Subdaily Rain-Rate Properties in Western Java Analyzed Using C-Band Doppler Radar

SOPIA LESTARI, ALAIN PROTAT, VALENTIN LOUF, ANDREW KING, CLAIRE VINCENT, AND SHUICHI MORI

Abstract: Jakarta, a megacity in Indonesia, experiences recurrent floods associated with heavy rainfall. Characteristics of subdaily rainfall and the local factors influencing rainfall around Jakarta have not been thoroughly investigated, primarily because of data limitations. In this study, we examine the frequency and intensity of hourly and daily rain rate, including spatial characteristics and variations across time scales. We use 6-min C-band Doppler radar and 1-min in situ data during 2009–12 to resolve spatial rain-rate characteristics at higher resolution than previous studies. A reflectivity–rain rate (Z–R) relationship is derived ($Z = 102.7R^{1.75}$) and applied to estimate hourly rain rate. Our results show that rain rate around Jakarta is spatially inhomogeneous. In the rainy season [December–February (DJF)], rain rate exhibits statistical properties markedly different from other seasons, with much higher frequency of rain, but, on average, less intense rain rate. In all seasons, there is a persistent higher hourly and daily mean rain rate found over mountainous areas, indicating the importance of local orographic effects. In contrast, for hourly rain-rate extremes, peaks are observed mostly over the coastal land and lowland areas. For the diurnal cycle of mean rain rate, a distinct afternoon peak is found developing earlier in DJF and later in the dry season. This study has implications for other analyses of mesoscale rain-rate extremes in areas of complex topography and suggests that coarse-grain products may miss major features of the rain-rate variability identified in our study.

Significance Statement: For many years, Jakarta and its surrounding regions have been repeatedly inundated by flooding triggered by short-duration heavy rainfall or rainfall accumulated over multiple days. Little is known about the distribution of local rainfall and how it differs between seasons. In this study, we used high-resolution C-band Doppler radar during 2009–12 to understand the characteristics of rainfall over this complex topography. The results demonstrate that the rainfall features vary spatially and seasonally. In the wet season, rainfall is more frequent but, on average, lighter relative to other seasons. In all seasons, the highest hourly and daily mean rain rate persistently occurs over the mountains, indicating the vital role of topography in generating rainfall in the region.

Keywords: Complex terrain; Maritime Continent; Tropics; Rainfall; Automatic weather stations; Data quality control; Radars/radar observations; Statistics; Diurnal effects; Seasonal variability; Topographic effects

1. Introduction

Indonesia, located in the central part of the “Maritime Continent” (MC), is a country of mountainous islands with complex coastlines. Indonesia often experiences dramatic hydrometeorological hazards. Jakarta, the most populated city in Indonesia, is particularly prone to floods (Firman et al. 2011; Holderness and Turpin 2015). Over the past century, these intense floods have been associated with extreme rainfall events (Liu et al. 2015; Siswanto et al. 2016, 2017; Trilaksono et al. 2011; Yusuf and Francisco 2009) and exacerbated by sea level rise (Satterthwaite 2008) and land subsidence (Abidin et al. 2008, 2011; Firman et al. 2011). In 2007, 2013, and 2020, flood disasters in Jakarta have resulted in multiple fatalities and economic losses of millions of dollars due to severe damage to infrastructure (BAPPENAS 2008; Badan Penanggulangan Bencana Daerah 2013; Wijayanti et al. 2017).

Despite the large impact of flooding in Jakarta, very few studies have investigated local factors such as the land–sea-breeze circulation and topography influencing mesoscale rainfall extremes in and around Jakarta. This makes model validation for weather prediction particularly difficult. Existing observational studies have examined the role of the local sea-breeze circulation during the dry season in Jakarta (Hadi et al. 2002) and the
effects of local drivers on rainfall intensity and frequency in the Indonesian Maritime Continent (IMC) (Mori et al. 2004; Yamanaka 2016) and East Java (Supari et al. 2012). By using the C-band Doppler radar (CDR), Automatic Weather Station (AWS), sounding, and reanalysis datasets during January–February 2010, analyses for the Jakarta area have concluded that around the city, rainfall intensity reaches a peak in the afternoon over the foothills of mountainous regions and in the night to morning in the coastal areas (Mori et al. 2018). Relative to lower topography, a higher rainfall frequency is observed in the high-elevation regions with a strong seasonal dependence (Lestari et al. 2019).

Rainfall over the IMC is not only influenced by local drivers but also related to large-scale circulations such as El Niño–Southern Oscillation (ENSO), the Indian Ocean dipole (IOD), Madden–Julian oscillation (MJO), and Cross Equatorial Northerly Surge (CENS). The ENSO (La Niña phase) and negative IOD tend to increase rainfall intensity in the transitional and dry seasons (June–November) (Hamada et al. 2012; Kurniadi et al. 2021; Lestari et al. 2019; Zhang et al. 2016), while the active phase of MJO and CENS are associated with increased rainfall during the wet season (November–April) (Hidayat and Kizu 2010; Matsumoto et al. 2017; Peatman et al. 2014). Although many studies examined the effect of large-scale variability on daily rainfall from long-term datasets or subdaily rainfall on short datasets, further work is required to investigate diurnal rainfall variations using subdaily data at high spatial resolution for a longer period to provide a baseline for model evaluation and analysis of local rainfall processes.

Over the IMC, few subdaily rain-rate gauge datasets are available. Weather radars are a valuable tool to characterize the spatial variability of rainfall (Overeem et al. 2010; Peleg et al. 2018; Smith et al. 2012), for flood monitoring (Seo et al. 2015; Smith et al. 2007), operational flood forecasting (Barge et al. 1979; Kouwen and Soulis 1994; Moore et al. 2005), and model evaluation (Mandapaka et al. 2013; Overeem et al. 2010). In earlier studies, several months of weather radar datasets have been used to investigate local variability in rainfall over Jakarta (Katsumata et al. 2018; Mori et al. 2018) and the western part of IMC (Kamimera et al. 2012; Paski and Permana 2018; Yokoi et al. 2019; Yokoi et al. 2017). These data have been applied to observing the development and propagation of convection during the 2007 (Wu et al. 2007) and 2013 (Wu et al. 2013) Jakarta floods. Despite the usefulness of radar rainfall estimates in a region where there is high spatial heterogeneity in rainfall, there has been a paucity of comprehensive statistical studies to date. In this study, we utilize multiyear time series of C-band radar observations to characterize the statistical properties of rainfall over Jakarta and the surrounding areas, with a focus on the spatial distribution, the regional and seasonal variability, the diurnal cycle, and the impact of local topography.

2. Methods

a. Study domain

This study area is focused on west Java including Jakarta. This region is characterized by complex topography, including

FIG. 1. The research area is taken from the inset map of the IMC. The radar coverage has four encircled range distances of 30, 60, 90, and 105 km from the center of radar. The two orange dots represent the high-resolution AWS stations at Balitklimat and Serang. The three violet dots show high mountains: Halimun Salak (1.9 km), Salak (2.2 km), and Pangrango (3 km). The red dot in the center of the radar domain indicates the location of the CDR at Serpong.

1) the urban area of Jakarta at low altitude in the northern part of the radar domain and bordered by coastal areas, 2) suburban areas at slightly higher altitudes, and 3) high mountains extending from the west to the east in the southern part of the radar domain. The radar also covers the bay of Jakarta in the northern part of the study area (Fig. 1). The determination of different regions (coastal waters, coastal land, lowland, and low and high mountains) is described in section 3c.

The analysis period was from January 2009 to December 2012, which covers multiple wet seasons [December–February (DJF)], transitional periods to the dry season [March–May (MAM)], dry seasons [June–August (JJA)], and transitional periods to the wet season [September–November (SON)].

The CDR and AWS data availability are shown in Table 1. The CDR, located at Serpong, runs a 6-min volumetric sequence, comprising 18 sweeps at elevations ranging from 0.6° to 50° and a maximum range of 105 km (volume scan mode). The beamwidth and range gate of the CDR are 0.98° and 0.125 km, respectively. It operated between January 2009 and early January 2013 (four years) except for March–May 2009; December 2011; and September 2012. In this study, we use local time (LT) for the whole analysis (+7 h from UTC).
b. Data

The 1-min AWS data were obtained from the last field campaign of the Hydrometeorological Array for Intraseasonal Variation–Monsoon Automonitoring (HARIMAU) project (2005–10) (Mori et al. 2018). The AWS data at Pramuka island were excluded from this study because it lies within a radar beam blocking area. We used rainfall AWS data from two stations at Serang (inland) and Balitklimat (mountain) together with the radar and disdrometer data to develop a relationship between reflectivity Z and rain rate R (the Z–R relationship) for the Jakarta area. The data availability covered the period January–December 2010 at Balitklimat and January–February 2010 at Serang. These datasets have been quality controlled by checking spurious data, such as negative or unrealistic rainfall intensity. As an additional check, the 1-min AWS data were aggregated to hourly and daily rainfall, for comparison with the annual maximum of hourly and daily data from previous work (Siswanto et al. 2016).

The small number of AWS stations with 1-min data and the short time series with missing months of AWS data might impact the derivation of our Z–R relationship. For example, the spatial and seasonal variability might not be fully captured. However, 1-min AWS data are most useful for estimating heavy rain rate over very short periods (Vuerich et al. 2009), which often occurs in the Jakarta area (Brinkman and Hartman 2008). Although the station data were accumulated to hourly values for the purposes of deriving the Z–R relationship, the 1-min resolution allowed us to conduct a sensitivity study of the impact of accumulation period on the derived Z–R relationship, as discussed in section 2d(4).

A Thies Clima disdrometer installed in Serpong (−6.3592°, 106.6735°) (red dot in Fig. 1) was also used during August 2016–May 2020 to investigate the statistical properties of the rain drop size distribution (DSD; the distribution of the number of raindrops per unit volume (m⁻³ mm⁻¹) as a function of droplet diameter) in this region and derive relationships between radar parameters (reflectivity and specific attenuation) and DSD parameters. The T-matrix method (Leinonen 2014) was also used to compute Z from the disdrometer DSDs and derive an alternative Z–R relationship that was compared with the Z–R formulated using nearly collocated radar and rain gauge measurements. To provide some large-scale context to our radar observations, the zonal u and meridional v winds at the 850-hPa level of ERA5 with 0.25° × 0.25° spatial resolution were used (Copernicus Climate Change Service 2017) with the same time step as the radar.

c. Definition of rain-rate indices

Global climate models are known to overproduce the frequency of occurrence of light rain rates (Nguyen et al. 2015). Rain-rate indices have therefore been selected to characterize the statistical properties of rain rate at different temporal resolutions, as well as their spatial and seasonal (DJF, MAM, JJA, and SON) variability (Table 2): mean hourly and daily rain rate (Rₘ and Rₚ respectively), hourly and daily rain rate [simple hourly intensity index (SHII) and simple daily intensity index (SDII), respectively], and rain frequency (W_freq). For the SDII, Rwj is the accumulation of daily rainfall for the rainfall intensity that is greater than or equal to 1 mm, w is a day with rainfall intensity that is more than or equal to 1 mm (w ≥ 1 mm), and “j” is the period of our observational data record, that is from 2009 to 2012. If W represents the number (sum or count) of w ≥ 1 mm in j, then the SDII = sum(Rwj)/W (WMO 2009). SHII is the same definition as SDII [SHII = sum(Rwj)/W], but now W in SHII is the number (sum or count) of hours w ≥ 1 mm h⁻¹ in j.

Analyzing rain-rate properties using these indices also provides a reference to which models can be compared in this region. We determine 1 mm as the minimum accumulated rain-rate threshold to estimate Rₘ, Rₚ, SHII, SDII, and W_freq following recommendations from the Joint Commission for Climatology (CCL)/CLIVAR/JCOMM Expert Team on Climate Change.

### TABLE 1. The CDR and AWS data availability for different years.

<table>
<thead>
<tr>
<th>Measure</th>
<th>Location</th>
<th>2009</th>
<th>2010</th>
<th>2011</th>
<th>2012</th>
</tr>
</thead>
<tbody>
<tr>
<td>1-min AWS</td>
<td>Balitklimat</td>
<td>—</td>
<td>Jan–Oct; Dec</td>
<td>—</td>
<td>—</td>
</tr>
<tr>
<td></td>
<td>Serang</td>
<td>—</td>
<td>Jan–Feb</td>
<td>—</td>
<td>—</td>
</tr>
</tbody>
</table>

### TABLE 2. Definition of rain-rate indices.

<table>
<thead>
<tr>
<th>Measure</th>
<th>Indices</th>
<th>Units</th>
<th>Definitions</th>
</tr>
</thead>
<tbody>
<tr>
<td>Mean rain rate</td>
<td>Mean daily rain rate R_day</td>
<td>mm day⁻¹</td>
<td>Total amount of daily rain rate ≥ 1 mm day⁻¹ divided by total no. of days with rain rate ≥ 0 mm day⁻¹</td>
</tr>
<tr>
<td></td>
<td>Mean hourly rain rate R_h</td>
<td>mm h⁻¹</td>
<td>Total amount of hourly rain rate ≥ 1 mm h⁻¹ divided by total no. of hours with rain rate ≥ 0 mm h⁻¹</td>
</tr>
<tr>
<td>Simple daily intensity index (SDII)</td>
<td></td>
<td>mm day⁻¹</td>
<td>Ratio of the total amount of daily rain rate ≥ 1 mm day⁻¹ to no. of days with rain rate ≥ 1 mm day⁻¹</td>
</tr>
<tr>
<td>Simple hourly intensity index (SHII)</td>
<td></td>
<td>mm h⁻¹</td>
<td>Ratio of the total amount of hourly rain rate ≥ 1 mm h⁻¹ to no. of hours with rain rate ≥ 1 mm h⁻¹</td>
</tr>
<tr>
<td>Rain frequency W_freq</td>
<td>×100 (%)</td>
<td>Frequency of hours where rain rate ≥ 1 mm h⁻¹ divided by frequency of hours where rain rate ≥ 0 mm h⁻¹</td>
<td></td>
</tr>
</tbody>
</table>

d. Quality control of CDR

Quality control of CDR has been performed to reduce uncertainty in the data. Initially, noise and nonmeteorological echoes were filtered out using the Py-ART algorithm (Helmus and Collis 2016). Further quality control of CDR data includes the reflectivity calibration, attenuation correction, and beam blocking analysis (Fig. 2), as described in sections 2d(1)–(3), respectively.

1) RCA TECHNIQUE

A calibration check is needed to ensure the best performance of radar in accurately estimating rain rates (Raghavan 2003). Numerous studies have applied calibration techniques using rain gauges, DSD measurements (Ulbrich and Lee 1999), neighboring radars (Ribaud et al. 2016), and fixed targets such as ground clutter echoes (Vaccarono et al. 2016). A robust result using the ground clutter method was found in previous research (Borowska and Zrnic 2012; Silberstein et al. 2008).

Detection of ground clutter was used to accurately track changes in calibration over the four years of radar observations using the so-called relative calibration adjustment (RCA) technique (Louf et al. 2019; Wolff et al. 2015). The main assumption of the method is that a change in ground clutter reflectivity can be attributed to a change in calibration of the radar. The principle of this technique is as follows. The first step is to identify permanent ground clutter (e.g., buildings, bridges, tall trees, and pylons) within a 10-km range. The 95th percentile of the \( Z_c \) distribution is then used to track changes in calibration. Since the technique does not provide an absolute calibration, we also use, as explained later, disdrometer observations to obtain the absolute calibration reference.

Figure 3 shows changes of the 95th percentile of clutter reflectivity for the March–July 2010 period, and Fig. S1 in the online supplemental material shows clutter reflectivity after correction. An abrupt change of ground clutter reflectivity is clearly observed (Fig. 3). These variations have been subtracted from the initial reflectivity to establish a baseline for further calibration of the whole radar dataset with a single reference.

The smoothing filter was applied prior to RCA adjustment. We removed 6-min points that were regarded as outliers, and then we computed the daily average of the 6-min 95th percentile of ground clutter reflectivity. We identified periods when the RCA of daily mean was stable (black lines on Fig. 3) and compared the value of that black line with the baseline to obtain the offset. The offset was then applied to all 6-min data of that period.

2) ATTENUATION CORRECTION

The reflectivity of the CDR is significantly reduced by attenuation through scattering and absorption by hydrometeors along the pathway of the radar beam, especially in strong convective cells (Hildebrand 1978). In this study, the effect of attenuation has been corrected following a modified version of Krämer’s technique (Jacobi and Heistermann 2016). The method uses the relationship between specific attenuation \( A \) and reflectivity \( Z \) to derive the path-integrated attenuation (PIA), which then can be used to correct \( Z \):

\[
A = cZ^d.
\]
PIA corresponds to the specific two-way attenuation accumulated over the whole propagation path and recursively corrects the reflectivity according to

$$\text{PIA}_i = c \left( Z_i + \sum_{j=0}^{i-1} c Z_j^d \right) 2\Delta r,$$

where $\Delta r$ is the radar range bin resolution. In Eqs. (1) and (2), $c$ and $d$ are empirical constants. The PIA, (dB) is then added to reflectivity $Z_i$ (dBZ) to obtain the corrected reflectivity $Z_{\text{corr},i}$:

$$Z_{\text{corr},i} = Z_i + \text{PIA}_i,$$

Parameters $c$ and $d$ are dependent on the DSD properties, temperature, number and shape of the drops, and the wavelength of the radar. Besides miscalibration, a significant error in the estimation of $A$ and nonrepresentative $c$ and $d$ parameters resulting from unknown rain characteristics can result in uncertainty in PIA. In our study, both parameters were estimated from the reflectivity ($Z$)-attenuation ($A$) relationship simulated from disdrometer DSDs at Serpong. The following power-law relationship between $A$ and $Z$ has been obtained: $A = 0.00018Z^{0.7408}$. These attenuation coefficients $c = 1.8 \times 10^{-4}$ and $d = 0.7408$ were used to correct $Z$ for attenuation.

To prevent instabilities in the attenuation correction, constraints were applied to the PIA. Past studies either suggested to limit the PIA between 4 and 5 dB (Harrison et al. 2006), or to apply constants on coefficients $c$ and $d$ (Krämer and Verworn 2008). It is also suggested that when $Z_{\text{corr},i}$ [Eq. (3)] reaches values greater than 59 dBZ, it indicates that the attenuation correction is unstable (Krämer 2008).

Our study follows all previous suggestions by implementing a criterion of “double constraint” to avoid unstable corrections (Jacobi and Heistermann 2016). It includes specifying the maximum threshold of PIA to 20 dB or $Z_{\text{corr},i}$ to 59 dBZ. Instead of using $c$ and $d$ as proposed by Krämer, we used $c = 1.8 \times 10^{-4}$ and $d = 0.7408$ derived from our disdrometer fitted to the local rain-rate properties around Jakarta, as mentioned above. Accordingly, we computed the PIA by successively correcting the reflectivity by using the wradlib open-source software (https://docs.wradlib.org/en/stable/notebooks/attenuation/wradlib_attenuation.html).

A comparison between reflectivity before and after the correction for a selected radar ray on 31 January 2010 is shown in Fig. 4a. The corrected and uncorrected reflectivity at 0001 LT at azimuth 273° with different ranges of radar is also presented in Fig. 4b. The scheme corrects the attenuated reflectivity as shown by an increase of corrected reflectivity with increasing distance from the radar.

3) BEAM BLOCKING ANALYSIS

Besides ground clutter, other permanent objects, such as mountains, nearby trees, and buildings, block all or part of the radar transmitted signal. This problem is particularly problematic for this radar where there is blocking by mountains (to the southeast and southwest), nearby tall trees, and high-rise buildings (to the north-northeast). Beam blocking reduces reflectivity and therefore leads to underestimated rain rates if not corrected for or masked out. Regions of total beam blocking have been identified and excluded from the analysis. Figure 5 shows the probability maps of the number of points with reflectivity $>10$ dBZ (Ref10). Ref10 is used as a proxy of minimum threshold to show the occurrence of the rainfall events. A digital terrain elevation map has been used to mask the areas where beam blocking is expected from terrain only to be greater than 3 dB. However, despite this first beam blocking analysis, the probability maps at different elevation angles show that for many areas located in the north-northwest–southeast sector, and also isolated areas to the southeast, southwest, and northwest, there is still evidence of strongly reduced probability (corresponding to beam blocking) in the first (0.97°) elevation (Fig. 5a), second (1.69°) elevation (Fig. 5b), and, to a lesser extent, the third (2.13°) elevation (Fig. 5c). The least radar beam blocking was found in the fourth elevation of the radar (3.19°) (Fig. 5d).

This result shows that higher-elevation data are needed to produce spatial rain-rate maps where the impact of beam blocking is minimal. The problem with using increasingly higher elevations is that at long ranges the radar beam reaches the melting-layer altitude (around 4-km height in this region).
measuring a mixture of liquid, melted, and ice particles that are not representative of the rain rate at the ground. The effect of snow above melting layer can be seen as a reduced probability of reflectivity greater than 10 dBZ at long range on the third and fourth elevations (Figs. 5c,d).

To mitigate the impact of beam blocking on our results, a composite map has been produced using the third and fourth elevations and an empirically determined Ref10 threshold of <0.03% to screen out beam blocking areas. Merging between the two elevations was done by taking the average reflectivity from the two elevations for each pixel. This approach also required us to restrict the maximum range for the third elevation to 89 km and the fourth elevation to 67 km from the radar, which is equivalent to around 3.7-km height, to exclude melting-layer contamination. The resulting spatial map using the beam blocking analysis is shown in Fig. 6a. Although some areas north from the radar are probably still slightly contaminated by beam blocking, the other parts of the domain appear to produce a continuous map of probability of precipitation.

4) THE Z–R RELATIONSHIP OVER THE JAKARTA AREA

Rain rate can be estimated from radar reflectivity data by developing an empirical relationship between collocated radar pixels Z and measurement data from stations R (herein, the Z–R relationship). Similar to the method applied for the beam blocking analysis, we use the average value of reflectivity from the third and fourth elevation angles to compare with the rain rates on the ground.

The corrected reflectivity Z is interpolated onto the horizontal map in the cartesian coordinate with 1-km pixel size. One-minute rain rate R from the AWS at Balitklimat and Serang and extracted Z in the same pixel as the AWS were used to determine the Z–R relationship. The 6-min radar reflectivity Z

---

**Fig. 5.** A probability map of frequency of reflectivity >10 dBZ. Frequency of reflectivity < 0.03 is masked since probability distribution of this frequency is constant during the study period (2009–12). The persistent high probability indicates the existence of the beam blocking. The beam blocking (masking) areas over the radar domain shows at the (a) first (0.97°), (b) second (1.69°), (c) third (2.13°), and (d) fourth (3.19°) elevations. Colored contour lines illustrate the altitude of the mountains: + 750 m (gray lines) and + 1500 m (black lines).
in linear units was averaged over 1 h. Meanwhile, the 1-min AWS $R$ was accumulated over 1 h. When the signal-to-noise ratio (SNR) is below the threshold, we did not use any value for $Z$; instead we indicated it as “not a number” (NaN). An SNR below the threshold is used to identify noise in data.

From the 1-h average $Z$ and 1-h accumulated $R$, we paired nonzero hourly $Z$ and $R$ values to derive coefficients $a$ and $b$ from the equation of $R$ (mm h$^{-1}$) = $aZ^b$ using a total least squares fitting procedure as for the radar data.

In this study, an independent $Z$–$R$ relationship ($Z = 103.6R^{0.76}$) was also derived using hourly averages from disdrometer data only, using $Z$ from T-matrix calculations, and applying the same total least squares fitting procedure as for the radar data.

There has been considerable research on using disdrometer data to calibrate rain-rate estimation from radar reflectivity measurements (Bringi et al. 2001; Hazenberg et al. 2014; Park et al. 2005). Since our disdrometer is approximately 1 km away from the radar, we can use reflectivity simulated from the disdrometer observations (using the T-matrix calculations) to statistically calibrate the radar through comparisons of $Z$–$R$ relationships from disdrometer and radar–rain-rate observations. It must be noted that the exponent $b$ of the $Z$–$R$ relationship obtained prior to calibrating the Serpong radar (but including RCA corrections) was very close to that obtained from disdrometer only (1.75 vs 1.76), which is a very encouraging result that indicates the robustness of the approach.

The calibration offset for the radar reflectivity was estimated based on the difference in the fit line intercepts, and it was found that the radar reflectivity obtained after correcting for changes in calibration with the RCA technique was 5.3 dB lower than the disdrometer reflectivity. Therefore, we corrected our whole radar reflectivity datasets by 5.3 dB to match the disdrometer $Z$–$R$ relationship.

Given the initial agreement between the radar and disdrometer $Z$–$R$ exponents, we obtain a $Z$–$R$ relationship using radar that is similar to the disdrometer after calibrating the reflectivity. A relationship between $Z$ (mm$^6$ m$^{-3}$) and $R$ (mm h$^{-1}$) around Jakarta was formulated ($Z = 102.7R^{1.75}$) that is based on 388 hourly points from radar reflectivity and rain rate from the station, with a significant relationship between $Z$ and $R$ (correlation coefficient $r = 0.6$) at the 95% confidence level as tested by Spearman rank correlation (Fig. 7). A total least squares fit was used in this study, which takes into account the errors that likely occur both from observed and predicted variables (Nievergelt 1994). Previous studies also found that the correlations between radar-derived rain rate and gauge rain rate ranged from 0.6 to 0.7 (Sebastianelli et al. 2013) and from 0.728 to 0.9 (Protat et al. 2019). In addition, the tropics (Protat et al. 2019) as pulled from the same total least squares fitting procedure as for the radar data.

Note that we also conducted sensitivity tests by deriving the $Z$–$R$ relationship from pairs of 6-min instantaneous radar and 6-min AWS rainfall, pairs of 12-min reflectivity radar and AWS rainfall, and pairs of 30-min reflectivity radar and AWS rainfall. Yet, the $Z$–$R$ derivation from the accumulated 1-h AWS rainfall and averaged 1-h radar reflectivity was closest to the disdrometer $Z$–$R$. This is because the higher the temporal resolution is, the higher the rain-rate detection level (mm h$^{-1}$) will be, yielding unstable fits that are not well constrained by lower rain rates (not shown). Only results using accumulation of 30 min or more were close to the hourly accumulation results. Therefore, we opted for the hourly averaging/accumulation approach. Previous studies also applied the sensitivity tests in deriving $Z$–$R$ using different time resolutions to find the optimum result of radar rain-rate estimates (Hasan 2016; Mapiam et al. 2014; Morin et al. 2001).
well as Indonesia from disdrometer at Sumatera, Kalimantan, Sulawesi, and Papua (Marzuki et al. 2013), and our Z–R from radar and disdrometer. The radar Z–R for the Jakarta region appears to systematically produce higher rain rates for lower reflectivity relative to other relationships. Although we do not have the appropriate aerosol observations or modeling tools to fully test this hypothesis, we speculate that this difference in Z–R relationship could be related to the high levels of pollution and associated aerosol–deep convection interactions potentially producing higher concentrations of small droplets and lower concentration of bigger droplets (Bréon et al. 2002; Dave et al. 2017), resulting in higher rain rate than usual in the low to moderate reflectivity range. Earlier studies also reported that high pollution might suppress very high rain rates not only by preventing the development of cold-precipitation processes in the cloud, but also by reducing the radius of cloud droplets. As a result, the coalescence of raindrops might be inefficient (Givati and Rosenfeld 2004; Jirak and Cotton 2006; Rosenfeld 2000). However, to our knowledge, over the IMC, no studies have examined the relationship between the size of the raindrops and the seasonal variation of aerosols.

5) GAUGE AND CDR COMPARISON

The reflectivity map was converted to a rain-rate map for further analysis of rain-rate distribution and a comparison between rain rate from station and radar. First, the 6-min reflectivity map was averaged to hourly resolution then the Z–R relationship was applied to obtain an hourly rain-rate map.

The probability distribution function (PDFs) of rain rate from the radar and station observations for 2010 are shown in Fig. 8. The PDFs were plotted from lighter (1–10 mm h⁻¹) to heavier rain rate (10–100 mm h⁻¹). The radar rain rate was selected based on the nearest pixels to the station coordinates. A comparison between PDFs of rain rate from radar and AWS at Balitklimat (Fig. 8a) and Serang (not shown) generally showed similar features.

For the low rain rates ≤ 10 mm h⁻¹, the radar captured relatively well the rain-rate distribution (Fig. 8a). In general, despite fewer occurrences of rain rates greater than 50 mm h⁻¹ captured by the radar, there is a good agreement between the radar and stations rainfall PDFs. Statistics of hourly rain rate from radar and AWS in 2010 show that the root-mean-square error (RMSE), a Spearman correlation, and mean bias between two datasets are 4.99, 0.58, and 0.07 mm h⁻¹, respectively. Figure 8b also shows similar values of frequency of rainfall detected (above 1 mm) and the 95th and 99th percentiles between collocated radar and stations. This indicates a good match of rainfall between radar and stations.

3. Results

Characterizing the statistical properties of rain rate provides an important baseline for improving model accuracy and understanding processes. Here we analyze statistical properties of rain rate in our region of interest, as well as the seasonal and diurnal variability of rain rate. Processes responsible for this variability are also discussed where possible, but a full process-based study is beyond the scope of this investigation.

a. Statistical properties of rain rate in the Jakarta region

The intensity of hourly and daily rain rates and hourly rain frequency are examined in this section to document the rain-rate climatology over the whole radar domain. Over our study period, the rain rate in the Jakarta region is characterized by a mean daily rain rate of 11.9 mm day⁻¹ (Fig. 9c), a 99th percentile of mean daily rain rate of 148 mm day⁻¹ (Fig. 9e), a daily rain rate of 12 mm day⁻¹ (Fig. 9e), a mean hourly rain rate of 0.58 mm h⁻¹ (Fig. 9d), a 99th percentile of mean hourly rain rate of 14 mm h⁻¹ (Fig. 9f), and an hourly rain rate of 9 mm h⁻¹ (Fig. 9f) (Table 3). The spatial map of $R_{day}$ (Fig. 9a) distribution shows that the highest $R_{day}$ in the domain peaks around ≥21 mm day⁻¹ over the mountainous region from the southwest to the southern part of the radar coverage area accompanied by the dominant westerly winds at 850-hPa level from the west coastal land with the average speed of around 1–2 m s⁻¹. In general, the $R_{day}$ has a low correlation ($r < 0.3$) with the westerly wind speed at 700 and 850 hPa (Fig. S2 in the online supplemental material). For the geographic location, relative to the southern part of the radar domain (lowland and low and high mountains), there is a higher positive correlation ($0.2 < r < 0.3$) between $R_{day}$ and westerly wind.
speed over the northern part of radar domain around the coastal water and coastal land as well as in the western part of the radar domain.

The $W_{\text{Freq}}$ (Fig. 9b) is also highest (around 10%) over the mountainous areas. The consistent spatial patterns in mean daily rain rate and hourly frequency indicate the highly localized nature of rain rate shown by large spatial variability.

The PDF of rain rates (Fig. 10) illustrates how the hourly rain rate observed from radar is distributed statistically. In all seasons, the highest probability of rain rate occurs at low rain rates of around 0.5 mm h$^{-1}$ and probability of occurrence decreases with increased rain rates.

### Seasonal variability of rain rate in the Jakarta region

The seasonal variability of hourly and daily rain rate is shown in Figs. 9 and 10. There is a well-defined seasonal pattern in each index, with the largest average rain rate consistently found in DJF and $R_h$ and $R_{\text{dail}}$ more than 2 times as great in DJF as in JJA (Figs. 9c and d).

The only lower rain-rate index in DJF is the hourly rain rate (SHII), which is below the annual average (Fig. 9f). The higher hourly rain rates in DJF are a result of a much higher frequency of rain occurrence (Fig. 10), but, on average, less intense rain rate (Fig. 9f). Although SHII in DJF is lower than all seasons, the 75th, 95th, and 99th percentiles of hourly rain rate are still higher than in the other seasons (Fig. 9f), indicating that the low value of SHII is mostly due to more frequent light rain rates.

The seasonal variability of the rain-rate PDF (Fig. 10) shows that the nature of SON rain rate is relatively different from other seasons, mainly in the tail of the distribution for higher rain rates $> 10$ mm h$^{-1}$. In the wet season (DJF), the frequency of light to heavy rain (rain-rate amount $\leq 60$ mm h$^{-1}$) is much higher than in all other seasons. The PDFs of very high rain rates ($>60$ mm h$^{-1}$) are based on fewer samples but show a higher proportional occurrence of very high rain rates in SON than other seasons. A high occurrence of rain rates in SON might be due to interannual variability.

To examine the influence of the interannual variability on seasonal rainfall, additional analysis was done by comparing the monthly rainfall climatology (1974–2016) and monthly radar rain rate during ENSO and IOD (2009–12). The rainfall climatology was computed on the basis of two station datasets.
at Kemayoran (coastal land) and Citeko (high mountain). It was found that during La Niña in 2011, in JJA and SON, the rain rate was higher than the climatology. The result is also consistent with Lestari et al. (2019), which found that a higher occurrence of heavier rain occurs in other La Niña seasons, particularly in SON. In contrast, during the El Niño event, the rain rate was lower than climatology. Interestingly, two events occur at the same time, La Niña and positive IOD in 2012. In this year, the rain rate is lower than climatology, indicating that the effect of ENSO and IOD varies depending on the local topography (Yamanaka 2016). In addition, in 2012, there was a negative anomaly of SSTs over the eastern Indian Ocean.

![FIG. 9. In all seasons: (a) Mean daily rain rate $R_{\text{day}}$ (mm day$^{-1}$). The daily average of wind vector and the speed of $u$ and $v$ components at 850-hPa level are also shown. (b) Frequency of hourly rain $\geq 1$ mm h$^{-1}$ $W_{\text{freq}}$ (%). Over radar domain: (c) mean daily rain rate $R_{\text{day}}$ (mm day$^{-1}$); (d) mean hourly rain rate $R_{h}$ (mm h$^{-1}$); (e) total amount of daily rain rate $\geq 1$ mm day$^{-1}$ divided by frequency of daily rain $\geq 1$ mm SDII (mm day$^{-1}$) and 75th, 95th, and 99th percentile of daily rain rate; and (f) total amount of hourly rain rate $\geq 1$ mm h$^{-1}$ divided by frequency of hourly rain $\geq 1$ mm SHII (mm h$^{-1}$) and 75th, 95th, and 99th percentile of hourly rain rate.]

Unauthenticated | Downloaded 12/31/23 08:33 PM UTC
near Java Island. These lower-than-normal SSTs seems also to result in a lower rain rate over Jakarta (Fig. S3 in the online supplemental material). However, to further explore the effect of high interannual variability on robust climatology of rain-rate characteristics, it would require a much longer radar data-set than we have available.

In terms of the diurnal cycle, the $R_h$ for all years (Fig. 11a: black line) is characterized by a broad afternoon peak between 1400 and 1800 LT, centered on 1500 LT with the amplitude of around 1.8 mm h$^{-1}$. This late-afternoon peak in the diurnal cycle is typical of continental rain rate in the tropics (Kumar et al. 2013; May et al. 2012; Nguyen et al. 2015). Note that the other typical peak usually found in the tropics around 0200 and 0400 LT, generally associated with oceanic convection, is not clearly present in our region of interest when considering all seasons together. Because of the beam blocking, the oceanic region covers only a small fraction of the domain. We described this region as coastal waters as explained in more detail in section 3c.

In turning our attention to the seasonal variability of the diurnal cycle of rain rate, it is seen that, although all seasons are characterized by a broad afternoon rain-rate peak, there are differences in timing and amplitude of diurnal variations of $R_h$ (Fig. 11a), with DJF and MAM rain rate peaking earlier (by about 1 h). In contrast, the JJA peak is delayed (by about 1 h) and of much smaller amplitude than the annual amplitude. The $R_h$ in DJF produces a slightly higher rain rate than the annual mean peak of the afternoon rain rate, whereas $R_h$ in SON has a much higher rain rate than the annual mean peak of rain rate. The PDFs also show that relative to the climatology the occurrence probability of heavy rain rate in SON is higher (Fig. 10b). The slightly higher $R_h$ in the early morning in the annual statistics is primarily associated with DJF, which is also characterized by the highest amount of early morning rain rate when compared with other seasons (2 times the average, and 4 times that for JJA and SON over the 0000–1000 LT period) (Fig. 11a).

Figure 11 also allows for a more in-depth understanding of the sources of seasonal variability of the diurnal cycle of rain rate. The much higher DJF $R_h$ between 0000 and 1000 LT (Fig. 11a) is mostly associated with a substantial increase in $W_{freq}$ (Fig. 11b) and a moderate increase in SHII (Fig. 11c). The previously discussed observation of DJF rain rate being of higher frequency of occurrence but lower average intensity (from Fig. 9f) is in fact due to a higher frequency across the whole diurnal cycle (Fig. 11b), but the lower intensity overall results from less intense rain rate at the wettest time of day (in the afternoon) and more intense rain rate in the morning relative to MAM, JJA, SON, and all seasons (Fig. 11c). The behavior of the JJA rain rate over the diurnal cycle is also markedly different, with the lower rain rate in the afternoon peak (relative to DJF, MAM, SON, and all seasons) being mostly driven by a much smaller frequency of more intense than average rain rate (Fig. 11). In addition, a distinct feature in diurnal rain rate in SON is observed, with both higher frequency and intensity relative to annual rain rate in the afternoon.

Table 3. Rainfall amplitude for whole seasons over the whole radar domain.

<table>
<thead>
<tr>
<th>Indices</th>
<th>Time increment</th>
<th>Amplitude</th>
</tr>
</thead>
<tbody>
<tr>
<td>Mean rain rate</td>
<td>Daily</td>
<td>11.9 mm day$^{-1}$ (Fig. 9e)</td>
</tr>
<tr>
<td></td>
<td>Hourly</td>
<td>0.58 mm h$^{-1}$ (Fig. 9d)</td>
</tr>
<tr>
<td>99th percentile of mean rain rate</td>
<td>Daily</td>
<td>148 mm day$^{-1}$ (Fig. 9c)</td>
</tr>
<tr>
<td></td>
<td>Hourly</td>
<td>14 mm h$^{-1}$ (Fig. 9f)</td>
</tr>
<tr>
<td>Rain rate</td>
<td>Daily</td>
<td>12 mm day$^{-1}$ (Fig. 9e)</td>
</tr>
<tr>
<td></td>
<td>Hourly</td>
<td>9 mm h$^{-1}$ (Fig. 9f)</td>
</tr>
</tbody>
</table>

Figure 10. PDFs of hourly rain rate in DJF (blue), MAM (orange), JJA (green), SON (red), and all seasons (black), derived from radar reflectivity for the (a) light rain rate ($\leq$ 10 mm h$^{-1}$) with bin size = 0.5 mm h$^{-1}$ and (b) high rain rate (>10–100 mm h$^{-1}$) with bin size = 5 mm h$^{-1}$. 

Unauthenticated | Downloaded 12/31/23 08:33 PM UTC
c. The modulation of rain-rate properties by topography in the Jakarta region

The spatial distribution of rain rate is analyzed in this section to better understand the variability within the radar domain related to local-scale effects over the different seasons. It has been shown in the literature that local drivers such as land–sea breeze and topography significantly contribute to rain-rate variability, including seasonal and diurnal variability (Yamanaka et al. 2018).

Spatial maps of $R_h$ (figure not shown) and $R_{day}$ (Fig. 12) show that the highest mean hourly and daily rain rate occur in the wet season (DJF). The western lowland area, mountainous sites in the southwest (Mount Halimun Salak) and southern part (Mount Salak) of the radar domain have the highest wet-season rain rate, of around $\approx 1.2$ mm h$^{-1}$ and $\approx 20$ mm day$^{-1}$. In JJA, only a small area around the Salak mountain receives a rain rate of around 18 mm day$^{-1}$. The fact that the highest rain rate is confined to the mountainous regions in the dry and transitional seasons suggests that the enhanced rain rate at these times of year is associated with a local orographic effect. The prevailing daily winds in DJF and MAM are westerly, with the highest wind speed found during DJF (5 m s$^{-1}$). In JJA and SON, the easterly wind is predominant with lower wind speed than that observed in DJF (Fig. 12).

In contrast to $R_{day}$ in Fig. 12, the SHII maps show that rain rate tends to be dominated by heavier rainfall in MAM, JJA, and SON (Fig. 13), even though the most rain occurs in DJF. Although DJF also experiences heavy rainfall, the SHII decreases because of frequent lighter rain rate (Fig. 11b). SHII and $R_h$ differ in spatial distribution of rain rate in all seasons, indicative of a substantial contribution of light rain rate ($< 1$ mm h$^{-1}$) to $R_h$ in MAM, JJA, and SON. It is also, and perhaps more, due to the substantial contribution of no rainfall hours on $R_h$.

The spatial patterns of $R_h$ at hourly resolution were analyzed in DJF, MAM, JJA, and SON to understand how the timing of the highest mean rain rate over different topography changes between seasons (Fig. 14). We filtered for days with a diurnal component to the rainfall by choosing days where the peak of the $R_h$ was $>1.5$ times as high as the mean $R_h$ within 24 h for a given grid point. Over the mountainous area, it is apparent that in DJF the timing of the strongest $R_h$ peak is identified between 1300 and 1500 LT whereas most of the lowland has a peak at around 1400–1600 LT (Fig. 14). In contrast, the maximum of $R_h$ is found in the night and morning between approximately 2000 and 1000 LT over the surrounding coastal waters and coastal land.

The main features of the seasonal variability of the timing of the diurnal cycle of rain rate (as shown in Fig. 11) are...
an earlier afternoon peak in DJF and a later peak in JJA. From the spatial maps in Fig. 14, it appears that DJF is characterized by much earlier peaks in the western part of the domain and over the coastal waters (other seasons are much more similar), and earlier peaks over the land and mountainous areas. The times of the diurnal peaks in the western part of the domain and coastal waters are significantly different than the times of the peak over the land and mountainous region. In contrast, JJA is characterized by a delayed afternoon peak over land (most obvious south of the radar), including over the mountains.

From Fig. 14, it is difficult to quantify which of these regions contribute most to modulating the diurnal cycle of $R_h$ within each season. To address this, we have classified the regions based on altitudes and defined the following categories: altitude $\geq 1500$ m as high mountains; altitude $< 1500$ m and $> 200$ m as low mountains; altitude $> 30$ m and $\leq 200$ m with a distance from the northern part of the coastal borders $\geq \sim 20$ km as lowland; altitude $\leq 30$ m with a distance from the northern part of the coastal borders $< \sim 20$ km as coastal lands; and northern part of the offshore region as coastal waters. For a detailed diurnal cycle analysis, we selected a $15$ km $\times 15$ km representative box each for coastal water (box 1), coastal land (box 2), lowland (box 3), low mountain (box 4), and high mountain (box 5). All boxes were the same size except for box 5, where the size is $9$ km $\times 9$ km (Fig. 6b) to better isolate the effect of the high mountains.

Considering all seasons together, it appears that the overall diurnal cycle with all regions combined (Fig. 15; black dash line) has a similar timing to the diurnal cycle over the mountains. This highlights the important role of mountainous regions that dominate the regional average of rain rate over the whole domain. In DJF, there is also a slight peak in $R_h$ around 0400 LT that is driven by the diurnal cycle of $R_h$ over coastal waters that might be influenced by the large-scale process of MJO (Hidayat and Kizu 2010).

In general, mountainous areas experience the highest $R_h$ in all seasons, with the largest amounts in DJF and SON (around
3–4.5 mm h⁻¹) and the lowest in JJA (around 2 mm h⁻¹) (Fig. 15). In DJF, over the high and low mountains, the timing of the afternoon peak in maximum \( R_\text{h} \) around 1400 LT might relate to rainfall generated over the high and low mountain regions, with some contributions from the land regions characterized by a peak at 1500 LT (lowland). Those three regions are systematically the main contributors of the afternoon peak for all seasons. Over the coastal land, the peak of \( R_\text{h} \) in DJF is observed at 1700 LT while over the coastal waters the peak occurs at night (2300 LT).

In JJA, the delayed peak in afternoon rain rate discussed previously is clearly associated with a delayed peak over the high and low mountains (1700 LT), as well as the peak rain rate over the lowland (1600–1700 LT). The afternoon peak rain rate during the transition seasons MAM and SON, are characterized by a double peak structure, centered at 1400 and 1700 LT in MAM and 1300 and 1700 LT in SON. Interestingly, over that season the earlier peak is driven by rain rate over the lowlands and the high mountain (MAM), while the later peak is associated with rain rate over the mountain areas.

The diurnal phase difference between the mountains and lowland exhibits a seasonal variation. In DJF, the diurnal peak over the mountains leads the peak over the lowland. In contrast, in JJA and SON, the diurnal peak over the lowland starts earlier than the diurnal peak over the mountains. In MAM, the diurnal peak of rain rate over lowland and high mountains appears simultaneously but a secondary peak of the mountain occurs. In JJA, the delayed peak of rain rate over the mountains might be related to the 850-hPa zonal wind. Relative to DJF, in JJA the background westerly wind weakens; as a result, land–sea breeze is stronger with onshore winds moving farther inland and reaching the mountains. As the sea breeze propagates farther inland, the structure of the sea breeze is then altered over the complex topography inducing cloud convection near the slopes of mountains later in the day (Hadi et al. 2002). Consequently, the diurnal peak of rain rate might also occur in late afternoon (Hadi et al. 2002).
d. Spatial distribution and diurnal cycle of rain-rate extremes in the Jakarta region

In this section, we explore where hourly rain-rate extremes (REs) are found in the region, at what time of day, and how the location and time of day of REs change with season. We only focused on the peak of rainy (DJF) and dry seasons (JJA) for the spatial analysis of the REs distribution. To characterize the REs, the SHII has been estimated and only the contours $95\text{th percentile of SHII}$ have been plotted as spatial maps for 3-h time intervals of the diurnal cycle (Fig. 16). Unlike spatial maps of the highest average $R_h$ (not shown), the REs are generally detected in the northern part of the radar domain over the coastal waters, coastal land, and lowland. We speculate that mesoscale convective systems induced a higher rainfall intensity over the windward side of the high mountains (Hamada et al. 2008), and in this study, this effect is observed in the northern part of the radar domain.

In general, the diurnal cycle of REs appears mostly in the northern part of the radar domain including over the coastal waters, coastal land, and lowland. We speculate that mesoscale convective systems induced a higher rainfall intensity over the windward side of the high mountains (Hamada et al. 2008), and in this study, this effect is observed in the northern part of the radar domain.

In general, the diurnal cycle of REs appears mostly in the northern part of the radar domain including coastal waters in the evening to morning (1800–0900 LT). REs starts to diminish over the coastal waters and move farther to the south over the land in the northeast in the afternoon (0900–1800 LT).

In DJF, in the evening (2100–0000 LT) and early morning (0000–0300 LT), the REs mostly occur in the western part of the radar domain including over the coastal waters, coastal land, and mountains. The REs then mainly fall in the coastal land between 0300 and 2100 LT. REs are more isolated over the northeastern part of the domain at 0900–1500 LT.

In JJA, at 0000–0600 LT the REs also mostly occur in the western part of the radar domain, but at 0300–0600 LT the REs also take place in the east. At 0600–1200 LT, the REs are observed in the north even though very few areas in the south also receive REs. At 1200–1800 LT, the REs move a little bit eastward. Then, at 1800–0000 LT, the REs start to occur in the northeastern part of the domain, mainly at 1800–2100 LT.

In JJA, at 1200–1500 LT, the mean easterly wind may contribute to the REs on the windward side of the mountain peaks and is consistent with SHII distribution (Fig. 13) that is concentrated upwind of the mountain peaks. In DJF, at 1200–1500 LT, a strong westerly flow might help to drive the REs in the northeast of domain like SHII distribution in Fig. 13.
For the north of the radar domain, REs occur mainly in the northern coastal land region in DJF (0300–0600 LT) and start to develop later over the north coastal land in JJA (0600–0900 LT). The REs form over the northeast of the radar domain at 1200–2100 LT for DJF and 1500–0000 LT for JJA. The morning REs over the coastal waters in the northwest begin to occur in early evening in DJF (2100–0300 LT), but later in JJA (0000–0600 LT).

These results demonstrate that the Jakarta region experiences substantial spatial, seasonal, and diurnal differences in REs. More work will be done in the future to investigate these signatures and the potential role of large-scale drivers by studying mean rainfall and REs relationships with intraseasonal variability such as the MJO.

4. Discussion and conclusions

The Serpong CDR has been used to study hourly and daily rain rates within the complex topography region around Jakarta. The generation of a radar dataset that could be used to quantitatively analyze rain-rate characteristics around Jakarta was more challenging than in many other areas of the world due to issues with the presence of beam blockage and length of in situ station observations for calibrating the radar and generating a robust $Z-R$ relationship. While we acknowledge that there are areas that were not used in the analysis due to beam blocking, we are satisfied, after thorough quality control such as correcting for calibration and attenuation and careful analysis, that this dataset may be used for understanding the nature of rainfall around Jakarta. This analysis also provides a method and framework for investigating subdaily rain-rate variability in a developing country in the tropics since very few studies using radar in the region due to the unique nature of the dataset.

This analysis shows that there is highly localized rain rate around Jakarta, as shown by the large spatial variability of the mean rain rate, rain frequency, and rainfall extremes. An important finding of this study is that the average $R_h$ over the Jakarta region peaks in DJF, but results from a much higher $W_{freq}$ but a surprisingly lower $SHH$. We note that during our study period (2009–2012), there were two La Niña (2011, 2012), one El Niño (2010), one negative Indian Ocean dipole (2011), and one positive IOD (2012) events (Lestari et al. 2019). This large-scale variability might contribute to a different behavior of general rainfall organization, such as higher occurrence of heavier rain in other seasons, particularly in SON.
FIG. 16. The SHII during wet season (DJF; blue) and dry season (JJA; red). Only ≥ the 95th percentile of SHII is plotted on the map. The 95th percentile of SHII varies every 3 h and each season.
Another prominent feature is identified in the seasonal spatial variability of $R_{day}$ (Fig. 12). Our results show that the highest $R_{day}$ over the mountainous region appears not only in the DJF but also in other seasons indicating that this increased rain rate is largely linked to the local orographic effect.

A comparison between radar and station rain rates shows that there is a difference in the rate distributions particularly for heavier rain rate $\geq 50$ mm h$^{-1}$ (Fig. 8a). Previous studies identified that the large negative bias between radar and gauge mostly occurs at more intense rain rate (Legates 2000; Sivasubramaniam et al. 2019). However, in this study, Fig. 7 shows that the $Z-R$ relationship underestimates rain rates particularly at higher $Z$. Future work is needed to establish the variability of the $Z-R$ relationship as a function of the local forcing (topography, large-scale conditions) and the type of rain (i.e., convective or stratiform) to improve the accuracy of rain estimation from radar especially for heavy rain. This will require collection of more observations than are currently available.

Analysis of $R_h$ also exhibits substantial seasonal variations over different topography (Figs. 14 and 15). The most noticeable result in the timing of the diurnal cycle of $R_h$ is an earlier afternoon peak in DJF than in JJA. The difference in time of the diurnal peak between DJF and JJA is statistically significant at the 95th-percentile confidence level. The prevailing background wind and local land–sea breeze might result in the difference in mechanism of diurnal cycle of rain rates and timing of the peaks (Oki and Musiakte 1994; Yanase et al. 2017). In DJF, as compared with JJA, the midlevel westerly wind (Figs. 12 and 13) might help to initiate an earlier peak of diurnal rain rate particularly over the mountains (Hadi et al. 2002). Additionally, it is evident that the timing of the peak is similar to the peak time of the diurnal cycle over the mountainous regions. Although it depends on the selection of the size of the region, it clearly shows that the mountains play a vital role in rain-rate development and timing.

In the dry season (JJA), the rainfall is more localized and generated by the land-based convection (Marzuki et al. 2016) due to the sea breeze. In JJA, when the prevailing westerly wind is weak, the magnitude of the onshore flow reaches a maximum, but it occurs in late afternoon between 1700 and 1800 LT (Hadi et al. 2002). This might be also one of the causes of a delay in the peak of rainfall that exists over the mountains.

The spatial map of REs (Fig. 16) reveals that REs predominantly exist in the northern part of radar domain over the coastal waters in the northwest in the evening to morning between 1800 and 0900 LT with the extremes in DJF commencing earlier than JJA. This result is different from the $R_h$ showing the maximum values in the afternoon mainly in the southern part of radar domain over the high–low mountains and the lowlands. We hypothesize that the regular systems occur and originate over the mountain producing high $R_h$, then might grow and move toward the windward side of mountains generating REs when the system is mature over the lowlands. A distinct difference in the spatial distribution of REs and $R_h$ indicates further research is necessary, especially with more focus on how the large-scale dynamics (such as MJO) interacts with local-scale drivers. To date studies in this region have been limited by a lack of quality-controlled observational data, but our work provides a basis for such analysis including numerical modeling.

Our study has highlighted the strong spatial variability in rainfall in a small area of the MC. This broad region is of exceptional importance to global climate (Neale and Slingo 2003) but limited in situ observations, a patchy radar network, coarse satellite data, and low-resolution climate models mean that our ability to monitor, simulate, and predict the climate of this area of the world is poor. Radar data can help us to understand weather and climate variability in the MC, and our study provides a framework for such analyses.

**Acknowledgments.** This research is supported by the Australia Award Scholarship (AAS), Hadi Soesastro Prize and the Australian Research Council (ARC) Centre of Excellence for Climate Extremes (CLEX) (CE170100023). We thank the Weather Modification Unit, the Agency for the Assessment, and Application of Technology (BPPT), Indonesia, together with the Japan Agency for the Marine-Earth Science and Technology (JAMSTEC), Japan, for providing disdrometer, C-band Doppler radar, and rainfall datasets from the Japanese Earth Observation System Promotion Program (JEPP)–Hydrometeorological Array for Intraseasonal Variation–Monsoon Automonitoring (HARIMAU) project (JFY 2005–2009). This research was conducted with the assistance of the resources and services from the National Computational Infrastructure (NCI), which is supported by the Australian Government. Author King is supported by the ARC DECRA Fellowship (DE180100638), and author Vincent is supported by the ARC CLEX (CE170100023).

**Data availability statement.** Supporting data are available at https://doi.org/10.5281/zenodo.6818345.

**REFERENCES**


