Microphysical Characteristics and Types of Precipitation for Different Seasons over North Taiwan

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Abstract

In this work, long-term (10 years) raindrop size distribution (RSD) measurements from the Joss–Waldvogel Disdrometer (JWD) installed at the National Central University (NCU) (24°58′6″N, 121°11′27″E), Taiwan, and the vertical profile of radar reflectivity were used to analyze the variations in the gamma parameters of six seasons (winter, spring, mei-yu, summer, typhoon, and autumn) and types of precipitation. The normalized gamma distribution of RSD revealed that the highest mean \(D_m\) (mass-weighted average diameter) values occurred in the summer, whereas the highest mean \(\log_{10} N_w\) (normalized intercept parameter) values were found in the winter. Furthermore, most of the rain falling at a rate of less than 20 mm h\(^{-1}\) occurs in Northern Taiwan. In this study, we used radar reflectivity to differentiate between convective and stratiform systems. It was revealed that the mean \(D_m\) values are higher in convective systems, whereas the mean \(\log_{10} N_w\) values are higher in stratiform systems. The structure of RSD in stratiform systems remains constant in all seasons; however, convection is similar
to maritime type. The microphysical characteristics that are responsible for the different RSD features in different seasons and types of precipitation are illustrated with the help of contoured frequency by altitude diagrams of radar reflectivity.

Keywords rain rate; raindrop size distribution; radar reflectivity

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1. Introduction

Raindrop size distribution (RSD) is a metric that is widely used in meteorology and hydrology (Calheiros and Zawadzki 1987), especially in understanding the precipitation microphysics (Rosenfeld and Ulbrich 2003). RSD varies with regard to the type of rain, location, and spatial distribution (Tokay and Short 1996; Bring et al. 2003; Gatlin et al. 2015). Thus, quantitative precipitation estimation (QPE) algorithms for radar measurements through radar reflectivity (Z) and rainfall rate (R) relations (Seliga and Bringi 1976; Ryzhkov and Zrnic 1995; Boodoo et al. 2015) and microphysical parameterization in WRF model (Gilmore et al. 2004; Cohen and McCaul 2006; Fadnavis et al. 2014; Wainwright et al. 2014; Tapiador et al. 2014; McFarquhar et al. 2015) strongly depend on the variations of the RSD (Lee and Zawadzki 2005; Chapon et al. 2008; Smith et al. 2009; Yoshioka et al. 2014), which means that it is important to analyze the RSD in various seasons as well as with various types of rain.

The size distribution of raindrops is the fundamental property of precipitation from the perspective of microphysical investigation and remote sensing using radars. In 1948, Marshall and Palmer (M-P; 1948) proposed the RSD. Ulbrich (1983) claimed that most observations did not fit the M-P method well, particularly when dealing with raindrops of a small size. Ulbrich and Atlas (1984) subsequently proposed using gamma distribution, which included \( N_0 \), \( \mu \), and \( \Lambda \). The unit \( N_0 \) represents the intercept as \( \text{m}^{-3} \text{mm}^{-1} \), \( \mu \) represents the shape, and \( \Lambda \) represents the slope of drop size distribution. Kozu and Nakamura (1991) reported that gamma distribution is more accurate than the M-P method in describing the characteristics of RSD. In the analysis of tropical rainfall events, Tokay and Short (1996) advanced the \( N_0 - R \) relationship to the classification of precipitation as convective and stratiform systems and also determined that most of the reflectivity in convective systems exceeds 40 dBZ. Moreover, they noticed that the effects of coalescence and evaporation are higher in convection systems than in stratiform systems, such that the gamma parameter \( (N_0) \) becomes smaller. By utilizing the data sets of the Joss–Waldvogel Disdrometer (JWD) and radars over Switzerland, Huggel et al. (1996) discovered that precipitation with a bright band is strongly correlated with large droplets and small \( N_0 \) and \( \Lambda \). Gamma parameters can be used to describe the RSD; however, the value of \( N_0 \) (\( \text{m}^{-3} \text{mm}^{-1} \)) is influenced by the value of \( \mu \). Testud et al. (2001) proposed the normalization of \( N_0 \), resulting in a redefinition of the intercept parameter, \( N_w \) (\( \text{m}^{-3} \text{mm}^{-1} \)). Bringi et al. (2003) used the method that was proposed by Testud et al. (2001) in the analysis of observational data from radars in various climatic regions. They demonstrated that \( D_m \) and \( N_w \) are strongly negatively correlated in stratiform systems, which may be due to the two characteristics of microphysical processes in stratiform systems. The first is an obvious bright band indicating the melting of large low-density snowflakes into rain, resulting in a large value of \( N_w \) and a small value of \( D_m \). The second is the lack of a bright band indicating the melting of small solid graupel and rimed snow particles into rain, causing a small value of \( N_w \) and a large value of \( D_m \). This phenomenon was actually called “\( N_0 \) jump” by Waldvogel (1974). Moreover, by collecting observational data, two types of convection systems, that is, a “maritime-like” system with large values of \( N_w \) and small values of \( D_m \) and a “continental-like” system with small values of \( N_w \) and large values of \( D_m \), were identified by Bringi et al. (2003).

Most previous research on RSD focused on different climatic regions worldwide (Bringi et al. 2003; Yin et al. 2011; Yin et al. 2013; Giangrande et al. 2014; Gatlin et al. 2015; Wang et al. 2015; Seela et al. 2017), seasons (Jayalakshmi and Reddy, 2014; Krishna et al. 2016; Wen et al. 2017), rainfall types (Tokay and Short, 1996; Tokay et al. 2008; Sharma et al. 2009;
Niu et al. 2010; Thurai et al. 2016), diurnal studies (Ushiyama et al. 2009; Suh et al. 2016), and case studies (Marzuki et al. 2010; Jung et al. 2012; Kumar and Reddy 2013; Kumari et al. 2014; Chen et al. 2016; Janapati et al. 2017; Zhang et al. 2017; Wen et al. 2017). In the region of Asia, comparing the gamma parameters of RSD ($N_w$ and $D_m$) with the center, leading, and trailing edge of the squall line, Chen et al. (2016) reported larger values of rain parameters in the central squall line. In addition, Zhang et al. (2017) noticed a distinct RSD in different squall-line stages and different precipitation types. In Southern Taiwan, Xu and Zipser (2015) concluded that heavy rainfall is mostly associated with a quasi-stationary mei-yu front and that upstream low-level jets and unstable upstream conditions constrained with moist neutral storm environments trigger heavy precipitation. Besides, storm initiation and evolution (Xu et al. 2012), as well as kinematic structures and microphysical characteristics (Chang et al. 2015; Jung et al. 2012), were observed by a Doppler radar. Chang et al. (2009) deduced a relationship between $D_m$ and height in typhoons using 2DVD and radar data during the period from 2001 to 2005. Chen (2013) analyzed the mei-yu front with southwesterly flow, which is responsible for flooding in Northern Taiwan.

So far, scarce research has been carried out on seasonal RSD over the Northwest Pacific region, especially over Taiwan, with long-term data. Hence, for the first time, using 10 years of RSD and radar reflectivity data, we tried to report the microphysical characteristics of different seasons (winter, spring, mei-yu, summer, typhoon, and autumn) over Northern Taiwan. Our findings provided an understanding of the seasonal variations in RSD and rain type and helped elucidate the factors affecting the RSD variations among the six seasons. The method and data used in this work are presented in Section 2. A comparison between the seasonal variation and precipitation type of microphysical characteristics is detailed in Section 3. Finally, a summary is provided in Section 4.

2. Data and methodology

Taiwan is located in a subtropical monsoon area in the Western Pacific Ocean, located off continental East Asia. Its climate is strongly influenced by the East Asian monsoon. The topography over Taiwan is dominated by a central mountain range, the north–south orientation of which causes large spatial variations in the island climate. The instruments (disdrometer, radar, and sounding data stations) used are shown in Fig. 1, and the methods used in the present work are outlined in this section. The rainy days that were recorded by the disdrometer during the passage of typhoons over Taiwan within Taiwan’s Central Weather Bureau (CWB) typhoon warning periods were considered as the typhoon (TY) season. The rainy days in December, January, and February were treated as the winter (DJF) season. Similarly, we considered the rainy days in March and April as the spring; in May and June as mei-yu; in July and August as the summer; and in September, October, and November as the autumn. In the current study, we ascertained that the disdrometer data does not include any solid precipitation in the winter (snow, graupel, etc.) or summer (hail) seasons. In order to avoid the solid precipitation from the analysis, we adopted the snowfall identification criteria of Chen et al. (2016) to the disdrometer data of winter seasons, and we found that there were no solid precipitation occurrences over

Fig. 1. Distribution of radar, disdrometer (JWD), radio sounding stations over Taiwan. Location of radar sites are marked with blue cross (“X”) [Wu-Fan San (RCWF), Hua-Lien (RCHL), Chi-Gu (RCCG), Ken-Ting (RCKT), Ma-Kung (RCMK) and Ching-Chuan-Kang (RCCK)]. JWD and automatic weather station of CWB (C0C520) are marked with red triangle (“▲”), and radiosonde launching stations are marked with black circle (“○”).
the observational site (National Central University: NCU) during the period from 2005 to 2014. More details were provided by Seela et al. (2018).

2.1 Calculation of drop size distribution

An outdoor RD-80 disdrometer was connected to an indoor analog-to-digital converter (ADA-90), which converts drop pulses into digital signals and then sends them to a personal computer. The disdrometer had a cross-sectional area of \( F = 0.005 \text{ m}^2 \). The range of drop sizes was divided into 20 categories, covering a range from 0.359 to 5.373 mm (listed in Table 1).

The JWD recorded the number of drops per minute in each category. The raindrop concentration, \( N(D_i) \), can be computed as follows:

\[
N(D_i) = \frac{n_i}{(F \times T \times V(D_i) \times \Delta D_i)},
\]

where \( n_i \) is the number of drops in the \( i \)th class, \( T \) is the sampling time (60 s), and \( V(D_i) \) is the terminal velocity of the raindrop \( D_i \) computed by Gunn and Kinzer (1949). As the location of the disdrometer (129 m a.s.l.) was not far above the sea level (Fig. 1), while computing the terminal velocity of the raindrops, we used the equation of Gunn and Kinzer (1949) without adopting any correction factors to the drop velocity equation.

The M-P formula can be used to convert the number of raindrops into drop size distribution \( N(D_i) \) by fitting the observational data to the gamma distribution (Kozu and Nakamura 1991) using \( N_0, \mu, \) and \( \Lambda \). The formula is derived by substituting the gamma distribution into the following equation:

\[
M_x = \int_0^\infty D^x N(D) dD,
\]

where \( x = 3, 4, \) and 6 for the 3rd \( (M_3) \), 4th \( (M_4) \), and 6th \( (M_6) \) moment of the drop size distribution, respectively. Parameter \( G \) can then be derived using the following equation:

\[
G = \frac{M_4 \mu}{M_3 \mu^2}.
\]

Equations (4), (5), and (6) are the three parameters of gamma distribution \( (\mu, \Lambda, \) and \( N_0) \). The RSD can be reconstructed using these three parameters, as follows:

\[
\mu = \frac{11G - 8 + \sqrt{G(G + 8)}}{2(1 - G)},
\]

\[
\Lambda = \frac{(\mu + 4)M_6}{M_4} = \frac{\mu + 4}{D_w},
\]
The radar reflectivity ($Z$, mm$^{-6}$ m$^{-3}$) and rainfall rate ($R$, mm h$^{-1}$) are computed using the following equations:

$$N_0 = \frac{\Lambda^{(\mu+4)} M_5}{\Gamma(\mu + 4)}, \quad (6)$$

$$D_0 = D_m \cdot \frac{3.67 + \mu}{4 + \mu}, \quad (7)$$

$$D_m = \frac{4 + \mu}{\Lambda}. \quad (8)$$

The radar reflectivity ($Z$, mm$^{-6}$ m$^{-3}$) and rainfall rate ($R$, mm h$^{-1}$) are computed using the following equations:

$$Z = \sum_{i=1}^{20} N(D_i) D_i^6 \Delta D_i, \quad (9)$$

$$R = 6\pi \times 10^{-4} \sum_{i=1}^{20} V(D_i) N(D_i) D_i^3 \Delta D_i, \quad (10)$$

$D_m$ (mm) (mass-weighted average diameter) can be calculated using the RSD. The three parameters in the gamma distribution can be used to provide an objective description of the RSD; however, $N_0$ ($m^{-3} \text{mm}^{2.5} \mu$) cannot be used in the same manner because it changes with the parameter $\mu$. $N_0$, a normalized parameter of $N_0$ that was proposed by Testud et al. (2001), will be used later on. In order to obtain reliable data, it was stipulated that any 1 min sample containing fewer than 10 drops or any 1 min sample with a rainfall rate of less than 0.5 mm h$^{-1}$ is regarded as noise and excluded. Limitations inherent to JWD observations leave the system susceptible to underestimating the number of small drops or omitting them altogether in cases of heavy precipitation. This situation is referred to as “dead time effect”. In this study, dead time correction was applied to the disdrometer data of Sauvageot and Lacaux (1995). In order to understand the data quality of the JWD, a scatter plot of JWD and CWB (C0C520) hourly accumulated rainfall for 10 years is shown in Fig. 2a. As the JWD can measure raindrops of up to 5.4 mm diameter, drops larger than 5.4 mm would be counted in the largest size bin, which leads to the underestimation of the rainfall rate in heavy-rain events (the data points above 45 mm h$^{-1}$ in Fig. 2a). This could be the possible reason for the variation between the rain gauge and disdrometer measurements (Tokay and Short 1996). Nonetheless, the linear fit applied to the scatter plot of the 10-year rainfall showed a high correlation coefficient between the JWD and CWB (C0C520) measurements. This clearly suggests that rain integral parameters derived from the JWD can be utilized in understanding the rainfall RSD characteristics of Northern Taiwan.

### 2.2 Radar data

The radar reflectivity mosaic used in the present study was deduced from six ground-based radars installed over Taiwan as shown in Fig. 1. Among the six radars, four Doppler radars, that is, Wu-Fan San (RCWF; 121.77°E, 25.07°N), Hual-Lien (RCHL; 121.619°E, 23.989°N), Chi-Gu (RCCG; 120.086°E, 23.1467°N), and Ken-Ting (RCKT; 120.849°E, 21.899°N), which have a beam width of 1° and wavelength of 10 cm (S-band), belonged to the CWB. Chang et al. (2009) mentioned more detailed information regarding these radars. The other two radars located at Ma-Kung (RCMK; 119.634°E, 23.563°N)
and Ching-Chuan-Kang (RCCK; 120.63°E, 24.25°N) are owned by Taiwan Air Force. The Nyquist velocity of RCMK and RCCK was 37.18 and 37.15 m s\(^{-1}\), respectively. The radar data from QPESUMS (from six operational Doppler radar units) can be used to differentiate between the various types of rain. Observations of reflectivity from the individual radars were combined to generate three-dimensional reflectivity mosaic grids (Zhang et al. 2005). These mosaic grids had a spatial resolution of 0.0125° on the latitude–longitude coordinate system and a 10 min update cycle. The degree of reflectivity calculated from the JWD was positively correlated with the data obtained from QPESUMS (Fig. 2b). The QPESUMS reflectivity is the mosaic of six ground-based radars. For further details on the QPESUMS reflectivity quality check, please refer to Zhang et al. (2009). As the reflectivity profiles are available with high-resolution (0.0125°) mosaic grids, the 10 min reflectivity profile over the NCU was obtained by averaging the profiles over the location of the NCU (24.55–24.6°N, 121.0875–121.1375°E). The reflectivity plotted on the ordinate of Fig. 2b is the lowest-level (around 1 km) reflectivity of QPESUMS with grid averaging over the NCU. By considering nearly 1 km height reflectivity with grid averaging, we could notice the discrepancy between the disdrometer-measured reflectivity and QPESUMS reflectivity. However, most of the data fell within two standard deviations of the regression line. This data was used as the vertical structure of reflectivity, which can be provided by contoured frequency by altitude diagrams (CFADs). The location of the NCU was the center of the domain (5 × 5 grid points) used to calculate the CFADs.

### 2.3 Classification of stratiform and convective precipitation

Various methods were used to classify stratiform and convective precipitation. Waldvogel (1974) found that \(N_0\) undergoes dramatic changes when the precipitation system changes. This phenomenon is referred to as the \(N_0\) jump. Gamache and Houze (1982) used a reflectivity threshold of 38 dBZ to differentiate between the types of precipitation. Tokay and Short (1996) found that the \(N_0–R\) distribution may change with the precipitation system, whereas Churchill and Houze (1984) used a 40 dBZ threshold to separate the precipitation into stratiform and convective type. Tokay et al. (1999) designated a rainfall rate exceeding 10 mm h\(^{-1}\) as convective precipitation. Bringi et al. (2003) applied a standard deviation smaller than 1.5 mm h\(^{-1}\) to stratiform precipitation and a standard deviation greater than 1.5 mm h\(^{-1}\) to convective precipitation. Using a dual-polarization radar and a dual-frequency profiler, Bringi et al. (2009) developed a procedure for distinguishing the precipitation into stratiform and convective type, and they deduced an equation (\(\log_{10} N_0 = -1.6D_T + 6.3\)) that separates the stratiform and convective regimes. Chang et al. (2009) reported a rainfall rate threshold of 10 mm h\(^{-1}\) to differentiate between the various types of precipitation. In this study, the method developed by Steiner et al. (1995) was adopted for the differentiation of stratiform and convective systems.

### 2.4 Sounding data

Sounding data plays an important role in defining the melting level and altitude where the temperature drops to \(-40^\circ\text{C}\). The radio sound data available twice daily (00:00 UTC and 12:00 UTC) for Banqiao (121.4420°E, 24.9976°N) and Hua-lien (121.6133°E, 23.9751°N) meteorological observation stations for the period from 2005 to 2014 were used in the present research. These meteorological stations are shown as black open circles in Fig. 1. The temperature profile from the radiosonde data was used to calculate the mean elevations at which the temperature dropped to \(0^\circ\text{C}\) over a period of 10 years.

### 2.5 ERA-Interim and remote sensing data

Along with the disdrometer and radar data, ERA-Interim, Moderate Resolution Imaging Spectroradiometer (MODIS), and Tropical Rainfall Measuring Mission (TRMM) data sets were used for the disdrometer-measured rainy days during a recording period of 10 years over the observational site (24–25.2°N, 121–121.9°E). The convective available potential energy (CAPE), vertical integral of water vapor (VISWV) from ERA-Interim, cloud effective radii (CERs) of ice and liquid particles from MODIS, and storm and bright band heights from TRMM were used to illuminate the causes for microphysical and thermodynamic discrepancies among the six seasons. For further explanation regarding these data sets, please refer to Seela et al. (2018).

### 3. Results and discussion

#### 3.1 Climatology of raindrop size distribution

Chen and Chen (2003) reported that monsoons greatly influence precipitation over Taiwan. In order to understand the rainfall patterns in the six seasons over Northern Taiwan, the rate of rainfall and radar reflectivity from the JWD are shown in Fig. 3. The frequency of rainfall occurrence and radar reflectivity
are represented using a logarithmic scale in Figs. 3a and 3b, respectively. As shown in Fig. 3a, the rate of rainfall is generally less than 15 mm h$^{-1}$. The most extensive rainfall rate distribution was observed during the typhoon and mei-yu seasons, whereas the narrowest rainfall rate distribution was observed during the winter. Results pertaining to reflectivity (Fig. 3b) were consistent with the rainfall rate distribution (i.e., the rainfall during the typhoon and mei-yu seasons tended to be heavier). The mei-yu front and orographic rain are the main sources of rainfall during the mei-yu season, and both are influenced by the southwest monsoon and subsynoptic scale of the disturbance from the Tibetan Plateau, causing considerable rainfall on the windward side of the western parts of the central mountain range. Typhoons and convective systems combined with the southwest monsoon bring about large amounts of rainfall in the summer (Chen and Chen 2003). These results indicate that Northern Taiwan has different precipitating cloud systems in different seasons, which are the major sources of QPE uncertainty through $Z–R$ relations. Figure 4 presents the RSD in the six seasons. After applying quality control to the RSD of the six seasons, a nearly equal number of samples for each season, except for the summer, are described in the legend of Fig. 4. A larger number of large drops and fewer small drops were observed during the typhoon, mei-yu, and summer seasons. The distribution in winter was closer to exponential, with a higher number of smaller drops compared to any other season. As most of the precipitating clouds in the winter are of a shallow type (Seela et al. 2018), winter has a greater number of small drops compared to other seasons through collision–coalescence processes. In every season, raindrops were observed with a diameter ranging from 0.3 mm to the limit of the instrument at 5.4 mm. Table 2 provides the occurrence percentage of gamma parameters ($D_w$, $N_w$, $\mu$, and $\lambda$) over a period of 10 years (combining all seasons). The formula for skewness (SK) is as follows: 

$$SK = \frac{E(x - \mu)^3}{\sigma^3},$$

where $\mu$ is the mean value of $x$, $\sigma$ is the standard deviation, and $E$ is the expectation of $(x - \mu)$. 

![Fig. 3. Overview of (a) rate of rainfall (mm h$^{-1}$) and (b) radar reflectivity (dBZ) calculated from JWD in different seasons over a period of ten years. The color bar represents the occurrence frequency in log scale.](image1)

![Fig. 4. Raindrop concentration [log$_{10}$ N(D), m$^{-3}$ mm$^{-1}$] vs. raindrop diameter (D, mm) for six seasons over a period of ten years. The number in the legend represents data points of 10-min sampling.](image2)
The occurrence percentage of gamma parameters quantitatively defines the RSD in the rainfall in Northern Taiwan using the following: \( D_m \) (1.16 mm), \( \mu \) (5.60 [-]), \( \lambda \) (9.82 mm\(^{-1}\)), and \( \log_{10} N_w \) (4.36 m\(^{-3}\) mm\(^{-1}\)). Precipitation in Northern Taiwan shows larger numbers of small- and medium-sized drops than what was observed in tropical Malaysia (Lam et al. 2015) and China (Wen et al. 2017), but larger drops and normalized concentration in Palau (Seela et al. 2017). Over Southern India, the \( D_m \) (\( \log_{10} N_w \)) values in the southwest monsoon (June–September) were found to be larger (smaller) than in the northeast monsoon (October–December) (Jayalakshmi and Reddy 2014). The climatological \( D_m \) value of Northern Taiwan was found to be higher (lower) compared to the northeast (southwest) monsoon rainfall of Southern India, whereas the climatological \( \log_{10} N_w \) values were larger compared to both southwest and northeast monsoon rainfall of Southern India.

Figures 5a and 5c present the normalized intercept parameter (\( N_w \)), mass-weighted mean diameter (\( D_m \)) (from the JWD), and CFADs of reflectivity (from the radar) from observations over a period of 10 years. Current \( D_m-N_w \) distributions are compared with the maritime and continental clusters (red- and green-colored rectangular boxes in Fig. 5a) of Bringi et al. (2003), and \( D_0-N_w \) distributions are compared with the stratiform and convective separation line (black dotted line in Fig. 5b) of Bringi et al. (2009). From Fig. 5a, it is apparent that the climatological \( D_m-N_w \) parameters of the six seasons are close to those of “maritime-like” systems. Moreover, Fig. 5b shows that most of the rain appeared to be stratiform, which is consistent with Fig. 3 and Table 2. The vertical structure of precipitation observed by the radar is related to CFADs (Rudolph and Friedrich 2014). The climatology of CFAD (Fig. 5c) depicts the deeper convection of 30 dBZ up to 5.7 km, with a spread of 7–37 dBZ below the melting level. It also displays the mean profile of vertical reflectivity (vertical white stars line) in combination with the melting level (horizontal white dotted line) calculated using sounding data. These results

| Parameter | Mean | SD  | Skewness | Kurtosis |
|-----------|------|-----|----------|----------|
| Normalized | \( \log_{10} N_w \) | 4.36 | 0.61 | 0.21 | 3.36 |
| Gamma | \( D_m \) | 1.16 | 0.41 | 0.76 | 4.72 |
| | \( \mu \) | 5.60 | 4.2 | 1.85 | 10.07 |
| | \( \lambda \) | 9.82 | 7.31 | 2.35 | 9.69 |

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Fig. 5. (a) Distribution of \( \log_{10} N_w \) (m\(^{-3}\) mm\(^{-1}\)) and \( D_m \) (mm) over period of ten years (all seasons). Green and red rectangular boxes represent continental-like and maritime-like convection, respectively. (b) Percentage distribution of \( D_0-\log_{10} N_w \) for all seasons with stratiform and convective separation line (inclined black dashed line) of Bringi et al. (2009). (c) Radar reflectivity of CFADs over period of ten years (all seasons) with mean reflectivity profile of six seasons in white star dotted line. The horizontal white dotted line represents the mean melting layer height of six seasons obtained from radiosonde.
show that the mean melting level is approximately at 4.2 km and the reflectivity decreases with an increase in elevation, thereby indicating that the particles are smaller and the icing process is more pronounced at higher elevations. Thus, the degree of tilting (−3.74 dBZ km\(^{-1}\) in climatology) around the melting level can be regarded as the intensity of vertical motion of hydrometeors associated with various microphysical processes. The accumulation of ice crystals at higher elevations causes a decrease in reflectivity with height. These outcomes are used as a reference in the discussion of seasonal variations in Section 3.2.

### 3.2 Seasonal variation

Thompson et al. (2015) proposed a conceptual model to describe microphysical processes using a scatter plot of \( D_m \) and \( N_w \). In both stratiform and convective precipitation, an increase of the liquid water content (LWC) with the increase in drop size was shown to enhance the process of icing. Conversely, weak convection with small drops was dominated by condensation and coalescence. Fig. 6a provides a scatter plot of \( D_m \) and \( N_w \) for the six seasons. Fig. 6b depicts the difference between the seasonal mean and the annual mean values of \( D_m \) and LWC for the six seasons. From Fig. 6a, it is apparent that, except for summer season, the remaining five seasons had mean \( \log_{10} N_w (D_m) \) values around (smaller than) maritime convective clusters of Bringi et al. (2003). Deviations in LWC and \( D_m \) can be seen in Fig. 6b, indicating that the ice processes differ according to the season. The distribution of \( D_m \) and \( N_w \) values in the six seasons (Fig. 6c) was shown to be below the stratiform–convective separation line of Bringi et al. (2009) (the black dotted line in Fig. 6c). This confirms that stratiform-type precipitation is prominent in all seasons, which is consistent with Fig. 3. In order to more fully elucidate the cloud formation and rain processes in each season, the CFADs of the six seasons were constructed in Fig. 7. On the basis of radar reflectivity of CFADs, the six seasons were then classified as high vertical structures (HVSs) and low vertical structures (LVSs). The mei-yu, summer, and typhoon seasons (Figs. 7c–e) are associated with HVSs, whereas winter, spring, and autumn (Figs. 7a, b, f) are associated with LVSs. The mean melting elevation associated with strong icing conditions is approximately 5 km for the HVS seasons (Figs. 7c–e), whereas the mean melting elevation associated with weak icing conditions is 4 km for LVS seasons (Figs. 7a, b, f).

The mean reflectivity profile (vertical red stars line) in each HVS season (Figs. 7c–e) is greater than the annual mean reflectivity profile (vertical white stars line). In contrast, the mean reflectivity profile in each LVS season (Figs. 7a, b, f) is lower than the annual mean reflectivity profile. This shows that HVS seasons were much warmer and humid, accelerating the growth of raindrops, compared to LVS seasons. Furthermore, the tilting of the LVS (HVS) around the melting level was about −4.50, −4.80, and ‐4.10 dBZ km\(^{-1}\) (−3.80, −2.75, and ‐1.68 dBZ km\(^{-1}\)) for winter, spring, and autumn (mei-yu, summer, and typhoon), respectively. The tilting of LVS or HVS around the melting level indicates the various mi-
crophysical characteristics that were affected by the vertical motions. In HVSs, the reflectivity profile of summer had 1% of 30 dBZ up to 10 km and a spread of reflectivity from 10 to 46 dBZ below the melting level (Fig. 7d). This shows that deeper convection occurs in the summer compared to any other seasons of HVSs. The 30 dBZ reflectivity is extended up to 7 km in the mei-yu season, with a spread of 10–45 dBZ below the melting layer (Fig. 7c), whereas in the typhoon season, the 30 dBZ reflectivity (~1%) is extended up to 7.7 km, and a spread of 10–40 dBZ can be seen below the melting layer. Although the well-organized is observed in the typhoon than mei-yu season, strong wind shear might break the raindrops to a smaller size below the melting layer, which results in relatively less spread of reflectivity (below the melting layer) in the typhoon than in the mei-yu season. The CFADs of HVSs clearly show that larger drops appeared in the summer, with a high concentration in the mei-yu season.

In LVSs, the reflectivity profile of winter had 1% of 30 dBZ up to 4.9 km, with a spread of 9–35 dBZ below the melting level (Fig. 7a). These consequences indicate that shallower convection takes place in the winter compared to any other seasons of LVSs. The reflectivity profile of spring and autumn had 1% of 30 dBZ up to 5.5 km and a spread of 5–39 dBZ below the melting level. Even though the spring and autumn appeared to have similar vertical structures, the different melting layer heights between these two seasons were responsible for the dissimilar results in RSD. According to Chen and Chen (2003), spring is a transitional period when the southwestern flow gradually increases, bringing warm air from the South China Sea above 850 hPa, which often meets a cold front from the leeward side of the Tibetan Plateau. This causes a transition from steady rainfall in the winter to convective-type rainfall in the spring. Autumn is not the rainiest season in Taiwan; however, typhoons and frontal systems during this period bring about considerable rainfall. Following the onset of the northeast monsoon, precipitation increases greatly on the windward side (over Northern and Northeastern Taiwan) and decreases greatly on the leeward side (Chen and Chen 2003). During this period, rainfall patterns change from convective to stratiform. However, both of these seasons are transitional, and the atmospheric condition alters from warm to cold (in the autumn) or from cold to warm (in the spring), which adjusts the height of the melting level. Both of the results in HVS and LVS correspond to the variation of seasonal RSD over Northern Taiwan. To sum up, vertical development is the dominant factor in RSD. In a later section, CFADs will be used to examine the types of rain in
In order to further emphasize the dynamic and thermodynamic differences among the six seasons, the convective available energy (CAPE, J kg\(^{-1}\)) and vertical integral water vapor (kg m\(^{-2}\)) values derived for the six seasons as well as for the HVS and LVS seasons are depicted in Fig. 8. The CAPE and water vapor values are taken from ERA-Interim for the disdrometer measurement site for the rainy days of each season. From the figure, it can be clearly seen that higher CAPE and water vapor values can be seen for the summer and typhoon seasons, followed by mei-yu and autumn. Among the six seasons, winter and spring have the lowest CAPE and water vapor values. The higher the CAPE value, the greater the convective activity with vigorous updrafts and downdrafts. Clouds with higher convective activity (or higher CAPE values) reach deeper altitudes than clouds with lower CAPE values. This was confirmed by the higher storm and bright band heights for the summer, typhoon, and mei-yu seasons compared to other seasons (Fig. 9). Figure 9a illustrates that the LVS (winter, spring, and autumn) seasons are typically accompanied with clouds of storm heights less than 4 km. In contrast, the HVS (summer, typhoon, and mei-yu) seasons have clouds with storm heights higher than 4 km. With
higher storm and bright band heights and intense convection (severe updrafts and downdrafts), clouds in the summer, typhoon, and mei-yu seasons are associated with larger ice and liquid particles compared to the winter, spring, and autumn seasons. This feature can be seen with higher CERs of ice and liquid particles in the HVS (summer, typhoon, and mei-yu) seasons compared to the LVS (winter, spring, and autumn) seasons (Fig. 10). The CER values of ice and liquid particles for the six seasons were obtained from the MODIS cloud data product for the disdrometer observational site. As the storm heights of the HVS seasons are greater than the mean zero-degree isotherm height of the six seasons (4.3 km computed from the radiosonde), the clouds in the HVS seasons were associated with cold rain processes. Conversely, most of the precipitating clouds in the winter, spring, and autumn had storm heights less than 4 km, and this implies that the clouds in the LVS seasons are mostly associated with warm rain processes, in which collision–coalescence is the dominant process. On the other hand, the HVS seasons had greater bright band heights, with a median value greater than 4.5 km. The LVS seasons had lower bright band heights, with a median value less than 4.5 km. The melting layer is considered as the transition zone for ice and rain processes, above which supercooled water and ice particles can be found. HVS seasons (summer, mei-yu, and typhoon) are associated with a warmer environment and stronger vertical motion (Fig. 8a), which means that a greater volume of water vapor (Fig. 8b) can be carried up to higher elevations, thereby increasing the size of the drops. As the equilibrium vapor pressure of water vapor with respect to ice is less than that with respect to liquid, at a given subfreezing temperature, the ice crystals initially formed through freezing of liquid droplets or by sublimation of water vapor will further grow to a larger size by vapor deposition at the expense of the supercooled water droplets (Rodgers and Yau 1989; Pruppacher and Klett 1997). When ice crystals pass through the melting layer (whereupon aggregation and accretion change to collision and coalescence), relatively large ice particles in the HVS seasons melt below the melting layer, resulting in larger drops at the ground compared to the LVS seasons.

Table 3 displays the mean, standard deviation, skewness, and kurtosis values of each parameter in the six seasons (winter, spring, mei-yu, summer, typhoon, and autumn). Using these gamma parameters (Table 3) allows reconstructing the RSD in gamma distribution form, which is similar to that in Fig. 4 but slightly smoother. From Table 3, it can be noticed that the summer has the maximum mean \( D_{\text{m}} \) value, followed by the typhoon, mei-yu, spring, autumn, and winter seasons. On the other hand, the winter has the maximum mean \( \log_{10} N_w \) value, followed by the autumn, spring, mei-yu, typhoon, and summer seasons. A very low positive skewness value for \( \log_{10} N_w \) in the winter indicates that most of the \( \log_{10} N_w \) values are distributed near its mean value. Besides, a comparatively lower (higher) mean \( D_{\text{m}} \) (\( \log_{10} N_w \)) value with very low positive skewness in the winter compared to other seasons specifies that the winter has a very large number of small drops compared to other seasons, which is typically accompanied by collision–coalescence and breakup processes, which are affirmed by...
lower storm heights (Fig. 8) and mean vertical profiles (Fig. 7a). The negative skewness of \( \log_{10} N_w \) in the summer depicts that most of the population is distributed to the right side of its mean. A higher mean \( D_m \) value in the summer compared to the rest of the seasons with negative skewness for drop concentration (\( \log_{10} N_w \)) confirms that the summer has a higher concentration of large drops compared to the rest of the seasons. Among the six seasons, the highest (lowest) mean \( D_m \) (\( \log_{10} N_w \)) values in the summer illustrate that the summer is accompanied by enhanced vapor deposition and aggregation above the melting layer, as well as collision–coalescence and evaporation processes below the melting layer, which are strongly supported by higher CAPE values (Fig. 8) and higher mean reflectivity profiles with a deeper extent of 30 dBZ reflectivity in the CFAD (Fig. 7d). This clearly demonstrates that more large drops are observed in the summer, whereas more small drops are observed in the winter. This result is in accordance with the previous study by Seela et al. (2018). In HVS seasons, the sequence of the mean \( D_m \) values (highest in the summer, followed by the typhoon and mei-yu seasons) and \( \log_{10} N_w \) (lowest in the summer, followed by the typhoon and mei-yu seasons) values are in accordance with their relative melting layer heights (Fig. 7), convective activity (Fig. 8), degree of mean reflectivity tilting around the melting layer, and differences in the icing process (Fig. 6b). Similarly, in LVS seasons, the order of \( D_m \) (highest in the spring, followed by the autumn and winter) and \( \log_{10} N_w \) (highest in the winter, followed by the autumn and spring) values is in agreement with the relative extent of clouds (Fig. 9), degree of mean reflectivity tilting (Fig. 7), and \( D_m \) and LWC distributions (Fig. 6b). The relatively lower mean slope parameter (\( \lambda \)) in the summer, typhoon, and mei-yu seasons compared to the rest of the seasons (winter, autumn, and spring) with higher positive skewness values confirmed that the breadth of the drop size distribution in the summer, typhoon, and mei-yu seasons is larger compared to the winter, autumn, and spring seasons, and this characteristic can be seen in Fig. 4.
3.3 Rainfall type in different seasons

As the RSD is strongly influenced by the type of rain, it is of paramount importance to investigate the stratiform and convective forms of precipitation. In the current study, the six seasons’ rainfall was classified into stratiform or convective type by implementing the reflectivity classification technique developed by Steiner et al. (1995).

a. Stratiform

The gamma parameters of the six seasons’ stratiform precipitation are provided in Table 4. The $D_m$ and $\mu$ values in the HVS seasons exceeded those in the LVS seasons; however, the values of $N_w$ and $\lambda$ were lower. The mean $D_m$ values in the stratiform precipitation of HVS seasons are higher compared to the LVS seasons, with the highest mean $D_m$ in the summer and the lowest in the winter. Further, the mean $\log_{10} N_w$ values in the stratiform precipitation of HVS seasons are lower compared to the LVS seasons, with the highest mean $\log_{10} N_w$ in the winter and the lowest in the summer. The comparatively low positive skewness values of $\log_{10} N_w$ in the LVS seasons indicated that most of the $\log_{10} N_w$ populations are distributed to the left side of their mean values. On the other hand, the negative skewness values of $\log_{10} N_w$ in the HVS seasons indicated that most of the $\log_{10} N_w$ values tended to be distributed to the right side of their mean values.

With higher (lower) mean $D_m$ ($\log_{10} N_w$) values and negative skewness values for $\log_{10} N_w$, the stratiform precipitation of the HVS seasons is associated with a large number of bigger drops compared to the LVS seasons. Because of the quantitative gamma parameters, the RSD can be constructed in gamma form (Fig. 11). From Fig. 11, it can be seen that, for every season, the drop diameter was shown to range between 0.3 and 4.2 mm, whereas the raindrop concentration ranged from 100 to 10,000 m$^{-3}$ mm$^{-1}$, and the mean rate of rainfall ranged from 3.46 to 7.05 mm h$^{-1}$. Moreover, the widest RSD was observed in the summer, which means that the summer is associated with a greater proportion of large drops than small drops.

Figure 12a presents the scatter plot for $\log_{10} N_w$ and

| Parameter | Mean | SD  | Skewness | Kurtosis |
|-----------|------|-----|----------|----------|
| Winter    |      |     |          |          |
| $\log_{10} N_w$ | 4.21 | 0.41| 0.29     | 3.37     |
| $D_m$     | 1.14 | 0.27| 0.51     | 3.90     |
| $\mu$     | 3.51 | 2.32| 0.53     | 3.34     |
| $\lambda$ | 6.87 | 2.59| 0.9      | 3.77     |
| Spring    |      |     |          |          |
| $\log_{10} N_w$ | 4.03 | 0.43| 0.15     | 4.33     |
| $D_m$     | 1.31 | 0.34| 0.83     | 4.34     |
| $\mu$     | 3.81 | 2.49| 1.43     | 10.10    |
| $\lambda$ | 6.33 | 2.41| 0.84     | 3.712    |
| Mei-yu    |      |     |          |          |
| $\log_{10} N_w$ | 3.95 | 0.43| -0.09    | 3.75     |
| $D_m$     | 1.41 | 0.33| 1.01     | 6.11     |
| $\mu$     | 4.58 | 2.24| 0.83     | 5.18     |
| $\lambda$ | 6.43 | 2.3 | 0.81     | 3.72     |
| Summer    |      |     |          |          |
| $\log_{10} N_w$ | 3.66 | 0.46| -0.65    | 4.75     |
| $D_m$     | 1.59 | 0.43| 1.56     | 8.31     |
| $\mu$     | 4.76 | 2.53| 0.8      | 4.88     |
| $\lambda$ | 6.03 | 2.68| 0.84     | 4.41     |
| Typhoon   |      |     |          |          |
| $\log_{10} N_w$ | 3.83 | 0.37| -0.47    | 5.34     |
| $D_m$     | 1.47 | 0.3 | 0.98     | 6.20     |
| $\mu$     | 5.92 | 3.06| 1.12     | 6.57     |
| $\lambda$ | 6.98 | 2.36| 0.62     | 3.71     |
| Autumn    |      |     |          |          |
| $\log_{10} N_w$ | 4.09 | 0.48| 0.11     | 3.06     |
| $D_m$     | 1.28 | 0.34| 0.78     | 4.45     |
| $\mu$     | 4.26 | 2.28| 0.71     | 4.34     |
| $\lambda$ | 6.84 | 2.48| 0.8      | 3.61     |
$D_m$ associated with stratiform precipitation for the six seasons. The climatological CFAD of the six seasons’ stratiform precipitation is depicted in Fig. 12b, and the distribution of $\log_{10} N_w$ and $D_0$ for stratiform precipitation of the six seasons is shown in Fig. 12c. The distribution of these $D_m$ and $N_w$ values (Fig. 12a) is consistent with the climatological distribution $D_m-N_w$ (as shown in Fig. 6), which emphasizes that most of the rainfall in Northern Taiwan is maritime-like. However, the stratiform $D_0-log_{10} N_w$ values in each season (Fig. 12c) run parallel to the stratiform–convective separation line defined by Bringi et al. (2009). As mentioned in the previous section, understanding the precipitation vertical structure by figuring the CFAD in stratiform type (Fig. 12b) is the method of knowing the RSD. Its reflectivity had 1% of 30 dBZ up to 5.5 km, with a spread of 8–35 dBZ below the melting level. Furthermore, the slope around the melting level was $-3.58$ dBZ km$^{-1}$.

Generally, a vertical structure in stratiform is weaker than in climatology. The stratiform precipitation CFADs of the six seasons are depicted in Fig. 13. The mean reflectivity profile of stratiform type (vertical red stars line) is greater than the annual mean reflectivity profile (vertical white stars line) in each HVS (Figs. 13c–e) season, and the mean reflectivity profile (vertical red stars line) of each LVS season (Figs. 13a, b, f) is smaller than the annual mean reflectivity profile (vertical white stars line), which corresponds with the climatology. Furthermore, the tilting of the LVS seasons around the melting level was about $4.65$, $-4.54$, and $-3.77$ dBZ km$^{-1}$ for the winter, spring, and autumn, respectively. Moreover, the tilting of the HVS seasons can also be quantified about $-3.74$, $-2.55$, and $-1.44$ dBZ km$^{-1}$ for the mei-yu, summer, and typhoon seasons, respectively. The different tilting values of the six seasons are an indication of the various microphysical characteristics among these seasons.

In HVS, the reflectivity of stratiform precipitation in the summer had 1% of 30 dBZ up to 7.5 km, with a spread of 6–40 dBZ below the melting level (Fig. 13d). In addition, mei-yu had 1% of 30 dBZ up to only 6 km, with a spread of 5–36 dBZ below the

![Fig. 11. Gamma distribution using the mean values of stratiform RSD in different seasons. The mean rate of rainfall is also shown in the legend.](image)

![Fig. 12. (a) Mean and standard deviation values of $D_m$ (mm) and $log_{10} N_w$ (m$^{-3}$ mm$^{-1}$) for stratiform precipitations of six seasons. (b) Total mean CFADs of stratiform precipitation of six seasons. The mean melting layer height of six seasons is represented with horizontal white dotted line and the mean vertical profiles of reflectivity for six seasons is represented with vertical white stars line. (c) $D_0-log_{10} N_w$ distribution for stratiform precipitations of six seasons with stratiform and convective separation line (inclined black dashed line) of Bringi et al. (2009).](image)
melting level (Fig. 13c). However, the deepest slope still occurred in the typhoon season (Fig. 13e), with 1% of 30 dBZ up to 7.5 km and a spread of reflectivity of 6–38 dBZ below the melting level. In general, the growth of RSD in the summer stratiform is larger than in the rest of the HVS stratiform precipitations. In LVS, the reflectivity of stratiform precipitation in the winter had 1% of 30 dBZ up to 4.8 km, with the distribution of reflectivity varying from 7 to 34 dBZ below the melting level (Fig. 13a). Although the stratiform precipitation CFADs of the spring (Fig. 13b) and autumn (Fig. 13f) show a similar mean reflectivity profile, they have different heights of 30 dBZ reflectivity extents. The 1% of 30 dBZ is extended to 5.5 km and 5.2 km in the autumn and spring, respectively. Moreover, the characteristic of the distribution of reflectivity below the melting level is slightly not the same for the autumn and spring. A comparison of CFADs in the summer (Fig. 13d) and winter (Fig. 13a) revealed a distinct difference in the vertical structure of reflectivity. As reported by Chen and Chen (2003), the weather condition in the winter leads to less rainfall and small drops. In contrast, a large amount of water vapor from the southwest monsoon and incoming solar radiation produce strong vertical motion in the summer. This tends to exacerbate the microphysical process, resulting in larger drops in the summer.

b. Convection

Table 5 lists the gamma parameters associated with the convective precipitation of the six seasons as well as the reconstructed RSD in these seasons. Analogous to the stratiform precipitation, the convective precipitation of the HVS seasons had higher (lower) mean $D_m$ ($\log_{10} N_w$) values compared to the LVS seasons. With resemblance to the stratiform precipitation, the negative skewness values of $\log_{10} N_w$ in the convective precipitation of the HVS seasons indicated that most of the $\log_{10} N_w$ values tended to be distributed to the right side of their mean values. In contrast, the low positive skewness values of $\log_{10} N_w$ in the convective precipitation of the LVS seasons indicated that most of the $\log_{10} N_w$ populations are distributed to the left side of their mean values. The lower (higher) mean $\log_{10} N_w$ ($D_m$) values and negative skewness values for $\log_{10} N_w$ in the convective precipitation of the HVS seasons indicated that most of the $\log_{10} N_w$ populations are distributed to the left side of their mean values. The lower (higher) mean $\log_{10} N_w$ ($D_m$) values and negative skewness values for $\log_{10} N_w$ in the convective precipitation of the HVS seasons confirmed that they are associated with a large number of bigger drops compared to the LVS seasons. The convective precipitation RSDs of the six seasons are depicted in Fig. 14, which illustrates that the maximum diameter varies from 4.5 mm up to the limitation.
of the instrument (5.3 mm), \( N(D) \) ranges from \( 10^2 \) to \( 10^5 \ \text{m}^{-3} \ \text{mm}^{-1} \), and the mean rate of rainfall ranges from 4.33 to 14.52 mm h\(^{-1}\). A larger number of large drops are observed in the summer than in the winter (Fig. 14), which is in general agreement with the results in Fig. 4 and Fig. 11. For the given same size of raindrops (i.e., 3 mm), the number concentration in the convection type is larger than in the stratiform type. These results indicate a wider RSD in the convection type of rainfall compared to the stratiform type.

Table 5. Statistics of convection RSD parameters derived from disdrometer data (Jan 2005- Dec 2014, 10-min rain data) in different seasons in Northern Taiwan.

| Season   | Parameter | Mean | SD  | Skewness | Kurtosis |
|----------|-----------|------|-----|----------|----------|
| Winter   | \( \log N_w \) | 4.26 | 0.5 | 0.21 | 4.05 |
|          | \( D_m \)  | 1.24 | 0.36 | 0.77 | 5.01 |
|          | \( \mu \)   | 3.34 | 2.08 | 0.45 | 3.51 |
|          | \( \lambda \) | 6.43 | 2.78 | 1.29 | 4.65 |
| Spring   | \( \log N_w \) | 3.99 | 0.44 | 0.23 | 5.62 |
|          | \( D_m \)  | 1.45 | 0.38 | 0.49 | 4.37 |
|          | \( \mu \)   | 3.7  | 2.24 | 2.05 | 14.77 |
|          | \( \lambda \) | 5.73 | 2.37 | 1.36 | 5.28 |
| Mei-yu   | \( \log N_w \) | 3.96 | 0.36 | -0.22 | 4.04 |
|          | \( D_m \)  | 1.58 | 0.35 | 0.08 | 2.88 |
|          | \( \mu \)   | 4.64 | 2.23 | 0.72 | 4.54 |
|          | \( \lambda \) | 5.81 | 2.16 | 0.98 | 3.86 |
| Summer   | \( \log N_w \) | 3.85 | 0.48 | -0.91 | 7.75 |
|          | \( D_m \)  | 1.75 | 0.43 | 0.92 | 4.65 |
|          | \( \mu \)   | 4.91 | 3.06 | 1.55 | 6.55 |
|          | \( \lambda \) | 5.4  | 2.37 | 1.59 | 6.64 |
| Typhoon  | \( \log N_w \) | 3.91 | 0.42 | -0.41 | 5.38 |
|          | \( D_m \)  | 1.52 | 0.38 | 0.6  | 4.75 |
|          | \( \mu \)   | 5.2  | 2.84 | 0.46 | 3.30 |
|          | \( \lambda \) | 6.46 | 2.57 | 0.64 | 3.21 |
| Autumn   | \( \log N_w \) | 4.03 | 0.46 | 0.23 | 3.62 |
|          | \( D_m \)  | 1.41 | 0.38 | 0.36 | 3.36 |
|          | \( \mu \)   | 4.28 | 2.31 | 0.93 | 5.25 |
|          | \( \lambda \) | 6.45 | 2.9  | 1.08 | 3.78 |

Figure 15a describes the \( \log_{10} N_w \) and \( D_m \) distribution with the maritime-like and continental-like convective clusters of Bringi et al. (2003). Most values of \( D_m \) and \( \log_{10} N_w \) in each season are similar, except for the winter and summer, which demonstrates distinct microphysical processes between these two seasons. Furthermore, the convection types in Northern Taiwan are near the maritime-like convection of Bringi et al. (2003). According to the preceding section, the vertical structure of radar reflectivity (CFAD, Fig. 15b) supports the understanding of the RSD of rain type. The reflectivity of convection had
1 \% of 30 dBZ up to 6.1 km, and the slope around the melting layer was $-4.36 \text{ dBZ km}^{-1}$. Further, below the melting layer, a reflectivity spread of $3-40$ dBZ can be seen in Fig. 15b. Thus, larger diameter of raindrops can be easily discovered in convective system than in stratiform and climatology. The CFADs of the six seasons’ convective precipitation are depicted in Fig. 16. The mean reflectivity profile of each season’s convection type (vertical red stars line) is greater than the annual mean reflectivity (vertical white stars line) in each HVS season (Figs. 16c–e), and each LVS season showed an opposite characteristic (Figs. 16a, b, f).
Moreover, the tilting of LVS (HVS) around the melting level was about −4.12, −5.62, and −5.26 dBZ km\(^{-1}\) (−4.01, −3.42, and −2.68 dBZ km\(^{-1}\)) for the winter, spring, and autumn (mei-yu, summer, and typhoon), respectively.

The convective precipitation CFADs of the six seasons are illustrated in Fig. 16. The summer season has 1% of 30 dBZ extending up to 10 km, with a spread of reflectivity of 21–49 dBZ under the melting layer (Fig. 16d), whereas the mei-yu season showed 1% of 30 dBZ up to 8.2 km, with a spread of reflectivity of 12–47 dBZ (Fig. 16c) below the melting layer. On the other hand, the minimum slope of −2.68 dBZ km\(^{-1}\) around the melting layer and the widest distribution of reflectivity (3–46 dBZ) below the melting layer appeared in the typhoon season (Fig. 16e). Summer is typically accompanied by thunderstorms, which are characterized by strong convection and vigorous updrafts and downdrafts, whereas the mei-yu season is more strongly associated with frontal systems with little convection, a wide stratiform area, and widespread rainfall. Among the LVS seasons, winter has 1% of 30 dBZ extending only up to 5 km, with a spread below the melting layer of 15–37 dBZ (Fig. 16a), and it shows relatively weakest convection not only in LVS but also in HVS seasons. The remaining two LVS seasons (spring and autumn) have a similar vertical structure with the same 1% of 30 dBZ height (6 km), but with different distributions below the melting layer, 17–43 dBZ and 3–40 dBZ, respectively, for the spring (Fig. 16b) and autumn (Fig. 16f).

c. Skewness between stratiform and convection

Positive and negative skewness are influenced by the location of long-tailed distribution. Positive (negative) skewness shows that long-tailed distribution is located above (below) the mean value. According to the statistics of log\(_{10}\) \(N_w\), skewness is positive (negative) for the LVS (HVS). Therefore, we consider the winter and summer, the largest and lowest value of skewness in log\(_{10}\) \(N_w\), for a detailed investigation.

Figure 17 shows positive skewness in the winter. However, negative skewness is observed in the summer. This is because the long-tailed distribution is located at different sides of the mean value in each season. Top (less) 10% of log\(_{10}\) \(N_w\) was selected in the winter (summer) to further investigate the physical meaning of skewness. For the winter, both convective and stratiform precipitation with positive skewness are associated with an LVS (Figs. 18b, d) when compared with mean fields (Figs. 18a, c). With the positive skewness of log\(_{10}\) \(N_w\), the CFAD corresponding to the top 10% of log\(_{10}\) \(N_w\) clearly shows more concentration of small drops in both stratiform and convective precipitation (of the shallow type) of the winter. On the other hand, convection and stratiform precipitation of the summer with negative skewness are linked with HVSs (Figs. 18f, h) when compared with the mean field (Figs. 18e, g). The vertical structure of less 10% of log\(_{10}\) \(N_w\) (Figs. 18f, h) is similar to the mean field (Figs. 18e, g).

Figure 19 clearly depicts the vertical structure with the mean field (red line) and selected data (blue line, top and less 10% of log\(_{10}\) \(N_w\)). Winter (Figs. 18b, d) clearly shows that a lower vertical structure occurred in the top 10% of log\(_{10}\) \(N_w\) with positive skewness. Although the reflectivity profiles in the summer below 4 km (around the melting layer) are similar to the
Fig. 18. CFADs of summer and winter with convection and stratiform. a, b, c, and d (e, f, g, and h) refer to winter (summer) seasons. Top 10 \% of log$_{10}$ $N_w$ is present in winter season (b, d) and less 10 \% is present in summer season (f, h). Figures 18a, and 18c, (e and g) depict the mean field of winter (summer) CFAD.
mean field, the icing processes are stronger than the mean field (red line). This indicates that strong aggregation and riming processes occur in the summer, even in the HVS. Therefore, the skewness of $\log_{10} N_w$ can also be clearly separated HVS and LVS by its sign in stratiform and convection.

4. Summery and conclusions

Taiwan is a subtropical island dominated by mountain ranges extending from the north to the south, with a maximum altitude of over 4,000 m. The significant land–sea distribution with complex terrain greatly hinders efforts to forecast the weather over this island. In this work, we sought to overcome some of these difficulties by elucidating seasonal characteristics that manifest in the type of rain. A disdrometer (JWD, installed at the NCU) and a radar (QPESUMS from CWB) were employed to observe the rain events in each season. Integral rainfall and gamma parameters were also calculated to identify the factors and microphysical processes that determine the RSD. The mean gamma parameters were listed to quantify the RSD in each season and rain type. Results from the JWD showed a high concentration of large drops and a low concentration of small drops in the summer. On the other hand, much more small raindrops occurred during the winter. This phenomenon originated from the different surrounding conditions, leading to distinct RSDs between the summer and winter in Northern Taiwan. Besides, alternative seasons (spring and autumn) seemed to have a similar vertical structure that produces slightly different RSDs. However, owing to the presence of strong wind shear and upward motions in typhoons, the typhoon season’s microphysics are different from those of the mei-yu season. These comprehensions can be figured out through the vertical reflectivity profiles, not only in seasons, but also in rain type. In the six seasons, the convective systems presented larger drops and a lower concentration compared to stratiform systems. A greater vertical motion in convective systems tends to carry water vapor to higher altitudes such that the

Fig. 19. Mean radar reflectivity profiles in summer (a and c) and winter (b and d). Convection and stratiform precipitation are shown in upper and lower panel, respectively.
growth of ice crystals is more rapid than of water droplets above the melting layer. The strong warm rain process below the melting layer is also responsible for the formation of larger raindrops in convective systems, compared to stratiform systems. Moreover, the skewness of $\log_{10} N_v$, separating the LVS and HVS, depicted that the completely different vertical structures within seasons lead to distinct statistical distributions. The seasonal variation of RSD can also be referred to the changeable atmospheric condition in Northern Taiwan.

This study provided a preliminary investigation of the subtropical RSD during the period from 2005 to 2014 (10 years) from radar data (QPESUMS) over Northern Taiwan. Furthermore, the use of reflectivity from Doppler radars is suitable only for inferring a rough approximation of particle size. Moreover, further details of RSD can be revealed from the parameters of the dual polarimetric radar, such as reflectivity ($Z_{hh}$), differential reflectivity ($Z_{dr}$), specific differential phase ($K_{dp}$), and zero lag cross-correlation of horizontal and vertical waves ($R_{hv}$). Besides, fuzzy logic analysis can also help in particle identification.

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