Characteristics of the Raindrop Size Distribution and Drop Shape Relation in Typhoon Systems in the Western Pacific from the 2D Video Disdrometer and NCU C-Band Polarimetric Radar

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ABSTRACT

The drop size distribution (DSD) and drop shape relation (DSR) characteristics that were observed by a ground-based 2D video disdrometer and retrieved from a C-band polarimetric radar in the typhoon systems during landfall in the western Pacific, near northern Taiwan, were analyzed. The evolution of the DSD and its relation with the vertical development of the reflectivity of two rainband cases are fully illustrated. Three different types of precipitation systems were classified—weak stratiform, stratiform, and convective—according to characteristics of the mass-weighted diameter $D_m$, the maximum diameter, and the vertical structure of reflectivity. Further study of the relationship between the height $H$ of the 15-dBZ contour of the vertical reflectivity profile, surface reflectivity $Z$, and the mass-weighted diameter $D_m$ showed that $D_m$ increased with a corresponding increase in the system depth $H$ and reflectivity $Z$.

An analysis of DSDs retrieved from the National Central University (NCU) C-band polarimetric radar and disdrometer in typhoon cases indicates that the DSDs from the typhoon systems on the ocean were mainly a maritime convective type. However, the DSDs collected over land tended to uniquely locate in between the continental and maritime clusters. The average mass-weighted diameter $D_m$ was about 2 mm and the average logarithmic normalized intercept $N_w$ was about $3.8 \log_{10} \text{mm}^{-1} \text{m}^{-3}$ in typhoon cases. The unique terrain-influenced deep convective systems embedded in typhoons in northern Taiwan might be the reason for these characteristics.

The “effective DSR” of typhoon systems had an axis ratio similar to that found by E. A. Brandes et al. when the raindrops were less than 1.5 mm. Nevertheless, the axis ratio tended to be more spherical with drops greater than 1.5 mm and under higher horizontal winds (maximum wind speed less than 8 m s$^{-1}$). A fourth-order fitting DSR was derived for typhoon systems and the value was also very close to the estimated DSR from the polarimetric measurements in Typhoon Saomai (2006).

1. Introduction

For a measured drop size distribution (DSD), we can calculate numerous rainfall integral parameters (e.g., rainfall rate, liquid water content, etc.) and the dual-polarimetric radar measurements (e.g., reflectivity, differential reflectivity, specific differential phase shift, etc.). It had been found that the DSDs show significant variation under different types of rainfall conditions (Ulbrich 1983). Further investigations of the characteristics of DSDs from disdrometers and polarimetric radars in different climatic regions have been performed by Bringi et al. (2003a). Their results indicate that the maritime convective precipitation has a higher concentration but smaller raindrops compared to continental convective precipitation.

The DSD studies have played a critical role in the development of quantitative precipitation estimation (QPE) algorithms via the forward scattering simulation of radar measurements (e.g., Seliga and Bringi 1976; Sachidananda and Zrnić 1987; Chandrasekar et al. 1990; Gorgucci et al. 1994; Gorgucci and Sarchilli 1997; Ryzhkov and Zrnić 1995; etc.). The most common and widely used algorithm for QPE is the $Z$-$R$ relation, which simply derives the rainfall rate from the reflectivity measurements $Z_H$ with the coefficients $A$ and $b$ in $Z = AR^b$. However, the diversity of DSDs in different rainfall conditions will correspond to different coefficients. With the various DSD data retrieved from
polarimetric radar observations, the coefficients for each corresponding rainfall condition can be acquired and the accuracy of QPE will consequently be improved (Bringi and Chandrasekar 2001).

In addition to the characteristics of the DSDs, the shapes of the raindrops (oblateness) are also of great importance. The unique shape of a drop is determined by the balance between the surrounding forces resulting from surface tension, hydrostatic pressure, and aerodynamic pressure from airflow. Usually, the axis ratio between the minor and major axes is used to describe the oblateness of the raindrops and the axis ratio of raindrops will decrease with increasing volume-equivalent diameters. The drop shape relation (DSR) plays an important role in the interpretation of polarimetric radar measurements and the development of QPE algorithms from radar measurements (e.g., Gorgucci et al. 2001). Tokay and Beard (1996) have also evaluated the oscillation of raindrops, from a field study. Furthermore, Gorgucci et al. (2000) demonstrated that the DSRs should vary with different weather systems because of the oscillation and canting effect of raindrops.

The DSD and DSR characteristics of different types of precipitation systems in different regions have been investigated. The DSD of the tropical cyclones also has been studied by various researchers and well summarized by Tokay et al. (2008). The DSRs of rain have been fully examined by numerical simulations (e.g., Beard and Chuang 1987), wind tunnel experiments (e.g., Pruppacher and Beard 1970), and laboratory experiments (e.g., Andsager et al. 1999). Furthermore, the DSR has also been derived by combining observations from various authors and then fitting a fourth-order polynomial equation (Brandes et al. 2002). However, the characteristics of DSDs and DSRs in typhoon systems in the west Pacific region are still poorly studied. It is important to characterize the DSD and DSR in typhoon systems, which have unique dynamic and microphysical characteristics and cause significant damage. In this present research, the authors focus on the DSDs and DSRs observed in typhoon systems during landfall near northern Taiwan in the western Pacific and compare these with the DSDs and DSRs retrieved from the C-band polarimetric measurements. The remainder of the paper is organized as follows: Section 2 describes data acquirement and data quality-control procedures. The characteristics of typhoon DSDs are discussed in section 3. In section 4, the DSRs for typhoon systems are compared with non-typhoon systems, and the DSR in different weather condition are studied. The DSD retrievals from the C-band polarimetric radar (C-Pol) are examined in section 5, and some conclusions are made in section 6.

2. Disdrometer data analysis

In this research, DSD data were collected by a 2D video disdrometer (2DVD; Schönhuber et al. 1997) at National Central University (NCU; 24°58′18″N, 121°11′3″E), as shown in Fig. 1. A 3D sonic anemometer and a tipping-bucket rain gauge were also installed in
the same location. From June 2001 to October 2005, there were 13 typhoons monitored by the NCU 2DVD and the total accumulated precipitation for these typhoon systems was 1210.5 mm. The name, duration, accumulated rainfall (AR), and counts for 6-min DSD of the NCU 2DVD observation for each typhoon system are listed in Table 1. All the data from the typhoon systems were identified by satellite and radar data to ensure that the DSD data were for typhoon cases. The average distance of the typhoon center from NCU 2DVD was about 60–150 km. Because the 2DVD data were collected during the landfall period and under the influence of complex terrain (as shown in Fig. 1), the horizontal wind speeds observed by the nearby 3D sonic anemometer and surface station were between 0 and 8 m s⁻¹ near the surface.

Much like previous studies, the NCU 2DVD was affected by the error of oversampling small raindrops that results from the wind-caused turbulence near the optical camera and the splash contamination. Kruger and Krajewski (2002) suggested that the oversampling error can be removed by a fall velocity–based filter as follows:

\[ |V_{\text{measured}} - V_{\text{ideal}}| < CV_{\text{ideal}}, \]

where \( V_{\text{measured}} \) represents the observed fall velocity, \( V_{\text{ideal}} \) represents fall velocity derived from the formula given in Brandes et al. (2002), and coefficient \( C \) was chosen as 0.4 [as recommended by Kruger and Krajewski (2002) and Thurai and Bringi (2005)]. Before any further calculations, we applied the same algorithm to remove the oversampling error from NCU 2DVD typhoon data to insure the accuracy of the DSD calculation.

Every 6 min, DSDs were derived after removing the biased data. Then, the moment method (Kozu and Nakamura 1991; Tokay and Short 1996) was applied to calculate the coefficients of the gamma distribution for each DSD. The gamma distribution of the DSDs is of the form

\[ N(D) = N_0 D^\mu \exp(-\Lambda D), \]

where \( N(D) \) (mm⁻¹ m⁻³) represents the number of raindrops per unit volume per unit volume–equivalent diameter, which can be characterized by intercept \( N_0 \) (mm⁻¹ m⁻³), shape \( \mu \) (dimensionless), and slope \( \Lambda \) (mm⁻¹). Through the calculation of the integration of the third, fourth, and sixth moments of the DSDs, the coefficients can be deduced.

The study by Smith and Kliche (2005) indicated that bias can be produced when deriving the coefficients of the gamma distribution by using the moment method. However, the bias problem can be reduced by integrating a sufficient number of raindrops. Smith and Kliche (2005) suggested that the raindrop sampling number should be more than a few thousand to prevent the bias from the moment method. In the following calculation, a 6-min DSD calculation was chosen to ensure a sufficient amount of raindrops. Furthermore, rainfall rates of less than 1 mm h⁻¹ were also removed to eliminate inadequate DSD cases.

For the clarity and the convenience of comparison with earlier documentation, the normalized gamma distribution (Sekhon and Srivastava 1971; Willis 1984; Testud et al. 2001), instead of the gamma distribution, was used to illustrate the DSDs. The normalized intercept coefficient \( N_w \) (mm⁻¹ m⁻³) can be derived as follows (Bringi et al. 2003b):

\[ N_w = \frac{(4.0)^4}{\pi \rho_w} \left( \frac{10^3 W}{D_m} \right), \]

where \( \rho_w \) (1.0 g cm⁻³) represents the density of rainwater and \( W \) (g m⁻³) represents the liquid water content for the corresponding DSD. Consequently, the gamma distribution of DSDs can be modified as a normalized gamma distribution with \( N_w \) in the form

\[ N(D) = N_w f(\mu) \left( \frac{D}{D_m} \right)^\mu \exp\left[-(4.0 + \mu) \frac{D}{D_m}\right], \]

where

\[ f(\mu) = \frac{6}{(4.0)^4} \frac{(4.0 + \mu)^{\mu+4}}{\Gamma(\mu + 4)}. \]
The normalized gamma distribution of DSD can be derived from the calculation of \( \mu \), \( \Lambda \), and \( D_m \) by using the moment method and the integration of \( W \) from the DSDs; the rainfall rate can be calculated as follows:

\[
R \text{ (mm h}^{-1}\text{)} = (0.6 \times 10^{-3}\pi)(3.78)N_w f(\mu) \\
\times \Gamma(5.0 + \mu) \frac{D_m^{4.67}}{(4.0 + \mu)^{3.0 + \mu}}. \tag{5}
\]

3. The characteristics of DSDs in a typhoon system

a. Statistical analysis of typhoon cases

Among these 13 typhoon cases, there were 1884 of the 6-min DSDs fit into the gamma distribution form with coefficients \( \mu \) and \( \Lambda \) via the moment method calculation. The mass-weighted diameter \( D_m \) (mm) and the total concentration \( N_T \) of each 6-min DSD had also been calculated. In Fig. 2, the variabilities of \( \mu \), \( \Lambda \), \( D_m \), and \( N_T \) were distinct in the values of rainfall rate less than 10 mm h\(^{-1}\); however, the range of variation decreased with increasing rainfall rate and became more uniform in higher rainfall rates. The values of gamma coefficients \( \mu \) and \( \Lambda \) decreased with an increase in rainfall rate. The values of \( \mu \) and \( \Lambda \) were about 5 (dimensionless) and 4.7 mm\(^{-1}\), respectively, when the rainfall rate was 20–40 mm h\(^{-1}\); and 3.6 (dimensionless) and 3.4 mm\(^{-1}\), respectively, when the rainfall rate was 40–60 mm h\(^{-1}\). In Fig. 2c, \( D_m \) increased with an increase in the rainfall rate; the average \( D_m \) values were about 1.9 and 2.2 mm when the rainfall rates were 20–40 and 40–60 mm h\(^{-1}\), respectively. In Fig. 2d, the total concentration of raindrops increased with the increasing of rainfall rate, and the average \( N_T \) values were about 783 and 944 m\(^{-3}\) when the rainfall rates were 20–40 and 40–60 mm h\(^{-1}\), respectively. In Figs. 2c,d, there were three heavy rain events (rainfall rate >60 mm h\(^{-1}\)) characterized by high \( N_T \) values (above 10\(^3\) m\(^{-3}\)) and \( D_m \) values that remained about 2.2 mm. The unique values of \( N_T \) and \( D_m \) implied that the heavy rain events were mainly composed of high concentrations of small to medium sized raindrops rather than large raindrops. In this research, raindrops were considered small or medium if the diameter of raindrop was less than 1 mm or between 1 and 3 mm, respectively. Raindrops greater than 3 mm were considered large.

As shown in Fig. 3, the mean \( Z-R \) relationship (\( Z = AR^b \)) was also derived by the fitting of the typhoon system DSD-obtained rainfall rate and the reflectivity for comparison. And there was no separation of the convection and stratiform systems for the calculation of the mean \( Z-R \) relation. Compared to the standard \( Z-R \) relationship with the coefficients \( A = 300 \) and \( b = 1.4 \), the fitted mean \( Z-R \) relationship of typhoon systems had a lower value of \( A = 206.83 \) and similar value of \( b = 1.45 \). In Fig. 3, the standard \( Z-R \) relationship
underestimates the rainfall rate in most of the cases. Compared to the $Z-R$ relations derived from hurricanes by Jorgensen and Willis (1982; $Z = 300R^{1.35}$) and Wilson and Pollack (1974; $Z = 350R^{1.35}$), the $Z-R$ relation of the typhoon precipitation system showed that $A$ and $b$ were lower and higher, respectively. The difference was expected with the unique DSD of the typhoon precipitation systems. Therefore, the following study of this research will focus on the analysis of the DSD characteristics within the typhoon systems rather than the rainfall estimation application regarding the $Z-R$ relationship.

In the five years of DSD data, Typhoon Nari in 2001 and Typhoon Haima in 2004 had the highest accumulations of precipitation (see Table 1). The characteristics of the DSDs and the corresponding reflectivity vertical profile observed from the radar code of Wu-Fanshan (RCWF) radar for Typhoons Nari (2001) and Haima (2004) are further discussed in the following sections. The RCWF radar is an operational Weather Surveillance Radar-1988 Doppler (WSR-88D) radar of the Central Weather Bureau of Taiwan. The positions of the RCWF radar and the NCU 2DVD are shown in Fig. 1. The distance between these two locations is about 60.47 km and the RCWF radar provides a volume scan of 9 elevation angles every 6 min. The detailed specifications are listed in Table 2.

b. Typhoon Nari

The time series of the corresponding vertical reflectivity profile from RCWF radar, system type, mass-weighted diameter $D_m$, rainfall rate, forward-calculated reflectivity, wind speed, and 6-min DSD from 0000 UTC 16 to 1200 UTC 17 September 2001 were calculated from the NCU 2DVD data (see Fig. 4). The evolution of the DSDs indicates that the DSDs varied considerably with different rainfall rates and the maximum diameter $D_{max}$ of each 6-min DSD could vary from less than 2 mm to nearly 5.1 mm, whereas the $D_m$ could vary from 1 to 2.2 mm (see Figs. 4a,b). From the reflectivity vertical profile (Figs. 4a,b), it can be seen that the 30- and 10-dBZ contours can reach 6 and 12 km, respectively. The average melting level (0°C) in northern Taiwan was about 5.2 km above mean sea level (MSL), as calculated from the sounding data collected during Typhoon Nari (2001). According to the surface reflectivity, the structure of the vertical reflectivity profile, and the rainfall rate, typhoon systems can be classified into three different types: weak stratiform, stratiform, and convective systems (shown in Figs. 4a,b). The corresponding criteria
of the surface reflectivity and rainfall rate for these three types are listed in Table 3.

In the first 8.5 h (0000–0830 UTC 16 September), there was only very light rain, with shallow reflectivity vertical profiles. During this period, this system was considered to be a weak stratiform system and the corresponding DSDs had a relatively small maximum diameter (about 3.0 mm) and low concentration of raindrops with small to medium diameters. Nevertheless, the system became stronger and the rainfall rate also increased slightly from 0830 to 1600 UTC. It was then classified as a stratiform system and the corresponding DSDs showed a larger maximum diameter (about 3.5 mm) and a higher concentration of small to medium diameter raindrops.

Afterward, the first significant precipitation appeared around 1600 UTC and persisted for almost two hours. The rainfall rates were around 20–40 mm h\(^{-1}\) and the DSDs had more small to medium diameter raindrops [the \(\log_{10} N(D)\) values were greater than 3.5–4] with the largest raindrops reaching 4.4 mm. Therefore, \(D_m\) could reach about 2 mm. At the same time, the vertical reflectivity profiles were stronger and the 30-dBZ contour stretched up to 6 km, with a distinct strong vertical reflectivity column. It was classified as a convective system. After the convective system passed by, there was another 3-h period stratiform system. The second period of convective precipitation appeared at 2100 UTC. The DSDs and rainfall rates were similar to those of the first period of heavy precipitation, but they were weaker and the maximum diameter was about 3.9 mm. The missing data between 2200 and 2300 UTC were due to the overwhelming of the hard disk of the control computer for NCU 2DVD.

The convective system with heaviest precipitation began at 0030 UTC 17 September and persisted for about 3.5 h. The maximum precipitation (MP) event recorded at 0230 UTC was about 73 mm h\(^{-1}\), whereas the maximum mass-weighted diameter \(D_m\) event (MD\(_m\)) recorded at 0200 UTC, 30 min before the MP, was about 2.2 mm (see Fig. 4d). Even though the largest raindrop could reach 5 mm in MP, the \(D_m\) was still confined to less than 2.2 mm because of the higher concentration of small and medium raindrops. In fact, the DSD of the MD\(_m\) was composed of fewer small to medium diameter raindrops than the MP event with a lower rainfall rate of about 41.5 mm h\(^{-1}\). Higher concentrations of small to medium diameter raindrops during the MP event limited the \(D_m\) to a smaller value, which also indicates that the large amount of small to medium raindrops was responsible for the heavy precipitation rather than larger raindrops. The microphysical processes of this highly efficient production of small to medium drops need further investigation.

The corresponding characteristics for different types of systems obtained by analyzing the DSDs from Typhoon Nari could be summarized as follows: 1) weak stratiform precipitation had a small \(D_m\) (1.0–1.5 mm), a smaller maximum diameter (less than 3.0 mm), and fewer small to medium raindrops; 2) stratiform precipitation had a medium \(D_m\) (1.5–1.9 mm), a maximum diameter of about 3–3.8 mm, and a higher concentration of small to medium raindrops; 3) convective precipitation had a higher \(D_m\) (1.9–2.2 mm), a maximum diameter greater than 3.8 mm, and the highest concentration of small to medium drops (summarized in Table 3).

### c. Typhoon Haima

The time series of the vertical reflectivity profile, system type, and DSD data of Typhoon Haima (2004) from 18 UTC 10 to 06 UTC 11 September 2004 were also derived (wind speed data was not available in this case). In Figs. 5c,d, it can be seen that the maximum diameter varied from less than 1.5 mm to nearly 5.4 mm and the \(D_m\) varied between 1 and 2.5 mm. During this period, the heaviest 6-min rainfall rate was about 60.6 mm h\(^{-1}\). The first convective precipitation (20–60 mm h\(^{-1}\)) appeared at 1930 UTC 10 September and persisted for about 1.5 h. The corresponding vertical reflectivity profile indicated a very strong convective tower, and the 30-dBZ contour could reach 8 km above MSL. At the same time, the DSDs showed that, for relatively high concentrations of small to medium raindrops, the values of \(\log_{10} N(D)\) could reach about 3.5–4. The values of \(D_m\) could extend up to 2.5 mm, which is higher than the values for Typhoon Nari (2001). Although the DSDs showed larger \(D_m\) values and a maximum diameter value of 5.3 mm, they showed fewer small to medium raindrops than the previously
Fig. 4. The time series of the DSDs calculated from the NCU 2DVD and the reflectivity vertical profile from the RCWF WSR-88D radar of Typhoon Nari case from 0000 UTC 16 to 1200 UTC 17 Sep 2001. (a),(b) The color shaded represents the reflectivity vertical profile observed by RCWF radar. The y axis represents the altitude. The weak stratiform, stratiform, and convection systems (classified according to Table 3) are illustrated by the colors pink, blue, and red on the top of the diagram. (c),(d) The black line represents $D_m$. The color shades represent the DSD in logarithmic units of $\text{mm}^{-1} \text{m}^{-3}$. The y axis indicates the equivolumetric diameter (mm) of raindrops. The MP and MD$D_m$ events are marked by black dashed lines. (e),(f) The blue and green lines represent the reflectivity (dBZ) calculated from the DSDs and the rainfall rate (mm h$^{-1}$) calculated from the DSDs. The pink line represents the wind speed obtained from the nearby 3D sonic anemometer. The right-hand side y axis represents the scale of rainfall rate and the left-hand side y axis represents the reflectivity (dBZ) and the wind speed (m s$^{-1}$).
Table 3. The characteristic values of the rainfall rate, reflectivity, $D_m$, and $D_{\text{max}}$ for the weak stratiform, stratiform, and convection systems.

<table>
<thead>
<tr>
<th></th>
<th>Weak stratiform</th>
<th>Stratiform</th>
<th>Convection</th>
</tr>
</thead>
<tbody>
<tr>
<td>Rainfall rate (mm h$^{-1}$)</td>
<td>&lt;10</td>
<td>&lt;10</td>
<td>&gt;10</td>
</tr>
<tr>
<td>Reflectivity (dBZ)</td>
<td>&lt;30</td>
<td>30–35</td>
<td>&gt;35</td>
</tr>
<tr>
<td>$D_m$ (mm)</td>
<td>1–1.5</td>
<td>1.5–1.9</td>
<td>&gt;1.9</td>
</tr>
<tr>
<td>$D_{\text{max}}$ (mm)</td>
<td>&lt;3</td>
<td>3–3.8</td>
<td>&gt;3.8</td>
</tr>
</tbody>
</table>

The best-fit values of $D_m$ were used to estimate the DSDs. In Fig. 6, the values of $D_m$ calculated from DSD data and the $H$ from the RCWF radar observations indicate a very good correlation. Generally, the $D_m$ increased with an increase of $H$. Meanwhile, considering that the stronger systems usually associated with the higher value of $D_m$, the reflectivity $Z$ also showed good correlation with $D_m$. The best-fit values of $D_m$ were included for $H$ and the surface DSD derived reflectivity $Z$ (dBZ). The regression fitting of $D_m - H - Z$ and the root-mean-square error (RMSE) can be derived as follows:

$$D_m (\text{mm}) = 0.10843H (\text{km}) + 0.022744Z (\text{dBZ}) - 0.008853$$  \tag{6}

$$\text{RMSE} = \left[ \frac{\sum_{i=1}^{N} (D_m^{2\text{DVD}} - D_m^{\text{est}})^2}{N} \right]^{0.5}$$  \tag{7}

where the RMSE [Eq. (7)] was about 0.18105 mm. The RMSE was sufficiently small enough to estimate $D_m$, and it can be beneficial to the situation in which the DSD information is lacking. For example, the $D_m$ value estimated from the relationship of $D_m - H - Z$ can reduce the uncertainty of the QPE from the $Z - R$ relationship because of the variability of the DSD from the conventional radar.

This relationship indicates that deeper convective systems can provide an environment that favors the production of DSDs with larger raindrops (higher values of $D_m$) through collision–coalescence processes. Concurrently, a deeper convective system could produce larger raindrops from the melting snowflakes or graupel. However, because of the lack of in situ measurements of the microphysical processes in typhoon systems, the cause of this relationship has to be determined in the future by using observations and an explicit cloud model for further investigation.

d. Relationship between DSDs and the depth of the convective systems

In the previous examination of the disdrometer DSD data and the vertical reflectivity profiles from the radar observation of Typhoons Nari (2001) and Haima (2004), we can see that the maximum diameter of the raindrops extended to higher values when the typhoon systems had deeper convective systems. The low $D_m$ values usually were associated with shallower vertical reflectivity profiles and more small to medium raindrops. This was especially obvious between 2100 UTC 10 and 0400 UTC 11 September in the case of Typhoon Haima (2004). To confirm this relationship, the altitude of the 15-dBZ contours of the vertical reflectivity profiles ($H; \text{km}$) were used as an indicator of the depth of the typhoon system and the $D_m$ was used to characterize the DSDs. In Fig. 6, the values of $D_m$ calculated from DSD data and the $H$ from the RCWF radar observations indicate a very good correlation. Generally, the $D_m$ increased with an increase of $H$. Meanwhile, considering that the stronger systems usually associated with the higher value of $D_m$, the reflectivity $Z$ also showed good correlation with $D_m$. The best-fit values of $D_m$ were included for $H$ and the surface DSD derived reflectivity $Z$ (dBZ). The regression fitting of $D_m - H - Z$ and the root-mean-square error (RMSE) can be derived as follows:

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e. DSD characteristics of Typhoons Nari and Haima

In Fig. 7a, the distribution of $D_m$ and $N_w$ from the DSD of Typhoons Nari (2001) and Haima (2004) showed a great variation when the rainfall rate was less than 10 mm h$^{-1}$, which indicates stratiform precipitation in the typhoon systems. The results were similar to those of the stratiform precipitation DSDs from Bringi et al. (2003a). However, by comparing the values of $D_m$ and $N_w$ of the convective system (rainfall rate greater than 10 mm h$^{-1}$) from the two typhoon cases with the results from Bringi et al. (2003a), we found that the DSDs of the convective system in typhoons during landfall were
FIG. 5. As in Fig. 4, but for the Typhoon Haima case from 1800 UTC 10 Sep to 0600 UTC 12 Sep 2004. The wind speed data were not available in this case.
actually neither typical maritime nor continental convective systems. The results from Bringi et al. (2003a) indicate that the maritime (continental) type of convective systems have $D_m \approx 1.5–1.75 \text{ mm} \ (2–2.75 \text{ mm})$ and logarithmic $N_w \approx 4–4.5 \ (3–3.5)$. In this study, the mean values of $D_m$ and logarithmic $N_w$ for the convective systems of Typhoons Nari (2001) and Haima (2005) were about 2 mm and 3.8, respectively. These results indicate that the DSDs of the convective systems in typhoons had a lower (higher) concentration of raindrops and larger (smaller) raindrops in comparison to maritime (continental) convective precipitation (see Fig. 7a).

Bringi et al.’s (2003a) results indicate that the DSD of the typhoon systems were characterized by maritime-type convective precipitation. Moreover, Tokay et al. (2008) reported that the mean $D_m$ of 40-dBZ DSD (rainfall rate $\sim 18.5 \pm 0.5 \text{ mm h}^{-1}$) from Atlantic tropical cyclones was $1.67 \pm 0.3 \text{ mm}$, which was also considered as maritime-type convective systems. In this research, the mean $D_m$ and rainfall rate of the 40-dBZ DSD were about 1.88 mm and 12.1 mm h$^{-1}$, respectively. The different characteristics of the DSD in this research and the previous two studies were mainly due to the unique topography of northern Taiwan (see Fig. 1). In this research, the DSD data were collected at the western side of the Central Mountain Ridge (CMR) of Taiwan, whereas Typhoons Nari (2001) and Haima (2004) both moved through the northern part of Taiwan from the northeast toward the southwest. These characteristics of the DSD, the relative higher values of $D_m$ and the lower values of $N_w$, could be considered the features of terrain-influenced convective systems during typhoon landfall. However, the DSD retrieval from Typhoon Saomai (2006) showed similar maritime-type convective characteristics to Bringi et al. (2003a) when it was observed on the ocean. The characteristics of the DSD retrieved from Typhoon Saomai (2006) are discussed further in section 5.

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**Fig. 6.** The scatterplot of $D_m$ and $H$ of the 15-dBZ contour of the vertical reflectivity profile. The plus signs represent the data from Typhoon Nari (2001) and the circles represent the data from Typhoon Haima (2004). The solid and dashed lines represent the best fit of all $D_m$, altitude of the 15-dBZ contour of the vertical reflectivity profile $H$ and surface reflectivity $Z$.

**Fig. 7.** (a) Scatterplot of $D_m$ (mm) and $N_w$ (mm$^{-1}$ m$^{-3}$). The thin dashed lines indicate the rainfall rate from 10 to 100 mm h$^{-1}$ and the interval is 10.0 mm h$^{-1}$. The dots represent the DSD data from Typhoon Nari (2001) and the circles represent the data from Typhoon Haima (2004). The gray shades represents the $D_m$ and log$_{10}(N_w)$ values of the retrieved DSDs from NCU C-Pol in Typhoon Saomai (2006). The two outlined squares represent (left) the maritime and (right) continental types of convective systems. (b) As in (a), but the evolution of the $D_m$ and $N_w$ derived from the DSDs of Typhoon Haima (2004). The circles represent the DSD data of the convection system from 1930 to 2100 UTC 10 Sep. The triangles represent the weak stratiform system from 2100 to 2300 UTC 10 Sep. The squares represent the stratiform system from 2300 UTC 10 Sep to 0324 UTC 11 Sep.
In Typhoon Haima (2004), the dramatic change of the DSD and the corresponding vertical reflectivity profile can be seen in Fig. 5. The evolution of the DSD on the $D_m-N_w$ plot is shown in Fig. 7b. According to the previous classification, the evolution of the DSD was tracked throughout the three types of precipitation systems. The convective system was characterized by the $D_m$ values around 1.75–2.5 mm and the logarithmic $N_w$ values around 3.5–5. A following weak stratiform system with $D_m$ values around 1–1.5 mm and logarithmic $N_w$ values around 3.6–4.1 was observed. A stratiform system with $D_m$ values around 1.4–1.9 mm and the logarithmic $N_w$ values around 3.1–3.5 was subsequently observed. These three different characteristics of the DSD revealed different microphysical processes. The unique $N_w-D_m$ distribution cluster of the deep convective systems was associated with both warm rain and ice processes. The stratiform systems with bright-band signatures produced large raindrops and fewer small to medium raindrops. The weak stratiform system was characterized by a shallow development system associated with only small to medium raindrops.

4. Drop shape relation of typhoon systems from disdrometer measurements and polarimetric radar estimations

a. DSR from typhoon systems

One of the advantages of the 2DVD is the capability of recording the axis ratio of each raindrop. Thurai and Bringi (2005) had already used a 2D video disdrometer (same as NCU 2DVD) to measure artificial data generated from a bridge 80 m above the ground. Their results were similar to those of Brandes et al. (2002), who determined the DSR by combining various data and fitting the data to a fourth-order polynomial. The DSR from Brandes et al. (2002) is outlined as follows:

$$b/a = -2.492 \times 10^{-4}D^4 + 5.303 \times 10^{-3}D^3 - 3.644 \times 10^{-2}D^2 + 2.51 \times 10^{-2}D + 0.9951.$$  (8)

However, the DSR in typhoon systems still has not been well studied. In this uniquely dynamical environment (different from the clam condition), we analyzed the DSR characteristics and further apply the DSR to retrieve the DSDs from polarimetric measurements in typhoon systems in a later section.

Different from the artificial data of Thurai and Bringi (2005), the axis ratios of each raindrop collected in a natural environment among 13 typhoon cases were derived. As in Gorgucci et al. (2000, 2001) and Bringi et al. (2003a), considering the oscillation and canting effect in a natural environment, the axis ratio and DSR derived in this research should be referred to the “effective axis ratio” and the “effective DSR.” In Fig. 8a, the frequency distribution of every 0.02 of axis ratio (dimensionless) and 0.2 mm of equivolumetric diameter were shown by a logarithmic scale. The mean axis ratio (black solid line) and the mean absolute deviation (white vertical lines) were calculated for each corresponding equivolumetric diameter after applying the fall velocity–based filter to remove the biased data. Because of instrumental limitation, the axis ratios of raindrops sized less than 1 mm were still highly biased. Because of this, the axis ratios of raindrops sized less than 1 mm were artificially set to nearly spherical raindrops (0.9999). The observed maximum diameter could reach about 5.4 mm, but the calculation of the DSR fitting was established to be within 3.42 mm in order to have sufficient numbers of raindrops. Consequently, the fourth-order polynomial DSR fitting was derived as follows:

$$b/a = -1.9223 \times 10^{-4}D^4 + 4.3402 \times 10^{-3}D^3 - 3.3439 \times 10^{-2}D^2 + 4.2514 \times 10^{-2}D + 0.98287.$$  (9)

The fourth-order polynomial DSR fitting, as shown in Fig. 8a (red line), can only be reasonably extended to 5 mm. In the five years of typhoon DSD data, the maximum sizes of the raindrops seldom exceeded 4 mm with very few raindrops. Furthermore, the DSR from Brandes et al. (2002) is also shown on Fig. 8a (black dashed line) to facilitate intercomparison with the DSR from typhoon cases. The results indicate that the DSR of raindrops sized less than 1.5 mm in these typhoon cases was similar to the DSR from Brandes et al. (2002). However, the DSR is more spherical when the diameter is greater than 1.5 mm. The difference in the DSR from
FIG. 9. The contour plot of the DSR of raindrops (mean coefficient $b$ of the first-order polynomial DSR) for each different rainfall rate (0–50 mm h$^{-1}$) and horizontal wind velocity (0–8 m s$^{-1}$) for (a) nontyphoon and (b) typhoon systems. At each point in the panels, the coefficient $b$ is shown and the percentage of raindrops of each coefficient $b$ is shown in parentheses for each 10 mm h$^{-1}$ rainfall rate and each 1 m s$^{-1}$ horizontal wind velocity, respectively. The total concentrations for (a) and (b) were 42 530 445 and 1 156 538 raindrops, respectively.
Brandes et al. (2002) and the typhoon cases was important in the following application of the polarimetric measurements.

To further examine the credibility of the measurement of the axis ratio from 2DVD, the mean axis ratio for low rainfall rates and horizontal wind velocity conditions in nontyphoon cases during the same observation period were also derived. In Fig. 8b, as in Fig. 8a but with the derived values from nontyphoon cases, a rainfall rate was constrained below 2 mm h⁻¹ and horizontal winds below 1 m s⁻¹, which was similar to the artificial environment in Thurai and Bringi (2005). In Fig. 8b, the mean axis ratio from nontyphoon systems (derived from the same procedure as for the typhoon cases) was close to Thurai and Bringi (2005) and Brandes et al. (2002). The mean axis ratios of the raindrops less than 2.1 mm were similar to the DSR from Brandes et al. (2002). However, a small deviation can be noticed with raindrops greater than 2.1 mm. The calculation of the mean axis ratio from natural environmental data was affected by the insufficient number of raindrops and the axis-ratio observation errors. Nevertheless, the result indicates that the instrumental measuring errors were limited and acceptable.

b. DSR characteristics at different rainfall rates and wind speeds

The previous results indicated that the DSR of the typhoon systems was slightly more spherical when the raindrops were greater than 1.5 mm. To understand the reasons for the variation of the DSRs, the rainfall rate and horizontal winds were used as indicators during different environmental conditions. Research of natural environmental influences on the DSRs observed by a 2D video disdrometer has not been well reported in the past.

The first-order DSR equation from Pruppacher and Beard (1970) was applied to describe the DSRs:

\[
\frac{b}{a} = 1.03 - \beta D,
\]

where \( D \) represents the equivolumetric diameter in millimeters. The advantage of using the above first-order polynomial fitting is that the DSRs can be roughly described by a single coefficient \( \beta \). The \( \beta \) value in Pruppacher and Beard (1970) was 0.062. The corresponding coefficient \( \beta \) for Brandes et al. [2002; as shown in Eq. (8)] and the new DSR for typhoon systems [Eq. (9)] were 0.0599 and 0.0477, respectively. Higher values of the coefficient \( \beta \) represent more oblate raindrops and lower values represent more spherical raindrops. Values of \( \beta \) for different conditions were then calculated for each DSD dataset, which was categorized by every 10 mm h⁻¹ rainfall rate ranging from 0 to 50 mm h⁻¹ and every 1 m s⁻¹ horizontal wind velocity value from 0 to 8 m s⁻¹.

In Fig. 9a, the results from nontyphoon systems indicate that the DSRs show a great variation under different rainfall rates and wind velocities. Generally, the DSRs had more spherical raindrops with a higher horizontal wind velocity and rainfall rate. The coefficient \( \beta \) could vary from 0.05 (0–10 mm h⁻¹ and 0–1 m s⁻¹) to 0.03 (40–50 mm h⁻¹ and 7–8 m s⁻¹). When the horizontal wind velocity was lower than 3 m s⁻¹, the coefficient \( \beta \) (0.05–0.046) was close to the value from Pruppacher and Beard (1970). However, when the wind speed was greater than 3 m s⁻¹, the \( \beta \) value (0.04–0.03) decreased with the increases of wind speed and the rainfall rate. Hence, the 45° orientation contours from the upper left to lower right corner indicate that the wind velocity and the rainfall rate both play important roles in the DSRs.

In Fig. 9b, the DSRs of typhoon systems show a similar relation to wind speed as those of nontyphoon systems; that is, the DSRs tend to be more spherical when there was higher wind velocity. It should be noticed that the value of coefficient \( \beta \) varied from 0.04 to 0.02, which is lower than that for the nontyphoon cases (0.05–0.03) in the same range of the wind speed. However, the differently orientated contours (from the upper right to lower left corner) indicate that the DSR tends to be more oblate with increasing rainfall rate at the same wind speed, especially when the wind speed was greater than 3 m s⁻¹.

Although there were large variations in the DSR with the different environmental conditions among typhoon and nontyphoon systems (coefficient \( \beta \) varied from 0.02 to 0.05), there were fewer raindrops of high horizontal wind velocity. The total percentages of the raindrops that had a horizontal wind velocity greater than 4 m s⁻¹ were only 23.87% and 2.17% for typhoon and nontyphoon cases, respectively. Therefore, the corresponding average coefficient \( \beta \) for typhoon DSR is 0.0477 [fitting from Eq. (9)]. Generally, the horizontal wind velocity was the major factor of influence for the DSRs. The DSRs were relatively more spherical in typhoon systems than in nontyphoon systems. These features could be attributed to the oscillation and canting effects within the precipitation systems. The higher wind speeds lead raindrops to have more oscillation and canting effects. Therefore, the more-spherical DSRs derived in this research should be regarded as effective DSRs. One thing should be emphasized: the DSRs calculated from the NCU 2DVD observation were the combination of the instrumental measuring errors and the nature facts. The calculation of the coefficient \( \beta \) was regarded as
an approach to understanding the variation of the DSR from the observation of a 2D video disdrometer. The DSR of a typhoon system should be considered as Eq. (9) in this research.

c. The DSR estimated from NCU C-band polarimetric radar

To further verify the DSR in typhoon systems, the estimated $\beta$ from the C-band polarimetric radar measurements of Typhoon Saimai (2006) was derived. In this research, the polarimetric measurements from the NCU Department of Atmospheric Sciences C-Pol were used to estimate the DSR. The NCU C-Pol was upgraded from a conventional Doppler radar in December 2004. The NCU C-Pol transmits horizontal and vertical electromagnetic waves simultaneously via a magnetron transmitter (see detailed specifications in Table 2). According to the statistics, the accuracy of the differential reflectivity measurement $Z_{\text{DR}}$ is about $\pm 0.24$ dB. The polarimetric measurements—reflectivity $Z_H$ and differential reflectivity $Z_{\text{DR}}$—from NCU C-Pol in Typhoon Saomai (shown in Fig. 12) had already been carefully calibrated to correct for system bias and the attenuation effect before being applied to any further application. For the C-band polarimetric radar attenuation correction in rain medium, the self-consistent attenuation correction method from Brinigi et al. (2001) was applied to both the reflectivity $Z_H$ and differential reflectivity $Z_{\text{DR}}$. The value of $K_{\text{DP}}$ was calculated from the observed differential phase shift $\Phi_{\text{DP}}$ by using the algorithm from Hubbert and Bringi (1995).

Considering the oscillation of the raindrops, Gorgucci et al. (2000) demonstrated that the coefficient $\beta$ of DSR can be estimated via the polarimetric measurements from an S-band radar system. In this research, the corresponding coefficients of the equation have been modified for the C-band radar system as following:

$$\beta = 3.08 Z_H^{-0.372} K_{\text{DP}}^{0.375} (\xi_{\text{DR}} - 1)^{-0.299},$$

where the horizontal reflectivity $Z_H$ is in $\text{mm}^6 \text{m}^{-3}$, $\xi_{\text{DR}}$ is the ratio of horizontal and vertical reflectivity, and $K_{\text{DP}}$ is in degrees per kilometer. In Fig. 10, the histogram of estimated coefficient $\beta$ from Typhoon Saomai (2006) showed that the most of the values of $\beta$ concentrate about 0.05. The result showed that the estimated coefficient $\beta$, via polarimetric measurements, is very close to the surface disdrometer derived first-order DSR (corresponding coefficient $\beta$ was about 0.0477). From these two different approaches (surface 2D video disdrometer and estimation from a polarimetric radar), they both exhibited a more spherical DSR in a typhoon system. This consistency result, similar to the research from Gorgucci et al. (2000), indicates that the DSRs should vary with different weather systems.
5. Raindrop size distribution retrieval from NCU C-band polarimetric radar

Estimating the DSDs from the polarimetric radar measurements by the constrained-gamma method was first demonstrated by Zhang et al. (2001) and Brandes et al. (2003, 2004). The constrained-gamma method basically uses the inherent advantages of polarimetric measurements (i.e., differential reflectivity $Z_{DR}$) from a polarimetric radar and an empirically constrained relationship between the shape factor $\mu$ and slope parameter $\Lambda$ of the DSDs. The $\mu$–$\Lambda$ relationship has been

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**FIG. 11.** The scatterplot of $\mu$ and $\Lambda$ from each 6-min DSD of typhoon systems after removing the DSDs with rainfall rate $<5$ mm h$^{-1}$. The solid line and the thick dashed line represent the fitting of the $\mu$–$\Lambda$ relationship and the $\mu$–$\Lambda$ relationship from Brandes et al. (2004), respectively. The thin dashed lines represent $D_m$ from $\Delta D_m = 4 + \mu$.

**FIG. 12.** The NCU C-Pol observation of Typhoon Saomai at 0018 UTC 10 Aug 2006: (a) $Z_{H}$ (dBZ), (b) $Z_{DR}$ (dB), (c) $\Phi_{DP}$ ($^\circ$), (d) $R_{HV}$, (e) $K_{DP}$ ($^\circ$ km$^{-1}$), (f) $K_{DP}$ ($^\circ$ km$^{-1}$), (g) $\log_{10}(N_{W})$ of retrieved DSDs, and (h) $D_m$ of retrieved DSDs. The corresponding altitude of each range ring is also labeled.
proven capable of providing useful information describing DSDs (Zhang et al. 2003). Consequently, the polarimetric measurements can provide the high temporal and spatial resolution for DSD estimation, which is a great advantage over surface disdrometers. Brandes et al. (2004) and Vivekanandan et al. (2004) used a video disdrometer dataset from east-central Florida, obtained during the summer of 1998, to derive the following \( \mu-\Lambda \) relationship:

\[
\Lambda = 0.0365\mu^2 + 0.735\mu + 1.935. \tag{12}
\]

In this research, the empirical \( \mu-\Lambda \) relationship for typhoon systems was derived from five years of NCU 2DVD typhoon systems data (as shown in Table 1). The best fitting \( \mu-\Lambda \) relationship in the parameter diagram (GPD), which is the same as in Ulbrich and Atlas (2007), is shown in Fig. 11. The \( \mu-\Lambda \) relationship was established by removing any DSDs in which the rainfall rate was less than 5 mm h\(^{-1}\), which was the same as in Brandes et al. (2004). The new \( \mu-\Lambda \) relationship for typhoon systems was derived as

\[
\Lambda = 0.0136\mu^2 + 0.6984\mu + 1.5131. \tag{13}
\]

Comparing the \( \mu-\Lambda \) relationship for typhoon systems with that of Brandes et al. (2004), the \( \mu-\Lambda \) relationship for typhoon systems has a smaller slope. The differences reveal that the DSD in typhoon systems have lower values of \( D_m \), indicating that the microphysical processes in typhoon systems under terrain influence is different than the system observed in east-central Florida. The higher values of \( D_m \) from Brandes et al. (2004) indicate that DSD were more predominated by the deep convection.

With the understanding of typhoon systems DSDs obtained from surface disdrometer data, the further investigation of DSDs through a polarimetric radar in typhoon systems becomes practicable. On 10 August 2006, Typhoon Saomai passed over the ocean near northern Taiwan and moved northwestward. It was observed by the NCU C-Pol located over northern Taiwan. Figures 12a–d show the 0.5° elevation angle polarimetric measurements, showing the horizontal reflectivity \( Z_H \), differential reflectivity \( Z_{DR} \), differential phase shift \( \Phi_{DP} \), and cross-correlation coefficient \( \rho_{HV} \), which revealed the structure of Typhoon Saomai (2006). The center of the typhoon was about 200 km northeast of NCU C-Pol. The maximum wind speed near the eyewall was about 48 m s\(^{-1}\), and the central pressure was about 935 hPa (according to Central Weather Bureau of Taiwan). Because of the distance between the NCU C-Pol and Typhoon Saomai (2006), the southwest (northeast) of the typhoon’s structure observed by the NCU C-Pol, was about 2–5 km (5–8 km) above MSL. The DSD for Typhoon Saomai could only be retrieved from those data below an altitude of 3.5 km and a correlation coefficient above 0.95 to avoid rain–ice mixing and ice phase data.

The retrieval of the DSDs for typhoon systems using polarimetric radar measurements via the constrained-gamma method requires a proper \( \mu-\Lambda \) relationship and DSRs for typhoon systems. The DSD retrieval was derived applying the typhoon DSR [Eq. (9)] and the empirical \( \mu-\Lambda \) relationship [Eq. (13)]. However, the accuracy of the DSDs retrieved from polarimetric measurements needed to be verified before using the retrieved DSDs to characterize typhoon systems. Thus, the polarimetric radar–retrieved DSDs were verified by the self-consistency of \( K_{DP} \) to ensure accuracy. The self-consistency was performed as follows: the retrieved DSDs from \( Z_H \) and \( Z_{DR} \) were first calculated to derive the specified differential phase shift \( K_{DP} \), which was then compared with observed \( K_{DP} \) (observed \( K_{DP} \) should be immune to system bias and attenuation and acknowledged as reference truth). Figure 12e presents the \( K_{DP} \) calculated from the observed differential phase shift \( \Phi_{DP} \) by using the algorithm from Hubbert and Bringi (1995). The values of \( K_{DP} \) calculated from the retrieved DSDs, via the measurements of \( Z_H \) and \( Z_{DR} \), are shown in Fig. 12f.

To quantify the improvement of the DSD retrieval, \( K_{DP-ty} \) —the \( K_{DP} \) forward calculated from the retrieved DSD via the \( \mu-\Lambda \) relationship [Eq. (13)] and the DSR [Eq. (9)] of the typhoon system (Fig. 12f)—and \( K_{DP-ZB} \) —the \( K_{DP} \) forward calculated from the retrieved DSD via the \( \mu-\Lambda \) relationship from Zhang et al. [2003; Eq. (12)] and DSR from Brandes et al. [2002; Eq. (8)]—were derived. The RMSE was introduced to evaluate the forward-calculated \( K_{DP} \) as follows:

\[
\text{RMSE} = \left[ \frac{\sum^N (K^{\text{DSD}} - K^{\text{obs}})^2}{N} \right]^{0.5}, \tag{14}
\]

where \( K^{\text{DSD}} \) represents the \( K_{DP} \) forward calculated from the retrieved DSD via two different sets of \( \mu-\Lambda \) relationship and DSR \( (K_{DP-ty} \text{ and } K_{DP-ZB}) \) and \( K^{\text{obs}} \) represents the \( K_{DP} \) observed by NCU C-Pol. The RMSE for \( K_{DP-ty} \) and \( K_{DP-ZB} \) were 0.3715 and 1.0755, respectively. The result indicates that the RMSE was greatly reduced by almost an order of 3 via applying the \( \mu-\Lambda \) relationship [Eq. (13)] and the DSR [Eq. (9)] of this typhoon dataset. That means that the retrieved DSDs were sufficiently accurate to characterize the typhoon systems.
The probability density function (PDF) of the DSDs retrieved from the measurements made by the NCU C-Pol radar in Typhoon Saomai (2006) was added in Fig. 7a (shading plot). In the stratiform region (rainfall rate less than 10 mm h\(^{-1}\)), the distribution of DSDs were as broad as those found in Typhoons Nari (2001) and Haima (2004). In the convective region (rainfall rate greater than 10 mm h\(^{-1}\)), the distribution of the retrieved DSDs from Typhoon Saomai (2006) revealed lower \(D_m\) and larger \(N_w\) than the disdrometer DSDs from Typhoons Nari (2001) and Haima (2004). This feature indicates that DSD retrieval from Typhoon Saomai (2006) on the ocean showed that the DSDs were closer to the maritime convective system. These differences between the DSDs illustrate that the typhoon system had different characteristics before and after landfall.

6. Conclusions

The main goal of this research is to explore the characteristics of DSDs and DSRs of typhoon systems. Through the understanding of DSDs and DSRs obtained from the 2D video disdrometer, this information was further extended to improve the accuracy of DSD retrieval from polarimetric measurements. Thus, the retrieval of DSD from a polarimetric radar can provide better spatial coverage and higher temporal resolution.

The results obtained from the time evolution of DSDs and reflectivity vertical profiles for Typhoons Nari (2001) and Haima (2004) indicate that there were three different types of precipitation systems: 1) weak stratiform precipitation (rainfall rate of less than 10 mm h\(^{-1}\) and a weak and shallow reflectivity vertical profile) had small \(D_m\) (1–1.5 mm) with smaller maximum diameter (3 mm) and a low concentration of small to medium raindrops; 2) stratiform precipitation (rainfall rate less than 10 mm h\(^{-1}\) and stronger reflectivity vertical profile than weak stratiform) with medium \(D_m\) (1.5–1.9 mm) had a maximum diameter of about 3–3.8 mm and a higher concentration of small to medium raindrops; and 3) convective precipitation (rainfall rate greater than 10 mm h\(^{-1}\)) had a higher \(D_m\) (greater than 1.9 mm) with a maximum diameter greater than 3.8 mm and the highest concentration of small to medium raindrops. Further investigation of the relation between the maximum altitude of 15-dBZ contours of typhoon systems \(H\), surface reflectivity \(Z\), and the mass-weighted diameter \(D_m\) showed a very good linear correlation. The \(D_m\) increased with corresponding increasing of \(H\) and \(Z\). The possible reason for the correlation is due to the fact that deeper systems may prepare an environment, which provides more opportunities for the collision–coalescence processes or the melting of snowflakes and graupel. However, more in situ measurements are needed to clarify just what dominating microphysical process is behind the unique characteristics of DSDs. After more long-term verification, the relationship of \(D_m-H-Z\) can be applied to quantitative precipitation estimations of conventional Doppler radars by providing information of \(D_m\). This application could reduce the uncertainty of the \(Z-R\) rainfall rate from the reflectivity because of the variability of the DSDs.

The effective DSR for typhoon systems obtained in a natural environment from the 2D video disdrometer observations was also examined. The fourth-order fitting DSR of typhoon systems had an axis ratio similar to that of Brandes et al. (2002) when the raindrops were less than 1.5 mm in size. Nevertheless, the axis ratio of raindrops tended to be more spherical when raindrops were greater than 2 mm and the coefficient \(\beta\) of a first-order fitting DSR for typhoons was about 0.0477. The value was also very close to the estimated coefficient \(\beta\) in Typhoon Saomai (2006) from polarimetric measurements. Moreover, a further comparison of the DSRs for different rainfall conditions indicates that raindrops were more spherical when there were high horizontal winds. The DSRs tended to be more oblate (spherical) with the increase in the rainfall rate in typhoon (non-typhoon) systems. The DSRs were primarily influenced by horizontal winds rather than the rainfall rate. Obviously, the rainfall rate and horizontal winds still cannot satisfactorily explain the variation between the DSR in typhoon and nontyphoon systems. Further sophisticated aerodynamics model study is needed to understand these factors.

From the analysis of the surface DSD observation, the values of \(N_w\) and \(D_m\) increased mainly with increase in the rainfall rate, when rainfall rate is less than 50 mm h\(^{-1}\). However, \(D_m\) tended to have a stable value of about 2.2 mm and \(N_w\) tended to increase with an increase of the rainfall rate when the rainfall rate was greater than 50 mm h\(^{-1}\). This conclusion was true for both Typhoons Nari (2001) and Haima (2004). Comparing the results from Bringi et al. (2003a) to this research, it can be seen that the typhoon system DSDs observed from NCU 2DVD were neither the typical maritime nor the continental types of convective systems. The distributions of \(D_m\) and \(N_w\) fell between the maritime and continental types of convective systems. The unique environment of the typhoon systems observed in the northern part of Taiwan, characterized by a terrain-influenced deep convective system, might be the reason for such characteristics.

On the contrary, the retrieved DSDs from NCU C-Pol polarimetric measurements for Typhoon Saomai
(2006) via the constraint-gamma method, with an empirical \( \mu - \Lambda \) relation and a fourth order fitting DSR of the typhoon systems, showed slightly different characteristics to the surface disdrometer observation. The accuracy of the retrieved DSDs had been verified by the self-consistency of \( K_{DP} \) and the results showed great improvement after applying the empirical \( \mu - \Lambda \) relation and the DSR of typhoon systems. In general, the retrieved DSDs from Typhoon Saomai (2006) reveal the similar DSD signatures of the maritime convective system from Bringi et al. (2003a) and Tokay et al. (2008). The results implied that the DSDs of the typhoon system on the ocean were characterized as a maritime convective system. In contrast, this study found that the DSDs of a landing typhoon system were very unique to the typhoon system observed on the ocean.

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