Relationship between Precipitation Core Behavior in Cumulonimbus Clouds and Surface Rainfall Intensity on 18 August 2011 in the Kanto Region, Japan

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Abstract

This paper analyzes the behavior of the precipitation cores (PCOs) in three cumulonimbus clouds that caused localized heavy rainfall on 18 August 2011 in the Kanto region, Japan, and their relationship with temporal variations in surface rainfall intensity. The 3D structure of the cumulonimbus clouds was observed at 2-min intervals using a research X-band dual-polarization radar. A PCO was defined as a 3D contiguous region that contained one local maximum of horizontal radar reflectivity \( Z_h \), and the PCOs were automatically detected using adaptive thresholds. Subjective tracking procedures using the PCO dataset, which was based on observations made every 2 min, identified 15 PCOs during the total lifespan of the three cumulonimbus clouds.

The PCOs generally descended toward the ground after their appearance aloft. The average appearance height of the PCOs of 5.25 km above sea level (ASL) was slightly above the ambient 0°C level (5.1 km ASL). The duration of each PCO was roughly proportional to its appearance height.

Of the 12 temporal peaks at maximum surface rainfall intensity (> 10 mm h\(^{-1}\)) recorded from the three cumulonimbus clouds, 10 were associated with the descent of PCOs. In each cumulonimbus cloud, the first PCO was detected 10–12 min before rainfall heavier than 10 mm h\(^{-1}\) was recorded. These results indicate that the behavior of PCOs is closely related to the onset of strong surface rainfall and subsequent fluctuations in surface rainfall intensity.

Keywords  precipitation core; localized heavy rainfall; dual-polarization radar

1. Introduction

Localized heavy rainfall can cause significant damage in the densely populated urban areas of Japan. For example, the sudden flooding of the sewers at Zoshigaya (Toshima district, Tokyo) and the flash flood along the Toga River (Kobe, Hyogo) in 2008 were caused by localized heavy rainfall (Kato and Maki 2009; Hirano and Maki 2010; Kim et al. 2012); this resulted in several people being swept away in each incident.

Such localized heavy rainfall can occur in association with horizontally small cumulonimbus clouds under the atmospheric situations that are not influenced by synoptic disturbances such as typhoons or frontal systems. For example, 179 cumulonimbus clouds occurred in the Tokyo metropolitan area on 5 August 2008 when the localized heavy rainfall caused damage in the Zoshigaya area (Ishihara 2012a). Their horizontal scale was only about 5.5 km on average. The mode of their lifetime was from 20 to 40 min. However, one third of the 179 cumulonimbus clouds produced a total rainfall amount more than 60 mm. It is still difficult to accurately predict when and where such cumulonimbus clouds develop and bring local-
ized heavy rainfall. Consequently, improving the temporal resolution of monitoring and the early detection of the cumulonimbus clouds that produce such localized heavy rainfall is of increasing concern, especially in urban areas.

Several previous studies have reported descending precipitation cores (observed using horizontal radar reflectivity ($Z_h$) and/or dual-polarization radar-derived liquid water content (LWC) data) in cumulonimbus clouds that caused localized heavy rainfall (Ishihara 2012b; Kim et al. 2012). Based on 3D radar data with an interval of 10 min, Ishihara (2012b) reported that a precipitation core (identified using $Z_h$ data) appeared at an altitude of 5 km, 20 min before the peak surface rainfall intensity from the cumulonimbus cloud that brought severe local rainfall to the Zoshigaya area. He concluded that the formation of the precipitation core aloft represented an effective nowcasting index for this cumulonimbus cloud. However, he also noted that the descent of a precipitation core could not be clearly observed using 3D radar data with an interval of 10 min. Kim et al. (2012) presented the structure of precipitation cores (identified using dual-polarization radar-derived LWC data) in a multi-cellular storm over the Zoshigaya area, which comprised 20 precipitation cells, using 3D dual-polarization radar data with an interval of 5 min. Of these, 17 precipitation cells were characterized by a single precipitation core (single-core cells). The other three precipitation cells consisted of several auxiliary precipitation cores (multi-core cells). They showed that the precipitation cores in both types of precipitation cell descended to the ground within 25 min of their first appearance. However, the relationship between descending cores and intense rainfall at the surface was not clearly demonstrated. It appears to be difficult to determine this relationship using radar data with an interval of 5 min and scattered rain gauges.

From the perspective of a feasibility study of the nowcasting of localized heavy rainfall based on precipitation core (PCO) data, as a first step, it would be beneficial to develop an automatic detection method for precipitation cores aloft using 3D $Z_h$ data. In addition, the behavior of automatically detected precipitation cores and their relationship with temporal variations in surface rainfall should be investigated using radar data with a high temporal resolution. Shimizu and Uyeda (2012) developed an algorithm for the identification and tracking of convective cells (AITCC) that uses the horizontal distribution of $Z_h$. The tracking performance of the AITCC was checked using data from numerical simulations with temporal resolutions of 1, 5, and 10 min, and was highest with a resolution of 1 min. This implies that a temporal resolution of 5 or 10 min is not sufficient to reveal the detailed behavior of convective cells, and even more so, of their internal precipitation cores.

On 18 August 2011, three cumulonimbus clouds developed and localized heavy rain fell over the northwest of Tokyo and the southwest of Saitama Prefecture, Japan. The National Research Institute for Earth Science and Disaster Prevention (NIED) of Japan was able to obtain the 3D structure of the cumulonimbus clouds at 2-min intervals using a research X-band dual-polarization radar (MP-X) located at Ebina in Kanagawa Prefecture (Fig. 1). Our aim here is to analyze the behavior of the precipitation cores in these cumulonimbus clouds using these 3D $Z_h$ data at 2-min intervals and to determine their relationship with temporal variations in surface rainfall intensity. A method of automatically detecting the precipitation cores in these cumulonimbus clouds using 3D $Z_h$ data was developed for the analysis. Some of the microphysical properties of precipitation cores are also...
discussed based on the dual-polarization radar information.

2. Radar observations and data

In this study, the MP-X radar at Ebina (see Table 1 in Park et al. (2005) for system details) was used to observe and analyze the 3D structure of the cumulonimbus clouds that developed on 18 August 2011. The observation range of the MP-X was 80 km, and it transmitted and received horizontally and vertically polarized signals simultaneously at 9.375 GHz. The pulse width used was 0.5 µs. The range and azimuthal intervals of the radar data were 100 m and 1°, respectively. The polarimetric variables obtained by the MP-X were the horizontal radar reflectivity (Z_h), differential reflectivity (Z_DR), correlation coefficient between horizontal and vertical polarization signals (ρ_hv), differential phase shift (φ_DP), Doppler velocity (V_D), and spectral width (W_S). The SIGMET RVP8 (SIGMET 2008) was used for signal processing of the MP-X data. For quality control, Z_h data that fulfilled the following three threshold criteria were allowed to pass the processing: a signal-plus-noise to noise ratio (LOG) > 0.75 dB, a clutter-to-signal ratio (CSR) < 18 dB, and a signal quality index (SQI) > 0.40.

In this study, our analysis was based mainly on the Z_h and Z_DR data. Rainfall attenuation in Z_h (Z_DR) was corrected using the relationship between the specific attenuation of Z_h (Z_DR) and the specific differential phase, K_{DP}, derived from Φ_DP (Maesaka et al. 2011). As described later, precipitation cores were identified in the 3D distribution of Z_h at grid points with a horizontal interval of 1 km and a vertical interval of 0.25 km. To create the grid-point radar data, a Cressman-type weighting function was used for data interpolation. The effective radius of influence of the weighting function was 1.5 km in the horizontal direction and 1.0 km vertically.

Rainfall intensity data from the Ministry of Land, Infrastructure, Transport and Tourism’s (MLIT) X-band polarimetric radar information network (XRAIN) were also used in the analysis to examine the horizontal and temporal distribution of surface rainfall. The XRAIN product of rainfall intensity was estimated from K_{DP} for intense rainfall as described in Maesaka et al. (2011). The composite map of rainfall intensities was provided every minute, with a horizontal data resolution of approximately 250 m (Maesaka et al. 2011).

Figure 1 shows the MP-X observation area and the horizontal distribution of maximum rainfall intensity for the period 1100–1320 Japan Standard Time (JST = UTC + 9 hours) based on rainfall intensity data from the XRAIN product. Localized heavy rain, with a maximum intensity greater than 90 mm h\(^{-1}\), was observed over the northwest of Tokyo and the southwest of Saitama Prefecture. The MP-X was operated with 12 plan position indicator (PPI) scans at elevation angles of 0.7°, 1.2°, 1.7°, 2.2°, 2.7°, 3.3°, 3.9°, 4.7°, 5.7°, 6.9°, 8.4°, and 10.4°, and two range–height indicator (RHI) scans at an azimuth angle of 93.3° every 5 min to conduct 3D surveillance for radar-echo appearance until the first detectable radar echo (the minimum detectable sensitivity of Z_h was 13.2 dBZ at a range of 60 km from the MP-X radar) appeared in this area at 1100 JST. Just after this first echo was detected, the surveillance mode was replaced with consecutive volume scans at 2-min intervals to observe the 3D structure of the radar echo at a higher temporal resolution. Each volume scan (every 2 min) comprised 17 sector plan position indicator (SPPI) scans with a 50° azimuthal width at elevation angles of 0.7°, 1.2°, 1.7°, 2.2°, 2.7°, 3.3°, 3.9°, 4.7°, 5.6°, 6.5°, 7.4°, 8.3°, 9.2°, 10.3°, 11.8°, 13.5°, and 15.6°. The azimuth directions of the sector scans were adjusted to include the whole radar-echo areas during the observation period.

3. Atmospheric conditions and the cumulonimbus clouds

The surface weather map for 0900 JST on 18 August 2011 (Fig. 2) shows a stationary front over the Tohoku District, approximately 350 km north of our observation area around Tokyo. Upper troughs were not present at 500 or 300 hPa around the observation
area (not shown), which indicates that the observation area was not influenced by a synoptic disturbance. At low levels, warm and moist air from the west–southwest was dominant around Tokyo.

Based on upper-air sounding data derived from a radiosonde launched at 0900 JST from Tateno (see Fig. 1 for location), the near-surface temperature was 30.8°C, and the 0°C level was at about 5.1 km above sea level (ASL). Precipitable water was 59.6 mm. Relative humidity below 5.5 km ASL was relatively high (60% to 95%), and it was low (<33%) above that level. The lifting condensation level (LCL) and the level of free convection (LFC) were 921 and 2381 m ASL, respectively. The convective available potential energy (CAPE) was 1311 J kg\(^{-1}\). The atmosphere around the observation area was convectively unstable enough for cumulonimbus clouds to develop.

Figure 3 shows the time–longitude cross section of maximum \(Z_h\) between 35.75°N and 36.0°N at 2 km ASL.

4. Detection and identification of precipitation cores

In this study, a precipitation core (PCO) was defined and detected on the basis of the 3D grid-point \(Z_h\) data \((dx = 1 \text{ km}, dy = 1 \text{ km}, dz = 0.25 \text{ km})\). A PCO was defined as a contiguous grid region with \(Z_h\) greater than or equal to an adaptive \(Z_h\) threshold, and which contained a single local maximum of \(Z_h\). The \(Z_h\) threshold was gradually decreased from the given maximum value (DBZmax) to the given minimum value (DBZmin) at a certain interval (DBZint). The detailed procedure for the detection of the PCOs is described below.

Figure 4 shows the PCO detection method schematically. DBZmax was set to 60 dBZ in this study, which was greater than the maximum value of \(Z_h\) (58.8 dBZ) associated with the storms considered here. The \(Z_h\) threshold was gradually decreased from DBZmax in steps of DBZint (1 dB here). When the threshold was reduced to \(Z_h = X1 \text{ dBZ}\), a region with \(Z_h \geq X1 \text{ dBZ}\) appeared. This region becomes a PCO candidate for PCO1 that contains one local maximum of \(Z_h\). The region of the PCO candidate for PCO1 expands three-dimensionally as the threshold decreases. When the threshold was reduced to \(Z_h = X3 \text{ dBZ}\), the region of the PCO candidate for PCO1 merges with that of an adjacent PCO candidate for PCO2. Once these two PCO candidates merge, the two regions with \(Z_h\) larger than or equal to the
previous step of the threshold, i.e., \((X3 + DBZint)\) dBZ, were detected as PCO1 and PCO2 (Fig. 4). If another PCO candidate appeared on the outer side of the closed region containing PCO1 and PCO2, and if they were connected by the threshold of \(X4\) dBZ, the region of the outer PCO candidate with \(Z_h \geq (X4 + DBZint)\) dBZ was detected as PCO3 (Fig. 4).

A PCO candidate whose volume just prior to connection with an adjacent closed region was less than 1 km\(^3\) was not regarded as a PCO (see the region in Fig. 4 with \(Z_h \geq X2\) dBZ containing a second maximum in PCO2). When the region excluded from the PCO candidates was embedded in the region of another adjacent PCO, it was considered to be part of the region of the adjacent PCO (see PCO2 in Fig. 4). The detection of PCOs continued until the threshold was reduced to \(Z_h = DBZmin\) (13 dBZ, near the minimum detectable \(Z_h\) value for the storms in this study). A PCO candidate that did not merge with any other PCO candidates until the threshold was reduced to \(Z_h = DBZmin\) was also detected as a PCO.

The above procedure was adopted for the 3D grid-point datasets (with an interval of 2 min) of the three storms that covered the period between 1100 and 1320 JST on 18 August. A total of 102 PCO regions were automatically detected above 2 km ASL, which is around the level free from ground-clutter contamination in the region of the cumulonimbus clouds. After detection of these PCO regions, the 3D tracking of the PCO regions was conducted by subjective analysis using the PCO information detected every 2 min. The rules applied to the subjective tracking were as follows: 1) continual change in the location and shape of the PCO region; and 2) gradual temporal variation in the maximum \(Z_h\) values of the PCO region. As a result, 15 PCOs were identified based on the 2-min interval PCO dataset acquired over the combined lifespan of storms A, B, and C. The number of PCOs in storms A, B, and C was 1, 7, and 7, respectively, and they were named in order of appearance in each storm.

The volume of the above 102 PCO regions ranged from 1.2 to 1302.6 km\(^3\). Of these PCO regions, 50 % were smaller than 50 km\(^3\), and 60.8 % were smaller than 100 km\(^3\). The PCO volume relates not only to the maximum \(Z_h\) value within the PCO region, but also to the distance between coexisting PCOs in the storms, because the definition and detection procedures for the PCOs used adaptive threshold values.

Figure 5 shows an example of the PCO identification procedure. Two PCOs (B4 and B6) were identified at 1200 and 1202 JST, respectively, in storm B. The PCOs were located around local \(Z_h\) maxima, which indicates that the automatic PCO detection algorithm (described above) functioned well in the present case. Both PCOs, B4 and B6, maintained their integrity from 1200 to 1202 JST with only slight changes in their maximum \(Z_h\) values, locations, and volumes. These results show that it was indeed possible to track these PCOs by subjective analysis using datasets with an interval of 2 min. In the following section, the behavior of the PCOs is presented in more detail.

5. Precipitation core behavior and surface rainfall intensity

5.1 Vertical behavior and temporal properties of precipitation cores

Figure 6 shows the time series of the height of the PCOs (HPCO) in storms A, B, and C. Here, the HPCO was defined as the height of maximum \(Z_h\) in a PCO region. Most of the PCOs descended towards the ground after their appearance aloft. Some PCOs, such as B3, C2, and C3, showed a relatively monotonous descent, while others, such as B2, C1, and C5, showed vertical oscillations during their descent. The mean falling velocity for each PCO ranged from 0 to 10.4 m s\(^{-1}\). The average mean falling velocity for the 14 PCOs (excluding C7, which was observed only once) was 4.3 m s\(^{-1}\).

Figure 7 summarizes the appearance height of the PCOs (AHPCO), which was defined as the first detected HPCO for each PCO, and shows that 40 % appeared between 5 and 5.75 km ASL, although the AHPCOs were widely distributed between 2.25 and 7.25 km ASL. The average AHPCO was 5.25 km ASL, which is slightly above the ambient 0°C level (5.1 km ASL).

Figure 8 shows a scatter plot of the AHPCO data and the duration of the PCOs (DPCO). The DPCO ranged from 2 to 28 min, with an average of 13.9 min. The DPCO is roughly proportional to the AHPCO. The correlation coefficient between the AHPCO and the DPCO values was 0.70.

5.2 Correspondence relationship between precipitation core behavior and surface rainfall intensity

To examine the correspondence relationship between the PCO behavior and the temporal change in surface rainfall intensity for storms A, B, and C, Figs. 9–11 show temporal changes in the vertical profile of maximum \(Z_h\), together with the HPCO and maximum surface rainfall intensity for each storm. Here, surface rainfall intensity data at 2-min intervals were used in
Fig. 5. Horizontal distributions at 3.5 km ASL (left) and vertical cross-sections (right) of $Z_h$ at 1200 (a and b) and 1202 (c and d) JST. The locations of the vertical sections are indicated by dotted lines in the horizontal sections. The areas covered by the precipitation cores (B4 and B6) are hatched.

Fig. 6. Time series of height of precipitation core (HPCO) in storms A, B, and C. The HPCO for A1 at 1102 JST was linearly interpolated using data at 1100 and 1104 JST under the assumption of continuance of A1 during that period.
the analysis because XRAIN's product for a particular time is created using observational data from the previous two minutes. For the three storms, 12 temporal peaks in maximum surface rainfall intensity with values greater than 10 mm h\(^{-1}\) were recorded, and they are indicated by arrows in Figs 9b, 10b, and 11b. With regard to 10 of the 12 temporal peaks in maximum surface rainfall intensity, the PCOs that descended from above were observed between 2.0 and 3.0 km ASL within 2 min (before or after) of the temporal peaks, as described in more detail below.

For storm A, the surface rainfall intensity has a single peak maximum value of 21.1 mm h\(^{-1}\) at 1112 JST (Fig. 9b). This peak corresponds to the descent of A1 (Fig. 9a). A1 was first detected at 1100 JST, 12 min before the corresponding temporal peak in rainfall intensity. For storm B, five temporal peaks in rainfall intensity greater than 10 mm h\(^{-1}\) were observed at 1124, 1132, 1138, 1154, and 1202 JST (Fig. 10b). The four peaks at 1132, 1138, 1154, and 1202 JST correspond to the descents of B2, B3, B5, and B4, respectively (Fig. 10a). The PCOs B2, B3, B5, and B4 were first detected at 22, 6, 6, and 24 min, respectively, before their corresponding temporal peak in rainfall intensity. For storm C, six temporal peaks in rainfall intensity greater than 10 mm h\(^{-1}\) were observed at 1220, 1232, 1238, 1258, and 1304 JST (Fig. 11b). The peaks at 1220, 1232, 1238, 1258, and 1304 JST correspond to the descents of C1, C2, C3, C5, and C6, respectively (Fig. 11a). The PCOs C1, C2, C3, C5, and C6 were first detected at 18, 8, 10, 18, and 10 min, respectively, before their corresponding temporal peak in rainfall intensity. In summary, these 10 PCOs appeared, on average, 13.4 min before the corresponding temporal peak in surface rainfall intensity.

In contrast to the above 10 temporal rainfall peaks, the descent of a PCO was not observed with the two temporal peaks in rainfall intensity at 1124 JST in storm B, and at 1248 JST in storm C. B2 was suspended aloft at 1124 JST, but a region of strong radar echo with Z\(_b\) ≥ 50 dBZ extended downwards, and this corresponded to the peak in rainfall intensity at that time. The peak at 1248 JST in storm C also corresponded to the downward extension of a strong radar echo region of C5. Both B2 and C5 moved up at around, or just after, the time when their strong echo regions extended downwards. It is reasonable to consider that the updrafts of these PCOs were re-intensified at upper levels, or that the new updrafts developed in the immediate vicinity of the existing updrafts of B2 and C5, although it is difficult to confirm the mechanism that generates such temporal peaks with no associated descending PCOs using only the present observational dataset.

For the onset of intense rainfall in each storm, the maximum rainfall intensity in storms A, B, and C exceeded 10 mm h\(^{-1}\) at 1112, 1114, and 1212 JST, respectively (Figs 9b, 10b, 11b, respectively). PCOs were first detected in each storm at 1100 (A1), 1104 (B1), and 1202 (C1) JST (Figs 9a, 10a, 11a, respectively), which was at 12, 10, and 10 min, respectively,
Fig. 9. Temporal change in (a) the vertical profile of maximum $Z_h$ and the height of the precipitation core, and (b) the maximum rainfall intensity for storm A. The height of the precipitation core is indicated by open circles and connected by the line in (a) for each precipitation core.

Fig. 10. As in Fig. 9, but for storm B.
before the onset of rainfall heavier than 10 mm h\(^{-1}\).

In summary, 10 of the 12 temporal peaks in maximum rainfall intensity greater than 10 mm h\(^{-1}\) recorded from storms A, B, and C were observed in association with the descent of PCOs. The 10 PCOs appeared, on average, 13.4 min before the corresponding temporal peak in surface rainfall intensity. In each storm, the first PCO was detected 10–12 min before the rainfall heavier than 10 mm h\(^{-1}\) was recorded. These results show that the behavior of PCOs is closely related to the onset of heavy surface rainfall and the subsequent fluctuations in its intensity. This indicates the possible benefits of detecting PCOs aloft for the short-term forecasting of localized convective rainfall.

5.3 Formation of precipitation cores

As noted above, the behavior of the PCOs was closely related to surface rainfall in the storms investigated here. In this section, the PCOs’ formation is presented with the aim of characterizing PCO structure.

Figure 12 shows the frequency distribution of the horizontal distance between the genesis locations of successive PCOs and the preceding adjacent PCOs in each storm. The horizontal distance between two PCOs was determined from the horizontal component of the 3D locations of maximum \(Z_h\) in the PCOs.

Fig. 11. As in Fig. 9, but for storm C.

Fig. 12. Frequency distribution of the horizontal distance between genesis locations of successive precipitation cores from the preceding adjacent precipitation core.
The genesis locations of 12 PCOs (excluding the first PCO in each storm; i.e., A1, B1, and C1) are included in Fig. 12. The horizontal distance ranged from 1.0 to 5.8 km, with an average of 3.2 km.

A typical example of PCO formation has already been presented in Fig. 5, which shows the genesis of B6. The PCO B6 appeared at a horizontal distance of 4.1 km from B4 at 1200 JST, and was horizontally separate from B4 at every altitude (Fig. 5). This horizontal separation of a new PCO from its preceding adjacent PCO was commonly seen in the PCOs in the 2- to 5-km categories in Fig. 12. When the cell-tracking algorithm AITCC was applied to these radar echo distributions, the echo regions of each PCO were identified as the individual convective cells that have been recognized as a fundamental element of convective precipitation by many previous studies (e.g., Chisholm and Renick 1972; Shimizu and Uyeda 2012).

Three successive PCOs (B2, B3, and B4) in the 1-km category followed a different formation process to the PCOs in the 2- to 5-km categories. To illustrate the formation characteristics of B4 as an example of the 1-km category, vertical cross-sections of $Z_h$ through B4 and the preexisting B3 are shown in Fig. 13a–c. At 1138 JST, B4 appeared at 6.75 km ASL and the height of the preceding PCO (B3) was 4.25 km ASL (Figs. 6, 13a). The horizontal distance separating B4 and B3 was only 1.0 km, which was less than the horizontal extent of B3 and B4 (Fig. 13). As shown in Fig. 13a, B4 was located immediately above B3, with horizontal overlapping of their regions. This horizontal overlapping and vertical separation continued while these two PCOs coexisted until 1142 JST (Fig. 13a, b, c). This positional relationship (i.e., a new PCO forming on the upper part of the preceding PCO with horizontal overlapping) was observed for all three PCOs in the 1-km category.

Figures 13d–f shows the horizontal distribution of $Z_h$ at 5 km ASL, together with the PCO regions detected (B3 and B4), at 1138, 1140, and 1142 JST. A single convective echo region, accompanied by $Z_h$ greater than 50 dBZ, was maintained in the field of $Z_h$ at 5 km ASL during this period. When the AITCC was applied to this echo distribution, a single persistent convective cell was indicated during this period.
period. However, a PCO associated with the strong echo at 5 km was replaced from B3 to B4 at 1140 JST (Fig. 13e). The value of maximum \( Z_h \) at 5 km ASL increased from 54 dBZ at 1140 JST to 57 dBZ at 1142 JST, which corresponded to the descent of the intensifying B4 to this level at 1142 JST (Fig. 13f). This re-intensifying feature in the horizontal distribution of radar echoes was commonly seen in association with the descent of the three successive PCOs (B2–B4). Consequently, the descent of these PCOs in the 1-km category, detected using 3D radar data, accounted for the re-intensification of the convective cells conventionally identified using the 2D radar data, and the temporal rainfall peaks observed here.

6. Discussion

As shown in Section 5, the behavior of the PCOs was closely related to temporal changes in surface rainfall in the storms investigated here. To consider the physical processes controlling the behavior of the PCOs aloft, the microphysical properties and the factors relating to the vertical motion of PCOs are discussed in this section.

Temporal changes in the vertical profile of ZDR, and the height of the PCO, for storms A, B, and C are presented in Fig. 14 to highlight the microphysical properties of the PCOs. ZDR values in the vertical profile were extracted from the location of maximum \( Z_h \) at each altitude. ZDR shows relatively little change below 6 km ASL and decreases with altitude above that level around the core regions in the three storms. A marked increase in ZDR was not seen around the ambient 0°C level (5.1 km ASL), which also supports that the PCOs in these storms were of convective precipitation origin. The isolines of 1 dB fluctuated up and down between 6 and 8 km ASL during the active periods of these storms. We suggest that periodic convective updrafts existed around these altitudes, and that some raindrops were carried above the ambient 0°C level. As shown in Fig. 7, most PCOs appeared at 5 to 7 km ASL and showed positive ZDR values. This indicates that raindrops contributed to the appearance of detectable PCOs even around and above the ambient 0°C level in the present case.

The average of available ZDR values at the location of maximum \( Z_h \) in the PCO regions at all altitudes was 2.79 dB. The median volume diameter (\( D_0 \)), calculated on the assumption that all precipitation particles were raindrops in the volume of radar observation, was 2.94 mm according to the equation in Kim et al. (2010): \( D_0 = 0.77Z_{\text{DR}} + 0.79 \). The average of \( Z_{\text{DR}} \) at the location of the PCO’s maximum \( Z_h \) that was located below the ambient 0°C level was 2.84 dB, which means \( D_0 \) was 2.97 mm.

As shown in Figs. 6, 9–11, and 14, the PCOs in this study generally descended towards the ground after their appearance aloft, with the average of the mean falling velocities being 4.3 m s\(^{-1}\). Two major factors that may control the vertical motion of PCOs are the terminal falling velocity of dominant hydrometeors and the vertical air velocity. The terminal falling velocities for raindrops (\( w_T \)) calculated from the \( D_0 \) values presented above (2.94 and 2.97 mm) are 7.88 and 7.92 m s\(^{-1}\) based on the empirical equation linking \( w_T \) (m s\(^{-1}\)) with the diameter of raindrops (\( D; \text{mm} \)) in Atlas et al. (1973): \( w_T = 9.65 – 10.3e^{-0.6D} \). These calculated \( w_T \) values are larger than the mean falling velocities of PCOs, which suggests that the raindrops around the center of the PCOs were, on average, suspended by upward air motion during their descending periods.

As stated in Section 5.1, the vertical behavior of the PCOs varied in each cloud. Several other factors, such as the formation, growth, and disappearance of hydrometeors in and around the PCOs, as well as the variation in \( w_T \) of hydrometeors and vertical air velocity, are also thought to be significant for the vertical motion of PCOs. However, it is difficult to ascribe the vertical motion of the PCOs to these separate factors here. One of the reasons for this is that the vertical air velocity around the PCOs could not be estimated accurately because the storms appeared in a single radar observation area. In the future, detailed analysis of 3D structures in the raindrop size distribution and the wind velocity derived from multiple dual-polarization radars should increase our understanding of microphysical processes, such as the formation, growth, and evaporation of raindrops in and around the PCOs. This would lead to a further understanding of the physical factors that control the behavior of PCOs aloft and also their relationship with temporal changes in surface rainfall.

In addition, precipitation type classification based on the polarimetric variables has the potential to extend the result of this study to a more effective technique in monitoring and nowcasting of localized heavy rainfall. For example, polarimetric melting layer signatures around the ambient 0°C level can provide the information regarding the stages of evolution of convective rainfalls (Bringi et al. 1986; Teshiba et al. 2009) and characterize the rainfall types (Shusse et al. 2011), which are closely related to the convective rainfall processes. The active utilization
of these dual-polarization radar information together with the PCO detection technique in this study should be promoted to develop a practical system for the monitoring and nowcasting of localized heavy rainfall.

7. Summary

This paper analyzes the behavior of the precipitation cores (PCOs) in three cumulonimbus clouds that caused localized heavy rainfall on 18 August 2011 in the Kanto region, Japan, and their relationship with temporal variations in surface rainfall intensity. The 3D structure of the cumulonimbus clouds was observed at 2-min intervals using a research X-band dual-polarization radar. A PCO was defined as a 3D contiguous region that contained one local maximum of $Z_h$, and the PCOs were automatically detected using adaptive thresholds. The $Z_h$ threshold was gradually decreased from the maximum value of 60 dBZ to the minimum value of 13 dBZ at an interval of 1 dB.
an average mean falling velocity of 4.3 m s$^{-1}$.

We investigated the behavior of the 15 PCOs, and their relationship with temporal variations in surface rainfall intensity. The PCOs generally descended towards the ground after their appearance aloft, with an average mean falling velocity of 4.3 m s$^{-1}$. The average appearance height of the PCOs (5.25 km ASL) was slightly above the ambient 0°C level (5.1 km ASL). The duration of each PCO was roughly proportional to its appearance height.

Of the 12 temporal peaks in maximum surface rainfall intensity (>10 mm h$^{-1}$) recorded from the three cumulonimbus clouds, 10 were associated with the descent of PCOs. These 10 PCOs were detected aloft, on average, 13.4 min before the corresponding temporal peaks in surface rainfall intensity. As for the onset of intense rainfall associated with each cumulonimbus cloud, the first PCO in each cumulonimbus cloud was detected 10–12 min before the rainfall heavier than 10 mm h$^{-1}$ was recorded. In addition, the descent of some PCOs appeared to account for the re-intensification of convective cells that were conventionally identified using 2D radar data and for the temporal rainfall peaks. We conclude that the behavior of PCOs is closely related to the onset of heavy surface rainfall and subsequent fluctuations in its intensity.

Polarimetric signatures supported the conclusion that raindrops contributed to the appearance of PCOs studied here. Future studies of the physical processes associated with PCOs, including their microphysical properties, in different precipitation systems and different climate regimes will help characterize the relationship between the behavior of PCOs and surface rainfall in various atmospheric environments.

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