Polarimetric Doppler Radar Analysis of Organization of a Stationary Rainband with Changing Orientations in July 2010

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Abstract

A stationary rainband brought heavy rainfall across Gifu and Aichi prefectures, Japan, on 15 July 2010. The orientation of the rainband was initially southwest–northeast, and then changed from west-southwest to east-northeast, before reverting to its original orientation, although the rainband remained stationary over the same area. This study analyzes the organization of the rainband during these three orientation periods using polarimetric Doppler radar. The rainband was maintained by south-southwesterly inflows of high equivalent potential temperature (≥ 340 K) below 2 km, while southwesterly winds prevailed at middle level during the rainfall. It is suggested that these rainband orientations were determined by the travel directions of the convective cells and positions of cell generation relative to the rainband, which in turn were governed by intensities of low-level inflow and cell-origin outflow. During the first orientation period, convective cells formed over a wide area within the rainband and traveled northeastward. Low-level outflows from deep convective cells in the northern section of the rainband shifted the cell-generation area southward, and enhanced south-southwesterly inflows caused the southwestern portion of the rainband to drift slightly to the north; hence, the rainband was oriented from west-southwest to east-northeast. The convective cells were deeper during the second period and low-level outflows were stronger than those in the first stage. The strong outflows formed a cell generation area on the southern lateral side of the rainband, while enhanced low-level inflows contributed to the north-northeastward motion of the generated cells crossing the rainband at an angle of 45° . The outflows and south-southwesterly inflows then weakened, and convective cells formed successively on the southwestern edge of the rainband and moved to the northeast. As a result, the rainband reverted to its original southwest–northeast orientation.

Keywords rainband; convective cell; X-band polarimetric Doppler radar

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1. Introduction

During the rainy season in Japan (the Baiu period), the precipitation system often brings heavy rainfall. In particular, stationary line-shaped precipitation

systems (rainbands) can maintain their intensity and linear structure for several hours over a given location and consequently deliver a large amount of rainfall to localized areas (Kato and Goda 2001; Kato 2006). These rainbands are often organized by convective cells, each of which produces intense rainfall (Yoshizaki et al. 2000; Kato 2006), and the movement of the convective cell is governed by the airflow distribution (Takeda 1981; Seko 2001; Schumacher and Johnson 2005).

Previous studies have examined line-shaped mesoscale convective systems in the United States (Bluestein and Jain 1985; Bryan and Parker 2010; Coniglio et al. 2011). Parker and Johnson (2000) classified these linear precipitation systems on the basis of the positional relationship between the convective and stratiform regions; i.e., the relative extent of stratiform precipitation ahead of, behind, and parallel to the convective lines. Their classification suggested that a generation area and movement of convective cells controlled the shape and maintenance of the precipitation systems. Some convective lines with parallel stratiform precipitation backbuilt to the right of their motion vectors (i.e., back-building formation; Bluestein and Jain 1985). In other convective lines with parallel stratiform precipitation, a location of persistent deep convection appears to give rise to the convective lines via parallel advection. Relationships among the low- and middle-level winds, generation area of convective cells, and orientation of precipitation systems have been particularly investigated in many studies (Lin et al. 1998; Weisman and Rotunno 2004; Schumacher and Johnson 2005). Schumacher and Johnson (2005) showed that a middle-level (925–500 hPa) wind shear vector in a linear mesoscale convective system, with cell motion parallel to convective lines, has a large component parallel to the convective line of the linear precipitation system. Furthermore, Parker (2007) used a numerical model to consider the advection of hydrometeors and elucidate the formation of line-shaped precipitation systems and development of convective cells. These line-shaped precipitation systems in the United States propagate by the generation of convective cells on the leading edge and the presence of a precipitation-origin cold pool (Weisman 1993; Corfidi 2003; Cohen et al. 2007; Coniglio et al. 2011).

Marshall et al. (2011) observed a nocturnal mesoscale convective system, which occurred in northwest–southeast-oriented lines in the initiation, and then evolved to a northeast–southwest-oriented rainband. Trier et al. (2011) revealed that as this meso-

scale convective system began to reorganize, the northeast–southwest-oriented rainband developed within enhanced water vapor mixing ratio zone associated with a mesoscale upward motion coincident with enhanced low-level jet, and the rainband was initially perpendicular to the lower-tropospheric vertical wind shear. During the early stage of the development, the rainband lacked a significant surface cold pool, whereas a cold surface outflow developed owing to penetrative convective downdraft behind the rainband in the later stage.

For assessing convective developments, such as up and downdrafts, signatures of graupel (or hail) are especially helpful (Wakimoto and Bringi 1988). Recent studies using polarimetric radar measurements revealed the presence of graupel particles in convective cells in association with stronger updrafts and downdrafts (Park et al. 2009; Dolan and Rutledge 2010). Dolan and Rutledge (2010) demonstrated that graupel generated within updraft area led the strong downdraft by several minutes in convective cells. Adams-Selin et al. (2013) simulated convective systems with and without graupel. The convective systems with graupel produced stronger convective updraft, stronger surface-based rear-to-front flow, more intense cold pool, and stronger downdraft compared with no-graupel simulations.

Rainbands bringing heavy rainfall during the Baiu period in Japan often remain for extended periods over a given area or move slowly. Many of these rainbands have been investigated with respect to the formation area of convective cells and environmental winds in order to demonstrate their formation mechanisms (Takeda and Seko 1986; Kato and Goda 2001; Adachi et al. 2004; Seko and Nakamura 2005; Kato 2006). These studies revealed that the rainbands were generated by back-building formation, characterized by the successive generation of new convective cells in the upstream region of the precipitation systems. These convective cells moved along the middle- or upper-level wind vectors. Takeda (1981) demonstrated that convective cells formed successively in the same area because of a low-level moist inflow that crossed the precipitation system and then traveled in the same direction as the southwesterly winds in the middle and upper levels. The direction of the low-level winds relative to the middle- or upper-level winds played a role in determining the location of the generation of new convective cells and the formation of rainbands. An advanced numerical modeling study of the formation processes of rainbands (Seko 2001) suggested that new convective cells form in not only the

upstream region but also the flank of the precipitation systems. According to this mechanism, a convective line forms parallel to the middle-level wind and convective cells are generated in the flank of the convective line because of a low-level inflow across the convective line.

Several observational studies have described the vertical structure of convective cells associated with rainbands, in addition to the formation areas and environmental winds. Ishihara et al. (1995) described the structure of a rainband accompanying the Baiu front on the basis of the dual Doppler radar analysis. The orientation of the rainband was directed from the northwest toward the southeast, being parallel to the vertical wind shear between middle and lower levels. Convective cells formed successively at its southwestern edge and moved along the rainband. Some of these convective cells had a radar reflectivity of 30 dBZ and extended above 10 km. Kato (2006) found that the cloud-top heights of convective cells were less than 7 km in the western part of a rainband; however, most of the convective cells extended up to 14 km. His analysis demonstrated the importance of a low-level moist inflow and middle-level dry air in the Baiu frontal zone in terms of development of convective cells accompanying the rainband.

On 15 July 2010, during the Baiu period, heavy rainfall occurred in the area across Gifu and Aichi prefectures, Japan. Figure 1 shows the 12-h integration of rainfall amounts observed by the Japan Meteorological Agency (JMA) radars between 1200 LST and 2350 LST (local standard time, LST = UTC + 9). A heavy-rainfall area formed a band shape across the border between Gifu and Aichi prefectures. A raingauge recorded a large integrated rainfall amount of >200 mm at Garan (Fig. 1a). According to the JMA radar images (obtained every 10 min), the localized heavy rainfall was brought by a rainband that remained over the area. Although the rainband remained stationary, the orientation of the rainband's line shape changed during the period of heavy rainfall. Although many cases of line-shaped and stationary precipitation systems have been reported in Japan, our understanding of a rainband that changes its orientation while remaining over the same location is incomplete. To determine the nature of the organization of such rainbands, this study used observational data to examine the airflow structures in the stationary Baiu rainband of July 2010, as well as the direction of movement, location of formation, and vertical structure of the convective cells associated with the rainband.

Fig. 1. Twelve-hour integration of rainfall estimated from the JMA radar between 1200 and 2350 LST on 15 July 2010. Squares represent locations of raingauges shown in Fig. 6 (a: Garan, b: Tajimi, and c: Nagoya).

The remainder of this paper is organized as follows. Section 2 presents the data used in this study. The environment in which the heavy rain occurred is described in Section 3. The shape of the rainband and airflow distribution around it are described in Section 4, and the movement and development of the convective cells that constitute the precipitation system are analyzed and interpreted in Section 5. Section 6 discusses how convective cells organized the rainband on the basis of the observed structure of convective cells and airflow distributions. Finally, a summary and conclusions are presented in Section 7.

2. Data and methods

To investigate the 3D structure of precipitation and airflow and the motion of the convective cells, multiple Doppler radar observations are advantageous (Heymsfield 1978; Peterson Jr. 1984). With a focus on heavy rainfall in urban areas, X-band polarimetric Doppler radar networks have been operated by the Ministry of Land, Infrastructure, Transport and Tourism (MLIT), Japan since 2010 (Maesaka et al. 2011). The X-band radars, which simultaneously transmit and receive horizontally and vertically polar-

ized waves, perform volume scanning at 5-min intervals. Time series of the 3D data (i.e., 4D data) from the X-band polarimetric Doppler radar networks are particularly suited to the analysis of detailed airflow structures and rapidly developing convective cells. This study takes advantage of the MLIT's X-band polarimetric Doppler radar data to investigate airflow distributions and motions, as well as the vertical structures of convective cells.

The rainband of 15 July 2010 was observed by three MLIT radars located at Anjou, Bisai, and Suzuka (hereinafter, the MLITX radars). The locations and observation settings of the radars are shown in Fig. 2 and listed in Table 1, respectively. Operational data from each radar consist of 15 plan position indicator (PPI) scans at 12 elevation angles every 5 min, with an observation range of 80 km in radius and a radial resolution of 150 m. The PPI scans alternately include the second- or third-lowest angles every minute.

In this study, the radar reflectivity (Z_h) is fully computed with respect to liquid water. Values of Z_h and differential reflectivity are corrected for rainfall attenuation by considering the relationship between the specific differential propagation phase (K_{DP}) and specific attenuation and specific differential attenuation (Jameson 1992; Bringi and Chandrasekar 2001). The Z_h and radial velocity in the PPI scans observed by the MLITX radars were interpolated to Cartesian coordinates (constant altitude PPI; CAPPI). Before the calculation of CAPPI, radial velocities were unfolded manually. The CAPPI was calculated every 5 min, with horizontal and vertical resolutions of 1 km and 0.25 km, respectively, using the 5-min PPI volume scans from the three radars and a weighting function proposed by Cressman (1959). The 5-min composite CAPPI data of Z_h and the CAPPI data of radial velocity from each of the three radars were used in the multiple Doppler radar analysis developed by Gao et al. (1999) and advanced by Shimizu et al. (2008) to investigate the spatial distribution of airflow in the rainband. Shimizu et al. (2008) examined the accuracy of the Doppler radar analysis and showed that the quantitative accuracy was available for heights equal to and below 4 km, whereas only qualitative characteristics of the 3D wind velocity were available for heights above 4 km.

To confirm the presence of graupel particles in deep convective cells, a hydrometeor classification was performed using polarimetric radar variables measured by the MLITX radars. The classification used here was based on a study by Liu and Chan-

Fig. 2. Locations of observation sites. The crosses mark the locations of the MLITX radars (Anjou, Bisai, and Suzuka). The filled square indicates the sounding observation site at Hamamatsu. The filled circle indicates the wind profiler site at Nagoya. The area encircled by the solid line represents the radar observation range covered by the three radars (80 km each). The open squares denote locations of the JMA radars. The shading represents the terrain.

drasekar (2000), except that the range of K_{DP} in which a hydrometeor category is present was estimated by multiplying the K_{DP} range, introduced in their study, by 3.0 to adjust for X-band radars (Kouketsu and Uyeda 2010; Kouketsu et al. 2011). In this classification method, it is assumed that there is only one dominant type of particle present in a radar range-bin volume. Each range-bin in each PPI was classified, and the classified data were translated into CAPPIs using the PPI for each 5-min interval. Each grid in the CAPPI data adopts the nearest range-bin data of the

Table 1. Settings for the MLIT radar measurements at Anjou, Bisai, and Suzuka.

Plan position indicator (PPI) settings of the MLIT radars	
Frequency	9755 MHz (Anjou) 9725 MHz (Bisai) 9795 MHz (Suzuka)
Altitude of radar	25 m (Anjou) 62 m (Bisai) 42 m (Suzuka)
Polarization	H & V, Simultaneous transmit/receive
Number of elevations	12
Elevation angles	1.0°, 1.7°, 2.7°, 3.8°, 5.1°, 6.5°, 8.1°, 9.9°, 11.9°, 14.2°, 16.9°, and 20.0°
Radar coverage in radius	80 km
Rotation rate	18.0° s ⁻¹ (3 rpm)
Pulse width	1.0 μs
Number of integration pulses	100
Pulse repetition frequency	1800 Hz
Beam width	≤ 1.2°
Range gate spacing	150 m
Beam spacing	1.2°
Nyquist velocity	13.8 m s ⁻¹

classified hydrometeor type in the PPI data.

The JMA radars, of which the observation range covers a larger area than the MLIT radars, captured the perspective of the rainband. The horizontal distribution of rain intensity obtained from the JMA radars was used to describe the location and shape of the entire rainband of 15 July 2010 and the change in its orientation direction. The JMA radar products provide gridded rain intensity data with a 1-km horizontal resolution at 2 km above sea level (ASL) every 10 min. The surface rainfall amounts and temperatures during the rainfall event were observed by the Automated Meteorological Data Acquisition System (AMeDAS) of the JMA every 10 min. The JMA mesoscale reanalysis data (MANAL) and soundings at Hamamatsu were used as reference profiles to represent the environment of the study area. The soundings were observed twice a day at 0900 LST and 2100 LST. Moreover, a wind profiler operated by the JMA at Nagoya was used to confirm the vertical profile of horizontal wind around the study area. The wind profiler data were provided every 10 min. The vertical resolution of the wind profiler data was approximately 0.3 km, and the data were collected from heights above 0.45 km.

For detection of a convective area, the threshold of 30 dBZ, which corresponds to a rain rate of approximately 3 mm h⁻¹, has been widely adopted (DeMott and Rutledge 1998; Machado et al. 1998; Lang

and Rutledge 2011). The present study also uses the threshold of 30 dBZ to define an area of intense convective rain. However, because 30-dBZ areas sometimes also occur in stratiform rain, each of the convective cells in this study includes a 45-dBZ area within the 30-dBZ area at least once during its lifetime. Furthermore, high Z_h peaks from the stratiform precipitation were excluded by use of the bright band fraction method proposed by Chen et al. (2003). The convective cells were identified from composite CAPPI data every 5 min. Each convective cell had a peak in Z_h , which was at least 2 dBZ greater than the nearest local minimum value, which is the same threshold used in a method for identifying and tracking convective cells proposed by Shimizu and Uyeda (2012). If more than two peaks were present within a radius of 3 km, only the peak with the maximum value was taken and the others were not considered to be convective cells. Manual tracking of each cell was conducted using the 5-min CAPPI data at 2 km ASL until the peak value fell below the 30-dBZ threshold or until the peak could no longer be identified at 2 km ASL.

3. Environmental conditions

The weather map for 0900 LST on 15 July 2010 is shown in Fig. 3, revealing a stationary front (the Baiu front) extending from the northeast of Japan over the Pacific Ocean to the Sea of Japan and lying

Fig. 3. Weather map for 0900 LST 15 July 2010.
The rectangle indicates the present study area.

along the coast of the Sea of Japan. The Baiu front also extends to the west, over the East China Sea. The area discussed here is located to the south of the Baiu front. The environmental fields of wind and equivalent potential temperature (θ_e) obtained from the JMA mesoscale analysis data on 15–16 July 2012 are shown in Fig. 4. At the 950-hPa level, high- θ_e (≥ 340 K) air was flowing into the south of Gifu prefecture from the southwest (Fig. 4a). Figure 4b shows a time–height cross section of the horizontal winds and θ_e averaged over the dashed rectangular area in Fig. 4a. The wind speed of the inflow was $5\text{--}10\text{ m s}^{-1}$ at low levels below 900 hPa between 1200 LST on 15 July and 0000 LST on 16 July (Fig. 4b). In the middle and upper levels above 850 hPa, southwesterly winds prevailed. After 0000 LST on 16 July, lower- θ_e (< 335 K) westerly winds broke into the middle level around 400–750 hPa, and the speed of low-level high- θ_e inflow was less than 5 m s^{-1} .

Soundings at Hamamatsu on 15 July recorded thermodynamic profiles (Fig. 5), which show that high- θ_e air (> 340 K) was flowing in at low levels below 750 hPa (approximately 2.5 km). In particular, the low-level θ_e was higher at 2100 LST than at 0900 LST. These sounding profiles indicate that warm and moist air was flowing into the south of the precipitation system. The convective available potential energy, lifting condensation level, and level of free convection, based on a parcel from the lowest 500 m at Hamamatsu at 2100 LST, were 1572 J kg^{-1} , 973 hPa, and 935 hPa, respectively. On the basis of these environmental conditions, it is likely that the low-level inflows of the high- θ_e air contributed to the maintenance of convective instability

between 1200 LST on 15 July and 0000 LST on 16 July. The sounding profiles also show the dominance of southwesterly winds above 850 hPa. In particular, a stronger southwesterly wind of 15 m s^{-1} is apparent at middle altitudes around 5 km ASL. The 0°C level, based on the sounding data from Hamamatsu at 2100 LST, was at 4.8 km ASL.

4. Stationary line-shaped precipitation system on 15 July 2010

4.1 Behavior of precipitation system

The 12-h integration of rainfall (Fig. 1) shows that heavy rainfall on 15 July 2010 occurred across Gifu and Aichi prefectures, with the maximum amount exceeding 200 mm. The AMeDAS stations at Garan and Tajimi (locations are marked in Fig. 1) recorded intense rainfall and a drop in surface temperature near this rainfall (Fig. 6). Rainfall rates of at least 1 mm per 10 min continued from 1600 to 2130 LST. The temperature at Tajimi, which was located within the heavy rainfall area, decreased when the rainfall began at around 1700 LST. The decrease in temperature from 1700 to 2130 LST was approximately 4.1°C at Tajimi, whereas the decrease in temperature was small at Nagoya (1.6°C), which was located to the south of the area of heavy rain. The difference in temperature between Nagoya and Tajimi was $2\text{--}3^\circ\text{C}$ during the rainfall. To show the presence of cold pool, distribution of temperature at the mean sea level estimated from surface temperatures from the AMeDAS is shown in Fig. 7. The mean sea level temperature was calculated with the temperature lapse rate of $6.5^\circ\text{C km}^{-1}$. An area of lower temperature is found to the north and northwest of the heavy rainfall area with the difference from the south area of $3^\circ\text{--}5^\circ\text{C}$. This decrease in temperature would result from a cold pool produced by the evaporative cooling associated with precipitation particles around the stationary rainband.

The motion of the rainband can be tracked using the distributions of rainfall intensity observed every 30 min by the JMA radars (Fig. 8). The first convective cell formed at a longitude of 136.6°E and latitude of 35.4°N at 1430 LST, and then moved toward east-northeast (denoted by thin arrows in Fig. 8). This convective cell would be generated by forced ascent due to the orographic effect. The similar mechanism of enhanced convection was reported by Morotomi et al. (2012). The motion of the convective cell stagnated at around a longitude of 137.1°E and latitude of 35.5°N , where high- θ_e southwesterly winds flowed in the low level (Fig. 4), as the rainband began to form (1600 LST).

Fig. 4. (a) Horizontal distribution of equivalent potential temperature (θ_e , shaded) and horizontal winds (vectors) obtained from JMA mesoscale reanalysis data (MANAL) at 950 hPa for 1800 LST 15 July 2010, and (b) time–height cross section of θ_e (shaded) and horizontal winds (barbs) averaged over the area indicated by the dashed rectangle in (a). Terrain above the 950-hPa level is masked by mesh in (a). Square represents the sounding observation site at Hamamatsu.

Fig. 5. Sounding profiles of potential temperature (θ , thick line), equivalent potential temperature (θ_e , dashed line), saturation equivalent potential temperature (θ_{es} , thin line), and horizontal winds (barbs) at Hamamatsu at (a) 0900 LST and (b) 2100 LST on 15 July 2010.

Areas of precipitation exceeding 5 mm h^{-1} were aligned from southwest to northeast at 1600 LST. Between 1600 and 1730 LST, the rainband can be distinguished as an intense area of rainfall ($>10 \text{ mm}$

h^{-1}) that is approximately 90 km in length from southwest to northeast. The width of the rainband, estimated from the average around its middle section, was 10 km. From 1630 LST, the line shape became

Fig. 6. Ten-minute rainfall (bars) and surface temperature (dashed lines) measured by AMEDAS at (a) Garan, (b) Tajimi, and (c) Nagoya. Locations of these sites are marked in Fig. 1.

clear. The rainband maintained its intensity and shape in the same area until 1800 LST. From 1800 LST, the orientation of the rainband changed to west-southwest to east-northeast, but it remained in approximately the

same area. The rainband rotated clockwise around a pivot point at a longitude of 137.0°E and latitude of 35.4°N , with the southwest portion drifting to north and the northeast portion to south. The length and width of the rainband were approximately 60 and 17 km, respectively, at this time; i.e., wider than before 1800 LST. The rainband maintained its intensity and shape until 2000 LST. While the orientation of the rainband defined by precipitation rates above 10 mm h^{-1} remained in a west-southwest to east-northeast orientation, the intense rain region ($>50 \text{ mm h}^{-1}$) began to change to a southwest–northeast orientation at 2000 LST. From 2000 LST, the westerly portion of the rainband from 136.9°E dissipates as the northeasterly portion drifts slightly to the north. The length and width of the rainband were approximately 43 and 12 km, respectively, by this point. After 2130 LST, the rainband dissipated completely and another line-shaped precipitation system, aligned from southwest to northeast, passed over the area from west to east. The latter precipitation system generated new convective cells on its frontal edge as it propagated eastward.

From this time series of the horizontal distribution of rainfall intensity observed by the JMA radars, it can be seen that the rainband that generated the intense rainfall between 1600 and 2130 LST became stationary over one area; nevertheless, the orientation of the rainband changed. Now focusing on this stationary rainband and its changing orientation and analyzing its development during three periods defined by this orientation: Period A covers 1600–1755 LST, during which the orientation was southwest–northeast; Period B is 1800–1955 LST, during which the orientation was west-southwest to east-northeast; and Period C is 2000–2130 LST, when the alignment of the intense band of precipitation reverted to the southwest–northeast orientation.

4.2 Airflow distributions

Airflow structure around the rainband was revealed in the horizontal winds and divergence estimated from a multiple Doppler radar analysis using the three MLITX radars. Figure 9 shows the time-averaged divergence and horizontal winds during Period A (1600–1755 LST) at 1.5 km ASL. Despite the average being taken over 2 h, a narrow line-shaped convergence zone, shown as an area of divergence of less than $-2 \times 10^{-4} \text{ s}^{-1}$, is well-defined in Fig. 9a. The line-shaped convergence zone was formed by the south-southwesterly and southwesterly winds. The location of the convergence zone corresponds to the rainband.

Fig. 7. Horizontal distribution of mean sea level temperature estimated from AMeDAS temperature, assuming that the temperature lapse rate is $6.5^{\circ}\text{C km}^{-1}$, at (a) 1700, (b) 1900, and (c) 2100 LST on 15 July 2010. The dots denote locations of AMeDAS sites.

Figures 9b and 9c show hodographs averaged over the north and south sides, respectively, of the convergence zone represented in Fig. 9a. On the north side, the hodograph indicates the predominance of the southwesterly winds in all layers (Fig. 9b). On the south side, south-southwesterly winds prevail below a height of 2 km with a wind speed of approximately 8 m s^{-1} (Fig. 9c). At the middle and upper altitudes ($>4 \text{ km ASL}$), on both the south and north sides, stronger southwesterly winds prevail compared with low altitudes. These hodographs indicate that the low-level inflows cross the convergence line and rainband, while the middle- and upper-level winds are parallel to the orientation of the rainband. A low-level vertical wind shear from 0.5 to 2.5 km is approximately normal to the rainband orientation, whereas a vertical wind shear vector between the low-level inflows and middle-level winds (0.5–5 km) has both a line-normal component and line-parallel component (Fig. 9c).

During Period B (1800–1955 LST), a convergence zone, shown as an area of divergence less than $-2 \times 10^{-4} \text{ s}^{-1}$, was aligned mostly parallel to the rainband (Fig. 10a). This convergence was maintained by inflows of strong south-southwesterly winds compared with the north side. Winds between southwest and south-southwest predominated on the north side of the convergence zone. Hodographs averaged over the north and south sides, shown in Fig. 10a reveal that the southwesterly winds prevail in the middle and upper levels ($>4 \text{ km ASL}$) on both sides (Figs. 10b and 10c). Meanwhile, below 2 km ASL on

the north side, wind directions were between southwest and south-southwest (Fig. 10b). On the south side, the south-southwesterly winds prevail below 2 km ASL (Fig. 10c). The strength of the low-level south-southwesterly winds was approximately 11 m s^{-1} , which is stronger than that in Period A. The low-level wind shear from 0.5 to 2.5 km during Period B is toward southeast, approximately normal to the rainband orientation that is from west-southwest to east-northeast. The wind shear vector between the inflows and middle-level winds (0.5–5 km) has both line-normal and line-parallel components, although it was tilted slightly to the south compared with that of Period A (Fig. 10c). Both low-level and deeper wind shears suggest stronger low-level inflows during Period B.

For Period C, a weak convergence zone (approximately $-2 \times 10^{-4} \text{ s}^{-1}$) formed on the Gifu prefecture side (Fig. 11a). On the north side of the convergence zone, south-southwesterly to south-southeasterly winds predominated below 2 km ASL (Fig. 11b). The northward component of the low-level winds on the north side was slightly stronger than that in Period B. The winds between 0.5 and 1.0 km flowed to the northwest on the north side. These wind directions seem to be affected by winds toward the latter fast-moving precipitation system on the west. On the south side of the rainband, the south-southwesterly winds prevailed below 2 km ASL, in common with the former two periods. The low-level south-southwesterly wind speeds were approximately 2 m less

Fig. 8. Time series of rain intensity observed by the JMA radar every 30 minutes from 1430 to 2300 LST on 15 July 2010.

than those for Period B (Fig. 11c). The distribution of low-level winds and vertical wind shears on the south side is similar to that for Period A. The low-level inflows on the south side through these periods correspond to high- θ_e inflows at low levels, shown in Fig. 4. From the hodographs and θ_e environment shown in Figs. 4 and 5, it appears that the rainband was maintained over the convergence zone by the high- θ_e south-southwesterly inflows below 2 km ASL.

Temporal variations in the low- and middle-level flows were examined using wind profiler measure-

ments from Nagoya, which is located 30 km south of the rainband (Fig. 12). Zonal and meridional components are represented by shading in Figs. 12a and 12b, respectively. Throughout the three periods, southerly to south-southwesterly winds predominated below 2 km ASL and southwesterly winds predominated above 2 km ASL. These winds are consistent with the multiple Doppler radar analysis shown in Figs. 9–11. The middle-level wind speeds above 3.5 km ASL increased to 15–17 m s⁻¹ after 1930 LST, the later phase of Period B. In particular, the zonal compo-

nent above 3.5 km ASL became stronger from 1930 LST and exceeded 12 m s^{-1} through Period C. On the other hand, the meridional component was intensified to values exceeding 9 m s^{-1} below 1.7 km ASL from 1750 to 1940 LST, with the south-southwesterly wind speed reaching 10 m s^{-1} . This indicates that the stronger low-level moist air flowed into the rainband

during Period B. After 1940 LST, the end of Period B, the low-level flow decreased to $5\text{--}8 \text{ m s}^{-1}$. Additional strong southerly winds after 2110 LST represent inflows toward the fast moving precipitation system in the west.

5. Convective cells within the rainband

5.1 Movement of convective cells

The moving direction and speed of convective cell within the rainband were estimated by tracking the cells, and Fig. 13 shows locations and tracks covering the lifetimes of the cells. To exclude convective cells leaving each period during their lifetimes, the period for selection of the convective cells was limited to 30 min in the middle of each period (i.e., 1645–1715 LST for Period A, 1900–1930 LST for Period B, and 2030–2100 LST for Period C). Each circle in Fig. 13 represents the location of the Z_h peak within each convective cell. Convective cells detected in the selected period were tracked for their lifetimes. Contours in this figure represent time-averaged Z_h greater than 40 dBZ during each period defined above.

Convective cells during Period A were aligned along the rainband shown as an intense area greater than 40 dBZ being directed from southwest to northeast (Fig. 13a), each of which traveled on the rainband toward northeast. Most of convective cells formed over a wide area within the rainband, and few cells formed on just the south side of the rainband (Fig. 13b). An intense rain area with reflectivity greater than 40 dBZ in Period B (Fig. 13c) indicates that the rainband rotated clockwise, with a pivot point around a longitude of 137.0°E and latitude of 35.4°N , at the

Fig. 9. (a) Time average of divergence (shade) and horizontal winds (vectors) at 1.5 km ASL, and (b), (c) hodographs of horizontal winds averaged over dashed-square areas represented as Na and Sa, respectively, in (a). The divergence and horizontal winds were estimated from the multiple Doppler radar analysis and averaged between 1600 and 1755 LST on 15 July 2010, which is composed of 24 sets of CAPPI data. Horizontal velocity in (b) and (c) is displayed from a height of 0.5 to 10 km every 0.5 km. Thin arrows in (b) and (c) denote wind vectors at 0.5, 2, and 5 km. Thick arrows in (c) represent vertical wind shear vectors from 0.5 to 2.5 km (black) and from 0.5 to 5 km (gray). The low-level winds equal to and below 4 km ASL have quantitative accuracy; however, the middle- and upper-level winds represent qualitative characteristics.

Fig. 10. As for Fig. 8, but averaged between 1800 and 1955 LST on 15 July 2010, which includes 24 sets of CAPPI data.

border between Aichi and Gifu prefectures. Unlike Period A, during Period B, an intense rain area within the rainband (represented by mean reflectivity greater

Fig. 11. As for Fig. 8, but averaged between 2000 and 2130 LST on 15 July 2010, which includes 19 sets of CAPPI data.

than 40 dBZ) was formed by the alignment of convective cells with $Z_h > 45$ dBZ from west-southwest to east-northeast (Fig. 13c). These convective cells trav-

Fig. 12. Time–height cross section of horizontal winds (barbs) measured by a wind profiler at Nagoya from 1600 to 2130 LST on 15 July 2010. Shading in (a) and (b) represents the zonal and meridional components, respectively, of the horizontal winds.

eled toward the north-northeast. The mean speed of the convective cells is shown in Table 2. Relative to Period A, the zonal component of the mean speed is comparable, but the meridional component is significantly larger for the convective cells during Period B. These convective cells formed successively in the southern part, or just to the southern lateral side, of the rainband (Fig. 13d). An area of intense precipitation during Period C shifted toward the Gifu prefecture side (Fig. 13e) as the orientation became south-

west–northeast. Convective cells formed on the southwestern part of the rainband and generally traveled to the northeast (Fig. 13f). This represents a back-building formation process. The mean speed listed in Table 2 indicates the moving direction shifts to the east (clockwise) owing to intensification of the zonal component.

5.2 Vertical structure of convective cells

Deep convective cells developed during Period A

Fig. 13. (a) Locations and tracks of convective cells between 1645 and 1715 LST. The convective cells are tracked beyond the period 1645–1715 LST using 5-minute CAPPI data at 2 km ASL. The position of the Z_h peak within each convective cell every 5 minutes is indicated by circles. Each of the same convective cells are connected by solid lines in time. Color scale represents Z_h . (b) Locations where the convective cells formed. The circle, triangle, and star indicate formation times of 1600–1640, 1645–1655, and 1700–1715 LST, respectively. Contours represent Z_h at 2 km ASL averaged between 1645 and 1715 LST. (c), (d) As for (a) and (b), respectively, but between 1900 and 1930 LST. The circle, triangle, and star in (d) indicate formation times of 1825–1855, 1900–1910, and 1915–1930 LST, respectively. (e), (f) As for (a) and (b), respectively, but between 2030 and 2100 LST. The circle, triangle, and star in (f) indicate formation times of 2000–2010, 2015–2030, and 2035–2050 LST, respectively.

Table 2. Moving speed of convective cells estimated from CAPPI data at 2 km ASL.

	Zonal component [m s ⁻¹]	Meridional component [m s ⁻¹]	Number of cells
Period A	5.0	5.1	17
Period B	3.9	6.7	18
Period C	6.9	7.3	9

(Fig. 13a) with reflectivities of 30 dBZ and echo-top heights exceeding 5.3 km. The mean echo-top height of 30 dBZ during Period A was 8.7 km and the maximum height was 13.9 km. To reveal the vertical structure of individual convective cells, horizontal and vertical cross sections of the rainband and convective cells are presented in Fig. 14. The selected convective cell presented in Fig. 14 is the largest Z_h stage of the convective cell shown in Fig. 13a, traveling over the range where Doppler radar analysis was available during its lifetime. The horizontal cross section at 2 km ASL (Fig. 14a) reveals low-level south-southwesterly inflows on the south side of the rainband. In a vertical cross section of the deep convective cell along the direction of the low-level inflow (Fig. 14b), cell-relative flows produce a weak convergence at low levels around 1 km ASL and divergence at heights of 7.5–10 km ASL. In another vertical cross section along the rainband (Fig. 14c), updraft areas greater than 4 m s⁻¹ were found between 4 and 11 km ASL around the convective cell (indicated by a bold arrow) and on the northeast side (leeward) of the convective cell drifting with the cell-relative flow (approximately 6 m s⁻¹) toward the southwest. Around the convective cell, an area of Z_h greater than 30 dBZ reached 10 km ASL, which exceeds the 0°C level significantly. The area of large Z_h also reached high altitudes of up to 11 km ASL, with updrafts on the northeast side of the convective cell. Meanwhile, a region of weak downdraft of approximately 1 m s⁻¹ was produced on the north side of the convective cell below 2 km ASL, around 15 km distant from point b (Fig. 14c).

The areas of large Z_h and updraft above the 0°C level suggest the presence of graupel particles generated by the updrafts associated with the deep convective cells. Regions of dry graupel particles estimated from the hydrometeor classification are illustrated in Fig. 15, which shows a horizontal cross section of Z_h at 2 km ASL and horizontal winds at 5 km ASL from the same CAPPI data displayed in Fig. 14. Dry graupel regions estimated using the hydrometeor classification at 1705 LST are horizontally projected. This figure indicates that graupel regions were distributed

within the rainband parallel to the rainband orientation due to advection by southwesterly winds in the middle and upper levels. A vertical cross section of Z_h along the rainband is shown in Fig. 15b (c–c' line). The echo-top height of the 30-dBZ zone was 10–11 km ASL, and that of 35 dBZ was just below 10 km ASL. The graupel regions were distributed between 6 and 9 km ASL within an area of Z_h greater than 30 dBZ on the leeward side of a region of Z_h greater than 50 dBZ. The graupel distributions suggest falling graupel particles generated by the stronger updraft produced the low-level downdraft on the north side of the convective cell (Fig. 14c).

For convective cells during Period B, the mean echo-top height of the 30-dBZ region was 8.1 km and the maximum height was 15.0 km. Figure 16 shows the horizontal and vertical distributions of the rainband for Period B at 1910 LST. A convective cell on the south side (indicated by a bold arrow) shows a large Z_h with a low echo-top height below 7 km ASL (Figs. 16b, c), but the convective cell eventually deepened on moving toward the north-northeast. Echo-top heights in the 30-dBZ zone became higher to the north (Fig. 16d) accompanying a stronger updraft (>5 m s⁻¹, approximately 14 km ASL). This distribution suggests that shallower convective cells forming on the south lateral side of the rainband deepened considerably on moving north-northeast. The shallower convective cells were not likely to prevent inflows toward the deep convective cells. Strong cell-relative southward flows were present below 2 km ASL in association with a weak downdraft area (≥ 1 m s⁻¹) around the northern high echo-top areas below 4 km ASL. These flows are outflows from deep convective cells to the northern side. The outflow speed was approximately 10 m s⁻¹ relative to the convective cells, which is stronger than seen in a deep convective cell from Period A.

Dry graupel regions at 1910 LST estimated from the hydrometeor classification are illustrated in Figs. 17 and 15. Graupel regions spread horizontally to the northern side of the rainband (leeward of the rainband at the middle level). A vertical cross section of

Fig. 14. (a) Horizontal cross section of Z_h (shading) and horizontal winds (vectors) at 2 km ASL from CAPPI data at 1705 LST. Contours show Z_h every 5 dBZ from 10 dBZ. (b), (c) Vertical cross sections along (b) a-a' and (c) b-b' lines indicated in (a). Solid lines in (b) and (c) represent updraft every 1 m s^{-1} from 2 m s^{-1} , and dashed lines represent downdraft every 1 m s^{-1} from 0 m s^{-1} . Vectors in (b) and (c) represent cell-relative winds along lines a-a' and b-b', respectively. Bold arrows in (b) and (c) indicate a targeted cell described in the text.

Z_h along the g-g' line (Fig. 17b) shows the 30- and 35-dBZ areas extending up to 15 and 12 km, respectively, and the graupel regions reaching higher altitudes than in Period A on the northern side, just leeward of the updraft areas (Fig. 16d). These figures suggest that hydrometeors, including graupel particles, generated by the stronger updrafts and advected by the middle- and upper-level southwesterly winds contributed to the intense rain in the rainband during the period of deep development.

During Period C, the mean echo-top height of convective cells shown in Fig. 13e was 7.9 km and the maximum was 12.1 km; these values are lower than those for Period B. Figure 18 shows horizontal and vertical cross sections of the rainband for Period C at 2050 LST, with a convective cell having maximum Z_h at a height of 2 km and a 30-dBZ echo-top height of 8 km. The updraft shown at heights between 5 and 11 km is weak compared with the two former periods

(approximately 2 m s^{-1}). Cell-relative flows indicate weaker outflows (i.e., southwestward flows in Fig. 18b) of less than 7 m s^{-1} , which probably generated little graupel (the number of estimated dry graupel grids is 3, which is equal to 0.75 km^3 ; not shown).

Temporal variations in the 30-dBZ echo-top height, wind, and graupel distributions between 1600 and 2100 LST, within an area of $136.70^\circ\text{--}137.30^\circ\text{E}$ and $35.11^\circ\text{--}35.61^\circ\text{N}$, are shown in Fig. 19. The period 2105–2130 LST was excluded because the echo tops and graupel regions within the study area were related to a different precipitation system that appeared in the west. During Period A, an echo-top height of less than 12 km predominates and the southward components are less well represented. On the other hand, during Period B, echo-top heights extend above 12 km ASL, which allowed the southward components to become larger. The intense southward components generated convergences with northward winds to just south of

Fig. 15. (a) Horizontal cross section of Z_h at 2 km ASL (contours), horizontal winds at 5 km ASL (vectors), and horizontal projection of dry-graupel regions (shading) estimated from a hydrometeor classification from CAPPI data at 1705 LST. Contours of Z_h in (a) are drawn every 10 dBZ from 30 dBZ. (b) Vertical cross section of Z_h (contours) and dry-graupel regions along line c-c' in (a). Contours of Z_h in (b) are drawn every 5 dBZ from 30 dBZ.

the high echo-top areas until 2000 LST. After 2000 LST, the southward component is not distinguishable as the 30-dBZ echo-top height decreases to less than 10 km (Period C).

The relationship between outflows and graupel particles within the deep convection was investigated by analyzing the temporal variations in the maximum height and total volume of dry graupel, shown in Figs. 19b and 19c, respectively. The maximum heights

and total volumes of graupel during Period A include higher values from 1700 to 1730 LST in the northern area above 35.4°N. After 1730 LST, the later phase of Period A, the southward component of low-level winds appears in the northern area. This indicates that low-level southward outflows occurred in association with the fall of graupel particles generated in the deeper convective cells. Corresponding to increasing echo-top height during Period B, the total volume of dry graupel increases to large peaks at 1805 and 1905 LST. This figure suggests that, during Period B, the stronger, southward outflows were produced in association with deep convective cells generating graupel particles and then formed convergences to the south of the rainband. At the end of Period B, after 1930 LST, the graupel height and volume decrease with time. Corresponding to the decreasing graupel height and volume, the southward component of the low-level winds weakens. During Period C, the graupel height and volume decrease continuously as the southward component disappears, and the 30-dBZ echo-top heights become lower than 10 km ASL.

6. Discussion

On the basis of our observational analysis, how the convective cells controlled the organization of rainband and influenced the orientation of rainband during each period is now considered.

The first convective cell generated in a mountain area to the west of the rainband area at 1430 LST (Figs. 2, 8) moved toward east-northeast with the low-level south-southwesterly inflows and then stagnated at a convergence area when the southwest-northeast-oriented rainband began to form at the area. Rotunno et al. (1988) suggested that a storm cell's outflow situated in low-level wind shear produces less-inhibited and deeper lifting on the downshear side of the outflow compared with that in a situation without low-level shear. The rainband in this study seems to develop in a similar situation with low-level wind shear.

The organization of the line shape with a southwest-northeast orientation during Period A can be described as follows. Convective cells were formed and developed over a wide area within the rainband because of the low-level high- θ_e inflows (≥ 340 K) from the south-southwest. These convective cells traveled northeast along the rainband with the middle-level southwesterly winds. The orientation of the rainband was parallel to the direction of cell motion and middle-level southwesterly wind. Consequently, convective cells were generally aligned in a south-

Fig. 16. (a) Horizontal cross section of Z_h (shading) and horizontal winds (vectors) at 2 km ASL from CAPPI data at 1910 LST. Contours show Z_h every 5 dBZ from 10 dBZ. (b), (c), (d) Vertical cross sections along (b) d-d', (c) e-e', and (d) f-f' lines indicated in (a). Solid lines in (b), (c), and (d) represent updraft every 1 m s^{-1} from 2 m s^{-1} , and dashed lines represent downdraft every 1 m s^{-1} from 0 m s^{-1} . Vectors in (b) represent cell-relative winds along line d-d'. Bold arrows in (b) and (c) indicate a targeted cell described in the text.

west to northeast direction and the rainband formed a long linear shape. Furthermore, the graupel particles in the deep convective cells were advected by the middle-level winds whose direction corresponded to the orientation of the rainband. The advected graupel particles also contributed to the narrow and linear shape of the rainband.

The organization of the line shape with an orientation of west-southwest to east-northeast during Period B can be described as follows. Each of the convective cells formed successively just to the southern lateral side of the rainband. These convective cells provided intense rain, with low echo-top heights to

the south of the rainband, but with higher echo-top heights on the northern side. Each convective cell moved north-northeast, at an angle of approximately 45° to the rainband orientation. These cell movements maintained the shape and orientation of the rainband during Period B. The stronger southerly winds at low levels contributed to the movement of convective cells to the north-northeast, particularly during the low echo-top stages. Moreover, the stronger, high- θ_e south-southwesterly inflows contributed to the development of deep convective cells and formation of graupel particles over larger areas. Adams-Selin et al. (2013), who simulated convective systems

Fig. 17. (a) As for Fig. 14a, but from CAPPI data at 1910 LST. (b) As for Fig. 14b, but along line $g-g'$.

with and without graupel, suggested that the graupel system produced intensified downdraft compared with the no-graupel simulations. In Japan, deep convective cells with solid hydrometeors (i.e., hail and/or graupel) could produce stronger downdrafts and outflows relative to convective cells with hydrometeors and downdrafts limited in lower altitudes (Ohno 1995; Shimizu et al. 2008). In the present study, therefore, downdrafts due to falling hydrometeor particles, including graupel, and stronger outflows were generated near the deep convective cells on the northern lateral side of the rainband. The weakness of the low-level southerly flow on the northern side of the rainband (Fig. 10b) probably represents a block of southerly winds through the rainband caused by the

strong outflows. The outflows formed new convective cells to the southern lateral side of the rainband by convergence with the low-level south-southwesterly winds.

For period C, the organization of the rainband with a southwest–northeast orientation can be described as follows. Convective cells were formed successively in a limited area on the southwest edge by the high- θ_e , low-level south-southwesterly inflows and were moved northeast by the stronger middle-level southwesterly winds ($15\text{--}17\text{ m s}^{-1}$). The orientation of the rainband was parallel to the direction of the cell motion and middle-level southwesterly wind. Convective cells developing with lower echo-top heights and little graupel particles than those during Period B produced weaker outflows. The low-level south-southwesterly winds went through the rainband with little blocking by outflows. This is shown by the stronger southerly winds on the northern side of the rainband compared with Period B (Fig. 11b).

The cell motions for Periods A and C, parallel to the convective lines corresponding to the middle- and upper-level wind directions, conform with the precipitation system discussed by Takeda (1981). In particular, the generation areas of the convective cells during Period C were limited to the southwest edge of the rainband, representing a back-building formation process (Bluestein and Jain 1985), whereas those during Period A were a wide area within the rainband. On the other hand, for Period B, convective cells were formed on the southern flank of the rainband by the low-level inflows across the rainband (named as the back- and side-building type by Seko 2001). A somewhat similar behavior has been observed for mesoscale convective systems that initially take on the “parallel stratiform” archetype and eventually evolve into the “trailing stratiform” archetype after the cold outflow strengthens (Parker and Johnson 2000; Parker 2007). While the current rainband does not possess the stratiform precipitation, the airflow observed during Periods A and C was very similar to that for the parallel stratiform archetype, whereas the airflow observed during Period B was similar to that for the trailing stratiform archetype.

To facilitate consideration of the factors that produce the different convective lines, which resulted from a change in the generation position and direction of movement of the convective cells, the intensities of the inflow and outflow were taken into account. Figure 20 shows horizontal winds at 950 hPa from MANAL. The south to southwesterly winds were enhanced at 1800 LST in the western area of Aichi

Fig. 18. (a) Horizontal cross section of Z_h (shading) and horizontal winds (vectors) at 2 km ASL from CAPPI data at 2050 LST. Contours show Z_h every 5 dBZ from 10 dBZ. (b) Vertical cross sections along line h-h' in (a). Solid lines in (b), (c), and (d) represent updraft every 1 m s^{-1} from 2 m s^{-1} , and dashed lines represent downdraft every 1 m s^{-1} from 0 m s^{-1} . Vectors in (b) represent cell-relative winds along line h-h'.

prefecture and the southern area of Gifu prefecture, where the rainband developed. The environmental enhancement contributed to the enhancement of southward component of the low-level wind shear (Fig. 10c). The vertical wind shear enhanced by low-level winds is frequently observed in association with mesoscale convective systems in the United States (Rotunno et al. 1988; Trier et al. 2011). The low-level winds at 2100 LST (Period C) were still stronger than those at 1500 LST, meanwhile the middle level winds were also intensified (Figs. 11, 12). These winds during Period C resulted in similar vertical wind shear as those in Period A (Fig. 11c). The line-normal low-level wind shears were found during Periods A, B, and C. Especially, the line-normal component during Period B, which resulted from the environmental enhancement of the mesoscale low-level winds (Figs. 10, 12, 20).

Figure 19 shows that the distinct outflows associated with deep convective cells, including graupel particles, begin to appear from the end of Period A, around 1730 LST in the northern area. The graupel leading the low-level downdraft was also reported by Dolan and Rutledge (2010). Meanwhile, intensification of low-level southerly wind begins to appear from 1750 LST (Fig. 12). The outflows in the northern area at the end of Period A moved the generation area of the convective cell south, and the strong southerly winds caused the southwestern portion

of the rainband to drift slightly to the north. Consequently, the rainband orientation became west-southwest to east-northeast during Period B. The cell-derived southward outflows and intensification of the low-level southerly winds would contribute to the continuous generation of new convective cells to the south of the rainband and also to the stagnation of the rainband. This formation mechanism is similar to that for long-lived squall lines presented in Rotunno et al. (1988) and Weisman and Rotunno (2004), suggesting that a cold pool triggers a new cell on its downshear side. The low-level inflows decreased slightly, and thereby development of convective cells and generation of graupel particles begin to decline before and during Period C. The weak outflow from each shallower convective cell probably formed weaker convergence within the rainband and limited cell generation areas to the southwest edge of the rainband compared with Period A. The stronger middle-level winds than those in Period B contributed to the north-eastward motion of the convective cells during Period C.

7. Summary and conclusions

A stationary rainband brought heavy rainfall across the Gifu and Aichi prefectures on 15 July 2010. The rainband persisted from 1600 to 2130 LST on that day although its orientation changed at around 1800 LST and at 2000 LST. The airflow distributions, and

Fig. 19. (a) Time–latitude cross section of maximum echo-top heights of 30 dBZ (shading) and maximum southward components of radar-estimated horizontal winds (arrows) at 0.5 km ASL. The maximum echo-top heights and southward components were extracted from an area between 136.7°E and 137.3°E in longitude every 1 km in latitude. Blue and red arrows represent the southward and northward (negative southward component) directions, respectively. (b) Temporal variation of maximum heights for which dry graupel was estimated. Solid and dashed lines represent estimations from the same area as (a) and the northern area above 35.4°N, respectively. (c) As for (b), but for the integrated volume for which dry graupel was estimated.

Fig. 20. Horizontal distributions of horizontal wind at 950 hPa from MANAL at (a) 1500, (b) 1800, and (c) 2100 LST. Terrain above the 950-hPa level is masked by mesh.

the motions and vertical structures of convective cells that controlled the orientation of rainband, were investigated using polarimetric Doppler radar analysis from three X-band radars.

High- θ_e , south-southwesterly (≥ 340 K) winds below 900 hPa maintained convective instability over the area around Gifu and Aichi prefectures, while southwesterly winds prevailed in the middle and upper levels above 4 km ASL during the period of heavy rainfall. The rainband orientation was southwest–northeast between 1600 and 1755 LST, but the orientation changed to west-southwest to east-northeast in the period 1800–1955 LST, while remaining over the same area as the rainband rotated clockwise around a pivot point at a longitude of 137.0°E and latitude of 35.4°N. Finally, between 2000 and 2130 LST, the orientation of the rainband reverted to its original southwest–northeast alignment. This observational study has demonstrated, from the perspective of the motion and vertical structures of convective cells and airflow distributions, the organization of a rainband as it changed orientation. The changing rainband orientation was controlled by the direction of movement of the convective cells and position of cell generation relative to the rainband. These motions were governed by intensities of low-level southerly inflow and cell-origin outflow.

Schematic representations of the airflow distributions and convective cells controlling the organization of the rainband during the three periods are shown in Fig. 21. The organization of the rainband when its orientation was southwest–northeast was as follows (Period A; Figs. 21a, b). The length and width of the rainband were approximately 90 and 10 km, respectively. Convective cells formed successively and developed over a wide area within the rainband due to the high- θ_e , low-level south-southwesterly inflows, and then traveled to the northeast with the middle-level southwesterly winds. The travel direction of convective cells was parallel to the orientation of the rainband (Fig. 21a). Deep convective cells developed with 30-dBZ echo-top heights of up to 13.5 km ASL due to the high- θ_e south-southwesterly inflows below 2 km ASL. Graupel particles were generated in the deep convective cells and distributed within the rainband parallel to the rainband orientation by the middle- and upper-level southwesterly advectations. Downdrafts (approximately 1 m s^{-1}) and weaker outflow areas were located on the northern side of the convective cells (Fig. 21b).

During the second phase, the orientation of the rainband was west-southwest to east-northeast, and

Fig. 21. (a), (c), and (e) Horizontal and (b), (d), and (f) vertical schematic representations of the rainband in each of (a), (b) Period A; (c), (d) Period B; and (e), (f) Period C.

it was organized as follows (Period B; Figs. 21c, d). The length and width of the rainband were approximately 60 and 17 km, respectively. Convective cells formed successively on the southern lateral side of the rainband and were moved to the north-northeast by environmental enhancement of the low-level winds crossing the rainband at an angle of 45° (Fig. 21c). Deep convective cells with 30-dBZ echo-top heights reaching up to 15 km ASL predominated in the north lateral portion of the rainband. Graupel particles were generated over large areas in the deep convective cells and horizontally distributed along the northern lateral portion by the middle- and upper-level southwesterly advective flows. Downdrafts (approximately 1 m s^{-1}) and stronger outflows were produced at the low level near the deep convective cells. The outflows formed new convective cells on the southern lateral side of the rainband by convergence with the low-level south-southwesterly winds (Fig. 21d).

During the final phase, the orientation of the rainband was southwest–northeast between 2000 and 2130 LST, and it was organized as follows (Period C; Figs. 21e, f). The length and width of the rainband were approximately 43 and 11 km, respectively. The high- θ_e south-southwesterly inflows contributed to the successive generation of convective cells on the southwestern edge of the rainband. Each convective cell moved parallel to the rainband with the stronger

middle-level winds (Fig. 21e). Echo-top heights were lower than 9.5 km ASL. Convective cells generated little graupel and had weak outflows. Compared with the first and second periods, these weak outflows from convective cells with lower echo-top heights produced weak convergence within the rainband and limited cell generation areas to the southwest edge (Fig. 21f).

The low-level outflows associated with the deep convective cells became clearly distinguishable in the northern area toward the end of Period A and stronger throughout Period B. Meanwhile, the southerly inflows were intensified from the end of Period A. The outflows in the north moved the cell's generation area to the south, and the coincident environmental enhancement of the low-level wind caused the southwest portion of the rainband to drift slightly to the north. Consequently, the rainband shifted its orientation and was aligned from west-southwest to east-northeast during Period B. In contrast to the transition to Period B, approaching the beginning of Period C, the development of convective cells and generation of graupel particles began to decline, and the southward outflows dissipated during Period C, as the orientation reverted to southwest-to-northeast.

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