Diurnal differences in the effect of aerosols on cloud-to-ground lightning in the Sichuan Basin

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Abstract. The effect of aerosols on lightning has been examined in many studies, but its mechanisms are complex and far from understood. This study investigated the influence of aerosols on cloud-to-ground (CG) lightning during both afternoon and night in Sichuan Basin, to understand the difference in the influences of aerosols under conditions with and without solar radiation. The results showed that the difference in lightning flashes between the clean and polluted subsets was not obvious in the afternoon (1200-1800 Beijing Time), while the lightning flashes in the polluted subset were significantly greater than that in the clean subset in the rest of the time. The relationship between lightning and AOD is nonlinear in the afternoon (1200-1800 Beijing Time) and at night (2300-0500 Beijing Time) with a turning point at AOD=~0.3. When AOD is less than about 0.3, increasing AOD will lead to an increase in lightning flashes in the afternoon and night. When the AOD exceeds about 0.3, the increase of AOD will reduce the lightning flashes in the afternoon but have no obvious effect on the lightning flashes at night. The different relationship between aerosol and lightning flashes in the afternoon and at night after AOD exceeds 0.3 is related to the changes in solar radiation in these two periods. In the afternoon, aerosols reduce the solar radiation reaching the ground through its direct radiation and microphysical effects lead to the decrease of the surface temperature, increase the atmospheric stability, inhibit convection activities and thus reduce lightning. At night, due to the absence of solar radiation, the influence of aerosols on surface temperature is weakened, so the inhibition of aerosols on lightning activity is weakened.

1 Introduction

Lightning is an intense discharge phenomenon in the atmosphere that threatens the safety of human life and property and leads to significant increases in NO_x in the middle and upper troposphere (Holle et al., 2019; Zhang et al., 2011; Choi et al., 2005; Yair, 2018). Many observational studies have shown the important role of aerosols in affecting lightning from the ocean to the continent (Westcott, 1995; Bell et al., 2009; Yuan et al., 2011; Altaratz et al., 2010, 2017; Thornton et al., 2017; Liu et al., 2020).

The effects of aerosols on lightning encompass two parts: microphysical and radiative effects (Altatatz et al., 2014; Fan

et al., 2016; Li et al., 2017b; Kuniyal and Guleria, 2019). The microphysical effects are related to aerosols serving as cloud condensation nuclei (CCN). Adding aerosols to clouds produces more small cloud droplets, inhibits coalescence, and delays the onset of warm-rain processes. For deep convective clouds, this process allows more liquid water to ascend into the mixed-phase region of the atmosphere through strong updrafts, forming more super-cooled water and ice particles. Freezing a large amount of liquid water releases more latent heat and invigorates convection (Rosenfeld et al., 2008; Khain and Lynn, 2009; Tao et al., 2012). Through radiative effects, aerosols can heat the atmospheric layer and cool the surface by absorbing and scattering solar radiation, which increases the stability of the atmosphere and thus likely suppresses the development of convection (Kaufman et al., 2002; Koren et al., 2004, 2008; Tan et al., 2016; Li et al., 2017a). The content of super-cooled water and ice particles in clouds as well as the strength of convection are thus modulated by aerosols, and they are tightly correlated with the occurrence and progression of lightning (Takahashi, 1978; Jayaratne et al., 1983; Mansell et al., 2005; Saunders, 2008).

Generally, the relationships between aerosol loading and lightning or convective activity are complex and variable (Li et al., 2018; Lal et al., 2018; Dayeh et al., 2021). Some studies revealed the transition between two opposing effects of aerosols on clouds and lightning. The first is the microphysical effect which increases the convective intensity and lightning, followed by the radiative effect that becomes dominant with the increase in aerosol loading suppressing the convective intensity (Altaratz et al., 2010). The aerosol type determines the ability of aerosols to affect cloud microphysical and solar radiation, which further affects the relationship between aerosols and lightning (Li et al. 2017b). In central China, dominated by absorbing aerosols, a negative relationship was found between aerosols and thunderstorms as well as lightning. In contrast, in southeast China, dominated by hygroscopic aerosols, a positive relationship was found between them (Yang et al., 2013, 2014, 2016). The environmental condition can also change the direction of aerosol effects on lightning. For example, in dry northern Africa, the aerosols invigorate and suppress lightning under low and high aerosol loading conditions, respectively. However, in moist central Africa, the aerosol invigoration effects are sustained under both low and high aerosol loading conditions (Wang et al., 2018). In summary, the complex relationship between aerosols and lightning results from the superposition of aerosol microphysical and radiative effects, which are altered by aerosol loading, type, and specific environmental conditions. Some model studies simulated the development of thunderstorms under different aerosol backgrounds and found increasing aerosol concentration leads to the enhancement of lightning even if the aerosol concentration reaches a high level (Zhao et al., 2015; Shi et al., 2015, 2019). However, these model results are difficult to verify using observations, because the radiative effects of aerosols will offset the microphysical effects when the aerosol loading is excessively high.

Many observational studies concerning the relationship between aerosols and lightning on a relatively coarse time resolution (such as a yearly, monthly, or daily time scale) have been published (Lal et al., 2011; Wang et al., 2011; Altaratz et al., 2017; Shi et al., 2020; Liu et al., 2021; Wang et al., 2021; Chakraborty et al., 2021). Few studies have investigated the linkage between aerosols and lightning (and convective activity) on an hourly time scale. Guo et al. (2016) investigated the

daily cycle of precipitation and lightning under clean and polluted conditions over the Pearl River Delta. Under the polluted condition, heavy precipitation and lightning occurrences were delayed in the day. Chen et al. (2021) reported convective clouds preferentially occur under polluted conditions in the morning but the situation is reversed in the afternoon through analysing data of two warm seasons in eastern China. These researches revealed a diurnal difference in aerosol effects. Still, changes in aerosol effects on lightning or convective activity on an hourly time scale are far from understood.

Based on the above discussion, the combination of aerosol radiation suppression and microphysical stimulation and the change of their relative strength related to various factors are the main reasons for the complicated relationship between aerosols and lightning. The aerosol radiation suppression effect depends on solar radiation, and some studies based on hourly data reveal that aerosol inhibited effect on lightning weakens after sunset (Guo et al. 2016). The current study aims to investigate the difference in the effects of aerosols on lightning under conditions with and without solar radiation. To achieve this goal, we selected the Sichuan Basin as the study region. The Sichuan Basin is one of the aerosol high-value centres in China (Liu et al., 2016; Ning et al., 2018). It also is one of the most lightning-active regions in China. Due to its unique topography and meteorological conditions, lightning over the Sichuan Basin occurs more frequently at night time during the warm season (Yang et al., 2015; Xia et al., 2015). Nine years (2010-2018) of measurements and reanalysis data over the Sichuan Basin were analysed to investigate the difference in the influences of aerosols, dynamics-thermodynamics factors, and cloud-related variables on cloud-to-ground (CG) lightning in day and night. This paper is organised as follows. Sect. 2 describes the datasets, region of interest, and study methods. Results and discussion are given in Sect. 3. Summary and conclusion are presented in Sect. 4.

2 Data and methodology

2.1 Lightning data

The time series of cloud-to-ground (CG) lightning (hereafter referred to as lightning) data used in this study were obtained from the China Meteorological Administration (CMA) and China National Meteorological Center (CNMC). The CG lightning data were detected by the China Lightning Detection Network (CLDN). This network is based on the ground-based Advanced Time of Arrival and Direction (ADTD) system and uses the Improved Performance through Combined Technology (IMPACT) method (Cummins et al. 1998). This system provides information about the time, latitude, longitude, polarity, and peak current of the CG flashes. CLDN comprises 357 sensors, covers most of central and eastern China, and has a detection efficiency of 80–90% (Yang et al., 2015; Xia et al., 2015). Some data quality controls have been performed. Positive CG flashes with a peak current of less than 15 kA are removed to eliminate the possible contaminations of intracloud flashes (Cummins and Murphy, 2009). Only the first stroke is retained if more than one strokes occur in the next second with 10 km of the first stroke and two strokes that occur within 0.5 seconds assumed to be from the same flashes (Cummins et al. 1998). In addition, it is a different

flash if the polarity of the stroke is different. The CG lightning data used in this study were aggregated to a 0.5°×0.5° spatial resolution. We compared the lightning data used in this paper with the lightning data of LIS/OTD. Overall, the spatial distribution and diurnal variation of the lightning data used in this paper are similar to that from LIS/OTD (Fig. S1 in the supplement).

2.2 Aerosol data

The aerosol loading in this study is characterised by aerosol optical depth (AOD). AOD (550nm) data are obtained from the Modern-Era Retrospective Analysis for Research and Application version 2 (MERRA-2). It provides dust, black carbon (BC), organic carbon (OC), and total extinction AODs at 550nm. The AOD of total aerosols is selected in this study to discuss the possible effects of aerosols on lightning activity. The spatial resolution of it is 0.5°×0.625°. We interpolated AOD data onto a 0.5°×0.5° spatial resolution grid using bilinear interpolation. MERRA-2 is the latest version of global atmospheric reanalysis for the satellite era produced by NASA Global Modeling and Assimilation Office (GMAO) using the Goddard Earth Observing System Model (GEOS) (Randles et al. 1980; Gelaro et al. 2017). This dataset assimilates AOD from Moderate Resolution Imaging Spectroradiometer (MODIS), Advanced Very High-Resolution Radiometer (AVHRR) over the ocean, and the space-based Multiangle Imaging Spectroradiometer (MISR) over bright surfaces, and ground-based Aerosol Robotic Network (AERONET) (Buchard et al. 2017). We compared the AOD data of MERRA-2 and MODIS data sets. The AOD of the two data sets in the study region correlated well (Fig. S2 in the supplement).

2.3 Dynamics-thermodynamics and cloud-related data

In this study, we mainly select hourly reanalysis data of five factors: 2m temperature (T; units: ℃), convective available potential energy (CAPE; unit: J kg^{-1}), vertical wind shear (SHEAR; unit: m s^{-1}), total column cloud liquid water (TCLW; unit: kg m^{-2}), and total column cloud ice water (TCIW; units: kg m^{-2}). We choose CAPE and SHEAR to characterise the dynamics and thermodynamics conditions of the study region. We selected SHEAR in the low and middle troposphere, respectively. The SHEAR in the low troposphere is calculated from the hourly wind field ((U, V); units: m s^{-1}) at 850 and 700 hPa and the SHEAR in the middle troposphere is calculated from the hourly wind field at 500 and 400 hPa as follows:

$$Low\ SHEAR = \sqrt{(U_{700} - U_{850})^2 + (V_{700} - V_{850})^2} \ , \tag{1}$$

$$Middle\ SHEAR = \sqrt{(U_{400} - U_{500})^2 + (V_{400} - V_{500})^2} \ , \tag{2}$$

The TCLW and TCIW are selected to represent cloud-related parameters that affect the development of the lightning activity. The TCLW represents the amount of liquid water contained within clouds. Raindrops are not included in this parameter. The TCIW represents the amount of ice contained in clouds. Snow is not included in this parameter. These data are collected from the European Centre for Medium-Range Weather Forecast (ECMWF) ERA-5 reanalysis product with a spatial resolution of 0.25°×0.25°. To match the CG lightning and AOD data, these variables were interpolated onto a 0.5° spatial resolution grid.

ERA-5 is the fifth generation ECMWF reanalysis for the global climate and weather. It combines model data with observations from across the world into a complete, consistent dataset through the law of physics (Hersbach et al., 2018, 2020).

2.4 Region of interest

The black lines in Fig. 1 outline the specific study region. Fig. 1a presents its terrain of it. The spatial distributions of AOD and lightning density (units: flashes hour^{-1}km^{-2}) for 2010–2018, in the summer months, are shown in Figs. 1b and 1c. The study region is composed of low hills and plains, mostly entirely encircled by mountains (as shown in Fig. 1a). It contains the large city of Chengdu and parts of Chongqing. Intense anthropogenic emissions, unique terrain and the climatological low-pressure system at 700hPa over the basin induce heavy air pollution in this region (Liu et al., 2016; Ning et al., 2018). The lightning density in the Sichuan Basin (especially the southern part of it) is marked larger than that in its surrounding region. Fig. 2 shows the diurnal variation in the occurrence frequency of lightning flashes. Lightning mainly occurs during nighttime (1800–0600 Beijing Time, hereafter referred to as BJT), accounting for about 72.9% of the total lightning flashes. The peak of lightning occurs near midnight (2400–0100 BJT). This result is consistent with previous findings that the peak of lightning in the Sichuan Basin mainly happens at night during the warm season (May–September), while that in the contiguous southeastern part of China occurs in the afternoon (Xia et al. 2015).

2.5 Data selection and Producing

This study investigates summer (June, July, and August) data from 2010 to 2018 over the study region. The data of CG lightning flashes, AOD, dynamics-thermodynamics, and cloud-related variables are all at a spatial resolution of 0.5°×0.5°. The first question is how to define a sample and capture the AOD value of a sample. We limit a sample's duration from 1200 BJT in the day to the following 24 hours, and the CG lightning flashes during this period are at least 1. The lightning activity in the study region is weakest at noon (as shown in Fig. 2), so we choose 1200 BJT as the beginning and end of the sample period. We only retain grids with CG lightning flashes records within the time period of a sample. The averaged AOD of these grids in the first hour of a sample was selected to characterise this sample's aerosol loading. To minimize the uncertainties related to the higher probability of misclassification of clouds and aerosols in high-AOD regimes and the influence of hygroscopic growth in a humid environment, the AOD range in this study is set between 0 and 0.8, following the work of Wang et al. (2018, AOD<1.0), Koren et al. (2008, AOD<0.8; 2012, AOD<0.3), Kaufman et al. (2005, AOD<0.6), and Altaratz et al. (2017, AOD<0.4). In addition, within six hours before the start of each sample, the number of grids with CG lightning flashes records in each hour is less than 10% of the entire study region. This is to ensure that thunderstorm is relatively weak before the start of a sample and reduce the potential impact of precipitation on aerosol loading. Finally, we got 532 samples. Subsequently, we determine the clean and polluted subsets according to the AOD value of each sample. All samples are sorted according to AOD and divided into three equal sample subsets where the top third of the AOD range is labelled as polluted, and the bottom third

is labelled as clean. The AOD value distribution of all samples and the AOD range of clean and polluted subsets are shown in Fig. 3. Pearson correlation is used in this study to examine the linear relationship between two factors. The correlation is significant when it passes the significance test at 0.05 level.

3 Results and Discussion

3.1 Diurnal variation of lightning flashes

Figure 4 shows the spatial distribution of total CG lightning flashes with an interval of 3 hours. It can be found that the spatial distributions of the CG lightning flashes are markedly different between different periods of the day. In the early morning, the CG lightning flashes mainly occur on the northwest side of the Sichuan Basin (Fig. 4g). By afternoon, the CG lightning flashes are distributed throughout the basin (Fig. 4a and b). After 1800 BJT, the CG lightning flashes in the southern Sichuan Basin gradually strengthened (Fig. 4c and d). After midnight, lightning concentrated on the southwest and northwest parts of the Sichuan Basin with increasing focus on the northwest side of the Sichuan Basin (Fig. 4e and f).

Some studies have investigated the causes of frequent convection and precipitation at night in the Sichuan Basin and provided some explanations. Many clouds in the Sichuan Basin during the day block the short-wave solar radiation from reaching the ground, which is not conducive to convection. At night, the cloud top radiates and cools, making it easy to form convective activity (Li et al., 2008; Yu et al., 2010). Jin et al. (2012) proposed a conceptual model to explain the phenomenon of frequent convective and precipitation at night in the Sichuan Basin. The southwest flow in the lower troposphere strengthens at night, flows around the southeastern edge of the Yunnan-Guizhou Plateau, and enters the Sichuan Basin, forming a strong cyclonic rotation conducive to ascending movement. This process also brings a large amount of water vapour into the basin. Meanwhile, the downdraft along the eastern slope of the Tibetan Plateau encountered the mass accumulation formed in the low altitude of the basin by the southwest warm and wet air transport in the late night, thus generating diabatic warming at a low level of the troposphere in the central basin. Late night, a cold advection from the Tibetan Plateau to the basin leads to a cooling in the middle troposphere over the central basin. In addition, a recent study found that the prominent diurnal inertial oscillations of the boundary layer south-southwesterly low-level jet into the Sichuan Basin may play an essential role in affecting the daily precipitation cycles in the Sichuan Basin (Zhang et al. 2019). The mechanisms behind the nocturnal convection and precipitation in the Sichuan Basin are complex and are not completely understood. However, the results of Figs. 2 and 4, as well as the findings of previous studies, indicate that convection and thunderstorms are more likely to occur in the study region at night than during the day. In the following content, we further investigate the potential influence of aerosols on CG lightning in the study region on an hourly scale.

3.2 Changes in lightning flashes associated with aerosols

The hourly variation of lightning flashes is first shown for the clean and polluted subset in Fig. 5a. The histograms of the

differences between polluted and clean days are also given (Fig. 5b). On the clean subset, the diurnal variation of lightning flashes shows apparent two peaks. One is in the 1700-1800 BJT, and the other appears in 2400-0100 BJT). In the polluted subset, the diurnal variation of lightning flashes only has one apparent peak, which appears in 2400-0300 BJT. Except from 1200 to 1800 BJT, the lightning flashes in the polluted subset are larger than that in the clean subset. The lightning flashes difference between the polluted and clean subsets is largest during 2300-0500 BJT and smallest during 1200-1800 BJT (as shown in Fig. 5b). These results indicate that the changes in aerosol loading may lead to changes in the diurnal variation of lightning activity. The responses of the lightning flashes to aerosol loading are also different at different times of the day. In the afternoon (1200-1800 BJT), adding aerosols tends to decrease lightning flashes slightly. The rest of the time, adding aerosols tends to increase lightning flashes. Note that the difference in the averaged lightning flashes between polluted and clean subsets is relatively small during the afternoon (1200-1800 BJT) but is relatively large during (2300-0500 BJT). Note that the results found in Fig. 5 are based on regional averages. In Fig. 4, we can find that the spatial distribution of lightning flashes in the study region is distinctly different at different times of the day. Therefore, the response of lightning flashes to aerosol loading may be spatially distinct in different periods. Fig. 6 presents the difference in the lightning flashes in polluted and clean subsets (polluted−clean) with a spatial resolution of 0.5°×0.5°. Some statistics are also given in Fig.6. The differences in most (70%-90%) study regions are positive between 1800-1200 BJT. Between 1200-1500 BJT, the differences in most (71.4%) study regions are negative. Between 1500-1800 BJT, the areas with positive (41.4%) and negative (58.6%) differences are almost large. The differences in the lightning flashes between polluted and clean subsets are relatively small between 0900 and 2100 BJT. The absolute value of the differences in most grids during this period is less than 1 flashes hour^{-1}gird^{-1}. During 2100-0900 BJT, the difference in the lightning flashes between polluted and clean subsets becomes larger in more areas. In this period, the most obvious change of lightning flashes is mainly concentrated in the south and northwest of the study region (Fig. 6d, i, j, and k). The above results indicate that the response of CG lightning flashes to aerosol loading has apparent spatial distribution differences at different times of the day. The response of total CG lightning flashes to aerosol loading is more significant between 2100 BJT and 0600 BJT but changes little in other periods. This is consistent with the results found in Fig. 5. We also check the diurnal cycle of lightning flashes in polluted and clean subsets, respectively (Figs. S3 and S4). In general, the spatial distribution of lightning flashes under polluted and clean subsets is similar, especially between 1800 BJT and 0600 BJT. We speculate that the spatial distribution of lightning flashes in the study region is mainly controlled by terrain and meteorological conditions, and aerosol may have little impact on its spatial distribution. The difference brought by aerosols may be mainly reflected in the time difference.

Based on the above results, we can find that the response of CG lightning flashes to aerosol loading in the study region is different at different times of the day. The most apparent difference exists between the afternoon and the rest of the time. In the next content, we select data in the afternoon (1200-1800 BJT) and in a period of time at night (2300-0500 BJT, hereafter simply referred to as night) when the difference in the lightning flashes between polluted and clean subsets is relatively large

to further investigate the relationships between lightning flashes and AOD, thermodynamic-dynamics, and cloud-related factors.

3.3 Correlation between lightning and AOD, dynamics-thermodynamics, and cloud-related factors

Fig. 7 shows the relationships between lightning flashes and AOD in the afternoon (1200-1800 BJT) and at night (2300-0500 BJT). The AOD and lightning flashes show different nonlinear relationships in the afternoon and night. The lightning flashes first increase with the increase of AOD and then decrease when AOD exceed about 0.3. At night, the lightning flashes also first increase with the increase of AOD but change little when AOD exceed about 0.3. Some previous studies reported a nonlinear relationship between aerosols and lightning (i.e., Wang et al. 2018; Altaratz et al. 2010). When the aerosol loading is below a threshold, lightning increases with the increase of aerosol loading, but when the aerosol loading exceeds this threshold, the increase of aerosol may no longer cause significant changes in lightning and even inhibit lightning. The explanation for this phenomenon is that when the aerosol loading is relatively low, the aerosol microphysical effects may play the dominant role in promoting convection. When the aerosol loading further increases, the aerosol directly radiative effects become more marked, entangled with the aerosol microphysical effects, making the relationship between aerosols and lightning unclear, or the aerosol radiative effects dominate, leading to a decrease in convective intensity. In those studies, using AOD to character aerosol loading, this threshold is about AOD=0.3 which is similar to the result found in Fig. 7.

Fig. 8 presents the relationships between lightning flashes and four dynamics-thermodynamics factors (T, CAPE, Low SHEAR, and Middle SHEAR). Positive relationships can be found in lightning flashes and T both in the afternoon (R=0.68) and at night (R=0.68). The higher the surface temperature is, the more unstable the atmosphere tends to be, which is more conducive to the occurrence of convective activities. However, the promotion effect of T on lightning seems to reach saturation when the T is high. This phenomenon is more obvious in the afternoon. Positive relationships are also found between CAPE and CG lightning flashes in the afternoon (r=0.91) and at night (r=0.75). CAPE is a thermodynamics parameter that describes the potential buoyancy available to idealised rising air parcels and can denote the instability of the atmosphere (Riemann-Campe et al., 2009; Williams, 1992). A higher CAPE means that the atmosphere is more unstable and more likely to form thunderstorms. Many studies have reported a positive relationship between CAPE and lightning activity (Dewan et al., 2018; Murugavel et al., 2014; Pawar et al., 2012). Low SHEAR is negatively correlated (r=−0.72) with CG lightning flashes in the afternoon, but a positive relationship (r=0.71) between them is found at night. When considering the Middle SHEAR, negative relationships between them were both found in the afternoon (R=−0.70) and at night (R=0.70). The SHEAR is the vertical shear of the horizontal wind. It affects the dynamic flow structures around and within a deep convective cloud (Coniglio et al., 2006). Some studies have reported a negative relationship between SHEAR and lightning (i.e., Wang et al. 2018; Zhao et al. 2020). Our result is consistent with their findings. However, at night, the relationship between Low SHEAR and lightning flashes is different to them. Vertical wind shear can suppress vertical cloud development for isolated convection (Richardson

and Droegemeier, 2007). However, low-tropospheric and mid-tropospheric wind shear is critical in organising mesoscale convection systems, especially for squall lines (Coniglio et al., 2006; Takemi, 2007). From the results found in Fig. 4, we can know that the spatial distribution of lightning flashes in the study area is significantly different at different times. The lightning distributions in the afternoon and at night are different, which may lead to differences in the structure of thunderstorm activities during the two periods. This may result in the different relationship between Low SHEAR and lightning in these two periods. For Middle SHEAR, the difference in the lightning distribution has no great influence on its relationship with lightning flashes.

Fig. 9 shows the relationships between CG lightning flashes and two cloud-related factors (TCLW and TCIW). A nonlinear relationship is found between lightning flashes and TCLW in the two periods (Fig. 9a and c). The CG lightning flashes increase with the increase of TCLW when the TCLW is relatively low ($<\sim0.1$kg m^{-2}), but decrease with the rise of TCLW when its value exceeds about 0.1kg m^{-2}. With the updraft, increasing cloud liquid water can provide more liquid water to the mixed phase region of the cloud to form more supercooled water and ice particles which fuels lightning activity. However, too much cloud liquid water may promote warm cloud precipitation rather than from convection and lightning activities. For TCIW, a nonlinear relationship (R=0.11) between lightning flashes and TCIW is also found in the afternoon (Fig. 9b). A positive relationship (r=0.85) between them is found at night (Fig. 9d). Lightning development mainly depends on the noninductive electrification of the collision and separation between graupel and ice crystals in the presence of supercooled water. The more ice particles, the stronger the lightning activity may be. It should be noted that when the TCIW is less than about 0.05kg m^{-2}, the positive relationship between TCIW and lightning flashes is robust (both in the afternoon and at night). When the TCIW is greater than 0.05kg m^{-2}, the relationship between TCIW and lightning flashes becomes more dispersed and the uncertainties in each bin become larger. But in general, lightning flashes are more under conditions with more TCIW.

3.4 Possible reasons for the difference between the effects of aerosols on lightning flashes in the afternoon and at night

Based on the above analysis, when the AOD is relatively large ($>\sim0.3$), a significant difference can be found in the relationship between AOD and lightning flashes in the afternoon and at night. This part further analyzes the possible causes of this difference.

Figure 10 shows the relationship in AOD-T, TCLW-T, TCIW-T, and AOD-TCLW. In the afternoon, AOD is negatively (R=−0.74) correlated with T (Fig. 10a). At night, no obvious relationship (R=−0.10) between them can be found (Fig. 10e). We speculate that high aerosol loading leads to strong aerosol direct radiative effects. Excessive aerosol loading reduces the solar radiation reaching the ground by absorbing and scattering solar radiation, thus reducing the temperature of the ground below the aerosol layer. At night, solar radiation is absent, and so does the direct radiation effect of aerosols, which has no significant impact on the surface temperature. In general, TCLW was negatively (afternoon: R=−0.94, night: R=−0.84) correlated with T in the afternoon and at night (Fig. 10b and f). However, it is worth noting that when TCLW is less than about 0.1kg m^{-2}, the relationship between TCLW and T is not significant in the afternoon, while at night, TCLW is positively

correlated with T. An increase in the amount of liquid water in clouds means thicker and wider clouds. The thicker and wider clouds will block more solar radiation from reaching the ground, thus reducing the surface temperature. On the other hand, the increase in surface temperature will strengthen the updraft, so that more water vapour will be transmitted upward to form more cloud liquid water. In the afternoon, the relationship between TCLW and T may contain the above two mechanisms, which leads to an insignificant relationship between them. At night, the absence of solar radiation reduces the reduction of clouds to surface temperature, and the promotion of surface temperature to cloud liquid water content is dominant. Therefore, a positive correlation between TCLW and T at night. However, too much liquid water in the cloud may promote a warm-rain process. The precipitation falling to the ground will significantly reduce the surface temperature. Therefore, when the TCLW exceeds a certain value ($>\sim 0.1 \mathrm{kg\ m^{-2}}$), it has a negative correlation with the T. In the afternoon, the relationship between TCIW and T is similar to that between TCLW and T (Fig. 10c). But at night, the TCIW has no obvious relationship (R=0.33) with T (Fig. 10g). This may be because the ice water content in clouds is related to more factors, and the conversion process from ice water content to precipitation is more complex, with a less direct impact of surface temperature. In the afternoon, the inhibition of TCIW on surface temperature is more reflected in reducing the solar radiation reaching the ground. At night, the absence of solar radiation, and the precipitation formed by the cloud ice water has no obvious influence on the surface temperature, resulting in no obvious relationship between TCIW and T. AOD is positively correlated (afternoon: R=0.91, night: R=0.91) with TCLW in the afternoon and at night (Fig. 10d and h). By acting as CCN, increasing aerosol concentration will produce more but smaller cloud droplets, thus delaying the warm-rain process, so that more liquid water can be retained in the cloud. In addition, the decrease in the cloud droplet size will increase the albedo of clouds, further enhancing the scattering effect of clouds on solar radiation. Therefore, we can speculate that the inhibition of aerosols on the surface temperature in the afternoon is not only through the direct radiation inhibition of aerosols, but also through the increase of cloud water content.

In summary, excessive aerosols will reduce the surface temperature in the afternoon, thereby increasing atmospheric stability and suppressing lightning. At night, however, the weakening effect of aerosols on surface temperature is dissipated, and further increase of aerosols will not significantly inhibit lightning. Figure 11 shows a schematic diagram to summarise the possible difference in the effects of aerosols between afternoon and night found in this study. When the aerosol loading is relatively low (AOD$<\sim 0.3$), the microphysical stimulation of aerosols on convection may be dominant. Increasing aerosols will promote the development of convective activities and generate more lightning. This effect is consistent in the afternoon and at night. When aerosol loading is relatively high (AOD$>\sim 0.3$), the microphysical stimulation of aerosols reaches saturation, and the direct radiation effect of aerosols gradually increases. Because of the difference in solar radiation in the afternoon and at night, the influence of aerosols on lightning is different in the afternoon and at night. In the afternoon, the direct radiation effect of aerosols will reduce the solar radiation reaching the ground. On the other hand, the microphysical effects of aerosols will also increase cloud albedo, cloud cover and cloud thickness, and reduce the solar radiation reaching the ground. Both of these two mechanisms will reduce the ground temperature, thereby increasing the stability of the atmosphere, inhibiting

convective activities and reducing lightning. At night, due to the absence of solar radiation, the indirect and direct effects of aerosol radiation and microphysical effects on solar radiation also become weak or absent. Thus, the inhibition of aerosols on lightning activity becomes weak at night. At this time, due to the saturation of the microphysical stimulation of aerosols on convection, further increase of aerosols will not cause significant changes in lightning.

4 Summary and conclusion

The influence of aerosols on lightning is quite complicated. Especially under the condition of high aerosol loading, the direct radiation effect and microphysical effect of aerosols overlap each other and have different effects on lightning in combination with different environmental conditions. This study attempts to determine whether aerosols' impact on lightning is different during the day and at night. Due to the special terrain and environmental conditions, the Sichuan Basin has the unique characteristics of frequent lightning at night, and the aerosol loading in this region is also high. Based on this region, we investigated the response of lightning flashes to aerosols on an hourly scale and selected two time periods of the afternoon (1200-1800 BJT) and night (2300-0500 BJT) to investigate the influences of aerosols, thermodynamic factors and cloud-related parameters on lightning. The data used for analysis in this paper include lightning flashes, AOD, T, CAPE, Low SHEAR, Middle SHEAR, TCLW and TCIW in summer seasons from 2010 to 2018. The main conclusions are as follows:

Under clean and polluted subsets, the diurnal variation and the number of lightning flashes are different. This signifies the potential impact of aerosols on lightning. The difference in lightning flashes between clean and polluted subsets is insignificant in the afternoon, and the difference is more obvious in the rest time with more lightning flashes in the polluted subset. The difference in lightning flashes between polluted and clean subsets is also different in the spatial distribution in different time periods. Between 2100 BJT and 0900 BJT, the most obvious differences were concentrated in the south and northwest parts of the study region, while the lightning differences in other time periods were not significant. We selected two time periods: the afternoon (1200-1800 BJT) when the solar radiation is relatively strong and the night (2300-0500 BJT) when the solar radiation is absent, to investigate the influences of aerosols, thermodynamic factors and cloud-related parameters on lightning in these two time periods, respectively. We found that the relationship between aerosols and lightning is nonlinear in both two time periods with a turning point at AOD=~0.3. As the AOD is less than this threshold, the increase of AOD in two periods leads to an increase in lightning flashes. As the AOD exceeds this threshold, the increase of AOD in the afternoon causes a decrease in lightning flashes, while at night, the increase of AOD does not cause significant changes in lightning flashes. We further investigated the effects of aerosol and cloud-related parameters on surface temperature in these two time periods. We speculate that when the aerosol loading is relatively low (AOD<~0.3), the microphysical effect of aerosols is dominant, and increasing aerosols will stimulate the development of convection and generate more lightning. When the aerosols loading is relatively high (AOD>~0.3), the microphysical stimulation of aerosols on convection reaches saturation,

and the direct radiation effect of aerosols becomes stronger. In addition, the microphysical effects of aerosols will also increase the ability of clouds (increase cloud thickness, cover, and albedo) to reduce the solar radiation reaching the ground. In the afternoon, the direct radiation effect of aerosols and the change of cloud properties will reduce the solar radiation reaching the ground, reduce the surface temperature, increase the atmospheric stability, and inhibit the development of convective activities. At night, the solar radiation disappears, and the influence of aerosols on the surface temperature is weakened. Therefore, increasing aerosols will not significantly inhibit lightning in this period.

Previous studies indicated that when the aerosol loading was relatively high, the aerosol types and environmental conditions would alter the impact of aerosols on lightning activity (i.e., Wang et al. 2018). In this paper, we highlight a new possible influencing factor: the diurnal variation of solar radiation. While this finding was discovered in the Sichuan Basin, the mechanism proposed might apply to many other regions in the world with similar environmental conditions.

Moreover, the analysis of this study is based on limited measurements and reanalysis data. The detailed mechanisms behind these phenomena still need further research through analysis based on more comprehensive data and model simulations.

Data availability

Processed data in the study and the CG lightning data are available from the first author upon request (whcis4@163.com). MERRA-2 aerosol data can be downloaded from https://disc.gsfc.nasa.gov/datasets?page=1&keywords=MERRA (last access: 5 August 2022) and ERA5 data can be downloaded from https://cds.climate.copernicus.eu/#!/search?text=ERA5 (last access: 5 August 2022.)

Author contributions

HW, YT designed the research ideas for this study. HW carried the study out and prepared the paper. ZS and NY provided the analysis ideas for the dynamics-thermodynamics and cloud-related parameters. TZ edited the paper.

Competing interests

The authors declare that they have no conflict of interest.

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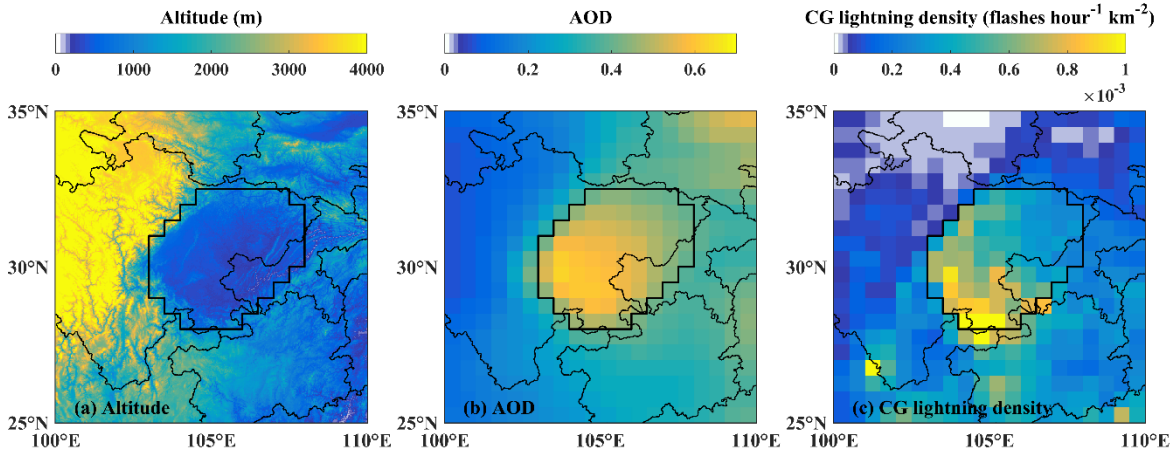


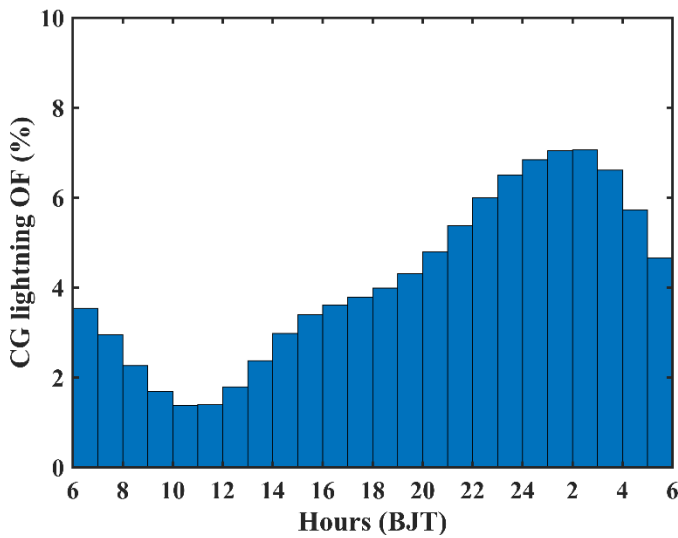
Figure 1: (a) The terrain of the study region with a spatial resolution of 0.02°×0.02°. Spatial distributions of (b) aerosol optical depth (AOD) and (c) CG lightning density (flashes hour⁻¹ km⁻²) at a spatial resolution of 0.5°×0.5° for the period 2010–2018 including the summer months (June, July, and August). The black lines in (a–c) outline the specific area investigated in this study.

555

Figure 2: The diurnal variation in the occurrence frequency of lightning flashes (units: %) over the study region during the summer months (June, July, and August) of 2010–2018.

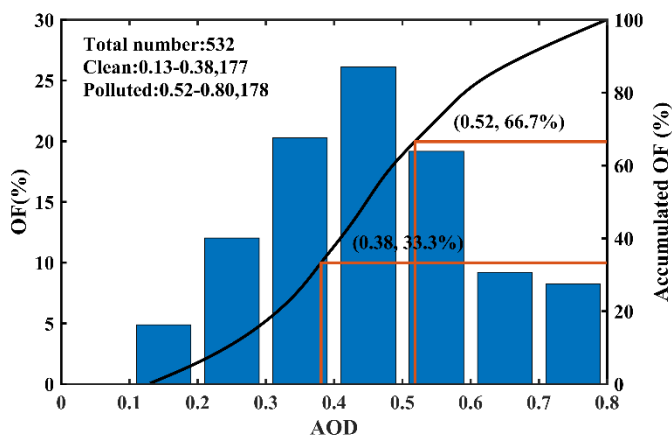



Figure 3: The probability density function of ranked AOD of 532 samples. Black solid lines denote accumulated occurrence

560 **frequencies for the AOD. Red lines show the top and bottom terciles. The AOD range and sample number in clean and polluted subset are also shown.**

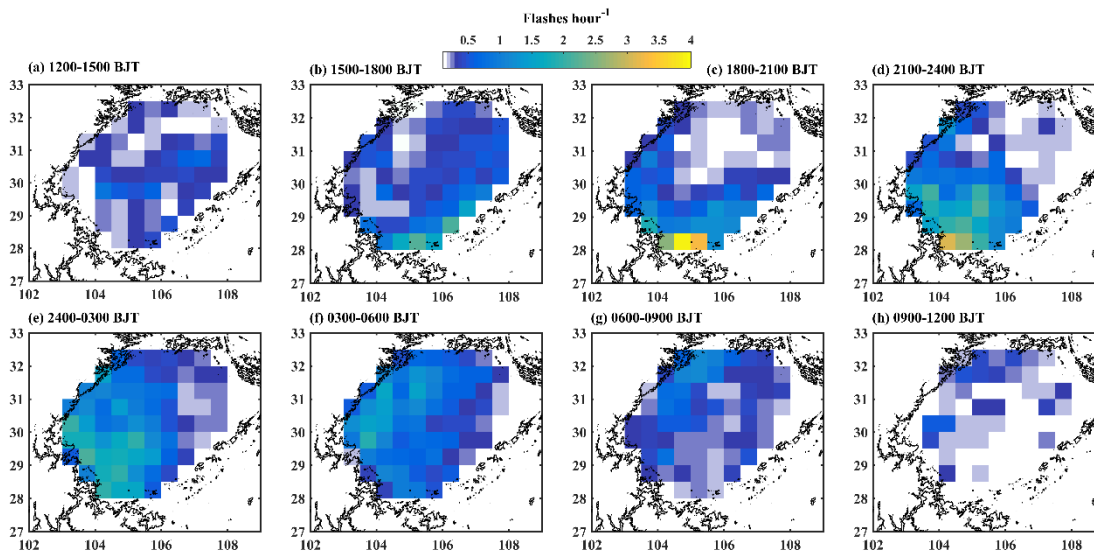



Figure 4: Diurnal cycle of lightning flashes (units: flashes hour⁻¹) on a 0.5°×0.5° grid with an interval of 3 hours (BJT) for 2010–2018 including the summer months (June, July, and August). The black lines represent the 1500m contour lines.

565

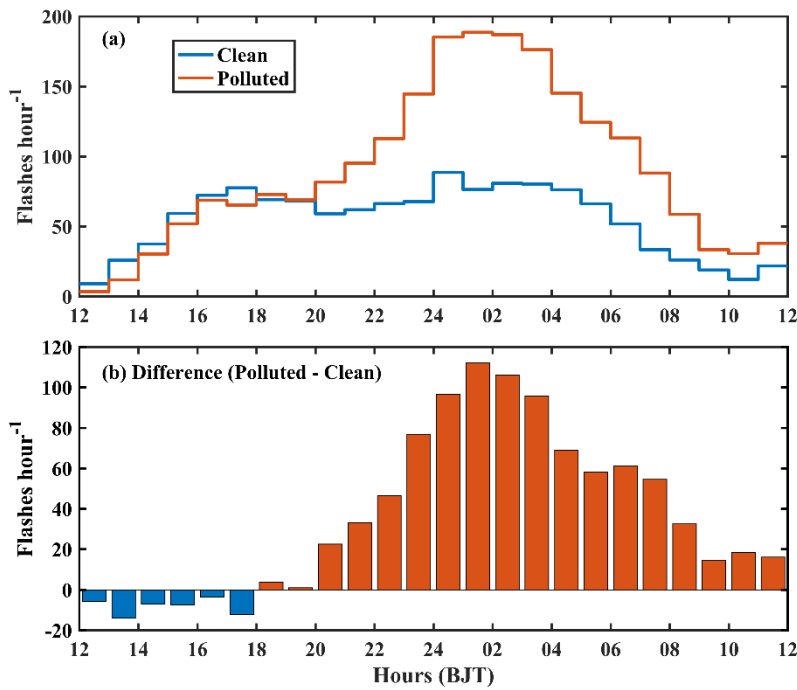



Figure 5: (a) The diurnal variations in lightning flashes (units: flashes hour⁻¹) under clean and polluted subset over the study region during the summer (June, July, and August) season of 2010-2018. (b) The histogram of the difference in CG lightning flashes between polluted and clean subset.

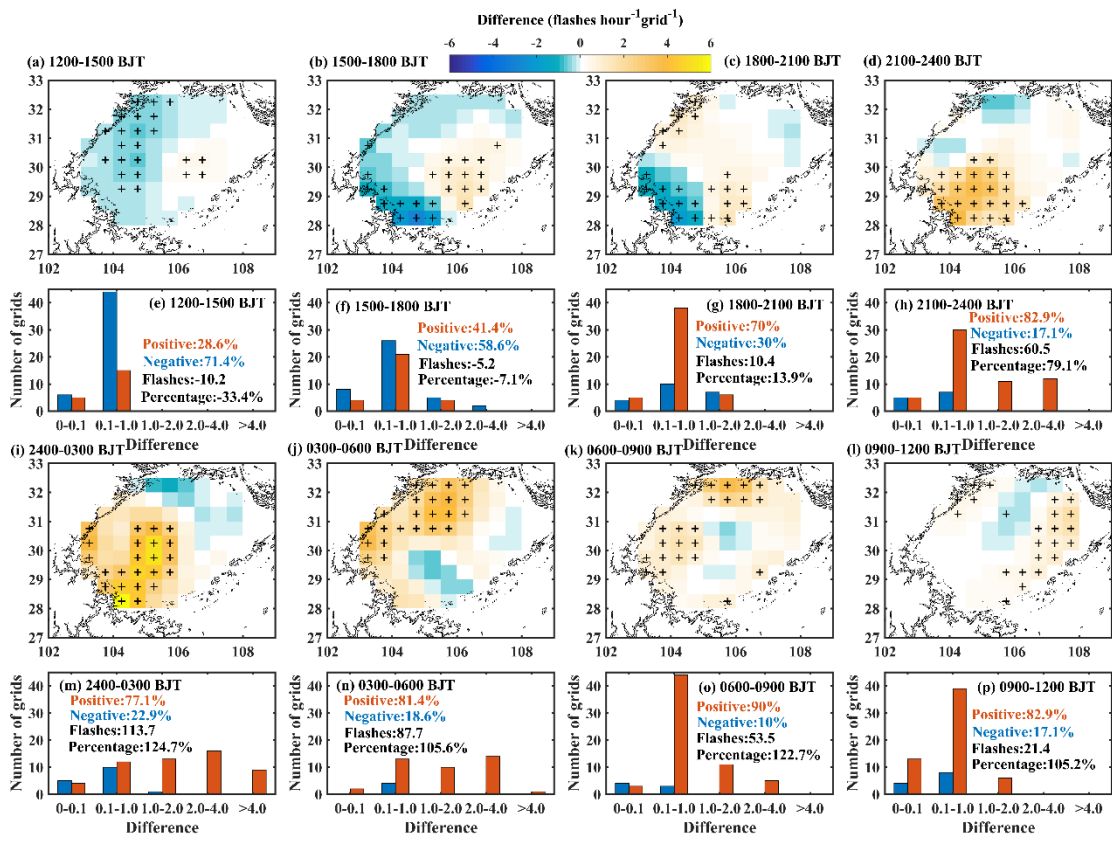



Figure 6: (a-d, i-l) Diurnal changes of total CG lightning flash differences (units: flashes hour^{-1}) between polluted and clean subset (polluted−clean) during the study period with an interval of 3 hours (BJT). Black lines represent the 1500m contour lines. The spatial resolution is 0.5°×0.5°. Warm (cold) colours in the figure mean more (less) lightning flashes in the polluted subset. Plus signs denote those grids with relatively large lightning flashes difference (the absolute value of lightning flashes difference ranks in the top third). **(e-h, m-p)** Histograms of the differences (red: positive, blue: negative) between lightning flashes in the polluted and clean subsets. The percentages of grids with the positive (negative) difference in the total grids, the total change of lightning flashes, and its percentage are also given.

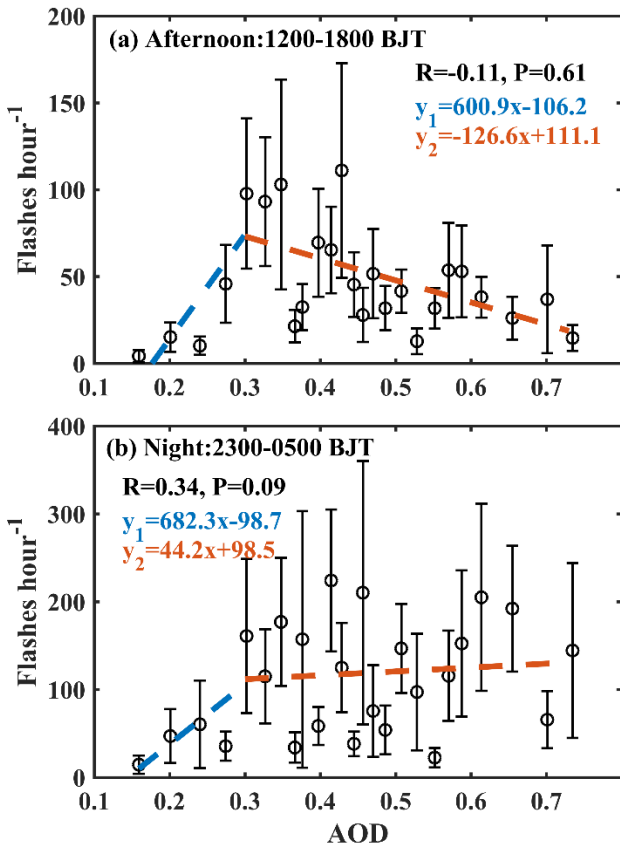



Figure 7: Relationships between lightning flashes and AOD in (a) afternoon (1200–1800 BJT) and (b) night (2300–0500 BJT). Note that samples are first sorted by AOD and then samples with similar AOD were averaged to create the presented scatter plot. The max number of samples in each bin is equal to or less than 20. The difference between the maximum and minimum AOD values of samples in each bin is equal to or less than 0.05. An estimation of the uncertainty was calculated from the standard deviation of each bin divided by the square root of the number of data points in the bin. Linear-fit lines, Pearson correlation coefficients (R), and significant level (P) are also shown.

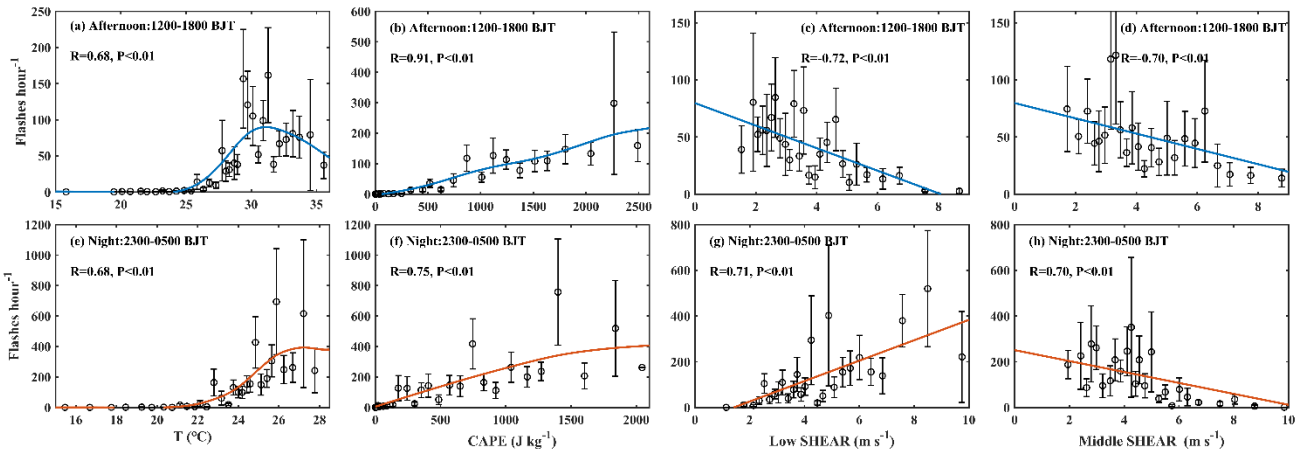



Figure 8: Relationships between lightning flashes and dynamics-thermodynamics factors: (a, e) 2m temperature (T, units: ℃), (b, f) convective available potential energy (CAPE, units: J kg⁻¹), (c, g) low troposphere (850hPa to 700hPa) vertical wind shear (Low SHEAR, units: m s⁻¹), and middle troposphere (500hPa to 400hPa) vertical wind shear (Middle SHEAR, units: m s⁻¹) in the afternoon (1200–1800 BJT) and night (2300–0500 BJT). Note that samples are first sorted by T, CAPE, or SHEAR and then samples with similar T, CAPE, or SHEAR were averaged to create the presented scatter plot. The max number of samples in each bin is equal to or less than 20. The difference between the maximum and minimum T, CAPE, Low SHEAR, and Middle SHEAR values of samples in each bin is equal to or less than 0.5, 200, 2, and 2. An estimation of the uncertainty was calculated from the standard deviation of each bin divided by the square root of the number of data points in the bin. Smoothing spline-fit curves, Pearson correlation coefficients (R), and significant level (P) are also shown in each panel.

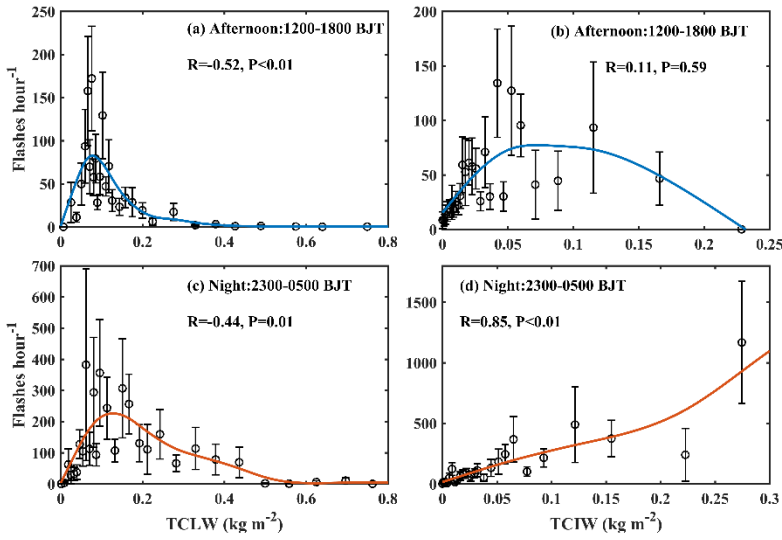



Figure 9: Same as in Fig. 8, but for the (a, c) lightning flashes-total column cloud liquid water (TCLW, units: kg m^{-2}) and (b, d) lightning flashes-total column ice water (TCIW, units: kg m^{-2}). The max number of samples in each bin is equal to or less than 20. The difference between the maximum and minimum (a and c) TCLW and (b and d) TCIW values of samples in each bin is equal to or less than 0.05 and 0.01. An estimation of the uncertainty was calculated from the standard deviation of each bin divided by the square root of the number of data points in the bin. Smoothing spline-fit curves, Pearson correlation coefficients (R), and significant level (P) are also shown in each panel.

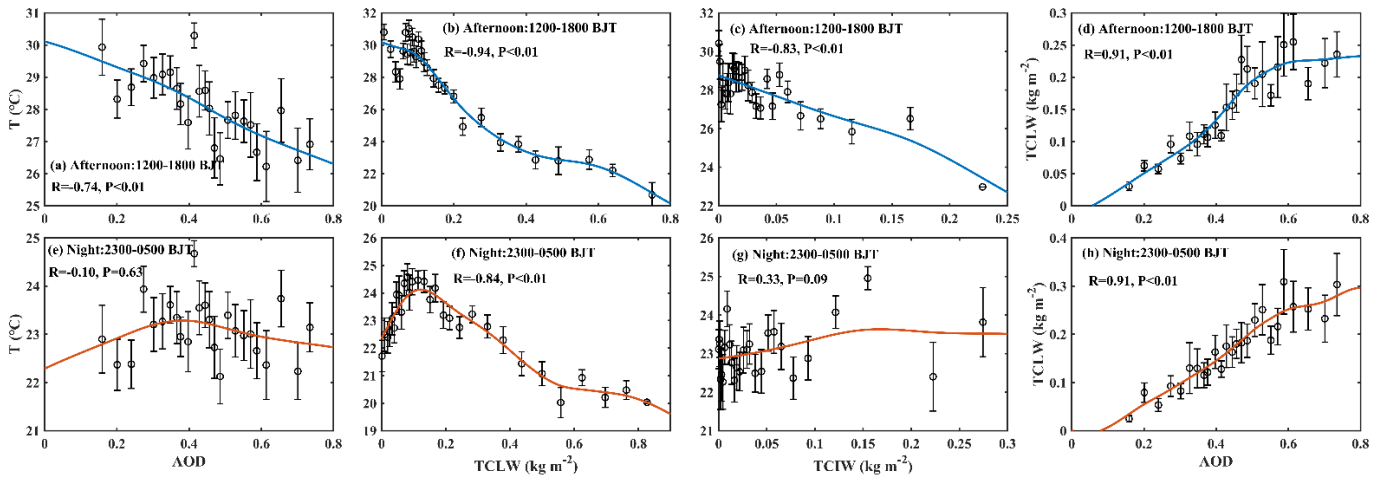



Figure 10: Relationships in (a, e) AOD-T, (b, f) TCLW-T, (c, g) TCIW and (d, h) AOD-TCLW in the afternoon (1200–1800 BJT) and night (2300–0500 BJT). Note that samples are first sorted by (a, e, d, and h) AOD, (b, f) TCLW, and (c, g) TCIW, and then samples with similar (a, e, d, and h) AOD, (b, f) TCLW, and (c, g) TCIW were averaged to create the presented scatter plot. The max number of samples in each bin is equal to or less than 20. The difference between the maximum and minimum AOD, TCLW, and TCIW values of samples in each bin is equal to or less than 0.05, 0.05 and 0.01. An estimation of the uncertainty was calculated from the standard deviation of each bin divided by the square root of the number of data points in the bin. Smoothing spline-fit curves, Pearson correlation coefficients (R), and significant level (P) are also shown in each panel.

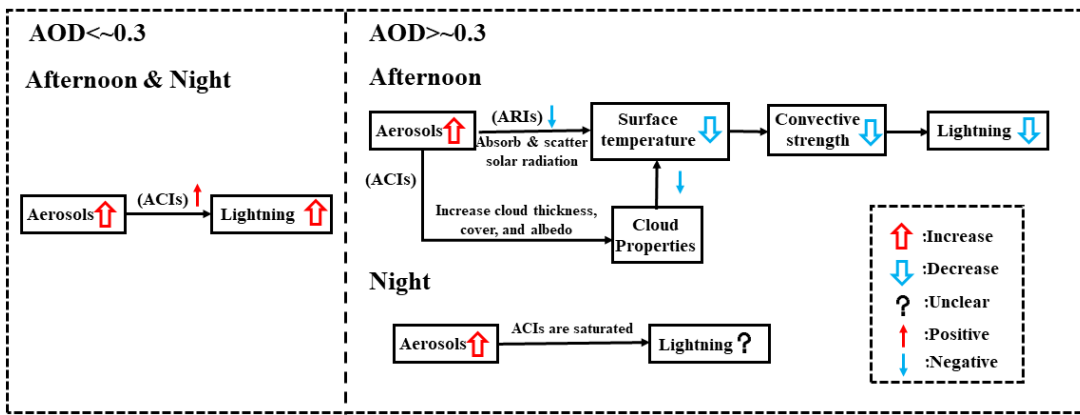



Figure 11: Schematic diagram illustrating the difference in effects of aerosols on lightning activity in the afternoon (1200-1800 BJT) and at night (2300-0500 BJT) over the study region.