Radar reflectivity and wind fields analysis by using two X-band Doppler radars at Okinawa, Japan from 11 to 12 June 2007

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ABSTRACT: During 11 June (CASE I) and 12 June (CASE II) 2007, several meso-β-scale convective systems (MβCSs) were generated along the Meiyu-Baiu front. In this study, the formation, evolution and dissipation of convective precipitation, kinematics structures, and features of the MβCSs developed over ocean were investigate using dual Doppler analysis. The analysis of synoptic conditions and atmospheric sounding suggested that in both cases the atmospheric environment was characterized by well-developed convective systems with warm and humid air-inflow and unstable conditions in addition to strong vertical wind shear at lower levels. In CASE I, an embedded-cell storm developed, approached the area, and moved slowly (≤ 2.5 m s⁻¹) northeastward. As the system advanced, strong dominant updraft could be observed at the core. After the system passage, the wind became weak at the centre of the cell, whereas the downdraft area extended up to 4 km in height on the front side. This downdraft flowed back into the convective cell and, maintained the convective activity. The evolution of the updraft and downdraft was in good agreement with the general life cycle of convective systems. In CASE II, line-shaped convective systems (LSCSs) passed from the southwest to east–northeast at 10 m s⁻¹. The dual Doppler analysis revealed that the wind at the rear of the band contributed to the cells merge and developed into a strong convective cell. In addition, the outflow from the new cell flowed into the rear of a merged convective cell and enhanced the vertical development of the cell.

KEY WORDS dual Doppler radar analysis; wind fields; convective cell; radiosonde; Meiyu-Baiu front

1. Introduction

The rainy season from June to July over East Asia is referred to Changma in Korea, Baiu in Japan and Mei-Yu in China. During this period, the west to east elongated rainfall area is known as the Meiyu-Baiu front (MBF) or Changma front to reflect regional differences in location, timing and duration in East Asia. Ninomiya and Akiyama (1992) showed that the intense precipitation around the MBF is usually relative to meso-β-scale convective systems (MβCSs). Movement of MβCSs developing around the MBF frequently leads to severe weather such as localized gusts and heavy rainfall, which results in life-threatening events and have effects on the security and economy of Korea, Japan, Taiwan and China. Most of the research in the field of MβCSs has been performed for the enhancement or maintenance of precipitation systems by conducting synoptic and meso-scale observations on land (Ishihara et al., 1995; Chen and Lin, 1997; Teng et al., 2000). However, an understanding of the formation mechanism, structure and variation of precipitation systems over the ocean is critical because MβCSs frequently develop over oceans and cause significant damage to the land.

Doppler radar is a considerably useful instrument for mesoscale meteorological observation because it can spatially and temporally estimate precipitation amounts and movement of the precipitation systems over a wide area with ease. High-resolution data indicate that the internal structure of storms can be observed at remote locations over the ocean. However, Doppler radar has restricted data of reflectivity, radial velocity, and spectrum width. Moreover, it is difficult to obtain three-dimensional wind fields using radial velocity of a single radar. If two or more radars are used, the two horizontal components (u and v) of the three-dimensional wind vector can be easily derived using the continuity equation, and the vertical component (w) can be determined by integrating the continuity equation upward (or downward) subject to the boundary condition w = 0 at the lower boundary (or upper boundary).

The above-mentioned dual-Doppler radar analysis methods are helpful for understanding a wide range of weather phenomena, that is from mesoscale convective complexes to bright bands at the melting layer (Armiño, 1969; Ray et al., 1981; Reinking et al., 1981; Carbone, 1983; Kessinger et al., 1987; Parsons and Kropfl, 1990; Atkins et al., 1995, Dowell and Bluestein, 1997; Kim et al., 1998; Nam et al., 2005). However, most of dual Doppler radar analyses based on the continuity
equation contain several notable deficiencies, including limitation due to the boundary conditions for vertical velocity, spatial interpolation errors, discretizations, uncertainties in radial wind estimates, and non-simultaneous measurements (Miller and Strauch, 1974; Ray et al., 1975, 1980; Gal-Chen, 1982; Testud and Chong, 1983; Chong et al., 1983a, 1983b; Ziegler et al., 1983; Shapiro and Mewes, 1999). Recently, a variational method has been used to minimize a cost function, defined as distance between the analysis and the observations to produce optimal $u$, $v$ and $w$ components (Ray et al., 1980; Sun and Crook, 1994; Shapiro and Mewes, 1999; Gao et al., 1999, 2004; Shimizu and Maesaka, 2006; Xue et al., 2006; Lee et al., 2007, You et al., 2010; Jeong et al., 2012).

To understand the structure, variation and development of MβCSs that form around MBFs, a joint intensive observation project was conducted from 2 to 15 June 2007 in the East China Sea, Pukyong National University and Hydrospheric Atmospheric Research Center (HyARC), Nagoya University contributed with two X-band Doppler radars in Taramajima and Miyakojima, in Okinawa, Japan. The observation area is the most suitable location for observing mesoscale convective systems (MCSs) associated with MBFs in the ocean, without considering terrain effects such as system enhancement or dissipation by orography (Figure 1). In this study, three-dimensional wind fields derived from the simplified variational method suggested by Shimizu and Maesaka (2006) were used to investigate the role of wind fields in the development and dissipation processes, kinematic structures and features of MβCSs, embedded-cell storm and LSCSs that develop around MBFs. The data and methodology are explained in Section 2. An atmospheric condition and a detailed description of three dimensional wind field structure (as obtained from dual-Doppler analysis) are explained in Section 3. Finally, a summary and concluding remarks of the results are presented in Section 4.

2. Data and methods

2.1. Data

Radar data were obtained from two X-band Doppler radars located in Taramajima (24.63° N, 124.72° E) and Miyakojima (24.80° N, 125.15° E), in Okinawa, Japan, during the intensive observation period. Taramajima and Miyakojima are isolated small islands near Okinawa’s main island. The location and characteristic of the two radars used for wind field analyses are shown in Figure 1 and listed Table 1. The radars had a 64-km detecting radius, 250-m interval resolution, and 1° azimuth resolution, and their volume scans were accomplished with 13 elevation angles every 6 min. The Nyquist velocity of the radar was 16 m s⁻¹, and the dealiasing correction of Doppler velocity was conducted manually considering temporal and spatial continuity of the wind fields (Shinoda et al., 2009). The distance between the two radars was 47 km, and wind fields were made in the region limited by angle $\beta = 30^\circ$, which was formed by lines that linked two radar points from the target.

To determine the weather condition of the upper layer, observed sounding data obtained at Ishigakijima (24.33° N, 124.16° E), located between Taiwan and Taramajima, were used. These data help to understand the patterns of MCS movement, characteristics of development, and dissipation of convective systems originating in south China through comparison with upper level data. In addition, the National Centers for Environmental Prediction/National Centers for Atmospheric Research (NCEP/NCAR) reanalysis data were used to determine the synoptic flow where radar and radiosonde observation are located.

The analysis period was from 0500 LST (local standard time; LST = UTC + 9 h) on 11 June 2007 to 1200 LST on 13 June 2007, during which time the MBF was located in the southern part of China, Taiwan, and Okinawa, and developed MβCSs which moved from the southern part of China to the analysis area.

| Location | Taramajima (red) | Miyakojima (blue) |
|----------|-----------------|-------------------|
| Range    | 64 km           | 64 km             |
| Nyquist velocity | 16 m s⁻¹ | 16 m s⁻¹ |
| Number of elevation angle | 13 (0.4–32.8°) | 13 (0.4–32.8°) |
| Scan interval | 6 min       | 6 min             |
| Wavelength | 3.2 cm (X-band) | 3.2 cm (X-band) |
| Bin spacing | 250 m         | 250 m             |
| Elevation angles | 13 (0.4–32.8°) | 13 (0.4–32.8°) |
| Number of elevation angles | 13 (0.4–32.8°) | 13 (0.4–32.8°) |
| Distance between the two radars | 47 km  | 47 km             |
| Location  | 24.63° N, 124.72° E | 24.80° N, 125.15° E |

Figure 1. Wind field analysis range for variational method at Taramajima and Miyakojima, Okinawa, Japan.
the sum of squared errors due to the misfit between observations and analyses subject to constraints. Each constraint is weighted by a factor that accounts for its accuracy, therefore different forms of $J$ that might be considered. In this study, the constraints used in Gao et al. (1999) are adapted. When the analysis is performed, the Plan Position Indicator (PPI) data observed in spherical co-ordinates is interpolated point location. This process is very costly, therefore the simplified Shimizu and Maesaka (2006) cost function is used. It is defined as:

$$ J = J_0 + J_B + J_D + J_S $$

(1)

$$ J_0 = \frac{1}{2} \sum_m \lambda_m (V_{rm}^m - V_{om}^m)^2 $$

(2)

$$ J_B = \frac{1}{2} \sum_{i,j,k} \lambda_{ob} (u - u_b)^2 + \sum_{i,j,k} \lambda_{ob} (v - v_b)^2 $$

$$ + \sum_{i,j,k} \lambda_{ob} (w - w_b)^2 $$

(3)

$$ J_D = \frac{1}{2} \sum_{i,j,k} \lambda_D D^2 $$

(4)

$$ J_S = \frac{1}{2} \sum_{i,j,k} \lambda_{ws} \left( \nabla^2 u \right)^2 $$

$$ + \sum_{i,j,k} \lambda_{vs} \left( \nabla^2 v \right)^2 $$

$$ + \sum_{i,j,k} \lambda_{ws} \left( \nabla^2 w \right)^2 $$

(5)

$$ D = \frac{\partial \varphi_u}{\partial x} + \frac{\partial \varphi_v}{\partial y} + \frac{\partial \varphi_w}{\partial z} $$

(6)

$$ V_r = u \sin \theta \cos \theta + v \cos \theta \cos \theta + (w + w_i) $$

(7)

Here, $J_0$ is the a sum of different radial velocity components ($V_{rm}^m$ : observations) interpolated to each grid point and radial velocity component ($V_{om}^m$ : analyses) derived from $u$, $v$, and $w$ at each grid point in Cartesian co-ordinates; $m$ is the number of radars; $u$, $v$ and $w$ are wind components in Cartesian co-ordinates ($x$, $y$, $z$); $w_i$ is terminal velocity of precipitation; $\theta$ and $\theta$ are azimuth and elevation angles; $J_B$ measures the distance of variational analysis to the background fields; $J_D$ imposes a weak and elastic mass continuity constraint on the on the analysed wind field; $\varphi$ is the mean of air density on the horizontal layer; and $J_S$ is a smoothness constraint that reduces the noise in the analysed field and also help to alleviate the under-determined nature of the problem (Gao et al., 2006).

The $\lambda$s represents the weight of each cost function.

To solve the above variational problem, the gradient of the cost function needs to be derived with respect to $u$, $v$, and $w$. The components of the gradient of $J$ suggested by Shimizu and Maesaka (2006) are applied, as follows:

$$ \frac{\partial J}{\partial u} = \lambda_D \left( -\sin \theta \cos \theta \right) \times \left( V_{om}^m - V_{om}^m \right) $$

$$ + \lambda_B \times (u - u_b) - \lambda_D \frac{\partial D}{\partial x} - \lambda_{ws} \nabla^2 (\nabla^2 u) $$

(8)

$$ \frac{\partial J}{\partial v} = \lambda_D \left( -\cos \theta \cos \theta \right) \times \left( V_{om}^m - V_{om}^m \right) $$

$$ + \lambda_B \times (v - v_b) - \lambda_D \frac{\partial D}{\partial y} - \lambda_{ws} \nabla^2 (\nabla^2 v) $$

(9)

$$ \frac{\partial J}{\partial w} = \lambda_D \left( -\sin \theta \right) \times \left( V_{om}^m - V_{om}^m \right) $$

$$ + \lambda_B \times (w - w_b) - \lambda_D \frac{\partial D}{\partial w} - \lambda_{ws} \nabla^2 (\nabla^2 w) $$

(10)

The limited-memory quasi-Newtonian conjugate-gradient algorithm (Liu and Nocedal, 1989) is applied to minimize through a number of iterations the gradients of the cost function $J$ shown in Equation (1) which can reduce the cost function $J$ to a smaller magnitude.

3. Results and discussion

3.1. Synoptic conditions and sounding profiles

Figure 2 shows the synoptic flow (as described by NCEP/NCAR reanalysis) for CASE I. The solid lines represent sea level pressure (hPa, contoured every 2 hPa, Figure 2(a)), the equivalent potential temperature (K, contoured every 3 K and greater than 333 K, Figure 2(b)), the relative vorticity ($10^{-5}$ s$^{-1}$, contoured every $2 \times 10^{-5}$ s$^{-1}$, Figure 2(c)), and the geopotential height (m, contoured every 50 m, Figure 2(d)). Arrows indicate wind direction and speed. The shaded areas represent wind speeds greater than 8, 12, 16, and 20 m s$^{-1}$ in Figures 2(a)–(d).

A convergence zone (Figure 2(a)) forms to a low in the southern part of China, a high in the Yellow Sea, and a low in southern part of the Japanese Islands (near 27° N, 140° E). Southwesterly, southerly and southeasterly winds greater than 8 m s$^{-1}$ flow into the analysis area (near 25° N, 125° E) due to the low-pressure system located in the southern part of China. The 850 hPa equivalent potential temperature (339 K) in the area of interest is higher than that in the surrounding areas and appear to be continuously supplied with warm and humid air from the tropics (Figure 2(b)). Relatively strong westerly winds (Figure 2(c)) delimit a positive vorticity and a negative vorticity area. Vertical strong wind shear occurs, as the horizontal wind veers westerly at 500 hPa, from southerly at the surface level. The East China Sea has weaker winds at 850 hPa, while at 500 and 300 hPa, winds are stronger than 20 m s$^{-1}$ in the eastern region (135° E and 25–30° N).

As shown in Figure S1(a) and (b), the area of interest is at the boundary of positive and negative vorticity zone enabling upward flow, and the wind speed is represented in a range of 10–30 m s$^{-1}$ in the upper layer. The strong convergence area has values of $-5 \times 10^{-6}$ s$^{-1}$ to $-10 \times 10^{-6}$ s$^{-1}$ under 300 hPa and $10 \times 10^{-6}$ s$^{-1}$ to $20 \times 10^{-6}$ s$^{-1}$ over 300 hPa (as shown in Figure S1(c) and (d)). The equivalent potential temperature is greater than 345 K at the surface, 330–335 K at 900 to 600 hPa, and more than 340 K above 600 hPa. These conditions favour convergence zones with inflows of humid and warm air from the south and strong vertical wind shear, therefore the formation of convective clouds is highly probable.

The synoptic flow of CASE II is shown in Figure 3 (variables, contours and shading as in Figure 2). The surface map shows the low-pressure system located in the southern part of China moving northeastward (near 27° N, 125° E) and the high-pressure area located in the southern part of the Japanese Islands. The convergence zone favours southwesterly wind and a southeasterly wind ($\geq 8$ m s$^{-1}$) into the analysed area. A southwesterly low-level jet (LLJ) appears over the analysed area at 850 hPa (Figure 3(b)). The LLJ transports warm moist air from low latitude at low levels and generates convective instability as well as lower the level of free convection (Jeong et al., 2012). In this case, a large amount of moist air is supplied from low latitude by the effect of LLJ. The 850 hPa equivalent potential temperature ranges between 339 and 345 K. The 500 hPa westerly and southwesterly winds (10 m s$^{-1}$) are weaker compared to those at lower levels (Figure 3(c)). The
Figure 2. (a) Pressure and wind vector at surface, (b) equivalent potential temperature and wind vector at 850 hPa, (c) relative vorticity and wind vector at 500 hPa, (d) geopotential height and wind vector at 300 hPa, at 0900 LST on 11 June 2007.

Figure 3. (a) Pressure and wind vector at surface, (b) equivalent potential temperature and wind vector at 850 hPa, (c) relative vorticity and wind vector at 500 hPa, (d) geopotential height and wind vector at 300 hPa, at 2100 LST on 12 June 2007.
300 hPa westerly and northwesterly winds have a wind speed of about 15 m s\(^{-1}\) (Figure 3(d)). Positive vorticity of \(2 \times 10^{-5} \text{ s}^{-1}\) from the surface to a height of 700 hPa is evident with negative vorticity on top (as shown in Figure S2(a)). In Figure S2(b), the analysed area is located at the boundary between the positive and negative vorticity zones below 600 hPa. Wind speed is below 10 m s\(^{-1}\) at mid-levels (600 to 200 hPa) but stronger (10–20 m s\(^{-1}\)) in the other areas. Strong convergence (\(-10 \times 10^{-6} \text{s}^{-1}\)) at low-level near the analysis area, whereas strong divergence (values of \(20 \times 10^{-6} \text{s}^{-1}\)) is located above 300 hPa (as shown in Figures S2(c) and (d)). In addition, the equivalent potential temperature from the surface to 900 hPa is 345 K, from 900 to 400 hPa is 335–345 K, and above 400 hPa more than 345 K.

Table 2 lists several environmental parameters calculated from upper sounding data recorded at Ishigakijima during the analysis period. The parameters are: lifting condensation level (LCL), level of free convection (LFC), convective available potential energy (CAPE), convective inhibition (CIN) and precipitable water (PW). At 0900 LST on 11 June 2007, horizontal wind direction veered with height from southerly to westerly, and the wind speed was 7–10 m s\(^{-1}\) below 2.5 km above sea level (ASL); however, the speed continuously increased to 20 m s\(^{-1}\) at a height of 5 km from 2.5 km ASL (as shown in Figure S3(a)). Such wind distribution is similar to that included in the NCEP/NCAR reanalysis data. The equivalent potential temperature \(\theta_e\) below 3 km ASL decreased with height, which indicates convective instability at lower level. The value of CAPE increased to approximately 1650 m\(^2\) s\(^{-2}\) (compared to 16.35 m\(^2\) s\(^{-2}\) at an earlier time), and CIN had a negative value of \(-5.54 \text{ m}^2 \text{s}^{-2}\). LCL was located at 947 hPa due to humid atmospheric conditions (PW, 58.9 mm). At 2100 LST on 10 June, LFC was located at 630 hPa but descended to 901 hPa at 0900 LST on 11 June when the precipitation system passed through the analysis area. The conditions were favourable for formation convection clouds. At 2100 LST on 12 June, wind veered from south to west–northwest with height. The southwesterly wind speed was 18 m s\(^{-1}\) at 2.5 km ASL but was reduced to under 10 m s\(^{-1}\) at 2.5–3 km ASL (as shown in Figure S3(b)). The atmosphere was convectively unstable and very humid below 2.5 km ASL. The CAPE was 3372 m\(^2\) s\(^{-2}\), greater than that of CASE I. This result indicates that the development possibility of severe convective storms was very high with this system.

3.2. Analysis on reflectivity and wind fields

The observed cases were convective systems embedded in the MBF; however, the shapes differed with single cell formation in the former and line-shaped convection in the latter. The cell of CASE I moved slowly (\(\leq 2.5 \text{ m} \text{s}^{-1}\)) through the analysis area from southwest to northeast, while the convective line of CASE II crossed the area from southwest to east–northeast (as shown in Figure 4).

To investigate the characteristics, structure, and variation patterns of the convective systems, the reflectivity, horizontal wind fields, convergence and upward and downward wind were closely inspected at the moving radar echoes.

Figure 4 shows radar reflectivity and retrieved horizontal wind field (vectors) at 2.5 km ASL from 0836 to 0936 LST on 11 June 2007 obtained using constant altitude plan position indicator (CAPPI) data. The convective system with a strong echo of more than 55 dBZ crossed the analysis area from southwest to northeast. Its movement was slow at \(\leq 2.5 \text{ m} \text{s}^{-1}\), and the top of echo (\(\geq 40 \text{ dBZ}\)) was at 7 km ASL. After 0912 LST, the convective system abated and altered to the dissipating stage. A different system with an echo of more than 45 dBZ formed near the southwest region of the convective system. The wind direction changed from south-southwesterly in the southern area of the system core to southwestery in the northern area after passing through the core. The wind direction at 2.5 km ASL, obtained from the sounding data (0900 LST on 11 June 2007, Ishigakijima), was southwest. This wind direction is similar to that derived using dual Doppler radar analysis and the speed was generally \(10–15 \text{ m} \text{s}^{-1}\).

Figure 5 shows the horizontal wind divergence and reflectivity at 0.5 km ASL, which was calculated using horizontal wind distributions retrieved by dual Doppler analysis. Areas of large divergence of approximately \(-3 \times 10^{-3} \text{ m} \text{s}^{-2}\) appeared in the west–southern region of the cell at 0836 LST. As the convective system moved, convergence areas grew in the centre of the cell (Figures 5(b) and (c)). Areas of larger convergence of \(-4 \times 10^{-3} \text{ m} \text{s}^{-2}\) were found at the southern region of cell in which a new cell of \(\geq 50 \text{ dBZ}\) developed at the low-level between 0912 and 0936 LST (Figures 5(d)–(f)).

In the vertical cross sections of the A–B line, as presented in Figure 4, the reflectivity fields were found to increase from 0836 to 0924 LST (Figure 6). In addition, the wind fields converged in the rear side and the lower level of the cell to trigger strong cyclonic circulation inside the cell, discussed in the previous section. After 0924 LST, the area of high reflectivity region (\(\geq 55 \text{ dBZ}\)) decreases and the maximum downward velocity in the cell is about \(2 \text{ m} \text{s}^{-1}\). Thus, the cell has characteristics of the dissipating stage in its life cycle. However, the downdraft flowed into the lower part of the cell enhancing and maintaining convergence at the lower level. To confirm these observations, the vertical wind speed and divergence were analysed along the A–B line. In agreement with the above observations, strong upward vertical motions up to \(4 \text{ m} \text{s}^{-1}\) were observed in the core area of the cell, indicating strong updraft and leading to the development of a matured cloud system. Vertical cross sections of divergence and reflectivity reveal the development of the

### Table 2. Environmental parameters determined from upper–sounding data obtained at Ishigakijima from 2100 LST on the 10 of June 2007 to 0900 LST on the 13 of June 2007: lifting condensation level (LCL), level of free convection (LFC), convective available potential energy (CAPE), convective inhibition (CIN), precipitable water (PW).

| Parameter | 10 June 2007 | 11 June 2007 | 12 June 2007 |
|-----------|-------------|-------------|-------------|
| LCL (hPa) | 924.7       | 947.5       | 941.4       |
| LFC (hPa) | 630.4       | 901.1       | 876.2       |
| CAPE (m\(^2\) s\(^{-2}\)) | 16.35 | 1674 | 2093 |
| CIN (m\(^2\) s\(^{-2}\)) | -190 | -5.54 | -19.5 |
| PW (mm)   | 58.87       | 58.94       | 61.71       |

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Figure 4. Distribution of reflectivity and wind (vectors) retrieved by variational dual-doppler radar analysis at 2.5 km ASL for (a) 0836 LST, (b) 0848 LST, (c) 0900 LST, (d) 0912 LST, (e) 0924 LST and (f) 0936 LST on 11 June 2007.

Figure 5. Distribution of horizontal wind divergence (shaded) and reflectivity (line) at 0.5 km ASL for (a) 0836 LST, (b) 0848 LST, (c) 0900 LST, (d) 0912 LST, (e) 0924 LST and (f) 0936 LST on 11 June 2007.
Figure 6. Vertical cross sections of reflectivity and vertical velocity (vector) along the A–B line in Figure 4 for (a) 0836 LST, (b) 0848 LST, (c) 0900 LST, (d) 0912 LST, (e) 0924 LST and (f) 0936 LST on 11 June 2007. Colour shade indicates reflectivity, and red (blue) line upward (downward) wind with 1 m s$^{-1}$ interval. The boundary of updraft and downdraft is shown as black line.

convergence from 0836 to 0924 LST in the lower level of the convective region (Figure 7).

Figure 8 shows radar reflectivity and retrieved horizontal wind (vector) at 2.5 km ASL from 2236 to 2312 LST on 12 June 2007. The propagation of the cells is marked with black squares (C1–C6) in Figure 8. Wind direction and speed did not vary dramatically around the cells. Since the strong wind blew continuously toward the northeast, the convective cells moved rapidly along. Three convective cells in the lower left cluster (C1) were merged into two cells (C2) and then into one (C5). As the cells merged, the strong echo area was deformed lengthwise (C3–C6) and became divided into two small systems. A small cell with an echo of more than 45 dBZ developed to the southwest of the band at 2254 LST and was maintained until 2312 LST.

As the convective system moved, convergence areas continued to grow within the cells (Figures 9(a) and (b)). At 2254

Figure 7. Vertical cross sections of divergence and reflectivity along the A–B line in Figure 4 for (a) 0836 LST, (b) 0848 LST, (c) 0900 LST, (d) 0912 LST, (e) 0924 LST and (f) 0936 LST on 11 June 2007. Colour shade indicates divergence and solid line means reflectivity.

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and 2300 LST, the convergence areas were enhanced to a maximum of $-5 \times 10^{-3}$ s$^{-1}$ at the rear side of the merged system and the new cell. After 2306 LST, areas of large convergence remained at the centre of merged cell, which means that the merged cell is in the developing stage. In the meantime, divergence areas were stronger between the merged cells and the new cell (Figures 9(d) and (e)).

Vertical cross sections of reflectivity and wind field on the A–B line shown in Figure 8 indicate that an upward wind of approximately 1–3 m s$^{-1}$ was dominant in the convective cells before they merged (Figure 10(a)). As the convective cells merged, the upward wind speed at the core of the cell was stronger and developed up to a height of 9 km (Figures 10(b) and (c)). At 2254 LST, a new precipitation cell developed at the southwestern edge on the upstream side of the merged cell (Figure 10(c)). Considering the location of the new cell and the weak downward wind in the pre-existing cells under the moist environment, the new cell generation conditions are similar to the LSCSs of the back-building type presented by Shinoda et al. (2009). An outflow from the new cell flowed into the rear of the merged system and rapidly ascended with the system (Figures 10(d) and (e)). At that time, the upward wind speed intensified to 7 m s$^{-1}$ from 18 to 22 km in the x direction, and the top of reflectivity of 45 dBZ in the cell occurred above a height of 9 km. At that time, the thickness of the convergence layer was deeper and convection was more active than that observed in CASE I. From 2300 to 2306 LST, outflow from the new cell flowed in the lower layer of the merged cell, which is indicated by significant intensification of convergence in the lower layer of the merged cell and a strengthening of divergence between the merged cells and the new cell (Figures 11(d) and (e)). As shown at 2300–2312 LST in Figure 10, the merged cell was vertically extended because the strengthened low-level convergence, which was caused by outflow from the new cell, played an important role in activating the ascending flow. However, the effect of outflow from the new cell was not detected in the LSCSs presented by Shinoda et al. (2009).

4. Summary and concluding remarks

To study the structure, variation and development of meso-$\beta$-scale convective systems (M$\beta$CSs) that form around the Meiyu-Baiu front (MBF), a joint intensive observation using two X-band Doppler radars was carried out on Taramajima (24.63° N, 124.72° E) and Miyakojima (24.80° N, 125.15° E), in Okinawa, Japan in 2007. During the observation period, the
two different types of MβCSs, embedded-cell storm (CASE I) and LSCSs (CASE II) developed around MBFs, observed on 11 and 12 June 2007. NCEP/NCAR reanalysis and upper sounding data were examined to investigate synoptic conditions and the possibility of convective system occurrence. Using the simplified variational method suggested by Shimizu and Maesaka (2006), three-dimensional wind fields were derived from data obtained by two radars in Taramajima and Miyakojima. Several meteorological variables of the convective systems such as horizontal and vertical wind fields, convergence, and divergence were closely investigated to determine their characteristics, structures and variation patterns.

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Figure 11. The same as Figure 7 but for (a) 2236 LST, (b) 2242 LST, (c) 2254 LST, (d) 2300 LST, (e) 2306 LST and (f) 2312 LST on 12 June 2007.

Figure 12. Schematic illustration of (a) CASE I and (b) CASE II. The three-dimensional distribution of precipitation is shown as grey shading. Upward arrows with solid lines indicate updraft stream lines. Downward arrows with broken lines indicate downdraft stream lines. Short white arrows indicate horizontal winds at 2.5 km ASL. The hatched area indicates the convergence region.

Low-level convergence areas formed in the analysed area due to a pressure pattern related to a strong low-pressure system in the southern part of China, which continuously transported warm and humid air from a tropical air mass at low latitudes to the convergence areas. The atmospheric unstable conditions and strong low-level vertical wind shear were revealed by results of sounding analysis, which provide severe convective storm development in both case. In particular, in CASE II, a high equivalent potential temperature inflow from the southwest at low-level suggests that the LLJ increased the supply of moist air into the lower atmosphere and generated convective instability.

The three-dimensional wind field and reflectivity observed in CASE I indicate that an embedded convection cell approached the analysis area and moved slowly ($\leq 2.5 \text{ m s}^{-1}$) northeastward. Due to the predominant southwesterly inflow, strong convergence areas were gradually enhanced in the lower convective region and the updraft was intensified inside the cell. At the dissipating stage, low level convergence areas were enhanced since the downdraft reflowed into the frontal side of the convective cell led to intensified convection. Therefore, the downdraft may have played a role in maintaining the precipitation system even in the dissipating stage. As a convergence region caused by downdraft and predominant southwesterly wind was located in the southwestern part of the cell, the generation and maintenance of the updraft caused by the convergence allowed the convective region to retain its intensity (Figure 12(a)). In addition, enhanced convergence caused by the downward and southwesterly winds created a new convective cell from the pre-existing cell.

The new precipitation cell from line-shaped convective systems (LSCS) in CASE II generated to the upstream side of the merged cell (pre-existing cells) and moved to the leeward. This structure and condition were similar to the LSCSs of backbuilding type described by Shinoda et al. (2009). However, unlike the LSCSs described by Shinoda et al. (2009), the LSCSs of CASE II were intensified by two factors including the...
merging of cells and outflow from the new cell. While the cells were merged due to the strong prevailing winds, the low-level convergence was enhanced at the rear side of the system and the low-level convergence in the merged cell was reinforced by outflow from the new cell. In addition, the updraft was intensified again at the convection regions, which indicates that the development of convective system was accelerated by the relationship between outflow from the new cell and the strong low-level convergence in the pre-existing cell (Figure 12(b)).

The obtained results explain successfully the evolution and structure of the two different types of MβCSs associated with the MBF over the ocean. The enhanced low level convergence by downdraft re-entrainment and the interaction between pre-existing cell and new developing cell in the convective region could develop and retain the convection under prevailing moist environment with unstable condition. However, further studies based on the statistical analysis and numerical simulations are required to prove the complex development mechanism of MβCSs developing over the ocean around the MBF.

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Supporting information

The following material is available as part of the online article:

**Figure S1.** Vertical cross section along the line A–A’ and B–B’ shown in Figure 2(c). (a) and (b) show wind speed and relative vorticity (10^{-5} \text{s}^{-1} \text{interval and contoured every} 2 \times 10^{-5} \text{s}^{-1}), (c) and (d) show equivalent potential temperature and convergence (10^{-6} \text{s}^{-1} \text{interval and contoured every} 5 \times 10^{-6} \text{s}^{-1}), at 0900 LST the 11 of June 2007. The shaded areas represent wind speeds greater than 10 m s^{-1}, and high equivalent potential temperature greater than 330 K.

**Figure S2.** The same as Figure S1 but for 2100 LST the 12 of June 2007.

**Figure S3.** (a) Vertical profiles of potential temperature \( \theta \), equivalent potential temperature \( \theta_{e} \), saturated equivalent potential temperature \( \theta_{e,s} \), and wind speed (b) direction (right side of the figure) obtained from sounding data at 0900 LST the 11 of June 2007 and 2100 LST the 12 of June 2007.

**Figure S4.** Horizontal distribution of reflectivity and system motion at 2.5 km ASL of (a) CASE I (0836–0936 LST on 11 June 2007) and (b) CASE II (2236–2312 LST on 12 June 2007).

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