Difference in the lightning frequency between the July 2018 heavy rainfall event over central Japan and the 2017 northern Kyushu heavy rainfall event in Japan

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Abstract
The causes for the differences in the lightning frequency between two heavy rainfall events, the 2017 northern Kyushu heavy rainfall event and the 2018 heavy rainfall event in central Japan, were examined using a numerical model coupled with an explicit bulk lightning model. These heavy rainfall events occurred near the Baiu frontal system of Japan in July, but the characteristics of rainfall differed. The former case was categorised as an extreme rainfall and extreme convective event by previous satellite observational studies, and the lightning frequency was high. Conversely, the latter case was categorised as an extreme rainfall without extreme convection event, and the lightning frequency was low. The numerical model used in this study successfully reproduced the differences in the lightning frequency between the two cases. Our analyses indicated that the differences in the lightning frequency between the two cases were attributed to the differences in the vertical structure of the charge separation rate and the charge density, which originated from the difference in the vertical distribution of graupel.

KEYWORDS
bulk lightning model, heavy rainfall events, lightning frequency, numerical model

1 | INTRODUCTION

Heavy rainfall is a major meteorological issue since it can lead to flash floods and landslides, and subsequently, cause extensive damage to life and property. According to climate prediction based on general circulation models, extreme precipitation events are expected to increase in the future (IPCC; Stocker et al., 2013); therefore, increased scientific understanding of rainfall is necessary. Previously, heavy rainfall was considered to be produced by tall convective clouds, which stimulated strong convection. However, a recent study by Hamada et al. (2015; hereafter H15) and Hamada and Takayabu (2018; hereafter HT18) reported that heavy rainfall is majorly produced by low convective clouds under convectively unstable but relatively stable and moist conditions compare to the climatology. H15 defined the heavy rainfall events produced from low convective clouds as extreme rainfall, but not extreme convection (R-only extreme event referred in H15). They also defined the heavy rainfall events produced by tall convective clouds as extreme rainfall and extreme convection (RH-extreme event referred in H15). HT18 indicated that heavy rainfall...
originating from R-only extreme events has low echo-top height and low lightning frequency. In addition, some previous studies indicated that heavy rain was not always accompanied by high lightning frequency (Kato et al., 2007). Therefore, these studies indicated that lightning is not always a good indicator of heavy rainfall. Further detailed understanding about the relationship between the lightning frequency and heavy rainfall is necessary since the information on lightning has been used for satellite-based rainfall prediction (e.g. Tapia et al., 1998; Xu et al., 2013).

Recently, two heavy rainfall events, which distinctly highlight the difference between R-only extreme and RH-extreme events, occurred in 2017 and 2018 in Japan (Tsuji et al., 2020). The heavy rainfall event in 2017 over Kyushu area, Japan, called the 2017 Kyushu northern heavy rainfall (hereafter referred to as Case 1), was produced by tall convective clouds generated under convectively unstable conditions with high lightning frequency (Kawano et al., 2018). This case was similar to RH-extreme events. Contrastingly, the heavy rainfall event in 2018 over the central part of Japan (hereafter referred to as Case 2), was stimulated by low convective clouds under moist and convectively less unstable conditions compared with climatology, and the lightning frequency in Case 2 was low (Kawano et al., 2018; Nakakita et al., 2019). This case was similar to R-only extreme events. Thus, detailed analyses on the differences between Case 1 and Case 2 enable us to increase our scientific understanding regarding the differences between the R-only extreme and RH-extreme events, and thus, about heavy rainfall events.

H15 and HT18 indicated that the reasons for low lightning frequency of R-only extreme events have a larger contribution to warm rain processes than those of RH-extreme events. However, numerical model results reported that graupel, which is closely associated with lightning (Takahashi, 1978), was generated in both the events (Hashimoto & Hayashi, 2020). This indicates that both cold and warm rain processes contributed to producing heavy rainfall in both events. Further knowledge regarding the reasons for the differences in the lightning frequency between the two events is required.

Multiple previous studies were conducted on Case 1 (e.g. Kato et al., 2018; Kawano & Kawamura, 2020; Takemi, 2018) and Case 2 (e.g. Kotsuki et al., 2019; Moteki, 2019; Sekizawa et al., 2019; Shimpo et al., 2019), but only a few study (Kawano et al., 2018; Nakakita et al., 2019) discussed the probable causes for the differences in the lightning frequency observed between the two cases. Numerical simulations coupled with the explicit lightning model (Barthe et al., 2005; Barthe et al., 2012; Barthe et al., 2016; Bovalo et al., 2019; Dafis et al., 2018; Fierro et al., 2013; Gharaylou et al., 2020; Hayashi, 2006; Helsdon et al., 2001; MacGorman et al., 2001; Mansell et al., 2002; Mansell et al., 2005; Mansell & Ziegler, 2013; Takahashi, 1978) are a powerful tool to examine the causes for the differences in lightning frequencies. However, studies using the lightning model have not been conducted for understanding the differences between R-only extreme and RH-extreme events.

Therefore, in this study, we aimed to investigate the causes for the differences in the lightning frequency between R-only extreme and RH-extreme events through numerical simulations coupled with an explicit bulk lightning model (Sato et al., 2019). Furthermore, the validity of the model in terms of the lightning frequency was examined by comparing observational results.

2 | DATA AND METHODS

2.1 | Model and experimental setup

Scalable computing for advanced library and environment (SCALE, version 5.3.6, Nishizawa et al., 2015; Sato et al., 2015) model was used in this study. An explicit bulk lightning model was implemented into SCALE by Sato et al. (2019) for calculating the charge density of the hydrometeor, the electrical potential, electrical field, and lightning. Non-inductive charge separation through the collision between graupel and ice/snow based on the look-up table (LUT) given by Takahashi (1978) was considered. Terrains following coordinate and map factor were applied to the lightning model. Neutralization (i.e. lightning) was calculated based on the scheme proposed by Fierro et al. (2013), in which the lightning initiated at a grid point where the electric field exceeded \(E_{\text{int}}\) and the charge of hydrometeors including the column with a cylindrical radius, \(r_{\text{cyl}}\), from the grid point was neutralized. Based on the procedure proposed by Fierro et al. (2013), we set \(E_{\text{int}}\) and \(r_{\text{cyl}}\) at 110 kV m\(^{-1}\) and 5 km, respectively. Information regarding the lightning component is shown in the study by Sato et al. (2019, 2021). A double moment bulk microphysical scheme given by Seiki and Nakajima (2014) and a radiation scheme based on k-distribution method (Sekiguchi & Nakajima, 2008) were used to calculate the effect of cloud microphysics and radiation, respectively. A Smagorinsky-type (Brown et al., 1994) and a Mellow-Yamada-type turbulence scheme (Nakanishi & Niino, 2006) were used to calculate the vertical and horizontal mixing directions by turbulences, respectively. Surface temperature and moisture over the land were calculated by a bucket-type land model and a
single-layered urban canopy model of Kusaka et al. (2001) for non-urban and urban areas, respectively.

Case 1 and Case 2 simulations were conducted from 0000 UTC on 5 July to 0000 UTC on 6 July 2017 and 2100 UTC on 6 July to 0000 UTC on 8 July 2018, respectively. The calculation domain for Case 1 and Case 2 respectively covered \(300 \times 300 \text{ km}^2\) and \(720 \times 600 \text{ km}^2\) (Figure 1) with 1 km horizontal grid spacing. Based on the report of Tsuji et al. (2020) the echo top height in Case 2 was lower than that of Case 1. The model top height in Case 2 was set lower than that in Case 1 to save computational resources. The vertical layer was divided into 60 layers from 40 m above the ground extending to 21,545 m (model top) and 57 layers from 40 m above the ground extending to 19,528 m (model top) for Case 1 and Case 2, respectively. The layer thickness ranged from 40 m (near the ground) to 683 m (model top) for Case 1 and 40 m (near the ground) to 651 m (model top) for Case 2. The initial and lateral boundary conditions were dynamically down-scaled from Mesoscale Analysis (MANL) produced by Japan Meteorological Agency (JMA) with horizontal grid resolution of 5 km and 50 vertical layers and with a temporal resolution of 3 h. The horizontal wind, air density, temperature, specific humidity, soil moisture, soil temperature, and sea surface temperature of the MANL were used. The interval of the model output was 1 min.

To simulate C-Band radar signals from the model output, a radar simulator of Masunaga and Kummerow (2005) included in Joint-simulator (Hashino et al., 2013) was used. The results of every 10 min of the model were used to simulate the C-Band radar signal and compare the results with the observation.

### 2.2 Observational data and analyses

Three types of observational data were used to evaluate the model performance. The radar/raingauge-analysed precipitation product (Makihara et al., 1996) produced by JMA was used for evaluating the simulated surface precipitation. The reflectivity of the C-Band radar standardized for constant altitude plan position indicator (CAPPI) with 1 km horizontal and vertical grid spatial resolution and 10 min temporal resolution produced by JMA was used to validate the simulated vertical distribution of the cloud hydrometeor. The cloud-to-ground (CG) and intra-cloud (IC) lightning detected by the lightning observation data of lightning detection network system (LIDEN; Ishii et al., 2014) operated by JMA was used to evaluate the simulated lightning. The total lightning in the model defined as the sum of CG and IC was calculated based on the bulk lightning scheme (Fierro et al., 2013). The flash origin density (FOD), which is defined by equation (9) in Fierro et al. (2013), simulated by the model was compared with the LIDEN data.

We conducted analyses only over the hatched area in Figure 1 to reduce the effects of the lateral boundary condition, and hereafter, we refer to this area as analysis area. The analysis period was set from 0600 UTC on 5 July to 0000 UTC on 6 July 2017 for Case 1, and 0300 UTC on 7 July to 0000 UTC on 8 July 2018 for Case 2 to exclude the results during spin up time from the analysis.
Validation of SCALE coupled with the lightning component

The validity of the precipitation simulated by the model was assessed by comparison with the observational data. Figure 2 indicates the geographical distribution of cumulative surface precipitation simulated by SCALE (Figure 2a,c) and the observational data (Figure 2b, d) during the analysis period over the analysis area. Large precipitation amounts exceeding 450 mm were observed in a narrow area (approximately 33.4°N 130.8°E) for Case 1 (Figure 2b). The model satisfactorily reproduced the large precipitation amounts over a narrow area, although the area with large precipitation amounts shifted southeast (Figure 2a). Large precipitation amounts exceeding 300 mm were observed at approximately 35.7°N, 137°E (Figure 2d) for Case 2 which was efficiently simulated by the model (Figure 2c). However, the model overestimated the precipitation over the windward area outside of the analysis area (figure not shown). Such a discrepancy of the model from the observation has been often seen during downscaling simulation especially over the windward area of the calculation domain due to several reasons such as mismatch of the meteorological field between the parent model (i.e. MANL in this study) and the child model (i.e. SCALE in this study) and differences in the microphysical model used in the parent model and the child model. However, the discrepancy was only seen outside of the analysis area. Thus, the model reproduced large precipitation amounts in both the cases over the analysis area.

Along with the horizontal distribution, the validity of the vertical distribution of the simulated hydrometeor was confirmed through the comparison with the observational data by the C-Band radar. Figure 3 indicates the contoured frequency by altitude diagram (CFAD) calculated from the results over the analysis area during the analysis period. Based on the CFAD derived from the observational data, larger reflectivities, corresponding to the hydrometer with a large size, were observed in

3 | RESULT

3.1 Validation of SCALE coupled with the lightning component

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Case 1 than Case 2. Additionally, small hydrometers above 9 km height \(z\) in Case 2 were observed as shown as a black solid rectangle in Figure 3d. Conversely, hydrometeors above \(z = 9\) km, which are shown as a black arrow in Figure 3b, in Case 1 were larger in size than in Case 2, and gradually reduced in size with increasing height. CFAD by the model (Figure 3a, c) well reproduced characteristics in the differences between Case 1 and Case 2, although some discrepancies from the observations were seen as described below. One discrepancy was that the model exhibited high reflectivity over the entire layer (Figure 3a and c). However, the trends of the differences between Case 1 and Case 2 represented by the black rectangles and arrow in CFAD of the observational data were also seen in the results of the model. Another discrepancy was observed in the upper layer (i.e. above \(z = 12\) km) corresponding to the overestimation of high stratiform clouds extending outside the convective clouds, which were mainly constructed by snow particles. Although the high clouds were overestimated in the model, lightning, which was the key investigation entity in this study, mainly occurred near the convective core. In summer thunderstorm, it is known that lightning from stratiform clouds outside the convective clouds does not occur frequently based on the observation (Carey, 2005; Peterson & Liu, 2011). The overestimation of the high clouds outside the convective clouds did not significantly affect lightning. Based on these results, we concluded that cloud field and precipitation were simulated well by the model to examine the differences in the lightning frequency between the two cases.

Figure 4 indicates the FOD simulated by the model (Figure 4a, c) and the flash number acquired through LIDEN (Figure 4b, d). Although heavy rainfall occurred in both the cases, observed lightning frequency in Case 1 was higher than that in Case 2 (Figure 4b,d). The
model well reproduced the differences in the lightning frequency between Case 1 and Case 2 over the analysis area (Figure 4a,c). High lightning frequency in the simulated model in Case 1 was located to the southeast, contradictory to the observations. The discrepancy in the location of high lightning frequency originated from the discrepancy of the location of the large precipitation amounts (Figure 2a). Discrepancies between the model and observational data can be improved by accurately simulating the location of the large precipitation amounts.

### 3.2 Difference of the lightning frequency between two cases

The vertical distributions of the mixing ratio (Figure 5a), charge density (Figure 5b), and charge separation rate (Figure 5c) provide suitable explanation to understand and discuss the causes for the differences in the lightning frequency between Case 1 and Case 2.

In the model, charge density above $z \sim 8 \text{ km}$ was critical for the lightning frequency since lightning majorly initiated at $z \sim 10 \text{ km}$ in the model (figure not shown). The charge density had a negative peak at $z \sim 9 \text{ km}$ ($z \sim 7.5 \text{ km}$) and a positive peak at $z \sim 13 \text{ km}$ in Case 1 (Case 2) as shown in Figure 5b. They originated from the non-inductive charge separation at $z \sim 10 \text{ km}$ (Figure 5c) induced by the collision of graupel with ice or snow. As the charge separation rate of graupel depended on the temperature based on LUT given by Takahashi (1978) (figure 8 of Takahashi (1978) and Figure 1 of Sato et al. (2019)), graupel was charged negatively, and the ice and snow were charged positively at $z = 10 \text{ km}$ for Case 1 (Figure 5c). The charge separation at $z \sim 10 \text{ km}$ also occurred in Case 2 over the convective core (Figure S1), and the negatively (positively) charged graupel (ice and snow) was generated in both cases. Note that the convective core is defined as the grid with the sum of the liquid water path and ice water path exceeding 5 kg m$^{-2}$. The negatively charged graupel fell by gravitational settling,
which produced the peak of negative charge at $z \sim 9 \text{ km}$ ($z \sim 7.5 \text{ km}$) in Case 1 (Case 2), as shown in Figure 5b. The positively charged ice and snow are transported upwards by the updraft around convective core and formed positively charged clouds at $z \sim 13 \text{ km}$. This tendency was seen in both the cases. However, the magnitude of the charge density above $z \sim 8 \text{ km}$ was higher in Case 1 than that in Case 2 (Figure 5b).

As the environment in Case 1 was unstable compared to that in Case 2 (Tsuji et al., 2020), the updraft in clouded grids in Case 1 was stronger than that in Case 2 (Figure S2), and the strong vertical velocity maintained below $z = 10 \text{ km}$, which corresponds to the peak of the charged separation rate of graupel (Figure 5c). In such a case, graupel was distributed at comparatively higher altitudes and graupel amount was larger in Case 1 than in Case 2 (Figure 5a). The larger amount of graupel at $z \sim 10 \text{ km}$ in Case 1 than that in Case 2 led to the frequent charge separation and a high magnitude of the charge separation rate in Case 1 compared to those in Case 2 at $z \sim 10 \text{ km}$ (Figure 5c). The magnitude of the positive and negative charge was respectively high at $z = 9 \text{ km}$ and $z = 13 \text{ km}$ in Case 1 because of the frequent charge separation at $z \sim 10 \text{ km}$ (Figure 5c). The high magnitude of the charge led to a large electric field at $z \sim 10 \text{ km}$, which resulted in frequent lightning in Case 1.

Contrastingly, in Case 2, graupel was distributed at comparatively lower heights (i.e. at $z \sim 7 \text{ km}$, Figure 5a) than that in Case 1 due to weaker updraft compared with Case 1 and lower peak height of upward velocity (Figure S2). As a result of the distribution of graupel at a low height, non-inductive charge separation at $z \sim 10 \text{ km}$
did not occur frequently in comparison to Case 1 (Figure 5c). The small charge separation rate at $z \sim 10$ km resulted in a small magnitude of the charge density at $z \sim 7.5$ km and $z \sim 13$ km. Based on the relationship between the charge density and the electric field, the small magnitude of charge density at $z \sim 7.5$ km and $z = 13$ km resulted in a small electric field at $z \sim 10$ km, consequently, leading to lower frequency of lightning in Case 2 than in Case 1.

Additionally, graupel was majorly distributed at $z \sim 6.5$ km in Case 2, and the non-inductive charge separation rate at this height was higher than that in Case 1 (Figure 5c). This large charge separation rate originated from low liquid water content (LWC) in Case 2 than Case 1 (blue lines in Figure 5a). Figure 5d,e indicate two-dimensional normalized frequency distributions of LWC and temperature in each grid in convective core at $z = 6.5$ km over LUT, which were used for determining the amount of charge obtained by graupel through collision (Takahashi, 1978). The high frequency area in Case 2 was mainly distributed over the area with a charge separation rate ranging from 20 to 30 fC. Contrastingly, the high frequency area in Case 1 was majorly distributed around the area with a charge separation rate ranging from 0 to 20 fC due to higher LWC (Figure 5a). The difference in the amount of charge separation through collision at $z = 6.5$ km resulted in a high charge separation rate at $z = 6.5$ km in Case 2 compared with that in Case 1. However, two peaks of the charge density at $z \sim 6$ km and $z \sim 7.5$ km, which were generated by the charge separation at $z = 6.5$ km, were smaller than that at $z > 8$ km in Case 1. Thus, the charge separation at $z = 6.5$ km did not have significant contribution to increase the lightning frequency compared to the charge separation at $z \sim 10$ km in Case 1.

Based on these results shown above, we can conclude that the difference in the lightning frequency between Case 1 and Case 2 originated from the differences in the vertical distribution and amount of graupel, which in turn originated from the differences in the strength of updraft in the two cases.

4 | CONCLUSION

In this study, we examined the causes for differences in the lightning frequency between two heavy rainfall events using a numerical model coupled with an explicit bulk lightning model. A heavy rainfall event (Case 1) was similar to the RH-extreme event as defined by H15, and was characterized by strong convection with high cloud-top height and high lightning frequency. Conversely, the other heavy rainfall event (Case 2) was similar to the R-only extreme event as defined by H15, and was characterized by clouds with low cloud-top height and low lightning frequency. Our results showed that the model well reproduced the observed lightning and the differences in the lightning frequency between the two cases over the analysis area.

The differences in the lightning frequency originated from differences in the vertical distribution of graupel in the two cases. In Case 1, graupel was distributed at higher altitudes due to strong updraft generated under the unstable environment compared with that in Case 2. In such conditions, the non-inductive charge separation at $z \sim 10$ km occurred frequently, resulting in a high magnitude of charge density. The high charge density resulted in a large electric field and high lightning frequency. Conversely, in Case 2, graupel was distributed at lower altitudes due to weak updraft compared with that in Case 1. Thus, the mechanisms that induced frequent lightning in Case 1 did not work effectively for Case 2, resulting in low lightning frequency.

The causes for differences in the lightning frequency between RH-extreme and R-only extreme events were explained by the vertical distribution of graupel. However, other mechanisms that could possibly explain the differences need to be assessed. One of these mechanisms is the importance of warm rain process in R-only extreme events. Based on the previous studies by H15 and HT18, the warm rain process is comparatively more significant in R-only extreme event (Case 2) compared with RH-extreme event (Case 1). However, hydrometeors were distributed above the height of 0 °C, that is, at an approximate height of 4.5 km in both the cases (Figure 5a). This implies the significance of the warm rain process cannot completely explain the differences in the lightning frequency.

This study focuses only on the cases with single simulation. However, our findings can serve as baseline studies for understanding the characteristics of R-only extreme and RH-extreme events and subsequently, assist deeper understanding of the relationship between the convective clouds and heavy rainfall, and assist in the prediction of heavy rainfall.

Numerical experiments to acquire further insights into the differences between the R-only extreme and RH-extreme events should be undertaken in the future.

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AUTHOR CONTRIBUTIONS

Yousuke Sato: Conceptualization; formal analysis; funding acquisition; investigation; methodology; project administration; resources; software; validation; visualization; writing – original draft. Syugo Hayashi: Data curation; methodology; writing – review and editing. Akihiro Hashimoto: Data curation; methodology; writing – review and editing.

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SUPPORTING INFORMATION
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