The oceanic mesoscale convective system and associated mesovortex observed 12 December 1992 during TOGA-COARE

By OLIVIER BOUSQUET* and MICHEL CHONG
Centre National de Recherches Météorologiques, France

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SUMMARY

This study documents the precipitation and kinematic structure of a mature/mature-to-decaying, westward propagating, near-equatorial oceanic mesoscale convective system (MCS) observed by airborne Doppler radars during TOGA-COARE, the Tropical Ocean/Global Atmosphere Coupled Ocean–Atmosphere Response Experiment. This system occurred on 12 December 1992 during the convectively active phase of an intraseasonal oscillation, and was followed, at its dissipating stage, by a redevelopment of convection that led to the MCS observed on 13 December 1992. These two successive MCSs were associated with a two-day atmospheric disturbance. Radar-deduced airflow at two time periods reveal many similarities with other tropical oceanic cloud clusters. In particular, a marked rear inflow jet associated with a midlevel cyclonic mesovortex was observed within the rear of the stratiform region.

The mesovortex was better defined both in size and depth during the mature-to-decaying stage and was located more inside the system within a region of convergence, due to the westward propagation of the rear inflow. Countergradient transports of momentum normal to the mean orientation of the system at midlevels, and downward transports at lower and higher levels helped to intensify this rear inflow. These transports were mostly accomplished by eddy structures. Cyclonic vorticity was concentrated at the rear of the MCS and peaked at the centre of the observed closed wind circulation. As previously observed, stretching of the pre-existing vertical vorticity was the dominant dynamical mechanism which helped to amplify the mesovortex at mid-to-upper levels, while tilting of the horizontal vorticity into the vertical was a lesser mechanism and had an opposite effect. In most respects, advective processes were negatively correlated with the stretching and tilting mechanisms, but not sufficiently to be balanced. The resulting tendency was an increase of cyclonic vorticity at low-to-mid levels and a net decrease above.

KEYWORDS: Doppler radar Observations Tropical dynamics

1. INTRODUCTION

From 1 November 1992 to 28 February 1993, the equatorial western Pacific ocean, commonly referred to as the ‘warm pool’, was the central region of the Intensive Operation Period (IOP) of the Tropical Ocean Global Atmosphere (TOGA) Coupled Ocean–Atmosphere Response Experiment (COARE). The overall objective of COARE was to achieve a better understanding of the role of the warm pool in the atmospheric general circulation (Webster and Lukas 1992) and to provide a better understanding of the processes that organize convection in this region. The experimental setting involved Doppler radars aboard three turboprop aircraft (the two WP-3Ds from the National Oceanic and Atmospheric Administration and the Electra from the National Center for Atmospheric Research), which were essential to provide detailed descriptions of the airflow and precipitation structure of Mesoscale Convective Systems (MCSs) that develop in that region.

According to Chen et al. (1996), who provided an overview of deep convection and large-scale flow over the warm pool region during TOGA-COARE, the convective activity was modulated by episodes of eastward propagating intraseasonal oscillation (ISO) and presented multiscale characteristics within the convective active phase of the ISO. During this phase, deep convection principally occurred in cloud clusters or groups of cloud clusters which were frequently organized into westward propagating 2-day disturbances and presented a distinct diurnal cycle. Three well-defined 2-day disturbances occurred in the vicinity of the Intensive Flux Array (IFA) observational

* Corresponding author, present address: University of Washington/JISAO, BOX 354235, Seattle, WA 98195-4235, USA. e-mail: bousquet@atmos.washington.edu
domain of TOGA-COARE from 11 to 15 December 1992, and two of these were sampled during four successive night aircraft missions from 12 to 15 December.

In a previous paper, Chong and Bousquet (1999) have analysed the characteristics of the 13 December MCS, which was the second of two cloud clusters associated with the 2-day disturbance that occurred on 12–13 December. In this study, we document the mesoscale organization of the first MCS, as derived from Doppler radar data collected by both WP-3D research aircraft on 12 December. The main objectives are to describe the kinematic structure of this MCS, which presented a well-defined mid-level mesovortex in the rear part of its stratiform precipitation region, and to provide some mechanisms involved in the mesovortex development. Section 2 gives an overview of the 12 December MCS and the large-scale flow over the warm pool region. The airflow structure deduced from airborne radar observations is presented in section 3, and mean characteristics are discussed in section 4. Section 5 investigates the physical processes involved in the mesovortex intensification, along with speculative remarks on the possible role of the mesovortex in the 13 December MCS generation.

2. SYNOPSIS OF THE 12 DECEMBER 1992 MCS

The observation of the 12 December 1992 MCS corresponded to the arrival over the IFA of the convectively active phase of the second of three intraseasonal oscillations that occurred during the 4-month IOP (Chen et al. 1996). This system was the first of four successive large and very active cloud clusters that were observed within or near the IFA between 12 and 15 December, and was also coincident with the beginning of a prominent period of anomalous westerly wind bursts (WWBs), which developed near the surface and gradually increased in depth and magnitude (Lin and Johnson 1996).

(a) Environmental conditions

A 6-hourly time series of infrared (IR) observations from the GMS (Geostationary Meteorological Satellite) between 2345 (all times are in UTC +11 h to give local time) on 11 December 1992 and 1745 on 13 December 1992 is shown in Fig. 1. At 2345 11 December (Fig. 1(a)) a field of growing convective clouds (arrow mark) began to develop near 3°N, 168°E, to the south and west of large cloud clusters that were entering dissipating stages. Deep convection then progressively intensified (Figs. 1(b) and (c)), and convective elements were organized into a north-west to south-east oriented band which moved at a speed of approximately 11 m s\(^{-1}\) toward the west-southwest. At 1145 12 December (Fig. 1(c)), the day before clusters dissipated, the present system developed a large cloud shield resulting from the development of growing cumulonimbus anvil. The associated cloud top IR temperature of <208 K in Fig. 1(d) shows that it turned into a large circular cloud cluster before it entered a dissipating stage (Fig. 1(e)). Dissipation continued until 0545 13 December (Fig. 1(f)) as the system progressively approached the IFA. From 0745 13 December (not shown), a renewal of the convective activity within the debris of this decaying cloud cluster led to the development of a new MCS (Figs. 1(g) and (h)), referred to as the 13 December MCS. This system, which was described by Chong and Bousquet (1999), lasted until 0300 14 December 1992 and dissipated near 0.5°S, 153°E. These two systems were associated with one of the three two-day disturbances that occurred near or over the IFA during 11–15 December 1992 (Chen et al. 1996). These authors considered them as distinct systems both having their own life cycle. Moreover, Chen and Houze (1997) found that MCSs during TOGA-COARE (including the 12–13 December cases) reached their maximum extent in the early morning and dissipated in the middle-to-late afternoon.
Figure 1. 6-hourly time series of Geostationary Meteorological Satellite (GMS) infrared radiation pictures between 2345 UTC 11 December and 1745 UTC 13 December. The colour key for cloud-top temperature is shown to the right of the panels. The TOGA-COARE Intensive Flux Array (IFA) domain is also shown. White arrows indicate the location of the 12 December 1992 mesoscale convective system (MCS).
(local time). Previous studies have suggested that this early morning activity was related to the mesoscale circulations inferred from the radiative heating difference between cloudy and clear columns of the atmosphere (Gray and Jacobson 1977), while the middle-to-late afternoon decline was due to the absorption of the solar radiations by upper-level clouds which acts to stabilize the atmosphere and subsequently reduce the convection (Randall et al. 1991). However, these cloud–radiation interactions did not explain the formation of these MCSs in the afternoon. Chen and Houze (1997) found that the ocean surface conditions (e.g. sea surface temperature, surface air temperature) which reached their maxima in the afternoon could favour the formation of cloud systems in this daytime period.

Figure 2 presents two IR satellite pictures within the two time periods during which Doppler radar observations were analysed. It shows that the observed system was first composed of several north-west to south-east oriented precipitation elements (Fig. 2(a)) which gradually decreased to form a large circular cloud cluster two hours later (Fig. 2(b)), while tracking south-westward. In both pictures, deep convection principally occurred within the central part of the MCS, where the area of coldest cloud-top IR temperatures could be observed at any time. According to the classification into five classes (Webster and Lukas 1992), it was a class 3 system (equivalent cloud-top blackbody temperature <208 K between 20 000 and 60 000 km²). One can also see that a convective line developed ahead of the MCS. This line had an evolution similar to the cloud cluster, i.e. it also decayed as it approached the IFA.

Figure 3 shows the flow charts at 850 and 500 hPa, as derived from the European Centre for Medium Range Weather Forecasts (ECMWF) re-analysis (Gibson et al. 1997) on 12 December 1992 at 1800 UTC. The low-level winds (Fig. 3(a)) reveal a north-westerly flow to the east of the IFA, where the cloud cluster was observed. This flow mainly originated from the northern hemisphere north-easterly trade winds that turned to north-westerlies as they progressed toward the equator. It contributed to an apparent cyclonic circulation north of a relatively extensive region of westerlies lying between the equator and 10°S. At midlevels (Fig. 3(b)), easterlies covered most of the domain of observation, with maxima concentrated at 10°N. They were incorporated into a broad cyclonic circulation to the south and south-east of the IFA (as in Chong and Bousquet (1999), ‘cyclonic’ is used here to mean positive relative vorticity which is strictly true only in the northern hemisphere.

Since sounding data were not available in the MCS region on 12 December, the environmental data we present in Fig. 4 were derived from a combination of various data from aircraft and wind profilers. These data were selected to provide the best representation of inflow air ahead of the system (see legend for construction). The thermodynamic profile (Fig. 4(a)) indicates that the atmosphere was convectively unstable and that an apparent layer of dry midlevel air could be observed between 600 and 500 hPa. Considering the mean conditions of the lowest layer of depth 100 hPa, the lifting condensation level was at 942 hPa (0.6 km) and the level of free convection (LFC) was at 900 hPa (1.0 km). The convective available potential energy was 1874 J kg⁻¹ and the convective inhibition that defined the amount of energy required to raise a parcel from the surface to the LFC was 5 J kg⁻¹. Then, once lifted up to the LFC, the parcel could acquire a positive buoyancy up to an altitude of 115 hPa (15.7 km) where the equilibrium temperature level was reached. The 0 °C level was at 564 hPa (4.9 km).

Consistently with the beginning of a WWB period, the wind hodograph (Fig. 4(b)) shows that the flow was north-westerly from the surface up to 850 hPa, with a maximum value of 7 m s⁻¹ near the surface. Easterlies prevailed above 850 hPa, up to 23 m s⁻¹.
Figure 2. As Fig. 1, but for (a) 1645 UTC, and (b) 1845 UTC 12 December, over a smaller area at greater resolution.
Figure 3. ECMWF re-analysed winds over the TOGA-COARE (see text) domain of observation at 1800 UTC 12 December 1992 at (a) 850 hPa, and (b) 500 hPa. The Intensive Flux Array (IFA) domain is indicated by the area enclosed by bold lines.
Figure 4. Composite sounding for the environment of the 12 December mesoscale convective system (MCS) based on the 1705 UTC N43RF aircraft sounding from the surface to 500 hPa, and 2320 UTC Kapingamarangi wind profiler data at higher levels. These soundings are courtesy of Drs LeMone and Trier. (a) Skew T-Log P diagram; and (b) composite wind hodograph, showing components $U$ and $V$ in eastward and northward directions, respectively.

near 110 hPa, with a local peak in the mid-troposphere (550 hPa), which is indicative of a favourable condition for long-lasting convective systems (e.g. Thorpe et al. 1982). Overall, Fig. 4(b) also indicates low- and mid-level shears, with a mean west–east orientation. According to LeMone et al. (1998), the radar observations hereafter discussed, exhibit main and secondary leading convective bands mainly perpendicular and parallel to the low-level shear, respectively.
(b) Radar observations

The two NOAA P3 Doppler-equipped (tail radar) research aircraft (referred to as N42RF and N43RF) were involved in the 12 December plane mission. Between 1600 and 2100, both aircraft collected dual- and quad-Doppler data relative to the convective and stratiform regions of the system. Figure 5 presents two composite reflectivity fields, as observed by the N42RF lower-fuselage radar at time intervals 1625–1705 (Fig. 5(a)) and 1845–1915 (Fig. 5(b)), within a domain of $500 \times 500 \text{ km}^2$. Reference will be made to a Cartesian coordinate system at $0.5^\circ \text{N}$, $160.5^\circ \text{E}$ with $x$ and $y$ axes pointing to the east and north. All data were corrected for advection effects which were estimated with respect to a system motion of $9 \text{ m s}^{-1}$ to the west and $2 \text{ m s}^{-1}$ to the south (as deduced from different sequences of radar reflectivity patterns), and a reference time of 1800 UTC. Note that this propagation speed is consistent with the one ($11 \text{ m s}^{-1}$ to the west and $3 \text{ m s}^{-1}$ to the south) that could be deduced from the successive IR GMS satellite images (Fig. 2). Between 1625 and 1705, the MCS was characterized by a $400 \text{ km}$ long area of stratiform precipitation (reflectivity $>$ 15 $\text{dBZ}$) with a well-defined weak-echo notch in its north-eastern part ($y > 50, x > 50 \text{ km}$). Convective precipitation principally occurred on the south-west of the extended area of stratiform precipitation. A main line of convection, oriented north-west to south-east ($315^\circ$), could be observed at the system leading edge, while secondary west–east oriented bands developed to the south and west of the MCS. From 1845 to 1915 (Fig. 5(b)), the reflectivity pattern presented a more circular structure quite consistent with the satellite IR observation (Fig. 2(b)). Although it still presented a leading region of convective precipitation to the south-west, the convective activity was less pronounced than earlier, suggesting that the system had entered its dissipating stage.

3. AIRFLOW STRUCTURE AND EVOLUTION OF THE MCS

(a) Data and analysis method

In this study, two sequences of approximately 80 and 110 minutes are used to describe the mesoscale structure of the cloud cluster. The first analysis (between 1607 and 1730, hereafter referred to as the T1 interval) was performed using the dual-Doppler data collected by the N42RF aircraft within a domain of $280 \times 240 \times 15 \text{ km}^3$, as reported in Fig. 5(a). For the second analysis, we considered multiple-Doppler data collected by both N42RF and N43RF aircraft between 1830 and 2020 (T2 time interval), within a domain of $280 \times 340 \times 15 \text{ km}^3$, as shown in Fig. 5(b). Both analyses were performed with horizontal and vertical grid resolution set to 4 km and 0.5 km, respectively. Reference time of 1800 UTC, advection speed, and Cartesian coordinate system are similar to those of Fig. 5. Three-dimensional wind fields were derived, following the multiple-Doppler synthesis and continuity adjustment technique (MUSCAT) proposed by Bousquet and Chong (1998), which is a variational approach that provides a simultaneous solution for the three wind components, combining the set of Doppler observations and the physical mass continuity equation.

(b) Horizontal wind fields

Figure 6 presents the system-relative horizontal wind fields and reflectivity patterns at 0.5, 4 and 7 km altitude for the T1 and T2 intervals, respectively.

At low levels (Figs. 6(a) and 6(d)), the general flow was eastward, i.e. from front to rear over all the domain of analysis. The system leading edge (to the west side of the domain) was a region of maximum inflow, while the rear stratiform region was
Figure 5. Flight level 4600 m radar reflectivity composite (dBZ) over the time periods (local time, see text) (a) 1625-1705, and (b) 1845-1915 from the N42RF lower-fuselage radar over a domain of 500 x 500 km². See text for a definition of the coordinate system. The grey scale key for radar reflectivity is shown to the right of each panel. The outlined boxes in (a) and (b) indicate the Doppler analysis domains for T1 and T2 analyses, respectively (see text for definitions).
Figure 6. System-relative horizontal wind vectors and reflectivity contours (from 15 dBZ at intervals of 7.5 dBZ) at 1 km, 4 km, and 7 km above mean sea level. Only every other wind vector is plotted. Panels (a), (b), (c) and (d), (e), (f) are valid for T1 and T2 time periods, respectively (see text). In order to compare the circulations associated with each time period, we consider an appropriate domain that overlaps the Doppler analysis domains presented in Fig. 5. NOAA aircraft flight tracks associated with both considered time periods are shown in (a) and (d). The black arrow in (d) shows estimated storm motion. Solid lines AA' and BB' in (b) and (e) indicate the position for which vertical cross-sections were calculated. In (e), arrows x' and y' define the orientation of cross- and along-line components presented in Fig. 9. Reference horizontal wind vectors are shown in the lower-right corner of each panel, and keys for reflectivities are given to the right of frames (c) and (f).
associated with a rearward divergent outflow resulting from a mesoscale downdraught that spread near the ground level (see later). At mid-levels (Figs. 6(b) and 6(e)), the flow at the leading side was still a front-to-rear (FTR) flow and progressed toward the rear of the system, except in the notch region where it opposed a westward propagating rear-to-front (RTF) inflow originating from the rear edge of the stratiform region. The location of this RTF inflow in the notch-pattern region suggests that strong evaporation of the precipitation probably occurred, as already observed in many previous studies (Stirling and Wakimoto 1989; Scott and Rutledge 1995; Chong and Bousquet 1999). Although a major part of this airflow contributed to the erosion of the north-eastern part of the stratiform region, it also incorporated an apparent closed cyclonic circulation to the south. At T1 (Fig. 6(b)), the mesovortex was centred at $x = 100$, $y = 30$ km, i.e. at the rear edge of the stratiform region, and had a horizontal dimension of 60 km. Its vertical extension was between 3 and 5.5 km altitude. Consistent with the westward propagation of the rear inflow, the mesovortex centre at T2 (Fig. 6(e)) shifted westward (near $x = 70$, $y = 30$ km) and the closed circulation increased both in depth and size. Indeed, it was twice as large (100 km) and its vertical extension was up to 7 km height, as revealed in Fig. 6(f), suggesting that the system dissipation had probably no negative effects on the vortex amplification. At higher levels (Figs. 6(c) and (f)), the flow at the western leading side was mainly from the south and underwent an anticyclonic rotation toward an eastward component as one progressed from south to north.

According to previous studies, mesovortices generally appeared to be better defined during the decaying stage of the MCSs (Zhang 1992; Brandes and Ziegler 1993; Keenan and Rutledge 1993; Chong and Bousquet 1999). Zhang (1992) found that the amplification during the dissipative stage was due to the increasing importance of the stretching as they moved toward the convergence zone along the interface between FTR and RTF flows. He also found that this process rather applies to a particular kind of circulation, referred to as a cooling-induced (or wake) mesovortex, which develops in response to the diabatic (sublimation, melting and evaporation) cooling that occurs in the rear descending portion of MCSs. Figure 7, which presents the vertical-velocity pattern at 4 km altitude associated with T1 and T2 analyses, suggests that the present mesovortex should be classified in this category, since it occurred in a region of subsidence mainly associated with the RTF inflow. On the other hand, its observation near the equator, as in the case of the 13 December 1992 MCS (Chong and Bousquet 1999), gives evidence that the formation of cooling-induced mesovortices does not require the planetary vorticity, as stated by Zhang (1992). One can also note that the intensification of the subsiding rear inflow highly contributes to reduce the horizontal extension of the convection and to concentrate upward motions at the leading edge of the system, as already observed in previous studies (e.g. Smull and Houze 1987; Rutledge et al. 1988).

(c) Vertical structure

Figures 8(a) and (b) present two vertical cross-sections of the system-relative flow and reflectivity pattern across the mesovortex along AA' (see Fig. 6(b)) at T1 and BB' (see Fig. 6(e)) at T2, respectively. Figure 8(a) shows a well-defined succession of small (at distance from origin $d = -40$ km), medium ($d = 0$ km) and deep ($d = 30$ km) convective cells in the leading edge of the MCS, followed by a more stratiform region of precipitation. This structure, which strongly recalls the typical oceanic cloud cluster description of Houze and Betts (1981), may be indicative of a discrete development of the system, as already suggested by Chen et al. (1996). The deepest cell exhibited reflectivity values greater than 25 dBZ up to 12 km altitude, and was associated with the
strongest updraughts that detrained both frontward and rearward at top levels. Maximum upward motions of 2.2 m s\(^{-1}\) occurred at a relatively high level (near 11 km altitude), which is consistent with other COARE observations (Jorgensen et al. 1997; Roux 1998). At T2 (Fig. 8(b)), the convection was less active and was concentrated at the leading edge \((x = -20 \text{ km})\), while the stratiform rain structure was more extensive. The stratiform region \((d > 50 \text{ km in Fig. 8(a); } x > 10 \text{ km in Fig. 8(b)})\) was characterized by a weak mesoscale updraught that transported air at higher levels, above a mesoscale downdraught originating near 6 km altitude which, in turn, contributed to the formation of a deep low-level rearward outflow. As compared to previous studies of such MCSs, one can note the absence of an RTF flow at low level. We think this particularity to be associated with the relative disorganization of the MCS, which is characterized by a fully three-dimensional structure. It is also probable that isolated gusts fronts, associated with convective cells, might have participated in the triggering of new convection at the system leading edge (leading to the apparent discrete propagation of the system). Due to the low resolution (4 km), the convective-scale (<1 km) motions were strongly smoothed through the analysis.

4. MEAN MOMENTUM TRANSPORTS

In order to determine the mean characteristics of the observed system, this section will focus on the vertical transport of horizontal momentum by the MCS during its mature and mature-to-decaying stages. As in LeMone and Moncrieff (1994), we define a right-handed coordinate system moving with the MCS, with the \(x'\) axis perpendicular to the mean orientation of the MCS (315\(^{\circ}\)) and positive in the direction of the system motion (257\(^{\circ}\)) as reported in Fig. 6(e). To obtain a comparable estimation of the mean vertical fluxes at times T1 and T2, we consider only the grid points having observed values at both instants.
Figure 8. Vertical cross-sections of the relative wind vectors and associated reflectivity (intervals of 10 dBZ) for: (a) T1 along A'A' (see Fig. 6(b)); and (b) T2 along B'B' (see Fig. 6(e)). Vectors are plotted at every other grid point. The scaling vectors for wind are indicated in the upper-right corner and a key for reflectivity is shown to the right of each panel. Horizontal distance, $d$, in (a) is obtained by considering a clockwise rotation of the $(x, y)$ coordinate system by an angle defined by the $x$ axis and AA' (see Fig. 6(b)). See text for further details.
The mean relative cross-line velocity component $\bar{U}$ (Fig. 9(a); parallel to the $x'$ axis, positive toward 225°) was negative in almost all the troposphere, consistent with the mean front-to-rear motion observed in previous section. Minimum values exceeding $-4 \text{ m s}^{-1}$ occurred near the surface, consistent with other observations of COARE mid-December MCSs (Roux, personal communication), which all showed stronger values of the FTR inflow at low levels. A local maximum, due to the RTF relative inflow, could be observed at midlevel. This maximum, which increased with time from $-3 \text{ m s}^{-1}$ at 4 km altitude to $-2 \text{ m s}^{-1}$ at 3.5 km altitude, confirms the substantial intensification of the RTF inflow as well as its downward extension. Overall, the mean relative along-line component $\bar{V}$ (Fig. 9(a)) was positive over all the analysis domain, except during T2 for which negative mean values could be observed between 8.5 and 12 km altitude. This component was relatively weak and did not undergo strong variations above 5–6 km, while it presented a marked sheared layer below. The profiles of the mean vertical velocity, $\bar{W}$, at T1 and T2 (Fig. 9(b)) confirm the gradual evolution of the system from
a convective to a rather stratiform stage. At T1, the mean vertical velocity was always positive with a peak value of 0.45 m s\(^{-1}\) near 11 km altitude. At T2, this maximum decreased to 0.25 m s\(^{-1}\) and shifted down to 10 km altitude, while mean subsiding motion occurred below 4.5 km.

As already observed in many previous studies, summarized in LeMone and Moncrieff (1994), the mean vertical flux of relative cross-line momentum \(\overrightarrow{\rho U W}\) in Fig. 9(c) was negative over most of the troposphere. Maximum values of the FTR momentum transport decreased with time from \(-0.8\) N m\(^{-2}\) at 6 km to \(-0.5\) N m\(^{-2}\) at 4 km, consistent with the general decrease of both FTR and vertical wind components. Overall, eddy momentum fluxes \(\overrightarrow{\rho U' W'}\) account for a large part of the mean total transport below the peak levels and were negligible above 7 km. This important contribution from the turbulent terms (especially for the second analysis) is probably related to the presence of the midlevel mesovortex which acts to exacerbate the local wind variations within a large part of the computation domain. One can note the negative correlation between the \(U\) component in Figs. 6(b) and 6(e) and \(W\) in Figs. 7(a) and 7(b), at 4 km altitude. The mean along-line momentum flux \(\overrightarrow{\rho V W}\) in Fig. 9(d) was positive over all the troposphere, except during T2 where negative transport could be observed at high levels. Peak values occurred around 3 km altitude and, similar to the cross-line momentum flux, eddy components had a major contribution.

With respect to the vertical wind shears that could be deduced from the averaged profiles of Fig. 9(a), the cross-line flux (Fig. 9(c)) indicates a down-gradient transport throughout the troposphere, except in the midlevel layer where it was counter-gradient. As a consequence, one can note a net time decrease of the shear of the cross-line component at low and high levels, while a net amplification could be observed at midlevel. These observations, which are consistent with the net amplification of the RTF inflow in the mid-troposphere (compare Figs. 6(b) and 6(e)), also suggest (if sustained) that the rear inflow will continue to increase with time, contributing to the intensification of the mesovortex circulation through stretching at midlevel (Zhang 1992).

5. VORTICITY ANALYSIS OF THE MESOVORTEX

In this section, we focus on the cyclonic mesovortex during the second period of observation.

(a) Mesoscale vorticity structure

The structure of the relative vertical vorticity, \(\zeta\), at 4 km (Fig. 10(a)) shows that positive (cyclonic) vorticity was concentrated at the rear of the MCS and extended over the major part of the stratiform region, while the convective region was mainly associated with anticyclonic motions. Overall, the region of closed wind circulation was associated with positive vorticity of \(1 \times 10^{-4}\) s\(^{-1}\) to \(2 \times 10^{-4}\) s\(^{-1}\), a value which is one order of magnitude higher than the large-scale vorticity that could be derived from the ECMWF winds at 700 hPa and 1800 UTC (Caillault 1998). The maximum vertical vorticity, reaching \(5 \times 10^{-4}\) s\(^{-1}\), was located to the south of the RTF inflow, and coincided with the centre of the closed circulation \((x = 70\) km, \(y = 30\) km), while a secondary local maximum also occurred to the north-west of the mesovortex in the region of weaker rear inflow. The divergence field at midlevel (Fig. 10(b)) shows that the flow was convergent over both stratiform and convective regions of the MCS. The divergence and vertical-vorticity structures are consistent with many previous studies of
tropical and mid-latitude MCSs presenting midlevel mesovortices (Gamache and Houze 1982; Brandes 1990; Scott and Rutledge 1995; among others).

Figure 11, which presents vertical cross-sections of the vertical-vorticity and divergence fields along BB′ (y = 40 km as shown in Fig. 6(e)) shows that the flow within the convective part of the MCS (x < 0 km) was anticyclonic at all levels (Fig. 11(a)) and convergent up to 7 km altitude (Fig. 11(b)). Above this level, the flow was divergent, due to the air detrainment in the upper portion of convective updraughts. The strongest divergence (convergence) occurred at the top (bottom) of the convective cells, as previously observed by Gamache and Houze (1982). The stratiform region (x > 10 km), was marked by positive vorticity, peaking at midlevel in the region of the RTF inflow. One can also note that the closed circulation (delimited by ζ > 2 × 10⁻⁴ s⁻¹, see below) mainly developed within the region of convergence.

(b) Budget

As in Chong and Bousquet (1999), we consider an approximate form of the vertical-vorticity equation within which the contribution of the planetary vorticity has been neglected due to the near-equatorial location of the MCS (0.5°N). Considering a reference frame moving at the speed of the system, the following form of the equation is used:

\[
\frac{\partial \zeta}{\partial t} = - \left( u \frac{\partial \zeta}{\partial x} + v \frac{\partial \zeta}{\partial y} \right) - \left( w \frac{\partial \zeta}{\partial z} \right) - \zeta \left( \frac{\partial u}{\partial x} + \frac{\partial v}{\partial y} \right) - \left( \frac{\partial w \frac{\partial v}{\partial z} - \frac{\partial w}{\partial y} \frac{\partial u}{\partial z}}{\partial x} \right) (1)
\]

where \( u, v, \) and \( w \) are the system-relative wind components (west to east, south to north, and vertical components, respectively) and \( \zeta \) is the vertical component of the
Figure 11. As Fig. 10, but for a vertical cross-section along BB' shown in Fig. 6(e).
relative vorticity \( (\zeta = \partial v/\partial x - \partial u/\partial y) \). Terms A, B, C and D describe vorticity changes due to horizontal advection, vertical advection, stretching (divergence or convergence), and tilting of the horizontal component of vorticity, respectively. They are evaluated from the radar-derived wind fields, while the residual is attributed to the net rate of change in the relative vertical vorticity \( (\partial \zeta/\partial t) \). All terms were calculated over a 80 km \( \times \) 80 km \( \times \) 12 km vortex-centred domain, reported in Fig. 10(a).

The vertical profiles of mean divergence and relative vertical vorticity, vertical velocity, and mean vorticity budget terms, are presented in Fig. 12(a), (b) and (c), respectively. Terms \( \partial \zeta/\partial t \), A, B, C and D of the budget equation (1) are labelled SUM, HADV, VADV, ST and TT respectively in Fig. 12(c). The averaged convergence profile (Fig. 12(a)) shows a layer of convergence at mid-to-upper (3–9.5 km) levels resulting from the strong opposition between the FTR and RTF inflows, while divergence can be found above and below. The corresponding mean vertical velocity (Fig. 12(b)) exhibits a typical stratiform profile with mesoscale downdraught below 5.5 km and mesoscale updraught aloft. Maximum downward and upward motions occurred at 3.5 and 10 km altitude, respectively. It is probable that this convergence would have been enhanced by latent cooling (warming) occurring within the negatively (positively) buoyant subsiding (ascending) mesoscale airflow. The averaged relative vertical-vorticity profile (Fig. 12(a)) was positive at nearly all levels, and had a peak value of \( 2.8 \times 10^{-4} \) s\(^{-1} \) at 4.5 km in the vicinity of the melting level. Above this level it gradually decreased and became negative near 11 km altitude. The vorticity budget (Fig. 12(c)) indicated a spin-up (positive tendency) of the vortex up to 6 km altitude, with a net spin-down aloft. Maximum amplification of the relative vertical vorticity occurred near 2 km at a rate of approximately \( 2.8 \times 10^{-8} \) s\(^{-2} \), and was mainly related to horizontal and vertical relative-vorticity advection which strongly exceeded the contribution from the other terms. As in previous studies (Verlinde and Cotton 1990; Brandes and Ziegler 1993; Chong and Bousquet 1999), mean horizontal and vertical advection terms were negatively correlated with stretching and tilting terms, respectively. Below 2 km, both advection and tilting terms, however, contributed to an overall increase of the vertical vorticity, while negative stretching, resulting from mean divergent motions near ground level, tended to oppose the relative-vorticity amplification. At mid-to-upper levels, tilting of horizontal vorticity into the vertical was negative and produced a vertical vorticity decrease up to 9 km. As in Brandes and Ziegler (1993), this negative tendency was, however, relatively weak and was strongly offset by the positive contribution of stretching, which appeared to be the main mechanism for the maintenance of the closed circulation at middle level. This observation confirms previous studies from many authors (e.g. Brandes and Ziegler 1993; Scott and Rutledge 1995; Zhang 1992; Johnson and Bartels 1992) who showed the amplification of pre-existing vertical vorticity by convergence to be the primary mechanism of mesovortex intensification during the mature-to-decaying stages of MCSs. This vorticity budget also confirms that vorticity advection played an important role in the local-vorticity budget of the mesovortex (Brandes and Ziegler 1993; Keenan and Rutledge 1993; Chong and Bousquet 1999). In the present case, one can note that horizontal advection tended to strongly offset the amplification by convergence above 6 km, which would act to limit the upward extension of the closed circulation. On the other hand, the vertical advection of relative vorticity was generally positive, and contributed to increase the magnitude of the cyclonic vorticity throughout the troposphere, due to the negative correlation between the mesoscale updraught/downdraught couplet and the mean relative-vorticity gradient within the mesovortex.
Figure 12. Vertical mean profile of: (a) relative vertical vorticity and divergence (ZETA, full line; DIV, dashed line) units are $1 \times 10^{-4}$ s$^{-1}$; (b) vertical velocity ($W$; m s$^{-1}$); and (c) vorticity tendencies ($1 \times 10^{-8}$ s$^{-2}$), averaged over the $(80 \times 80$ km$^2)$ domain shown in Fig. 10(a). HADV, VADV, ST, and TT in (c) are contributions of the horizontal vorticity advection, vertical vorticity advection, stretching and tilting, respectively, to the net rate of change in the relative vertical vorticity denoted by SUM.

(c) Possible impact of the mesovortex

From the present study, the 12 December 1992 MCS exhibits many structural and dynamical similarities with tropical oceanic cloud clusters that have been observed in previous studies (e.g. Gamache and House 1982), such as a rear inflow jet and a midlevel mesovortex that developed within the rear part of the stratiform region. However, this MCS presents some particularities such as the large-scale context in which it occurred, i.e. a two-day westward propagating disturbance resulting from the equatorial inertio–gravity wave, the westward propagative mode of which would be privileged (Takayabu 1994), and its long life cycle over 24 hours. Recently, Chen and Houze (1997) proposed a conceptual model that demonstrates the interplay between the
diurnal cycle of large convective systems and large-scale disturbances (including twoday disturbances) over the warm pool region. They showed that two-day disturbances could be interpreted as a combination of 2-day inertio-gravity waves, and a concept referred to as ‘diurnal dancing’, which accounts for the boundary-layer recovery time after an intense convective event.

According to this conceptual model, the second-day convection in two-day disturbances tends to occur hundred of kilometres to the west of the first MCS. Following Chen et al. (1996), this seems to apply to the 12–13 December disturbance. However, we believe that the cloud top IR temperature threshold, used by these authors to identify these MCSs, tends to exacerbate the distance between them, since this tends to track deep convection structure. Indeed, Fig. 13, which presents a two-hourly series of IR
satellite pictures between 0045 and 1045 UTC 13 December, clearly indicates a smooth transition between dissipation of the old system and development of new convection, if one considers the global structure of the cloud shield. Moreover, it shows that individual convective cells continuously developed within the decaying cloud cluster before the 13 December MCS appeared (Figs. 13(e) and (f)).

Following Zhang (1992), cooling-induced mesovortices provide a source of dry air that enhances diabatic cooling, and act to maintain a cold pool within the stratiform region of MCSs. Under particular conditions (Rotunno et al. 1988), these processes can also help sustain convection, and consequently mesovortices could increase the lifetime of MCSs, or assist the generation of new convection in the vicinity of the decaying cloud cluster (e.g. Fritsch et al. 1994). Although the lack of a consistent set of thermodynamic data did not allow us to study the thermodynamic structure of this MCS, we suggest that the mesovortex observed in the present study probably played a key role in the continuous regeneration of convective cells observed from Fig. 13; i.e. we believe that it acted to maintain convective activity, through modification of the nearby environmental conditions, until the MCS propagated in a large-scale environment favourable to the regeneration of deep convection (Chen and Houze 1997). This hypothesis could be supported by the 13 December MCS observations, which also revealed a mesovortex circulation at midlevel (Chong and Bousquet 1999). Although the connection between the two cases is not clear, these vortices would have formed a unique feature, similar to the one already observed by Fritsch et al. (1994) in mid latitudes. In the absence of continuous observation, a method to analyse the potential role of the 12 December mesovortex would consist of using numerical simulation, in which the present radar observations could be included as the initial state in a mesoscale cloud model. This work, which is in progress (Nuret et al. 1999), will also provide information on the thermodynamic structure of the mesovortex.

6. Conclusions

This paper has documented the mesoscale structure of the 12 December 1992 near-equatorial MCS, using airborne Doppler radar observations collected during the mature/mature-to-decaying stages of the system. This MCS was the first of two MCSs associated with a 2-day disturbance occurring within the convectively active phase of the second of three intraseasonal oscillations observed during TOGA-COARE.

The radar-deduced airflows have revealed strong similarities with previously observed tropical oceanic cloud clusters, such as a westward subsiding rear inflow jet, and a midlevel cyclonic mesovortex that developed within the rear part of the stratiform precipitation region. On the other hand, this MCS was characterized by the lack of a well-marked RTF flow at low levels. The mesovortex, which presented a well-defined closed circulation, occurred to the south of the westward propagating rear inflow which opposed a rearward westerly inflow, and was coincident with a marked ‘notch’ region.

The mesovortex appeared to be better defined in the dissipative stage, while it progressed more toward the convective region of the system. This dissipative stage was accompanied by a reduction of the convective activity, which concentrated at the leading edge due to the westward propagation of the rear inflow; radar-derived vertical motions have clearly shown the evolution toward a stratiform-like profile, with mesoscale subsidence at low-to-mid levels and mesoscale ascent above.

Consistently with previous studies, mean vertical transports of momentum normal to the mean orientation of the system (cross-line momentum) were negative throughout the troposphere, with down-gradient transports at low and high levels between which
counter-gradient transports existed. The observed increase of the rear inflow was consistent with cross-line momentum fluxes. A large part was accomplished by eddy structures, probably due to the presence of the midlevel vortex which enhanced the local wind variations. The along-line momentum flux was positive at all levels and eddy components also made a major contribution.

In regions of closed wind circulation associated with the mesovortex, the vertical vorticity was positive at all levels and peaked above the downdraught maximum. The vorticity budget has revealed the stretching of pre-existing vertical vorticity as the dominant dynamical process that amplified the positive vorticity at mid-to-upper levels, while tilting of horizontal vorticity had a negative effect. As in previous studies, there was a negative correlation between stretching and tilting, and horizontal and vertical advection, respectively. However these budget terms did not oppose sufficiently to be balanced, and their combined effects (net time tendency) led to an increase of positive vorticity at low-to-middle levels and a decrease aloft.

Finally, this study has examined some properties of an MCS at its mature-to-decaying stage but, in the absence of further observation, it was not possible to identify the role of the mesovortex in the redevelopment of the convection which led to the observation of the next day MCS. A possible method to accomplish this would be the use of numerical simulations that could take the present radar observations as input to a mesoscale cloud model, such as that jointly developed by the Centre National de Recherches Meteorologiques and Laboratoire d’Aérologie (Lafore et al. 1998).

REFERENCES


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