}} + \langle \mathbf{R} \rangle\,, \end{aligned} \tag{10}$$

where $\langle\ \rangle \equiv \frac{1}{g}\int_{P_{10}}^{P_S}(\ )dp$. Large uncertainty exists for the momentum flux above the tropopause level. It has been shown that tropical convection can trigger vertically propagating gravity waves that carry vertical momentum flux into the lower stratosphere (e.g., Fovell et al. 1992; Alexander and Pfister 1995; Alexander and Holton 1997). Piani et al. (2000) argue that convection-triggered gravity waves may account for 15–30% of the required forcing of the quasi-biennial oscillation (QBO).

Assuming that the momentum flux is ultimately absorbed in the lower stratosphere, we set $\mathbf{F}(P_{10}) = \overline{\rho\mathbf{v}'w'}|_{P_{10}} = 0$ in (10). Then the column integrated error is given by

$$\langle\mathbf{R}\rangle = \langle\mathbf{X}_{\rm o}\rangle - \mathbf{F}(P_{\rm S}) . \tag{11}$$

To adjust the value of $\mathbf{X}$ at each data level, we consider two methods. Method 1 applies constant adjustment of $\mathbf{X}$ with height,

$$\mathbf{X}_1 = \mathbf{X}_{\rm o} - g\frac{\langle\mathbf{R}\rangle}{(P_{\rm S} - P_{10})} . \tag{12}$$

Method 2 assumes errors to increase linearly with decreasing pressure,

$$\mathbf{X}_1(p) = \mathbf{X}_{\rm o} - 2g\frac{\langle\mathbf{R}\rangle(P_{\rm S} - p)}{(P_{\rm S} - P_{10})^2} . \tag{13}$$

Because both adjustments are made at every observation time independently, they can disrupt the continuity of data in time; consequently, the auto-correlations of the original time series may be modified. Therefore, in section 3, we choose to apply the power spectra and the second-order structure functions to the time series of $\mathbf{X}_{\rm o}$. When the physical constraint in vertical is required, as in section 4, we emphasize the results obtained from $\mathbf{X}_1$ while those from $\mathbf{X}_{\rm o}$ are also shown to confirm the credibility of our analyses.

3. Time-series analysis of the momentum budget residual

a. Overview of the atmospheric phenomena in the IFA during IOP

Figures 3a-c are the IOP time-height sections of the OBA$_{\rm UCLA}$ $\bar{u}$, $\bar{v}$, and Q_1, averaged over the IFA region. Figure 3d illustrates the 5-day CMAP and the $I_{\rm TBB}$ data averaged in the same region. The $I_{\rm TBB}$ and the CMAP data are consistent with each other. Except for the 5-day CMAP data, a two-day running mean has been applied to all the time series shown in these figures for the purpose of clearer illustrations. In Fig. 3a, three major westerly wind bursts (WWBs) associated with the Madden-Julian oscillation (MJO) (Madden and Julian 1971) events are identified. The first one is only partially sampled from early to mid November at the end of a MJO event that started in late October. The second one is observed from late December to early January, and the third is around early February. The WWB events initiate from the lower troposphere and gradually extend to the middle and upper troposphere. At the same time, strong baroclinic structures are observed as the easterlies above the 300 hPa level are also enhanced. When the WWBs occur, the signals in the meridional wind are also amplified, as seen in Fig. 3b. Following the third WWB, the Australian monsoon flow moves in and the zonal wind remains westerly below 500 hPa till the end of February. The MJO and the associated convective events during the IOP have been extensively studied (e.g., Nakazawa 1995; Lin and Johnson 1996a, b; Yanai et al. 2000).

In Fig. 3c, deep convective events are generally represented by maxima of Q_1 in the middle to upper troposphere. The validity of the computed Q_1 is confirmed by the rainfall and I_{TBB} data shown in Fig. 3d. From Figs. 3c and d, we see that the IFA is often convectively active during the IOP. In the IFA, deep convection tends to develop before the WWBs (see Fig. 3a). As the WWBs peak, deep convection becomes less significant; instead, a shallower type of convection dominates. Detailed investigations have shown, however, that besides deep convection, shallow cumulus and cumulus congestus that detrains at the freezing level also occur during the IOP (Johnson and Lin 1997; Johnson et al. 1999). Tung et al. (1999) found at least three basic modes of convective heating profiles characterizing different convective regimes in this region.

b. Time series and spectra of the momentum budget residual

From Figs. 3c and d, we recognize that deep cumulus convection induces significant heating in the troposphere between 600-300 hPa. In view of the common mechanisms inducing the heating and the acceleration/deceleration by convective processes [see Eqs.(1) and (3)], we will examine the dynamic fields at 500 hPa. Even though we limit our discussions to the variables at 500 hPa, the observations are also qualitatively similar at 900, 850, and 300 hPa. To find evidence of the convection-motion coupling, we also study the time series of I_{TBB} in the IFA region.

Figures 4a and b show the time series of U_{500} and V_{500}, i. e., $\bar{u}$ and $\bar{v}$ of OBA_{UCLA} at 500 hPa averaged in the IFA region. Figures 5a and b present the time series of the raw estimates of momentum budget residuals X and Y from OBA_{UCLA} averaged over the same region and between 525 and 475 hPa (called X_{500} and Y_{500}). No time filtering has been applied to these time series. X_{500} and Y_{500} contain more variance with higher frequencies than U_{500} and V_{500} do. In addition, X_{500} and Y_{500} involve more burst events. The time series of U_{500} (Fig. 4a) appears to be the most non-stationary among all, due to the dominant signals of the three MJO events.

The visual observations of the time series can be made quantitative with power spectral analysis. Figure 6a shows the power spectral densities (PSDs) of U_{500} and V_{500} and Fig. 6b those of X_{500} and Y_{500}. The PSDs were obtained by the fast Fourier transform (FFT) algorithm. The PSD of V_{500} is one order of magnitude smaller than that of U_{500}, while the PSDs of X_{500} and Y_{500} are on the same order. We note that in Fig. 6a, the signals of MJO in V_{500} are not much stronger than those near 4-5 days or near 10 days corresponding to various types of tropical waves. On the other hand, the MJO is associated with the strongest power in U_{500}, X_{500} and Y_{500} around the 30-day period. As inferred from the original time series, U_{500} and V_{500} have most power concentrating in the periods longer than 4 days. We can detect only very small spectral peaks corresponding to the diurnal cycle and the two-day oscillation in U_{500} and V_{500}.

In contrast, X_{500} and Y_{500} have power extending into much shorter periods. In Fig. 6b, in addition to the ~20–60-day band of the MJO, we can identify the peaks around 10, 4-5, 2-3 days, and that of the diurnal cycle. This multi-scale feature has also been found in the deep convection over the tropical western Pacific represented by satellite imagery (e.g., Nakazawa 1988; Sui and Lau 1992). Figure 6c shows the PSD of I_{TBB}, a measure of deep convection, over the IFA region during the IOP. The PSD of the I_{TBB} has very similar multi-scale features with nearly identical spectral peaks as the PSDs of X_{500} and Y_{500} in Fig. 6b. The similarity of the PSD of I_{TBB} with those of X_{500} and Y_{500} suggests a link between deep convection and the acceleration/deceleration of large-scale momentum field.

c. Fractal characteristics in the COARE dataset

The multi-scale time series of X_{500} and Y_{500} are further shown to have fractal properties, i.e., they are characterized with certain power-law scaling behaviors (e.g., Mandelbrot 1982). The second-order structure function,

$$D_2(\Delta t) = \{|A(t+\Delta t) - A(t)|^2\}, \tag{14}$$

is often used to explore the fractal properties (e.g., Monin and Yaglom 1975). In (14), $A(t)$ is the time series of interest, Δt a possible time interval, and $\{\ \}$ the ensemble average of all possible pairs of $|A(t+\Delta t) - A(t)|^2$ in which $A(t+\Delta t) - A(t)$ is an increment in $A(t)$. $A(t)$ is fractal if the following power-law behavior exists:

$$D_2(\Delta t) \sim \Delta t^{\zeta_2}, \tag{15}$$

with ζ_2 the scaling factor.

Figures 7a and b, respectively, show the logarithmic values of the second-order structure functions of X_{500} and Y_{500} plotted against those of Δt ranging from 1/4 day to 64 days. In both figures, we find power-law scaling regions in which (15) is satisfied. The scaling regions are characterized by the best-fit lines with slopes of ζ_2 values. It is noted that scaling properties can not be ascertained beyond 16-20 days because of too few samples. There are power-law scaling regions starting from 1/4 to about 20 days and longer in X_{500} and 1/4 to 16 days in Y_{500}, with $\zeta_2 = 0.09$ and 0.17, respectively. Besides the X_{500} and Y_{500} time series, most variables shown in this paper contain certain fractal features. For example, Fig. 8 shows the log-log plot of the $D_2(\Delta t)$ of U_{500}, which has scaling region from ~1 day to ~16 days with a scaling factor of 0.68. Furthermore, one may notice that the $\Delta t = 1\text{day}$ is a scaling break for the $D_2(\Delta t)$ of U_{500}, which is not clear in those of X_{500} and Y_{500}.

The fractal properties in Figs. 7 and 8 suggest the existence of processes which are common in all temporal scales as long as the power laws hold. Through the Wiener-Khinchin theorem (e. g., Monin and Yaglom 1975), ζ_2 is related to the decay in frequency of the PSD of a time series that has stationary increments. When $0 < \zeta_2 < 2$, the time series $A(t)$ is nonstationary with stationary increments, as discussed by Davis et al. (1994). Therefore, the time series U_{500} is nonstationary with stationary increments from about 1 to 16 days. X_{500} and Y_{500}, on the other hand, are weakly nonstationary with stationary increments from about 1/4 day up to about 16-20 days. Furthermore, X_{500} and Y_{500} are found to be "multifractal", i.e., their structure functions of various order satisfy different types of power laws.

4. The IOP mean characteristics of the momentum budget residual

a. IOP mean and variance of the momentum budget residual

Figures 9a-e show, respectively, the IOP mean vertical distributions of the zonal components of the local time change, horizontal advection, vertical advection, Coriolis, and pressure gradient terms in (6a) obtained from the OBA_{UCLA}. The mean values are shown together with the standard deviations. In the same fashion, Fig. 9f shows the sum of all terms, X, and its standard deviation. Notice that the horizontal scales of Figs. 9e and f are twice as large as those for the other figures. The most noticeable features in these diagrams are the large standard deviations as opposed to the small mean values, which also exist in the meridional components (not shown). Such features, however, are not surprising. As shown in Figs. 5a-b, time series of X and Y are highly variant reflecting various modulating processes such as the MJO, and their standard deviations are one to two orders of magnitude larger than the mean values. Hence, we take a view that not only the mean but also the variance contain physically meaningful signals.

Figure 10a, b, c illustrate the IOP-mean profiles of $\bar{u}$ and the impact of the adjustments described in section 2c on the mean profiles of X and F_x. Similarly, the IOP-mean profiles of $\bar{v}$, Y, and F_y are shown in Fig. 11. The profiles are plotted from sea level to 10 hPa. The open circles show the values before the adjustment, triangles are values using the adjustment Method 1, and closed circles are values resulted from Method 2. As seen in Fig. 10b and 11b, Method 2 attributes more errors in X and Y to the higher tropospheric levels and keeps the surface values intact, which better reflects the uncertainty in the geopotential gradient in (6a). Moreover, in Fig. 10c, Method 1 causes a change of sign in F_x while Method 2 tends to preserve the original sign before bringing $F_x(P_{10})$ to zero. In Fig. 11c, values of F_y produced by Method 1 are larger than those by Method 2. In overall, Method 1 seems too drastic an adjustment; therefore, Method 2 is adopted in this section. Because both $\bar{\mathbf{v}}$ and $\mathbf{X}$ vary during the IOP, detailed comparisons of their mean profiles are not meaningful unless further stratification of data is made.

b. Mean frictional deceleration

In Fig. 12 we show the IOP-mean vertical distributions of the quantities $\bar{u}X/|\bar{u}|$ and $\bar{v}Y/|\bar{v}|$ with their standard deviations. Positive (negative) $\bar{u}X/|\bar{u}|$ measures the acceleration (deceleration) of the zonal component of environmental flow resulting from X along the flow. $\bar{v}Y/|\bar{v}|$ has a similar meaning for the meridional component. $\bar{u}X/|\bar{u}|$ and $\bar{v}Y/|\bar{v}|$ were computed at each observation time before the IOP means are taken. Values of $[\bar{u}X/|\bar{u}|]$, where [ ] represents the IOP mean, using the unadjusted as well as adjusted momentum budget residuals (sections 2c and 3a) are negative throughout the troposphere, indicating that the momentum budget residual is acting to *decelerate* the zonal component of environmental flow (Fig. 12a). Furthermore, the values obtained from the adjusted residual show large deceleration (~1 m s^{-1} day^{-1}) near the surface and its gradual upward decrease to the minimum (~ 0) at 300 hPa. Then large deceleration (~ 2 m s^{-1} day^{-1}) occurs in a layer immediately below the tropopause (~ 125 hPa), suggesting the role of detrainment of excess momentum from deep cumuli [see Eq. (3)]. The values of the IOP mean of $[\bar{v}Y/|\bar{v}|]$ are also negative in most of the troposphere (Fig. 12b). The mean deceleration rate in the lower troposphere is ~ 0.5–1.0 m s^{-1} day^{-1} in both the zonal and meridional directions.

We conclude that as far as their IOP means are concerned, the momentum budget residuals in both directions decelerate the environmental flow, thus acting as friction. Mapes and Wu (2001) reached a similar conclusion based on the results of two-dimensional cloud-resolving model simulations. However, as shown in Fig. 12, there are large standard deviations in both $\bar{u}X/|\bar{u}|$ and $\bar{v}Y/|\bar{v}|$. A detailed inspection of time series of these quantities during IOP reveals that the momentum budget residuals often act to accelerate the environmental flow especially during the initial phase of the WWBs and with prominent squall lines.

c. Downscale transfer of kinetic energy

The quantity, $E = -(\bar{u}X + \bar{v}Y)$, measures the loss of kinetic energy from the mean flow to subgrid-scale eddies. In some early work E was equated to the rate of frictional dissipation (e. g., Kung 1967). However, in a convective atmosphere under study, E as well as buoyant production of eddy kinetic energy (not estimated in this study) must be considered to evaluate the dissipation. E can be rewitten as

$$\begin{aligned} E &= -\mathbf{X} \cdot \bar{\mathbf{v}} = \frac{\partial(\overline{\mathbf{v}'\omega'})}{\partial p} \cdot \bar{\mathbf{v}} \\ &= S + \frac{\partial(\overline{\mathbf{v}'\omega'} \cdot \bar{\mathbf{v}})}{\partial p}, \end{aligned} \tag{16}$$

where

$$S = -(\overline{\mathbf{v}'\omega'}) \cdot \frac{\partial \bar{\mathbf{v}}}{\partial p} = g\left(F_x \frac{\partial \bar{u}}{\partial p} + F_y \frac{\partial \bar{v}}{\partial p}\right) \tag{17}$$

is the mechanical (or shear) production of kinetic energy of subgrid-scale eddies. S is positive when the vertical eddy transport of momentum is downgradient, and it is negative when the momentum transport is upgradient. (16) with (17) shows that E may be interpreted by the conversion term S and a redistribution term. E and S were computed at every observation time before taking the IOP means.

In Fig. 13 we show the vertical distributions of IOP-mean values of E estimated using the unadjusted and adjusted momentum budget residuals and the standard deviations using the adjusted residual. The mean values of E estimated from the two sets of data show similar distributions. $[E]$ is nearly uniform (~ 1×10^{-4} m^2 s^{-3}) from the surface to 725 hPa. The value of $[E]$ estimated using the adjusted residual decreases from this level to the minimum (~ 0) at 250 hPa. There is a sharp increase of $[E]$ from this level to the maximum (~ 5×10^{-4} m^2 s^{-3}) near the tropopause, reflecting the large frictional force (Fig. 12a) and the large zonal wind velocity (Fig. 10a). This profile of $[E]$ has some resemblance to the mean summer (April-September) profile of E at 0000 UTC (evening) obtained by Kung (1967, Fig. 2) over North America for a 5-year (1958-1962) period, although Kung's summer profile shows more pronounced large values near the ground surface. $[E]$ estimated in the lower troposphere for the COARE IOP is on the order of 1.0×10^{-4} m^2 s^{-3} and approximately one half of the mean value of dissipation (1.9×10^{-4} m^2 s^{-3}) obtained by Hocking and Mu (1997, Fig. 13). The IOP-mean vertical integral $[\langle E \rangle]$ estimated from the adjusted residual is found to be 0.8 W m^{-2}. This value can be compared to some previous estimates of dissipation: 3-5 W m^{-2} over North America (Kung 1967) and 2.0 W m^{-2} for the globe (Peixoto and Oort 1992, p. 385).

Figure 14 illustrates the vertical distributions of the IOP-mean values of S using the unadjusted and adjusted momentum fluxes. The distributions are similar to each other except very large negative values above the 200-hPa level for the unadjusted case. S estimated using the adjusted momentum flux is positive (downgradient) throughout most of the troposphere with double peaks near 450 hPa and 250 hPa, although negative (upgradient) values are noted near the tropopause. The results shown in Figs. 13 and 14 indicate that the IOP-mean momentum transport is downgradient and kinetic energy is transferred from the large-scale motion to subgrid-scale eddies. However, as suggested by the standard deviations in Fig. 13, this conclusion does not exclude the occurrence of upgradient momentum transport and upscale energy transfer in certain situations. In Part II of the present paper, we will show examples of upgradient momentum transports associated with the squall lines and WWBs observed during the TOGA-COARE IOP.

5. Summary, discussion, and conclusions

The TOGA-COARE IOP provides a 4-month long observation of convectively active tropical atmosphere over the western Pacific warm pool. We utilized the sounding data taken during the IOP to study the convective momentum transport (CMT) in the vicinity of the COARE IFA. The effects of CMT are deduced from the residual of the large-scale momentum budget, $\mathbf{X} = (X, Y)$. The major points of this paper are summarized as follows:

- The time series of raw estimates of X and Y sampled at various levels from the lower to upper troposphere exhibit highly transient behavior, showing the modulations by the MJO and other disturbances.
- The MJO (20-60-day) signal is the most significant in the PSDs of $\bar{u}$, X and Y, while it is not in the PSD of $\bar{v}$. In addition, much of the variance in X and Y extends into higher frequencies showing spectral peaks near 10, 4-5, 2-3 days, and at the diurnal cycle. These spectral peaks are also evident in the PSD of I_{TBB}, an index of deep convective activity.
- The temporal behaviors of X and Y are characterized as fractal from 1/4 to ~ 20 days and 1/4 to ~ 16 days, respectively. The second order structure functions of X and Y time series at 500 hPa have the power law scaling properties with $\zeta_2 = 0.09$ and 0.17, respectively. These time series fall into the category of being weakly nonstationary but with stationary increments. The temporal behavior of X and Y manifests itself as the very large standard deviations of their IOP-mean values.
- The IOP-mean accelerations of the large-scale motion due to (X, Y) are studied by examining $\bar{u}X/|\bar{u}|$ and $\bar{v}Y/|\bar{v}|$. These quantities are mostly negative in the troposphere, i.e., the large-scale motion is decelerated by the subgrid-scale processes. The mean rate of deceleration in the troposphere is ~ 0.5–1.0 m s^{-1} day^{-1} in both directions.
- The IOP mean rate of kinetic energy transfer, $[E] = -[\bar{\mathbf{v}} \cdot \mathbf{X}]$, is positive in the troposphere. $[E]$ is uniform from the surface to ~ 700 hPa (~ 1×10^{-4} m^2 s^{-3}), then decreases from this level to ~ 250 hPa, but a sharp increase is observed in a layer between 250 hPa and the

tropopause.

- The examination of the product of the momentum flux and the wind shear shows that, on average, the vertical momentum transport is downgradient, i.e., kinetic energy is converted from the large-scale motion to subgrid-scale eddies.

We note that the peaks found in the spectra of X, Y, and I_{TBB} correspond well to those known for the tropical waves and diurnal variation (e.g., Yanai et al. 1968; Wallace 1971; Zangvil and Yanai 1980; Takayabu 1994; Chen and Houze 1997). The remarkable similarity of the spectra suggests a dynamical link between cumulus convection and the large-scale tropical motion through CMT.

This study shows that the frictional deceleration and downscale kinetic energy transfer take place in a deep tropospheric layer in the cumulus convective atmosphere over the warm pool, in contrast to the mixed layer over land and ocean without deep clouds (e.g., Lenschow 1970; Pennell and LeMone 1974; LeMone 1980; Stevens et al. 2001). The study also suggests the role of momentum detrainment from the top of deep cumulus clouds affecting the large-scale flow in the tropical upper troposphere. This effect had been inferred from the wind deceleration and the generation of vorticity couplets above mesoscale convective systems (e.g., Tollerud and Esbensen 1983; Sui and Yanai 1986; Wu and Yanai 1994; Tung and Yanai 2000).

Recent studies show that the transport of momentum during the COARE IOP is not only highly variant in time but also clearly case-dependent (e.g., Lewis et al. 1998; Roux 1998; Bousquet and Chong 2000; Tung and Yanai 2000). Accelerations and upgradient momentum transports are also observed in association with the WWBs and prominent squall lines. As suggested by previous studies (e.g., Wu and Yanai 1994), the pattern of spatial organization of cumulus convection is crucial in determining the directions of CMT for individual cases. A statistical study of the X and Y time series, as reported in Part I, does not clearly distinguish the effects of cumulus convection and other sugbrid-scale processes. Hence, it is important to isolate the CMT by examining significantly intense convective events, These will be reported in the forthcoming Part II of this paper.

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Figure Captions

FIG. 1. The TOGA-COARE observation arrays. LSA denotes the Large-scale Soundings Array, OSA the Outer Soundings Array, IFA the Intensive Flux Array. IMET indicates the buoy belonging to the Improved Meteorological Instrumentation operated by the Woods Hole Oceanographic Institution. In this paper, the $152.5°–157.5°$ E, $5°$ S – $0°$ box is referred to the IFA region.

FIG. 2. The vertically staggered grid for calculating the heat and momentum budget residuals. The thin solid lines show the standard isobaric levels in the $\mathrm{OBA}_{\mathrm{UCLA}}$. The dashed lines are the "half-levels" where the budget residuals, Q_1 and $\mathbf{X}$, are calculated. The tropopause level (P_{T}) is determined at each observation time. The geopotential height $\bar{\phi}$, the vertical p-velocity $\bar{\varpi}$, and the eddy momentum flux, $\mathbf{F}$, are calculated at the standard levels. At the surface level (P_{S}), $\mathbf{F} = -\tau_{\mathrm{s}}$ and τ_{s} is the surface wind stress.

FIG. 3. The IOP time-height sections of (a) $\bar{u}$ (m s^{-1}), (b) $\bar{v}$ (m s^{-1}), (c) Q_1 / c_p (K day^{-1}), and (d) the 5-day CMAP data (mm day^{-1}) and the TBB index (I_{TBB}), averaged over the IFA region. For clearer illustrations, a two-day running mean has been applied to all the time series except the CMAP rainfall rate.

FIG. 4. The original time series of the $\bar{u}$ and $\bar{v}$ (m s^{-1}) at 500 hPa averaged in the IFA region, denoted as U_{500} and V_{500}, respectively.

FIG. 5. The time series of the raw estimates of X and Y (m s^{-1} day^{-1}) averaged between 525–475 hPa in the IFA region, denoted as X_{500} and Y_{500}, respectively.

FIG. 6. The power spectral densities (PSDs) of (a) U_{500} and V_{500}, (b) X_{500} and Y_{500}, and (c) I_{TBB}. In (a), the left ordinate indicates the PSD of U_{500}; the right ordinate shows the PSD of V_{500}.

FIG. 7.The log-log plots of the second-order structure functions, $D_2(\Delta t)$ of (a) X_{500} and (b) Y_{500} plotted against Δt. The scaling regions are shown by the best-fit lines with a slope of $\zeta_2 = 0.09$

for X_{500} and $\zeta_2 = 0.17$ for Y_{500}, as written on the lower-right corner of each plot (see text).

FIG. 8. Same as Fig. 7, except for the $D_2(\Delta t)$ of U_{500}, with $\zeta_2 = 0.68$.

FIG. 9. The vertical profiles of (a) the local time change, (b) horizontal advection, (c) vertical advection, (d) Coriolis, and (e) pressure gradient terms and (f) X [see Eq. (6a)] in the IFA region (units: m s^{-1} day^{-1}). The center dark lines depict the IOP mean values, which are shown together with their standard deviations.

FIG. 10. The IOP mean profiles of (a) $\bar{u}$ (m s^{-1}), (b) X (m s^{-1} day^{-1}), and (c) F_x (N m^{-2}) in the IFA region. In (b) and (c), the open circles show the raw estimates before adjustment; triangles are the values after the Method 1 adjustment; the closed circles are the results of the Method 2 adjustment.

FIG. 11. Similar to Fig. 10, except for (a) $\bar{v}$ (m s^{-1}), (b) Y (m s^{-1} day^{-1}), and (c) F_y (N m^{-2}).

FIG. 12. Vertical profiles of the IOP mean acceleration of (a) zonal flow $[\bar{u}X/|\bar{u}|]$ (m s^{-1} day^{-1}) and (b) meridional flow $[\bar{v}Y/|\bar{v}|]$ (m s^{-1} day^{-1}) and their standard deviations in the IFA region. The closed (open) circles are the values obtained from the adjusted (unadjusted) momentum budget residual; standard deviations are obtained using the adjusted values.

FIG. 13. Similar to Fig. 12 but for the IOP-mean rate of downscale kinetic energy transfer $[E]$ and its standard deviation (10^{-4} m^2 s^{-3}) in the IFA region.

FIG. 14. Vertical profiles of the IOP mean S (10^{-4} m^2 s^{-3}) [see Eq. (17)] obtained from the unadjusted (open circles) and adjusted (closed circles) momentum budget residuals in the IFA region.