Preliminary Study of Land–Sea Microphysics Associated with the East Asian Summer Monsoon Rainband and Its Application to GPM DPR

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ABSTRACT

Raindrop size distribution (DSD) characteristics during the East Asian summer monsoon (EASM) were studied, using measurements from three OTT Particle Size Velocity (Parsivel) disdrometers in Nanjing, Chuzhou, and the northwestern Pacific (NWP), respectively. Western and eastern parts of the monsoon rainband were separated for a comparative study of the DSD variability. Along with disdrometer data, GPM Dual-Frequency Precipitation Radar (DPR), *Fengyun-2E (FY-2E)*, MODIS, GPCP, ERA-Interim, and in situ radiosonde datasets are combined to illustrate the possible microphysical mechanisms for the significant DSD variability in two parts, in terms of convective intensity, cloud structure, and aerosol effects. The DSD characteristics of six rain-rate classes and two rainfall categories (convective and stratiform) were studied. The western part has larger mass-weighted mean diameter D_m while smaller normalized intercept $\log_{10}(N_w)$ than the eastern part, and the convective clusters of the western part (land) could be identified more maritime-like than continental-like due to moisture transport from the tropical ocean, while that of the eastern part (sea) is between maritime-like and continental-like. Cross validation of GPM rainfall products are implemented based on surface disdrometer observations. DPR products manifest better performance over sea than land areas of the EASM rainband. Empirical D_m – Z_e and N_w – D_m relations were also derived preliminarily to improve the GPM rain-retrieval algorithms in the EASM season.

1. Introduction

The East Asian summer monsoon (EASM) rainband, a zonally elongated monsoon rainband extending eastward from southwestern China to the northwestern Pacific (NWP) through the Japanese archipelago (Fig. 1a), is one of the most significant rainfall systems affecting the hydrological cycle in the East Asian region (Sampe and Xie 2010; Kuwano-Yoshida et al. 2013). Rainstorm associated with EASM is crucial for summertime water resources and sometimes results in severe natural

disasters in East Asian countries (Ogura et al. 1985; Ninomiya 2000; Ninomiya and Shibagaki 2003; Yoshizaki et al. 2000; Kato and Aranami 2005; Kato 2006). Its extended also intimidates the security of naval ships on sea. Therefore, an urgent need to improve the performance of quantitative precipitation estimation (QPE) within the rainfall system for flood forecast, urban waterlogging prediction, and ocean storm monitoring is required.

Raindrop size distribution (DSD) is of great significance for the understanding of microphysical processes in generating rainfall particles and ultimately for improving radar QPE algorithms. Over the last decades, DSD characteristics in different continental and marine locations from tropics to midlatitudes along with different disdrometers have been well documented and are known to vary spatially and temporally

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FIG. 1. (a) GPCP precipitation during mei-yu season (16 Jun–15 Jul) [contour interval: 2 mm day^{-1} , with light (dark) shading for 4–8 (>8) mm day^{-1}] and locations of two observational regions in EASM rainband. (b) Probability density function (PDF) of TBB during the rainy days in two parts of EASM rainband. The superimposed rectangles in (a) correspond to the assumed western part and eastern part of this rainband. The distribution of TBB in (b) are obtained from *FY-2E* data products.

(e.g., Rosenfeld and Ulbrich 2003; Koizu et al. 2005; Kirankumar et al. 2008; Chakravarty and Raj 2013; Williams et al. 2014; Wen et al. 2016; Zhang et al. 2017; Wu et al. 2019a). The entire EASM rainband locates from China to Japan and extends to NWP, which is an ideal precipitation system for studying microphysical properties of different parts within the same weather system, since it covers both continental and oceanic areas. Apart from the underlying surface variability in the EASM rainband, the land–sea atmospheric patterns also differ from each other with equivalent barotropy in its continental part whereas typical midlatitude baroclinicity in its oceanic part (Chen and Chang 1980). Although studies have been reported to address the DSD of EASM, they mainly focused on its continental part in East Asia (e.g., Wen et al. 2016, 2017; Chen et al. 2013; Brongi et al. 2006). DSDs of the monsoon rainfall over oceans have received little investigations.

Moreover, atmospheric aerosols, formed from natural and anthropogenic processes, can also have an effect on precipitation through aerosol–cloud interactions (ACI) (e.g., May et al. 2011; Ramanathan et al. 2001). It has been well documented in the literature that the aerosols can act as cloud condensation nuclei (CCN), increases the number of cloud drops and reduces the effective size of the cloud particles (e.g., Ramanathan et al. 2001; Breon et al. 2002). The aerosol effects on microphysics of EASM rainfall should not be neglected due to the rapidly expanding economic and industrial developments in East Asian countries, causing a prominent increase in aerosol loading over this area (Streets et al. 2008; Liu et al. 2011; Jiang et al. 2017; Zhang et al. 2019).

Various radar QPE algorithms, including the recent Global Precipitation Measurement (GPM) Dual-Frequency Precipitation Radar (DPR) highly depend on surface-based DSD observations (e.g., Liao et al. 2014; Radhakrishna et al. 2016; Chen et al. 2017; Liao

and Meneghini 2019). GPM DPR satellite is expected to improve our knowledge of precipitation processes by providing greater dynamic range, more detailed information on microphysics, and better accuracy values in rainfall retrievals (Le and Chandrasekar 2013). Hydrological validation of DPR with surface DSD measurements helps to improve its physically based algorithms in EASM season.

Based on the above analyses, one specific subject herein is to obtain comprehensive understanding of DSDs in different parts of the EASM rainband including the aerosol effects, and ultimately improves the recent GPM DPR rainfall estimations in the EASM season. For a first attempt, measurements from surface Particle Size Velocity (Parsivel) disdrometers, along with satellite data such as Global Precipitation Climatology Project (GPCP), *Fengyun-2E* (*FY-2E*), Moderate Resolution Imaging Spectroradiometer (MODIS), and the European Centre for Medium-Range Weather Forecasts (ECMWF) interim reanalysis (ERA-Interim) data, in situ radiosonde data, are united to illustrate the dynamical and microphysical characteristics associated with the EASM rainband. And DSD-based relations are further derived to improve the accuracy of GPM DPR precipitation estimations in the EASM season.

The rest of this paper proceeds as follows: the data and methods are described in section 2, an outline of DSD and the possible reasons for the variations in overall DSD characteristics among two parts of the EASM rainband are analyzed in section 3, the relations of DSD parameters are obtained in section 4, GPM rainfall retrieval methods are discussed in section 5, and finally section 6 presents a summary and conclusions.

2. Data and methods

a. Observational sites and datasets

During June to July 2014, observations from in situ disdrometers and satellites are collected at Nanjing

FIG. 2. Field view of (top right) the observational Chuzhou site (CZ) and Nanjing site (NJ), as well as (bottom) the northwestern Pacific site (NWP). (top left) The local topography is also presented.

(32.0°N, 118.5°E; 15 m MSL), Chuzhou (32.3°N, 118.3°E; 18 m MSL) and NWP (27°–32°N, 143°–153°E; 22 m MSL) sites (Fig. 2). NWP is located in the southeast ocean area of Japan Sea, while Nanjing and Chuzhou are located in the East China continents. The thermal effect of land–sea contrast can contribute to the onset of EASM (Ding and Chan 2005). Due to different underlying surface properties and environmental conditions, the precipitation microphysical variability in different parts of the monsoon rainband could be of great significance. For further comparison, we separated the main EASM rainband and adopted Nanjing and Chuzhou data jointly as the western part of the EASM rainband (hereinafter the western part), while NWP data as the eastern part of this rainband (hereinafter the eastern part). And the multisource data products utilized in this work are presented as below:

1) IN SITU PARSIVEL² AND RADIOSONDE DATA

The DSD data selected for the analyses were measured by Parsivel² disdrometers manufactured by OTT Hydromet, Germany (Tokay et al. 2014; Löffler-Mang and Joss 2000). Parsivel² disdrometer herein is a second-generation optical disdrometer with a near-IR (780 nm) rectangular single beam of 27 mm wide, 180 mm long (that is a sensor horizontal area of 48.6 cm²), and 10 mm

high. It archives equivalent drop diameters sorted into 32 classes of diameter from 0 to 25 mm and 32 classes of fall speeds from 0 to 22.4 m s^{−1}. The time resolution can be selected and is set to 1 min in current study. The 1-min DSD data in NWP were measured over western Pacific by onboard Parsivel² disdrometer during a marine survey (Zhang 2018). Meanwhile, two Parsivel² disdrometers simultaneously measure the rainfall in Nanjing (in Jiangsu Province) and Chuzhou (in Anhui Province), China, respectively. Following the data quality control method in Wu et al. (2019a), as well as the definition of a rain event proposed by Tokay and Bashor (2010), 10 rain events incorporating 3688 one-minute effective DSD samples are identified for the western part, 11 rain events incorporating 975 samples are identified for the eastern part (Table 1). Meteorological variables (such as pressure, temperature, relative humidity, and horizontal winds) from in situ radiosonde (twice a day) are used.

2) GPM DPR LEVEL-2 PRODUCTS

GPM DPR consists of two precipitation radars (Ku band and Ka band), under cooperation of the National Aeronautics and Space Administration (NASA) and the Japan Aerospace Exploration Agency (JAXA). It has an orbital inclination angle of 65° and an altitude of approximately 407 km. In the present study, the

TABLE 1. Precipitation events used for the present study in two parts of monsoon rainband.

Regions	Event No.	Date	Time (LST)	1-min samples (min)	Accumulated precipitation (mm)	Max rain rate (mm h ⁻¹)
Western part	1	15 Jun 2014	0454–0615	62	13.7	19.6
	2	15 Jun 2014	1757–2212	152	15.2	18.4
	3	16 Jun 2014	0711–1701	216	18.1	16.8
	4	25 Jun 2014	0551–2347	364	29.2	39.4
	5	26 Jun 2014	0002–1817	545	38.6	43.1
	6	1 Jul 2014	0553–1551	263	12.3	32.5
	7	1–2 Jul 2014	2049–0937	417	19.4	37.2
	8	4 Jul 2014	1100–2359	631	39.8	101.1
	9	5 Jul 2014	0001–1253	725	41.1	115.3
	10	12 Jul 2014	0659–2258	313	71.9	145.2
Eastern part	1	23 Jun 2014	0318–0427	67	2.8	17.7
	2	26 Jun 2014	1915–2016	52	0.7	1.9
	3	28 Jun 2014	0234–0827	91	0.4	0.8
	4	1–2 Jul 2014	2156–0759	107	3.5	3.9
	5	2 Jul 2014	0955–2136	102	5.9	2.9
	6	4 Jul 2014	0458–1848	79	0.2	0.6
	7	5 Jul 2014	0520–0730	58	0.1	0.4
	8	5 Jul 2014	1758–1905	67	0.1	0.2
	9	6 Jul 2014	0159–1011	120	2.5	6.9
	10	6 Jul 2014	1212–2342	104	1.1	6.8
	11	8 Jul 2014	0251–1500	128	1.8	2.8

attenuation-corrected radar reflectivities and rain rate at the estimated surface are obtained from level-2A products of DPR and are matched with the Parsivel disdrometer data, as is already described in [Wu et al. \(2019b\)](#). Overall, there are 4 (7) effective observations from 8 (11) instantaneous cases over the western part (eastern part) chosen for the comparison between DPR and Parsivel. [Table 2](#) gives the DPR-observed

rain events during the EASM period from June to July 2014.

3) FY-2E RADIATION PRODUCTS AND MODIS CLOUD PRODUCTS

Blackbody temperature (TBB) products from FY-2E satellite available at $0.1^\circ \times 0.1^\circ$ spatial resolution are used with 1 h interval are used as in [Wu et al. \(2019a\)](#).

TABLE 2. Precipitation events observed by GPM DPR when it overpasses the two parts of monsoon rainband.

Regions	Pass No.	Date	Time (LST)	Rainfall observations (Yes/No)	Max rain rate (mm h ⁻¹)
Western part	1	15 Jun 2014	0256–0429	Yes	10.2
	2	15 Jun 2014	1211–1344	No	0
	3	23 Jun 2014	0041–0213	No	0
	4	25 Jun 2014	2339–0111	Yes	24.3
	5	26 Jun 2014	0854–1027	Yes	32.1
	6	2 Jul 2014	0650–0823	Yes	19.5
	7	6 Jul 2014	2021–2154	No	0
	8	12 Jul 2014	0424–0556	Yes	62.4
Eastern part	1	15 Jun 2014	0124–0256	No	0
	2	16 Jun 2014	0947–1119	Yes	0.8
	3	16 Jun 2014	2340–0112	No	0
	4	25 Jul 2014	0641–0813	No	0
	5	25 Jul 2014	2206–2339	No	0
	6	26 Jul 2014	2115–2247	Yes	0.6
	7	1 Jul 2014	2002–2135	Yes	0.4
	8	2 Jul 2014	1911–2043	Yes	0.2
	9	4 Jul 2014	1900–2032	Yes	5.4
	10	6 Jul 2014	1849–2021	Yes	4.9
	11	8 Jul 2014	0312–0445	Yes	1.7

Cloud and aerosol data, available at $1^\circ \times 1^\circ$ spatial resolution, from level-3 daily global products of MODIS (MOD08_D3) are also used (Platnick et al. 2003). MOD08_D3 provides atmospheric parameters associated with aerosol particle properties, water vapor, as well as cloud optical and physical properties (Remer et al. 2005).

4) GPCP MERGED PRECIPITATION DATA AND ERA-INTERIM DATA

Daily GPCP merged precipitation with $1^\circ \times 1^\circ$ spatial resolution are used (Huffman et al. 2012). ERA-Interim data, available four times a day with a horizontal resolution of $0.125^\circ \times 0.125^\circ$ and 38 pressure levels in the vertical direction, are used (Dee et al. 2011).

In this study, the data products from GPM, FY-2E, MODIS, GPCP, ERA-Interim, and in situ radiosonde are daily averaged to match with the Parsivel² measured rainy days over two parts of the EASM rainband.

b. Parsivel² disdrometer data processing

The DSDs are computed from the Parsivel² disdrometer counts and the integral precipitation variables, such as the rain rate R (mm h^{-1}), radar reflectivity Z ($\text{mm}^6 \text{m}^{-3}$), mass-weighted mean diameter D_m (mm), generalized intercept parameter N_w ($\text{mm}^{-1} \text{m}^{-3}$), and standard deviation of the mass spectrum σ_M (mm) are further estimated from measured DSDs (e.g., Wen et al. 2016; Wu et al. 2019a).

So far, the most commonly used mathematical model in formulating and analyzing measured DSDs is the gamma distribution proposed by Ulbrich (1983):

$$N(D) = N_0 D^\mu \exp(-\Lambda D), \quad (1)$$

where $N(D)$ represents the particle concentration, three parameters N_0 , μ , and Λ represent the intercept, shape, and slope of DSDs, respectively. The truncated moment fitting (TMF) method proposed by Vivekanandan et al. (2004) is used to construct the gamma model from disdrometer data.

c. GPM DPR variables calculated from DSD

In terms of DSD observations, the effective radar reflectivity factor Z_e for both Ku band and Ka band can be formulated as following:

$$Z_e = \frac{\lambda^4}{\pi^5 |K_w|^2} \sum_{i=3}^{32} N(D_i) \sigma_b(D_i, \lambda) \Delta D_i, \quad (2)$$

where λ is the wavelength, $N(D)$ is the DSD, K_w^2 is defined as the dielectric factor, and $\sigma_b(D_i, \lambda)$ is the back-scattering cross section for a particle with diameter D_i .

To avoid the nonindependence problem of the parameters of gamma distribution, the normalized gamma model based on D_m and N_w is adopted for DSDs as in Wu et al. (2019a).

The difference between the Ku- and Ka-band reflectivity measurements can be well described by the dual-frequency ratio (DFR), which is defined as the ratio of Z_e at Ku band (Z_{Ku}) to Z_e at Ka band (Z_{Ka}). According to Seto et al. (2013), the DFR (dB) is uncorrelated with N_w and is only a function of D_m :

$$\text{DFR} = 10 \log_{10} \left(\frac{Z_{\text{Ku}}}{Z_{\text{Ka}}} \right) = f(D_m). \quad (3)$$

Note the calculations of Z_e and DFR based on disdrometer observations are the same as in Wu et al. (2019b).

3. Outline of DSD

a. Drop size spectra

Before a detailed discussion of DSD property, we examined the average DSDs for the western part and the eastern part of the EASM rainband. To discern the precipitation differences between the two parts, the DSD samples are stratified into six classes on the basis of rain rate: $R \leq 2 \text{ mm h}^{-1}$, $2 < R \leq 5 \text{ mm h}^{-1}$, $5 < R \leq 10 \text{ mm h}^{-1}$, $10 < R \leq 20 \text{ mm h}^{-1}$, $20 < R \leq 40 \text{ mm h}^{-1}$, and $R > 40 \text{ mm h}^{-1}$. Large (small) drops are assumed to be $D > 4$ ($D < 1$) mm.

Figure 3 presents the average drop spectra for two parts in terms of six rain-rate classes, along with the remained data samples N of each rain-rate class. Note that the DSDs exhibit similar peak structures, and the content of large drops has an increasing tendency with rain rate. Particularly, the drop size spectra of the western part is narrower than that of the eastern part when $R \leq 5 \text{ mm h}^{-1}$, whereas it becomes significantly broader than that of the eastern part when $R > 5 \text{ mm h}^{-1}$. The concentration of small drops in the western part stays nearly constant when $R \leq 20 \text{ mm h}^{-1}$, while it increased significantly when $R > 20 \text{ mm h}^{-1}$.

b. Analysis

To explain the unique characteristics of drop size spectra, some environmental variables are collected and analyzed. The distribution of TBB for the rainy days in two parts of the EASM rainband are described with probability density function (PDF) in Fig. 1b. CAPE values and surface temperature for the Parsivel² measured rainy days over two parts are obtained from ERA-Interim, and are depicted with box-and-whisker plot in Figs. 4a and 4b. The average wind field, as well as relative humidity profile in Figs. 5a and 5b are all obtained

FIG. 3. The average DSDs at the six indicated rain-rate classes over two parts of EASM rainband. The remaining data samples N of each rain-rate class are also presented.

from the in situ radiosonde, and are also drawn for the rainy days in two parts. The cloud structure, cloud effective radii (CER) of liquid and ice particles, and aerosol optical depth (AOD) for the rainy days over the observational regions are acquired from MODIS data products and are provided in Figs. 5a, 6a, and 6b.

In general, an increase in rain rate could be associated with the enhancement of convection activity. TBB reflects the developing intensity of convection activity and is considered as an proof for convective development when $TBB \leq -32^{\circ}\text{C}$ (241.15 K) (e.g., Maddox 1980). Figure 1b suggests more intense development of

FIG. 4. Box-and-whisker plots of (a) CAPE (kJ kg^{-1}) and (b) surface temperature ($^{\circ}\text{C}$) from ERA-Interim over the western (red box) and eastern (blue box) parts of EASM rainband. The center line of the box indicates the median, and the bottom and top lines of the box indicate the 25th and 75th percentiles, respectively. The bottom and top of the solid vertical lines indicate the minimum and maximum values, respectively.

convections in terms of frequency of occurrence in the western part. Strong convective activity extends the clouds to deeper altitudes, which enhances cold rain processes. And this can be confirmed by the relatively larger CAPE value, higher surface temperature (Fig. 4b) and cloud top (Fig. 5a) in the western part. Deeper the extent of clouds, more ice/liquid particles grow to larger size. Yet we can notice larger CER of ice and water over observational site of the eastern part than the western part (Fig. 6a). This could be due to the higher AOD over the western part than the eastern part (Fig. 6b), causing an increase in the number of cloud drops and an decrease in the effective size of the cloud particles. Large

size of cloud droplets are favorable for the warm rain process (collision-coalescence) when rain rate is low. Thus, the relatively lower concentration of CCN and cloud droplet over the eastern part could contribute to larger CER (Fig. 6a), which results in broader drop size spectra through the collision-coalescence process between cloud drops. Whereas the higher aerosol concentration over the western part (Fig. 6b) suppresses the collision-coalescence processes, increases the number of cloud droplets and decreases their effective radius (Fig. 6a), leading to narrow drop size spectra (Wallace and Hobbs 1977; Twomey 1977; Albrecht 1989; Rosenfeld 1999; Andreae et al. 2004).

FIG. 5. Vertical profiles of (a) cloud structure (heights of cloud base, level of 0°C , cloud top; all in km) and vertical wind field (m s^{-1}) and (b) relative humidity (%) from MODIS and in situ radiosonde observations in two parts of EASM rainband. The green circles indicate cloud base, the red rectangles indicate the level of 0°C , and the blue triangles indicate cloud top. A full (half) wind barb represents a wind speed of 10 (5) m s^{-1} .

FIG. 6. As in Fig. 4, but for (a) cloud effective radius (CER, μm) of liquid and ice particles and (b) aerosol optical depth (AOD) from MODIS.

As rain rate increases, small drops are carried to higher altitudes by strong updrafts. Thus, more CCN due to higher AOD in the western part enter the clouds above the melting layer. And this strengthens more freezing of cloud drops and associated latent heat release (Rosenfeld and Woodley 2000) through cold rain processes (Bergeron process, riming, aggregation), which can upturn the convective activity and bring more large droplets, contributing to broader drop size spectra in the western part. Based on the above analyses, the aerosols could help inhibit warm rain process while enhance cold rain process in the western part, which partly explains the finding that the drop size spectra of the western part could be narrower than that of the eastern part when $R \leq 5 \text{ mm h}^{-1}$, whereas it becomes significantly broader than that of the eastern part when $R > 5 \text{ mm h}^{-1}$.

As the spectra broadens in the western part, more large drops would collect small drops, which is favorable for the collision-coalescence processes with a consumption of small drops. And this helps to explain the negligible variation of small drops in the western part with increasing rain rate when $R \leq 20 \text{ mm h}^{-1}$. Compared to the eastern part, the stronger cold rain processes in the western part conduce to the collision-breakup processes in strong rainstorm. Besides, the eastern part shows lower humidity above 0°C layer (Fig. 5) and the melting behavior of graupel can be suppressed by such lower humidity (e.g., Rasmussen and Heymsfield 1987), while the moister conditions in the western part would help ice-phase particles melt, resulting in more efficient collision-breakup processes for raindrops. Accordingly, more small drops would be produced and that would help explain the large sum of small drops in the western

part when $R > 20 \text{ mm h}^{-1}$. Therefore, the dominant microphysical processes in the western part vary with increasing rain rate.

The above findings suggest different microphysical mechanisms within the EASM rainband. To further understand the distinct variability of DSDs, the observation samples from the western part and the eastern part, according to the particle diameter, are stratified into six classes and the diameter values corresponding to each size classes are shown in Fig. 7.

It is obviously shown in Fig. 7 that the small drops dominate the number concentration ($>72\%$) for the total dataset in the EASM rainband and contribute to about 27% of rain rate in the eastern part. Moreover, the small and midsize drops ($D < 4 \text{ mm}$) contribute more than 99% of the number concentration, which indicates the small and midsize drops jointly dominate the precipitation during the monsoon. Compared to the eastern part, the contribution of midsize drops to number concentration as well as rain rate is both larger in the western part due to the significant coalescence and breakup processes, which has been explained in the above analysis of average drop spectra. Compared to the western part, the contribution of small drops to number concentration or rain rate is both larger in the eastern part, which indicates a pronounced effect of small drops on the total precipitation in the eastern part. This could be partly attributed to the stronger collision-breakup process caused by larger wind velocity (Fig. 5a) in the eastern part, which further reflected the typical baroclinicity in the eastern part (Chen and Chang 1980). Similar result was also demonstrated by Testik and Pei (2017) that the raindrop collisional breakup process is expected to intensify with an increase in wind speeds,

FIG. 7. Relative contribution of each size classes to (a) total drop concentration N_t (m^{-3}) and (b) rain rate R (mm h^{-1}) over two parts of the EASM rainband.

which have a statistically significant effect of increasing the number of small drops and decreasing the number of large drops.

4. DSD parameters

a. Statistical properties

The DSD samples in two parts of the EASM rainband derived from Parsivel² are further used to derive gamma model via the TMF method adopted by Vivekanandan et al. (2004). Table 3 presents the statistical values of obtained gamma parameters (N_0 , μ and Λ) in two parts, which indicates all larger mean value of three parameters in the western part. The larger μ in the western part could be attributed to the significant coalescence and breakup processes, both processes acting together increase the concentration of midsize drops in the western part (Fig. 7) and thereby result in an substantial increase of μ value (Rosenfeld and Ulbrich 2003). It is also reported by Rosenfeld and Ulbrich (2003) that stronger

coalescence process decreases both N_0 and Λ , while stronger breakup process increases both N_0 and Λ . The larger N_0 and Λ in the western part may indicate that both processes are significant with breakup process in predominance.

Figure 8 shows the coefficients μ and Λ with rain rate. The values of μ and Λ tend to decrease with increasing rain rate. While rain rate is larger than 25 mm h^{-1} , the PDF indicates that μ and Λ hold steady at about 6.0 and 7.0 mm^{-1} for the western part and 5.0 and 6.0 mm^{-1} for the eastern part, respectively. Notably, the variation of μ and Λ are both marked once rain rate is lower than 10 mm h^{-1} , which indicates the complicated microphysical mechanism of stratiform rain during EASM, where the cold cloud process, the warm cloud process, melting, and evaporation all have a significant impact on rain formation; gradually, the range of variation decreases with increasing rain rate and becomes more uniform at larger rain rates. The rain rate of 10 mm h^{-1} can be used as a reference for precipitation categorization. Tokay and Short (1996) noted that a rain rate

TABLE 3. Statistical results of gamma model parameters in two parts of the EASM rainband. SD and SK represent standard deviation and skewness, respectively.

Parameter	Region	Total precipitation			Convective precipitation			Stratiform precipitation		
		Mean	SD	SK	Mean	SD	SK	Mean	SD	SK
$\log_{10}(N_0)$	Western	5.38	1.46	0.70	5.40	1.40	1.28	5.37	1.48	0.52
	Eastern	4.93	1.81	2.14	4.91	1.10	0.57	4.93	1.84	2.14
μ	Western	5.91	3.23	0.98	5.84	3.41	0.90	5.94	3.17	1.02
	Eastern	4.72	3.04	1.17	5.03	2.34	0.07	4.71	3.07	1.20
Λ	Western	7.20	3.35	0.79	6.12	3.13	1.18	7.59	3.34	0.70
	Eastern	6.95	4.17	3.68	5.13	2.29	0.60	7.03	4.22	3.68

FIG. 8. Scatterplot of μ (dimensionless) and Λ (mm) vs R (mm h^{-1}) for each 1-min DSD observed in two parts of the EASM rainband. The PDF of two parameters observed at two parts are given in each panel as well.

greater than 10 mm h^{-1} is mostly determined by convective clouds. Therefore, the rainfall differences in rain rates below and above 10 mm h^{-1} may indicate the DSDs variability among different types of monsoon rain, which further helps us to categorize the precipitation into two types. Here, 1-min samples with $R > 10 \text{ mm h}^{-1}$ are considered convective rainfall (also in Wu et al. 2019a). As a result, the western part and the eastern part consist 83% (17%), 92% (8%) of stratiform (convective) rainfall samples, respectively.

b. Distribution of D_m and N_w

Previous studies (e.g., Bringi et al. 2003; Chen 2009; Chen et al. 2013) showed that the behavior of DSD parameters was distinctly different for the maritime and continental convective storms. Bringi et al. (2003) used a threshold of $\sigma_R = 1.5 \text{ mm h}^{-1}$ for the classification of stratiform and convective rain and found that maritime convective rain is characterized by larger $\log_{10}(N_w)$ and smaller D_m , while continental convective rain is characterized by smaller $\log_{10}(N_w)$ and larger D_m . To further discuss the DSD parameters for the EASM rainband, we also obtained $\log_{10}(N_w)$ and D_m of two classified rain types. The results from two classification methods including the method proposed by Bringi et al. (2003) as well as the method used in section 4a are both presented in Table 4 for comparison. It is notable that the method of Bringi et al. (2003) yields larger D_m and smaller $\log_{10}(N_w)$ in the western part, and slightly smaller $\log_{10}(N_w)$ in the eastern part. Apparently, different rainfall classification methods could influence the DSD parameters to some extent.

The observation of convective rain in Bringi et al. (2003) are presented with two black boxes in Fig. 9. For comparison, the $\log_{10}(N_w)$ and D_m obtained for the

EASM rainband by using the same rainfall classification methods are also shown in Fig. 9. Note that the convective precipitation in the western part is identified closer to maritime-like than continental-like convective precipitation as defined by Bringi et al. (2003), while the convective precipitation in the eastern part is identified between maritime-like and continental-like.

The convective precipitation in the western part (land area) shows a character of “maritime-like,” which could be related to the abundant moisture transported from tropical ocean during EASM. Figure 10 shows average horizontal moisture flux at 925-hPa calculated from reanalysis data during the EASM season. Moisture fluxes from the tropical Indian Ocean and North Pacific Ocean meet and turn northward around the South China Sea (SCS), then converge into the EASM rainband to sustain copious precipitation. Interestingly, it is noted that the $D_m(N_w)$ is even smaller (larger) in the convective rain of the western part than the eastern part, which means the convective precipitation in the western part is closer to maritime-like than the eastern part based on the definition in Bringi et al. (2003). We believe

TABLE 4. Derived parameters from two different rainfall classification methods in Bringi et al. (2003) and Wu et al. (2019a).

Region	Parameter	Rain type	Bringi et al. (2003)	Wu et al. (2019a)
Western	D_m	Convective	1.74	1.67
		Stratiform	1.18	1.38
	$\log_{10}(N_w)$	Convective	3.76	3.85
		Stratiform	3.60	3.35
Eastern	D_m	Convective	1.95	1.95
		Stratiform	1.22	1.38
	$\log_{10}(N_w)$	Convective	3.70	3.75
		Stratiform	3.28	3.16

FIG. 9. Distribution of $\log_{10}(N_w)$ and D_m for convective precipitation, stratiform precipitation, and total precipitation in two parts of EASM rainband (along with ± 1 standard deviation). The black diamonds indicate comparative studied regions. The two outlined rectangles correspond to the maritime and continental convective clusters reported by Brangi et al. (2003). The black dashed line indicates the characteristics of stratiform rain (Brangi et al. 2003). The letters C, S, and T denote the results from convective precipitation, stratiform precipitation, and total precipitation, respectively.

this is because of the following two reasons: first, the vapor transport from tropical oceans to the western part of the rainband increases its humidity and thereby inhibits its evaporation process, causing more small drops being preserved, thus the D_m decreases; second, as we analyzed in section 4a, the predominant breakup process in heavy rain of the western part may lead to an decrease of large drops while a significant increase of small drops, thus the D_m decreases, too. Therefore, the D_m is even smaller in the convective rain of the western part than the eastern part. We think it a very interesting and meaningful finding, which indicates that the identification method of the “maritime-like” and “continental-like” precipitation in Brangi et al. (2003) may have some obvious limitations, especially when it comes to the precipitation of some typical weather systems like EASM.

On the other hand, the convective precipitation of the eastern part (subtropical ocean) is not identified “maritime-like,” but between “maritime-like” and “continental-like,” and is characterized by $D_m = 1.95$ and $\log_{10}(N_w) = 3.70$. Also, recent research work of oceanic DSD by Protat et al. (2019) suggested that the averaged D_m and N_w values (indicated by angle brackets) of convective rain over the global oceans are approximately characterized by $\langle D_m \rangle = 1.8\text{--}2.2$ mm and $\log_{10}\langle N_w \rangle = 3.0\text{--}4.0$. Both of this research and the research by Protat et al. (2019) indicate that the definition of “maritime-like” precipitation in Brangi et al. (2003) shows obvious limitations due to the lack of rainfall observations at different latitude ocean areas. As is known that the definition of “maritime-like” convective

FIG. 10. The average horizontal moisture flux at 925 hPa ($\text{g kg}^{-1} \text{s}^{-1}$) calculated from reanalysis data.

precipitation proposed by Brangi et al. (2003) is based on the observations from tropical island and sea, and a few coastal areas in subtropics, which includes no subtropical ocean area or higher-latitude oceans. Whereas oceans at different latitudes could have significant DSD variability (Protat et al. 2019). The above analysis inspires us to propose another rainfall identification method that better represents the DSD variability globally, we leave that in our future works.

c. σ_M – D_m relations

Previous studies have shown that gamma function parameters (N_0 , μ and Λ) are not statistically independent but are correlated, with high Pearson correlation coefficients (Ulbrich 1983; Ulbrich and Atlas 1985; Chandrasekar and Brangi 1987; Illingworth and Blackman 2002; Moisseev and Chandrasekar 2007). Therefore, mathematical artifacts will appear in relationships once a DSD is assumed to follow a gamma function (Williams et al. 2014). Without an assumption of a gamma-shaped DSD a priori, raindrop mass-weighted mean diameter D_m and mass spectrum standard deviation σ_M are used to describe the DSD shape, and relationships between D_m and σ_M are investigated (Petersen and Jensen 2012; Williams et al. 2014; Thurai et al. 2014).

A relation of $\sigma_M = 0.263D_m^{1.60}$ was acquired by Petersen and Jensen (2012) during the Midlatitude Continental Convective Clouds Experiment (MC3E), conducted in central Oklahoma in early summer 2011. Based on long-term two-dimensional video disdrometer (2DVD) observations, Thurai et al. (2014) obtained $\sigma_M = 0.266D_m^{1.65}$ whereas Williams et al. (2014) obtained $\sigma_M = 0.30D_m^{1.36}$ for precipitation in Huntsville, Alabama. Cloud physics mechanisms may be regionally dependent and lead to variations in these statistical relationships. Thus, an investigation of DSD physical attributes during monsoon is crucial. To further reveal the microphysical distinctions between the western and eastern parts.

Figure 11 presents the σ_M - D_m distribution of observed DSDs in two parts of the EASM rainband. The distribution is described in terms of the frequency of occurrence, as two-dimensional PDF where the colors represent the number of cases with the corresponding σ_M - D_m pairs. In two parts, a similar variation already reported in Wu et al. (2019b) is found that σ_M increases with the increasing D_m . The σ_M - D_m relationship for each part rainfall samples is fitted by power law and is also shown in Fig. 11. The relationship obtained for the western part is as below:

$$\sigma_M = 0.285D_m^{1.28}. \quad (4)$$

While for the eastern part is given by

$$\sigma_M = 0.311D_m^{1.20}. \quad (5)$$

Compared to Oklahoma and Alabama on the same latitude (Petersen and Jensen 2012; Thurai et al. 2014; Williams et al. 2014), both the western part and the eastern part have smaller exponents for σ_M - D_m relations, causing smaller σ_M with an increase of D_m . And due to that, the spectral width of intense convective precipitation (usually having large D_m) would be narrower in monsoon rainfall than the other regions on the same latitude, which reflects the microphysical variability of precipitation in different climate regimes. Comparing the western part with the eastern part of the same EASM rainband, the western part has larger exponent for σ_M - D_m relations, resulting in broader spectral width with more sufficient large droplets of the convective precipitation in the western part. And this further implies the microphysics variability within the EASM rainband.

5. GPM-Parsivel comparison

a. Validation of GPM DPR products

With its onboard DPR and the Microwave Imager (GMI), the GPM Core Observatory satellite provides a global surveillance of precipitation since 2014 (e.g., Skofronick-Jackson et al. 2017). As a successor of Tropical Rainfall Measuring Mission (TRMM), GPM satellite faces its challenge in well behaving not only at tropics but more importantly over midlatitudes. It is therefore significant to evaluate how accurately GPM can determine precipitation over midlatitude EASM regions. To assess and improve the ability of the GPM satellite, we have compared concurrent DPR measurements with Parsivel observations in terms of attenuation-corrected radar reflectivity and rain rate at estimated surface.

FIG. 11. The σ_M - D_m two-dimensional PDF distributions and fitting results based on Parsivel disdrometer observations in two parts of the EASM rainband. The resolution of PDF is 0.1 mm $\times$ 0.1 mm and the size of the color mark represents the percentage. The dashed lines in corresponding color represent the empirical σ_M - D_m relations from Petersen and Jensen (2012), Thurai et al. (2014), and Williams et al. (2014), respectively.

Following Wu et al. (2019b), we choose DPR_MS product to obtain the satellite-observational variable of rain rate, while we choose Ku-PR and Ka-HS products to obtain the effective reflectivity factor at both Ku and Ka bands due to the advantages of each product. For validation, the Parsivel² observational data are used to calculate the Ku-band (Ka-band) effective radar reflectivity factor Z_{Ku} (Z_{Ka}) as well as rain rate over two parts of the EASM rainband using the method in section 2. Accordingly, the DFR can also be obtained via Eq. (3). The relationship between DFR and rainfall helps improve our understanding of microphysics regarding precipitation retrieval algorithm (Zhang and Fu 2018; Wu et al. 2019b). Thus, the statistical relationship between DFR and rain rate R in both the western part and the eastern part of the EASM rainband is analyzed and compared given that the results from GPM and Parsivel² observations are close to each other both spatially and temporally.

Figure 12 shows the distribution of DFR and rain rate at estimated surface from both GPM DPR measurements (Fig. 12, left) and disdrometer derivations (Fig. 12, right) for the western part (Fig. 12, red) and the eastern part (Fig. 12, blue). Similar to the results in Wu et al. (2019b), when high rain rates occur, the DFR may achieve a state of equilibrium with the effective reflectivity of two bands correlated linearly. Specifically, the DFR values from DPR measurements and DSD derivations approach to approximately 0.0 and 0.5 dB

FIG. 12. Scatterplots of DFR and rain rate from (left) DPR observations and (right) DSD derivations over two parts of the EASM rainband. The PDF of DFR value are also given in each panel.

in the western part, while 0.1 and 0.5 dB in the eastern part, respectively. The DPR–Parsivel normalized mean bias (BIAS) is estimated with the following equation:

$$\text{BIAS} = \frac{\sum(\text{DPR} - \text{DSD})}{\sum \text{DSD}}, \quad (6)$$

where DPR means GPM DPR observational data and DSD means Parsivel measurement values. The comparison results ($\text{BIAS} = 1$ for western, $\text{BIAS} = 0.8$ for eastern) indicate that GPM underestimate the DFR more in the western part than in the eastern part, which suggests that GPM may have better performance over sea than land.

It is additionally suggested in Fig. 12 that compared with Parsivel observations, GPM DPR measurements show distinct lower-frequency ratio of strong precipitation samples. Besides the attenuation caused by instrumental effect or multiple scattering, the possible reason causing such discrepancy could also come from two important aspects: one is the influence of nonuniform beam filling (NUBF) on GPM DPR estimates of path

attenuation by surface reference technique (SRT), which can be clearly noticed for heavy precipitation (Seto et al. 2015). And because of the comparatively stronger convective activity and broader DSD spectral width over the western part than the eastern part, DPR would underestimate more in the western part. The other can be related to the relatively complicated topography over land area, causing a poorer performance of SRT on land than sea.

b. Improvement of GPM DPR retrieval algorithm

The development and refinement of physically based GPM DPR algorithm requires an improved understanding of precipitation microphysics. To fulfill the need for better converting the measurements of GPM DPR to rainfall amounts in midlatitude EASM region, Parsivel² disdrometers are preliminarily employed over this region to develop the optimized constraints for GPM DPR retrieval algorithm.

In the latest GPM DPR product, DSD parameters (i.e., mass-weighted diameter D_m and generalized intercept parameter N_w) are retrieved at each range bin by

FIG. 13. Scatterplots of (a),(b) D_m (mm) and Z_{Ku} (dBZ), (c),(d) D_m (mm) and Z_{Ka} (dBZ), and (e),(f) $\log_{10}(N_w)$ ($\text{mm}^{-1}\text{m}^{-3}$) and D_m (mm) for EASM rainband for (left) the western part and (right) the eastern part. The overlaid red lines represent the fitted curves.

utilizing a dual-frequency algorithm, where DFR are computed and expressed as a function of D_m (Iguchi et al. 2018). It is, however, difficult to retrieve D_m uniquely due to the well-known “dual value” problem already reported by many researches (e.g., Meagher and

Haddad 2006; Wu et al. 2019a,b). That is, at a fixed μ , DFR increases monotonically with D_m only given $D_m > 1$ mm, whereas there exist dual solutions for D_m when DFR is a negative value (Meagher and Haddad 2006). This is because the DFR of a single drop is not a

TABLE 5. Second-degree polynomial relations of D_m - Z_{Ku} , D_m - Z_{Ka} , and $\log_{10}(N_w)$ - D_m derived in two parts of the EASM rainband.

Relation	Data	a	b	c
$D_m = aZ_{Ku}^2 + bZ_{Ku} + c$	Western	0.000 103 5	0.0446	0.0171
	Eastern	0.000 673 3	0.0129	0.5477
$D_m = aZ_{Ka}^2 + bZ_{Ka} + c$	Western	0.001 057 2	0.0159	0.1079
	Eastern	0.000 542 1	0.0185	0.5447
$\log_{10}(N_w) = aD_m^2 + bD_m + c$	Western	0.2878	-2.1643	6.1353
	Eastern	0.0784	-0.9911	4.4200

monotonic increasing function of drop size, but decreases first at small drop diameter before increasing at larger drop diameters (Munchak and Tokay 2008). Thus, a DSD consisting of small drops may have the same DFR value as a DSD including larger drops. To prevent the dual-value phenomenon, we followed the method in Wu et al. (2019b), by utilizing simply the effective radar reflectivities of Ka and Ku bands estimated from disdrometers to obtain their statistical relationship with D_m .

The scatterplots and fitted curves of D_m - Z_{Ku} (Z_{Ka}) and $\log_{10}(N_w)$ - D_m are obtained from disdrometer observations in two parts of the EASM rainband and depicted in Fig. 13. Similar to the findings in Wu et al. (2019b), we also notice a highly correlated monotonic increasing relationship between D_m and Z_{Ku} (Z_{Ka}) and a nonlinear inverse relationship between D_m and $\log_{10}(N_w)$ from Fig. 13. Using a least squares method, the quadratic polynomial relationships of D_m - Z_{Ku} (Z_{Ka}) and $\log_{10}(N_w)$ - D_m are fitted and shown in Table 5. Based on these relationships, D_m can be first derived with given Z_{Ku} or Z_{Ka} , then N_w can be estimated by substituting D_m into $\log_{10}(N_w)$ - D_m relationship. According to the normalized gamma distribution function already presented in section 2b, the DSD can be determined from the derived D_m and N_w after

constraining μ value, and once DSD is derived, the corresponding precipitation variables like rain rate can be calculated eventually. Note that the μ value should be constrained by different rain types, herein $\mu = 5.8$ (5.0) for convective rain and $\mu = 5.9$ (4.7) for stratiform rain in the western (eastern) part as shown in Table 3.

The performance of the derived new relations is evaluated by plotting the rain-rate values from both DPR and Parsivel together in two parts of the EASM rainband as shown in Fig. 14. It is notable that the green dots fit well with black line at low rain rates, while the dots gradually deviate from the line with an increase in rain rate. That indicates Ka band shows better performance at lower rain rates. Quite different from that, Ku band (orange dots) has better performance at higher rain rates. Thus, instead of using DFR for rain retrieval, we propose to combine two bands by estimating light rains with Ka-band relation (D_m - Z_{Ka}) while heavy rains with Ku-band relation (D_m - Z_{Ku}), which is preliminarily shown to have an advantage without any “dual value” problem. Meanwhile, since the performance of Ka and Ku differentiated by low and high rainfall intensity and rainfall microphysics shows large differences between convective and stratiform rain types. The relations between D_m and Z_{Ku} (Z_{Ka}) are also derived for different

FIG. 14. Scatterplots of GPM estimated R_{DPR} vs Parsivel-observed R_{DSD} for two parts of EASM rainband. The green crosses indicate results from Ka band. The orange circles indicate results from Ku band.

TABLE 6. As in Table 5, but for two different rain types.

Region	Relation	Rain type	a	b	c
Western	$D_m = aZ_{Ku}^2 + bZ_{Ku} + c$	Convective	0.000 589 8	0.0209	−0.0264
		Stratiform	0.000 428 7	0.0261	0.2949
	$D_m = aZ_{Ka}^2 + bZ_{Ka} + c$	Convective	0.001 107 1	0.0074	−0.1147
		Stratiform	0.000 557 2	0.0281	0.1913
Eastern	$D_m = aZ_{Ku}^2 + bZ_{Ku} + c$	Convective	0.000 918 1	0.0049	0.0050
		Stratiform	0.000 029 2	0.0403	0.2077
	$D_m = aZ_{Ka}^2 + bZ_{Ka} + c$	Convective	0.001 871 9	−0.0279	−0.0034
		Stratiform	0.000 037 5	0.0426	0.1908

rain types as shown in Table 6. Similarly, it is recommended to use D_m-Z_{Ku} for the retrieval of convective rain while D_m-Z_{Ka} for the retrieval of stratiform rain in both parts of the EASM rainband.

6. Summary and conclusions

Utilizing measurements from OTT Parsivel² disdrometers, as well as multisource satellite observations, we compared the DSD characteristics in the western and eastern parts of the EASM rainband and discussed the recent GPM rainfall retrievals in the EASM season. The study mainly reveals:

- 1) The spectral width of the western part is larger than the eastern part, which is caused by the stronger convection activities in the western part. The aerosols could inhibit warm rain process while enhance cold rain process in the western part. Overall, the small and midsize drops jointly dominate the rainfall in the EASM season, and the convective precipitation of the western part could be identified more maritime-like, while that of the eastern part is between maritime-like and continental-like.
- 2) The power-law σ_M-D_m relations are derived. The western part has larger exponent for σ_M-D_m relations, resulting in broader spectral width with more sufficient large droplets of the convective precipitation in the western part.
- 3) GPM underestimate the DFR more in the western part than in the eastern part, which suggests that GPM may have better performance over sea than land. Empirical D_m-Z_e and N_w-D_m relations were derived preliminarily to improve the GPM DPR rainfall retrieval algorithms in the EASM season.

Notably, the seasonal migration of the EASM rainband was not investigated but worth our future research with effective data collected. Besides, the GPM validation results are obtained during the EASM season, more observations are required to achieve robust validations beyond a specific meteorological regime. Although it is yet to be investigated, the performance of the other used

instruments and products (FY-2E, MODIS, GPCP, and ERA-Interim) under different surface condition are likely to be significant as well. In addition, more detailed evaluation work on the performance of GPM precipitation product will be carried out in future research, like in terms of different rain types, or even different rain-rate classes. The aerosol effects on DSDs also need further investigation, since the effect of the other factors cannot be excluded. From the rainfall identification results in both the western part and the eastern part, it should be noted that the classification method of Bringi et al. (2003) is easily affected by climate conditions (humidity, air pollution, etc.) and geophysical locations (tropics, midlatitudes, etc.). Hence an improved identification method of “maritime-like” and “continental-like” convective precipitation is required with various climate conditions under consideration as well as more marine precipitation data collected.

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