Modification of raindrop size distribution due to seeder–feeder interactions between stratiform precipitation and shallow convection observed by X-band polarimetric radar and optical disdrometer

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Abstract
The seeder–feeder interactions (SFIs), where raindrops from upper clouds grow by accreting cloud droplets in the lower clouds, have been extensively studied. However, there are few studies on the modification of raindrop size distribution (DSD) through this process. In the present study, rainfall from the landfalling rainbands of a typhoon was observed using an optical disdrometer and an X-band polarimetric radar. Rainfall was classified into the following three types based on the DSD characteristics at the surface and the existence of a $\rho_{HV}$ minimum in the upper layer: convective rainfall accompanied by a melting layer (type SF), convective rainfall without a melting layer (type C), and stratiform rainfall with a melting layer (type S). Type SF rainfall was regarded as having undergone SFIs between stratiform precipitation and shallow convection. The DSD for SF type rainfall was characterized by more small- to medium-sized raindrops and a larger normalized intercept parameter than rainfall types C and S. An analysis using vertical profiles of radar-derived DSD parameters for type SF rainfall suggested that the median-volume diameter of raindrops increased by accreting cloud droplets in the lower clouds, and that small- to medium-sized raindrops were produced by a warm rain process and breakup of raindrops.

KEYWORDS
disdrometer, polarimetric radar, raindrop size distribution, seeder–feeder

1 | INTRODUCTION
In general, the amount of precipitation in mountains is larger than that over plains when synoptic disturbances such as tropical and extratropical cyclones pass over land areas. This phenomenon is explained by seeder–feeder interactions (SFIs), where precipitation from an upper-level precipitating cloud (seeder) falls through a lower-level orographic stratus cloud (feeder) capping a hill or small mountain (https://glossary.ametsoc.org/wiki/Seeder-feeder). Bergeron (1965) analyzed raingauge data and proposed SFIs as a hypothesis. This hypothesis has...
been supported by radar observations (Hill et al., 1981; Harimaya and Tobizuka, 1988; Iwanami et al., 1988; Misumi, 1996; Smith et al., 2009; DeHart and Houze Jr., 2017) and numerical simulations (Gocho, 1978; Carruthers and Choularton, 1983; Richard et al., 1987; Kikuchi et al., 1988; Lee et al., 2018).

The concept of SFIs has been extended to precipitation over plains. Herzegh and Hobbs (1980) conducted aircraft observations of warm frontal rainbands and found that precipitation from generation cells (seeders) was strengthened by accreting cloud droplets in the low-level warm clouds (feeders). Fabry et al. (1993) reported a case study in which stratiform echoes washed out shallow convective echoes. Shusse et al. (2009) and Oue et al. (2010) studied convective echoes embedded in stratiform precipitation using polarimetric radars but did not mention SFIs.

As explained above, precipitation enhancement through SFIs has been studied in detail using both observations and numerical simulations. However, there is a lack of knowledge about how SFIs modify raindrop size distributions (DSDs). For example, Arulraj and Barros (2019) pointed out that the estimation of DSDs over mountains is the main source of errors in the global precipitation measurement/dual-frequency precipitation radar. For an accurate precipitation estimation based on remote sensing data, an appropriate modeling of DSDs is essential.

According to classical concepts of SFIs, raindrops become larger in size by collecting cloud droplets. Herzegh and Hobbs (1980) showed that the number density of raindrops smaller than 0.13 mm decreased whereas that of raindrops larger than this size increased through SFIs. Wilson and Barros (2014) observed that a large number of small raindrops (<0.2 mm) were supplied by low-level fog and clouds, and that small to medium-sized raindrops (<2 mm) increased through the collision coalescence of raindrops in the precipitation over the South Appalachian Mountains. Zwiebel et al. (2016) reported that the mass-weighted mean diameter of raindrops was smaller in the mountains than in plains. In the Olympic Mountains Experiment (OLYMPEX), extensive studies on precipitation over the mountains were conducted using disdrometers (Zagrodnik et al., 2019), ground-based and airborne radars (McMurdie et al., 2018; Zagrodnik et al., 2019), and numerical simulations (Morales et al., 2018; Naeger et al., 2020). In these studies, precipitation was strongly enhanced when the warm sector of extratropical cyclones passed over the mountains, and small- to medium-sized raindrops increased while the concentrations of large raindrops were varied. Although observations of DSDs are still limited, the findings of these studies suggest that not only large raindrops but also small drops increase via SFIs.

Typhoon Hagibis, which made landfall in Japan in October 2019, caused heavy rainfall over eastern Japan. Analysis of environmental conditions around Hagibis revealed that a moist absolute unstable layer (MAUL) was formed in the lower atmosphere (Takemi and Unuma, 2020), which provided favorable conditions for shallow convection acting as feeder clouds to develop. In the present study, the DSDs of rainfall from Hagibis were observed using an optical disdrometer and an X-band polarimetric radar to study the effect of SFIs on DSDs. The DSDs of landfalling tropical cyclones have been obtained using disdrometers (Chen et al., 2019; Janapati et al., 2020) and polarimetric radars (Chang et al., 2009; Wang et al., 2016), but the present study focuses on the effect of SFIs. Japanese Standard Time (JST; UTC + 9 h) is used in this paper.

2 | OBSERVATIONS

2.1 | Instruments

An X-band polarimetric radar, installed in Ebina City (35.40°N, 139.30°E) by the National Research Institute for Earth Science and Disaster Resilience, was used in the present study (EBN radar; Table S1). The radar observes radar reflectivity at horizontal polarization ($Z_H$), Doppler velocity, Doppler spectrum width, differential radar reflectivity ($Z_{DR}$), specific differential phase ($K_{DP}$), and copolar correlation coefficient ($\rho_{HV}$) by transmitting a 1.15° beam at pulse widths of 32 and 1.0 μs within a maximum observation range of 80 km. A set of 13 plan position indicator scans and one range height indicator (RHI) scan at an azimuthal angle of 47° (along the red line in Figure 1) were repeated every 5 min. The methods used for the quality control of radar parameters and attenuation corrections for $Z_H$ and $Z_{DR}$ are described in Maesaka et al. (2011).

For a measurement of DSDs from the ground, a laser precipitation monitor (LPM; Adolf Thies GmbH & Co. KG) was installed on the rooftop of a six-story building in Tokyo (point A in Figure 1; altitude: approximately 20 m). The LPM site is located at a distance of 51.3 km and an azimuth angle of 47° from the EBN radar site. The LPM forms a beam that is 20 mm wide and 228 mm long, and measures the size and velocity of particles falling across the beam at 1-min time intervals. The particle and the velocity classes are shown in Table S2. According to Friedrich et al. (2013), optical disdrometers give an unrealistically small fall velocity under strong wind conditions because the raindrops cross the sampling area at
large incident angles and water splashes the equipment. To remove such false data, only raindrops within ±60% of their empirical terminal velocity (Atlas et al., 1973) were analyzed. Observations by the EBN radar and the LPM were conducted from the onset to the end of the precipitation from Typhoon Hagibis (0000–2359 JST on October 12, 2019).

2.2 Derivation of DSD parameters

In general, a DSD can be approximated by a gamma distribution as (Ulbrich, 1983):

\[ N(D) = N_0 D^\mu \exp(-D^\lambda), \tag{1} \]

where \( N(D) \) represents the number density of raindrops with diameters between \( D \) and \( D + dD \). Here, \( N_0 \), \( \mu \), and \( \lambda \) are the intercept, the shape, and the slope parameters, respectively. The \( n \)th moments of \( N(D) \) are defined as follows:

\[ M_n = \int_0^\infty D^n N(D) dD. \tag{2} \]

A mass-weighted mean diameter, \( D_m \), and rainwater content, \( W_r \), can be respectively written using the moments of \( N(D) \) as

\[ D_m = \frac{M_4}{M_3}, \tag{3} \]

and

\[ W_r = \frac{\pi}{6} \rho_w M_3, \tag{4} \]

where \( \rho_w \) is the density of water and \( \pi \) is the geometric constant. The median volume diameter of raindrops, \( D_0 \), can be derived from Equation (1) as follows:

\[ D_0 = \frac{3.67 + \mu}{\lambda}. \tag{5} \]

The parameter \( N_0 \) in Equation (1) highly fluctuates depending on \( \mu \). Testud et al. (2001) introduced the following normalized intercept parameter that is independent of \( \mu \):

\[ N_{wm} = \frac{4^4}{\pi \rho_w} \left( \frac{W_r}{D_m^4} \right)^\frac{1}{4}. \tag{6} \]

For polarimetric radars, \( D_0 \) is sometimes used instead of \( D_m \) for the normalized intercept parameter (e.g., Anagnostou et al., 2008):

\[ N_{w0} = \frac{3.67^4}{\pi \rho_w} \left( \frac{W_r}{D_0^4} \right)^\frac{1}{4}. \tag{7} \]

Here, \( N_{wm} \) and \( N_{w0} \) are identical when \( \mu = 0 \), but slightly different when \( \mu \neq 0 \).

We used the method of Chen et al. (2016) for calculating \( N(D) \) from the LPM data. After \( N(D) \) was obtained, \( D_m, W_r \), and \( N_{wm} \) were calculated using Equations (3), (4), and (6), respectively. For the derivation of \( D_0 \) in Equation (5), the parameters \( \lambda \) and \( \mu \) were estimated by fitting \( N(D) \) to Equation (1) using the method of Kozu and Nakamura (1991). For EBN radar data, the rainfall intensity, \( R \) (mm·h\(^{-1}\)), was estimated using \( K_{DP} \) when \( Z_{HH} > 30 \text{ dBZ} \) using the method of Maesaka et al. (2011) as:
For the detection of a melting layer, a dataset of $\rho_{HV}$ was created by averaging in a 2-km distance over the LPM site and every 0.25 km in the vertical direction using RHI scan data obtained at 5-min intervals. Then, the dataset was interpolated into 1-min intervals. The interpolation was not conducted when the data interval was more than 5 min due to missing data. If a layer with $\Delta \rho_{HV} \geq 0.02$ was found between the 3 and 6 km levels (within 1.5 km above and below the 0°C level observed at Tateno), it was regarded as a melting layer. For the separation of convective and stratiform rainfall, $N_{wm}$ and $D_0$ obtained at 1-min intervals by LPM were used. Using this method, rainfall was classified into three types: convective rainfall accompanied by a melting layer (type SF), convective rainfall without a melting layer (type C), and stratiform rainfall accompanied by a melting layer (type S). As will be discussed in Section 4, type SF was regarded as precipitation associated with SFI.

3 | RESULTS

3.1 | Precipitation from Typhoon Hagibis

Typhoon Hagibis formed over the western Pacific Ocean at 0300 JST on October 6, 2019, and made landfall in Japan at around 1900 JST on 12 October before transitioning into an extratropical cyclone on 13 October. A strong southeasterly wind from the typhoon supplied warm air over the land (Figure S1), and heavy rainfall was observed in eastern Japan before and after landfall. Distances from the typhoon center to the LPM site were 600, 468, 267, and 22 km at 0300, 0900, 1500, and 2100 JST on 12 October, respectively, and precipitation in the observation area was mainly attributed to rainbands associated with the typhoon. According to the upper sounding at Tateno at 0900 JST on October 12, 2019 (Figure S2), a MAUL formed between 750 and 1,600 m. Takemi and Unuma (2020) showed that the MAUL spread over the northwestern quadrant of Typhoon Hagibis. The atmosphere was conditionally unstable from 1,600 to 3,800 m, but was stable above this level. A convective available potential energy of 167.0 J kg$^{-1}$ and the lifted condensation level of 981.58 hPa were calculated by the University of Wyoming (http://weather.uwyo.edu/upperair/sounding.html).

Figure 2 shows the distribution of the precipitation intensity derived from the Japan Meteorological Agency (JMA) radar composite map (left panels) and the RHIs of $Z_H$ derived from the EBN radar (right panels). At 0230 JST, the observation area was covered with
widespread rainfall, and bright band of $Z_H$ was observed at around the 4.5-km level in the RHI at 0229 JST. At this time, rainfall at the LPM site was classified as type S. At 0920 JST, widespread echoes greater than 20 mm·h$^{-1}$ were distributed in the observation range of the EBN radar. In the RHI at 0919 JST, a bright band was observed at around the 5-km level, while cellular echoes stronger than 40 dBZ appeared below the 3-km level. The rainfall type was SF at the LPM site. At 1930 JST, strong precipitation associated with the eyewall of the typhoon was observed near the EBN radar site, and type C rainfall was observed at the LPM site at 1929 JST.
3.2 DSD obtained from LPM data

Figure 3a,b shows the time-height cross-section of $\rho_{HV}$ and $Z_{HI}$ averaged in a 2-km horizontal length over the LPM site. A band of small $\rho_{HV}$ is observed in the levels between 4 and 6 km. Figure 3c shows the time variation in $N(D)$ and $D_m$ observed by the LPM. The distribution of $N(D)$ tends to be broad when rainfall intensity is strong (Figure 3e). $D_m$ fluctuates almost along the contour line of $N(D) = 10^2$, and sometimes shows large values (e.g., between 1940 and 1950 JST).

FIGURE 3 Time-height cross section of (a) $\rho_{HV}$ and (b) $Z_{HI}$ above the LPM site (point A in Figure 1) observed by the EBN radar. Time variations of (c) $N(D)$ (shadings) and $D_m$ (solid line), (d) radar reflectivity, and (e) $R$ (solid line) observed by LPM. Colored dots in (e) represent three rainfall types.
Since radar reflectivity is the sixth moment of $N(D)$, it varies together with the number concentration of large raindrops (Figure 3d). Rainfall intensity and rainfall types are shown in Figure 3e. The total number of S, C, and LF rainfall types estimated by the LPM data was 348, 108, and 33 min, respectively. The rest of the data were not classified, either because precipitation was too weak or because radar data were missing due to attenuation. The rainfall intensities (mean ± standard deviation) for rainfall types S, C, and SF were 4.7 ± 2.8, 13.3 ± 7.8, and 17.8 ± 8.0 mm h$^{-1}$, respectively.

The mean $N(D)$ obtained for the rainfall types (Figure 4a) indicates that the number density for all raindrop sizes was larger in convective rainfall (types C and SF) than in stratiform rainfall (type S). In type SF rainfall, the number concentration of raindrops smaller than 2.8 mm in diameter is larger than that in type C rainfall, while the number concentration of raindrops larger than this size was slightly smaller. The shapes of the $N(D)$ curves are similar when $R$ is large, even though they were observed at different times (Figure 4b). These curves are close to the equilibrium DSD (broken line), where the rates of collision coalescence and breakup of raindrops are balanced (McFarquhar, 2004). Figure 4c compares the $D_m$ versus $N_{wm}$ plots obtained in the present study to those of various climatological regimes summarized by Bringi et al. (2003). The mean value for type S is on the empirical line for stratiform precipitation (broken line). The mean values for types SF and C are near the area of maritime-like and continental-like convective rainfall, respectively. The mean slope parameter ($\lambda$) and the shape parameter ($\mu$) are the smallest for type SF rainfall (Figure 4d).

The relationship between radar reflectivity and rainfall intensity obtained from the LPM data was $Z = 349.3 R^{1.38}$. It is much different from that obtained by Chen et al. (2019), who suggested that $Z = 189 R^{1.38}$ for tropical cyclones around Tokyo. Our $Z$-R relationship gives a larger $Z$ than that of Chen et al. (2019) for a given $R$. This difference is likely due to the season. Chen et al. (2019) studied typhoons in summer (July–August) when Japan was covered with a subtropical airmass, while our observations were conducted in October when Japan was under the influence of frontal zones.

**FIGURE 4**  (a) Mean DSDs for each rainfall type. Values in parentheses in the legend indicate the sample numbers. (b) DSDs for the three strongest precipitation events of type SF rainfall. For comparison, equilibrium DSD (E-D) at 54 mm h$^{-1}$ calculated by McFarquhar (2004) is shown by the dashed line. (c) Plots of the mean values (dots) and standard deviations (error bars) of $D_m$ and $N_{wm}$ for each rainfall type. The dashed line indicates stratiform rain and squares are clusters of maritime-like and continental-like convective rainfall given by Bringi et al. (2003). (d) Mean values (dots) and standard deviations (error bars) of $\lambda$ and $\mu$. 

3.3 DSD parameters estimated from the EBN radar data

In order to examine radar data above the LPM site, a dataset was created by averaging the data in a 2-km horizontal direction over the LPM site and at 0.5-km intervals in the vertical direction using the RHI scan data. The data below the 1.2-km level were not used because they were contaminated by ground clutter. The created dataset was then averaged according to rainfall type. Types S, C, and SF rainfall were recorded 53, 7, and 34 times during the observation period, respectively. Figure 5 shows the vertical profiles of the mean and the standard deviations of the radar and DSD parameters for each rainfall type. Gray bands in the panels indicate the range of fluctuation in the 0°C levels (4.4–5.5 km) during the observation period.

FIGURE 5 Vertical profiles of mean values (dots) and standard deviations (error bars) of (a) $Z_H$, (b) $Z_{DR}$, (c) $K_{DP}$, (d) $\rho_{HV}$, (e) $D_0$, (f) $\log_{10}(N_0)$, and (g) $W_r$ for the three rainfall types. Gray bands indicate the range in the fluctuation of the 0°C levels during the observation period.
period; these levels were estimated using the JMA Meso-scale Model.

In type S rainfall, the mean $Z_{H}$ shows characteristic variations depending on altitude (Figure 5a). At altitudes above the 0°C level, $Z_{H}$ is relatively small, reflecting the existence of ice particles with a low dielectric constant. The large $K_{DP}$ and positive $Z_{DR}$ above the 0°C level suggest that the shape of ice particles was oblrate. Above the 0°C level, the presence of positive peaks in $Z_{H}$ and $Z_{DR}$ and a negative peak in $P_{HV}$ suggested the existence of a melting layer. Below the 0°C level, $Z_{H}$ and $W_{r}$ increase slightly toward the lower layer, while $D_{0}$ and $N_{w0}$ do not change markedly.

In type C rainfall, the echo-top height was approximately at the 7.5-km level, which was lower than the other two rainfall types. Above the 0°C level, the mean $Z_{H}$ and $Z_{DR}$ were relatively small. These findings suggest that the amount of ice particles was lower in this rainfall type than in type S and SF, and the shape of the particles was more spherical. The $P_{HV}$ minimum was above the 0°C level, likely indicating that a mixed-phase layer was created by transportation of super-cooled drops by updrafts. Below the 0°C level, $N_{w0}$ in type C rainfall was larger, but $D_{0}$ was smaller than in type S rainfall between the 1.5 and 4 km levels. This suggests that the number concentration of raindrops was larger in type C rainfall but the mean raindrop size was smaller than it was in type S rainfall.

In type SF rainfall, the variations in the radar parameters above the 0°C level were similar to those in type S rainfall, except that $Z_{H}$ and $K_{DP}$ were slightly larger. However, below the 0°C level the variations of radar parameters of type SF rainfall differed from those observed in type S rainfall. The mean values of $Z_{H}$, $Z_{DR}$, and $K_{DP}$ increased toward the lower layer, and $D_{0}$ and $W_{r}$ were larger than the other rainfall types. The mean $N_{w0}$ was close to that observed in type S rainfall at around the 4-km level, but it approached that observed in type C rainfall around the 1.5-km level.

4 | DISCUSSION

In this section, we discuss the effects of SFIs on the variation of DSDs. In type SF rainfall, signals of melting layers were found around the 0°C level, and the radar parameter characteristics were similar to type S rainfall above the 0°C level. These findings indicate that stratiform precipitation prevailed in the upper layer. However, below the 0°C level $D_{0}$ and $W_{r}$ increased rapidly toward the lower layer, suggesting that the growth of raindrops occurred by accreting cloud droplets. Considering the existence of a MAUL (Takemi and Unuma, 2020) and cellular echoes in the lower layer (Figure 2d), shallow convection developed below the 3-km level and SFIs occurred between stratiform precipitation and the shallow convection. In addition, the increase in $N_{w0}$ toward the lower layer suggests that small- and medium-sized raindrops were created in this process. One possible explanation for the formation of such raindrops is the warm-rain process in shallow convection, since a large $N_{w0}$ was also observed in type C rainfall (Figure 5f). Another possible explanation is breakup of raindrops with large $D_{0}$ just below the 0°C level (Figure 5e). Such large raindrops are likely to be seeded from the upper layer, since $Z_{H}$ and $K_{DP}$ were larger in type SF than other rainfall types. The findings that the shape of $N(D)$ in type SF rainfall was similar to the equilibrium size distribution (Figure 4b) also suggest that small- to medium-sized raindrop formation occurred by the collisional breakup of large raindrops.

The modifications of DSD through SFI between stratiform precipitation and shallow convection can therefore be summarized as follows: raindrops formed by melting ice particles in the upper stratiform clouds grow by accreting cloud droplets in the lower convective clouds. During this process, the mean diameter of raindrops increases. At the same time, small- to medium-sized raindrops are formed by the warm-rain process and breakup of raindrops, and the normalized intercept parameter of DSD also increases.

Although the increase in the number density of small raindrops through SFIs has been reported previously (Wilson and Barros, 2014; Zwiebel et al., 2016; Zagrodnik et al., 2018), the underlying mechanisms have not yet been clarified and further observations and numerical simulations on SFI are required.

5 | CONCLUSIONS

To study the modification of DSDs through SFIs, we observed precipitation from Typhoon Hagibis using an optical disdrometer and an X-band polarimetric radar. The conclusions are as follows:

1. Three types of rainfall were classified: convective rainfall accompanied by a melting layer (type SF), convective rainfall without a melting layer (type C), and stratiform rainfall with a melting layer.
2. In type SF rainfall, radar parameters exhibited similar characteristics to stratiform precipitation above the 0°C level, while the median volume diameter of raindrops increased rapidly toward the lower layer. These observations suggest that the raindrops formed in the upper stratiform clouds increased in size by accreting cloud droplets in shallow convective clouds in the lower level.
3. In this process, small- to medium-sized raindrops also increased and the DSDs of type SF rainfall showed large normalized intercept parameters. The DSD of type SF rainfall approached the equilibrium size distribution as rainfall intensity increased.

4. Small- to medium-sized raindrops in type SF rainfall were likely to form through the warm-rain process and breakup of raindrops in the lower convective clouds.

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