



Measurement report: Dependence of Aerosol–Sea Fog Interactions on the Cloud-Aerosol Mass Mixing Ratio (M_{Cr}) and Liquid Water Content (LWC) in Advection Sea Fog over the East China Sea

Xirui Ma^{1,2,3}, Yun Zhang^{2,3}, Yanqiong Xie^{2,3}, Panyan Ge^{1,2,3}, Lingbing Bu¹, Yan Yin¹

5 ¹Key Laboratory for Aerosol-Cloud-Precipitation of China Meteorological Administration/Collaborative Innovation Centre on Forecast and Evaluation of Meteorological Disasters (CIC-FEMD), Nanjing University of Information Science & Technology, Nanjing 210044, China

²College of Meteorology and Oceanography, National University of Defense Technology, Changsha 410073, China

³High Impact Weather Key Laboratory of CMA, Changsha 410073, China

10 Correspondence to: Yun Zhang (zhangyun.1207@163.com)

Abstract. Aerosol effects on sea fog microphysics are nonmonotonic with aerosol concentration, and complex aerosol-sea fog interactions leave substantial uncertainty in microphysical responses. Five sea fog events observed over the East China Sea during March-May 2025 are analyzed with the aerosol-cloud interaction index (ACI) and cloud-aerosol mass mixing ratio (M_{Cr}). Positive ACI time fraction generally decreases with fog duration. M_{Cr} ranges from 2.17 to 2.93. The M_{Cr}-ACI transition range is 2.50-2.63, and 2.63 is selected as an empirical grouping threshold. Under liquid water content (LWC) constraints, ACI_{re} at high-M_{Cr} range increases from 0.02 to 0.14 with LWC, consistent with a progressively stronger Twomey effect in droplet-size reduction. At low-M_{Cr}, ACI_{re} changes from -0.09 and -0.03 at low and intermediate LWC to 0.21 at high-LWC, indicating that the droplet-size response can change direction with LWC. Fog droplet size spectra show that, in the high-M_{Cr} range, the response of the small-droplet mode to increased aerosol loading strengthens with LWC. Under high-LWC, the small-droplet modal peak for high aerosol loading at low-M_{Cr} is about four times the corresponding high-M_{Cr} value. This pattern exhibits pronounced small-droplet accumulation and relatively constrained subsequent growth. This difference indicates M_{Cr} = 2.63 can distinguish relatively complete and constrained activation-growth microphysical states. The autoconversion threshold function (T) response further indicates that aerosol effects on droplet coalescence also depend on the M_{Cr}-LWC state. These results show that jointly considering M_{Cr}-LWC better distinguishes the direction and strength of aerosol effects on sea fog microphysics.

1 Introduction

Fog is a weather phenomenon in which water droplets or ice crystals are suspended in the near-surface layer, reducing horizontal visibility to less than 1 km (Gultepe et al., 2006, 2007; Niu et al., 2010b). Sea fog mainly forms when warm, moist air is advected over a colder sea surface and cools to saturation, and advection cooling fog is the predominant type of sea fog (Yang et al., 2024; Pilić et al., 1979; Tardif and Rasmussen, 2007; Koračin et al., 2014). Sea fog often persists over



extensive coastal and offshore areas for long periods, causing abrupt reductions in visibility and threatening the safety of marine navigation, port operations, and cross-sea transportation (Oliphant et al., 2021). In addition, sea fog can alter the radiation budget of coastal regions (Fernando et al., 2021). The northwestern Pacific is one of the major sea-fog-prone regions worldwide, and springtime advection sea fog is particularly frequent over the Yellow Sea and East China Sea (Zhang et al., 2009; Dorman et al., 2020). The formation of such sea fog is associated with factors such as the advection of warm, moist air over colder sea surfaces, sea surface temperature fronts, and large-scale atmospheric circulation (Gao et al., 2007; Yang et al., 2018; Liu et al., 2021; Huang et al., 2015). Meanwhile, anthropogenic aerosol emissions are strong along the eastern coast of China, and polluted air masses are frequently transported offshore by southerly flows, causing sea fog to often occur under high-aerosol backgrounds (Yan et al., 2019; Zhou et al., 2025). Li et al. (2025) used long-term marine observations, reanalysis data, and CMIP6 model simulations to attribute decadal changes in sea fog frequency along the Chinese coast. They concluded that aerosols were an important factor contributing to the increase in sea fog frequency in the early 21st century. Aerosols can enhance near-surface atmospheric stability through radiative cooling and can also act as cloud condensation nuclei to promote water vapor condensation, thereby increasing sea fog frequency. However, this attribution is based on statistical relationships in decadal sea fog frequency and model simulations and has not yet revealed the specific microphysical processes through which aerosols affect sea fog. Therefore, the microphysical processes by which aerosols influence sea fog still require further investigation through field observations.

Fog is essentially a cloud near the surface, and the fundamental processes of aerosol–cloud interactions (ACI) also apply to fog (Boucher et al., 2013; Fan et al., 2016; Boutle et al., 2018). Aerosols affect the macrophysical and microphysical properties of clouds through aerosol–radiation interactions and aerosol–cloud interactions, with important impacts on the global radiation budget and hydrological cycle (Ramanathan et al., 2001; Lohmann and Feichter, 2005; Rosenfeld et al., 2008). Aerosol particles act as cloud condensation nuclei (CCN) during the activation of cloud or fog droplets, and their activation capability depends on particle size distribution, chemical composition, and ambient supersaturation (Köhler, 1936; Twomey, 1959; McFiggans et al., 2006; Dusek et al., 2006; Petters and Kreidenweis, 2007; Reutter et al., 2009). At similar liquid water content, an increase in aerosol concentration leads to a larger number of activated cloud or fog droplets with smaller droplet sizes, known as the Twomey effect (Twomey, 1977; Kaufman and Fraser, 1997; Feingold et al., 2001). The reduction in cloud or fog droplet size decreases coalescence efficiency, thereby prolonging the lifetime of clouds and fog (Albrecht, 1989; Liu et al., 2004, 2005, 2006). Observational and numerical simulation studies of polluted continental fog generally show that an increase in aerosol concentration leads to an increase in fog droplet number concentration, a decrease in effective radius, enhanced light scattering, a further reduction in visibility, and prolonged fog duration (Niu et al., 2010; Quan et al., 2011, 2021; Yan et al., 2020; Arun et al., 2025). Jia et al. (2019) found over the North China Plain that, as aerosol concentration increased, fog droplet number concentration and liquid water content increased nonlinearly, while effective radius decreased. Fog duration, fog-top height, and liquid water path also increased accordingly. Yan et al. (2021) further confirmed through observations and simulations that, under polluted conditions, more fog droplets with smaller droplet sizes are formed, which suppresses fog droplet sedimentation and thereby



65 prolongs fog lifetime. Thus, in highly polluted environments, aerosols generally tend to enhance and prolong fog. However, aerosol effects on fog microphysics do not vary monotonically with aerosol concentration, but instead exhibit complex responses in both direction and magnitude. Some studies have found that the increases in fog droplet number concentration, liquid water content, and macroscopic fog characteristics may gradually weaken and approach saturation as aerosol concentration increases further. Once aerosol loading exceeds a certain level, the original enhancement may even turn into suppression (Quan et al., 2011; Jia et al., 2019; Mazoyer et al., 2019). From the perspective of microphysical processes, an increase in the number of activated fog droplets intensifies competition for water vapor, reduces environmental supersaturation, and inhibits the subsequent condensational growth of fog droplets (Stolaki et al., 2015; Poku et al., 2021; Wainwright et al., 2021). For aerosols containing absorbing components such as black carbon, the absorption of shortwave radiation can cause local heating and promote fog droplet evaporation, thereby shortening fog lifetime through the semi-
 70 direct effect. The direction of this effect is opposite to the enhancement effects described above (Maalick et al., 2016). In addition, processes such as gravitational sedimentation of fog droplets can weaken the microphysical effects resulting from aerosol activation, causing the effects of aerosols on fog to vary with environmental conditions. Under some conditions, lower aerosol concentrations may instead correspond to a longer fog lifetime (Pope and Igel, 2025). In polluted boundary layers with abundant water vapor, however, increased aerosol concentrations can promote an increase in fog droplet size near
 80 the fog top and thicken the fog layer (Arun et al., 2026). In summary, the direction and strength of aerosol effects on fog are not determined by aerosol concentration alone, but depend on the relative relationship between water vapor supply and aerosol loading, as well as the physicochemical properties of aerosols. Therefore, the nonmonotonic behavior of aerosol effects on fog and the coupling of multiple processes make it difficult to quantitatively describe the effects of aerosols on fog microphysical characteristics in a reasonable way.

85 To quantitatively describe the effects of aerosols on cloud or fog microphysical characteristics, previous studies have commonly used the aerosol-cloud interaction index (ACI) (Chen et al., 2021; Kaufman and Fraser, 1997; Feingold et al., 2001). However, traditional ACI calculations based on cloud or fog droplet effective radius or optical thickness generally assume that liquid water content remains approximately constant, and the calculated ACI values are comparable only under specific liquid water content conditions (McComiskey et al., 2009). In contrast, ACI calculated from cloud or fog droplet
 90 number concentration does not rely on the assumption of a fixed liquid water content (McComiskey et al., 2009). Even so, limitations in satellite retrievals and the fact that the conversion of aerosol particles to cloud condensation nuclei is influenced by factors such as particle size, chemical composition, hygroscopicity, and supersaturation make the accurate quantification of aerosol effects on clouds and fog one of the most challenging and uncertain issues in this field (Andreae and Rosenfeld, 2008; Quaas et al., 2009; Carslaw et al., 2013; Rosenfeld et al., 2014; Gryspeerd et al., 2016; Seinfeld et al.,
 95 2016; Rosenfeld et al., 2019; Bellouin et al., 2020; Boucher et al., 2013). As discussed above, the direction of aerosol effects depends on the relative relationship between cloud water content and aerosol loading. Therefore, a reasonable quantitative analysis requires an indicator that can reflect both quantities simultaneously. Lin et al. (2026) proposed the cloud-aerosol mass mixing ratio (M_{Cr}), defined as the liquid water content per unit aerosol mass, to characterize the relative state between



cloud water and aerosol loading. However, this framework has so far been applied to free-tropospheric warm clouds and has not yet been applied to sea fog. Sea fog differs physically from the radiation fog and warm clouds discussed above. Sea fog is mainly governed by advective cooling, relative humidity remains close to saturation for long periods, and it forms and persists within a strongly stable boundary layer near the sea surface. Its thermodynamic and dynamic structures differ from those of continental radiation fog dominated by surface radiative cooling and warm clouds (Koraćin et al., 2014; Huang et al., 2015; Koraćin et al., 2001; Lewis et al., 2003). Therefore, the MCr threshold established for warm clouds cannot be directly transferred to sea fog. In-depth studies based on simultaneous observations of aerosols and fog droplet microphysics in sea fog remain limited. Existing sea fog observations mainly focus on the study of droplet microphysical properties, and rarely simultaneously obtain aerosol information (Gultepe et al., 2021; He et al., 2022). Satellite remote sensing can provide large-scale identification and statistical information on sea fog, but it is difficult to directly analyze the microphysical characteristics of near-surface fog droplets (Liang and Griswold, 2024). In contrast, simultaneous and continuous ground-based observations of aerosols and fog droplet microphysics mainly come from continental radiation fog and mountain fog rather than sea fog in the marine boundary layer (Mazoyer et al., 2019; Liu et al., 2025; Haeffelin et al., 2010; Mazoyer et al., 2022). Overall, because observations of sea fog remain limited, it is still difficult to determine whether the MCr framework is applicable to sea fog and what its threshold value should be, which requires targeted field observations.

This study is based on spring observations over the East China Sea from March to May 2025. During the observation period, five sea fog events were recorded with complete simultaneous measurements of aerosols and fog droplet microphysics. To address the research gaps described above, we use a fog droplet spectrometer, an automatic weather station, a forward-scatter visibility meter, and a particulate matter monitor to investigate in detail the effects of aerosols on sea fog from a microphysical perspective during these five events. This study applies the MCr framework from free-tropospheric warm clouds to sea fog in the marine boundary layer for the first time and attempts to determine the transition range of the MCr-ACI response in sea fog, as well as an empirical grouping threshold (MCr) for subsequent grouping analysis. This study mainly addresses three questions. How does the ACI evolve with time during different life-cycle stages of sea fog, and how is it related to MCr? Can MCr provide an empirical grouping threshold for different response states? Under different combinations of MCr and LWC, how does aerosol loading affect fog droplet effective diameter, droplet size spectra, and the autoconversion threshold? The remainder of this paper is organized as follows. Section 2 introduces the observation site, data, and methods. Section 3 presents the fog droplet microphysical and aerosol characteristics of multiple sea fog events, the temporal evolution of ACI and its relationship with MCr during the five sea fog events, and the differences in aerosol effects on fog droplet effective diameter, droplet size, and the autoconversion threshold under different MCr and LWC conditions. Section 4 presents the conclusions.

2 Data and Methodology

2.1 Observation site and instrumentation

The ground-based observation platform used in this study is located at Xifeng Bay on Qingbang Island in the Dongji Islands of the Zhoushan Archipelago, East China Sea (approximately 30.21° N, 122.70° E; about 10 m a.s.l.). It is situated at the easternmost part of the Zhoushan Archipelago, about 60 km from the nearest point on the mainland coastline. The island is far from the mainland and has few local anthropogenic emission sources. Thus, the observed aerosol and fog droplet microphysical characteristics can reasonably represent the conditions of the marine boundary layer over the East China Sea. At the same time, advection sea fog occurs frequently in this region during spring, providing an ideal natural experimental field for investigating aerosol–sea fog interactions (see Fig. 1 for the site location). The observation period covered spring from March to May 2025, and the instruments, observed parameters, and temporal resolutions used in this study are summarized in Table 1.

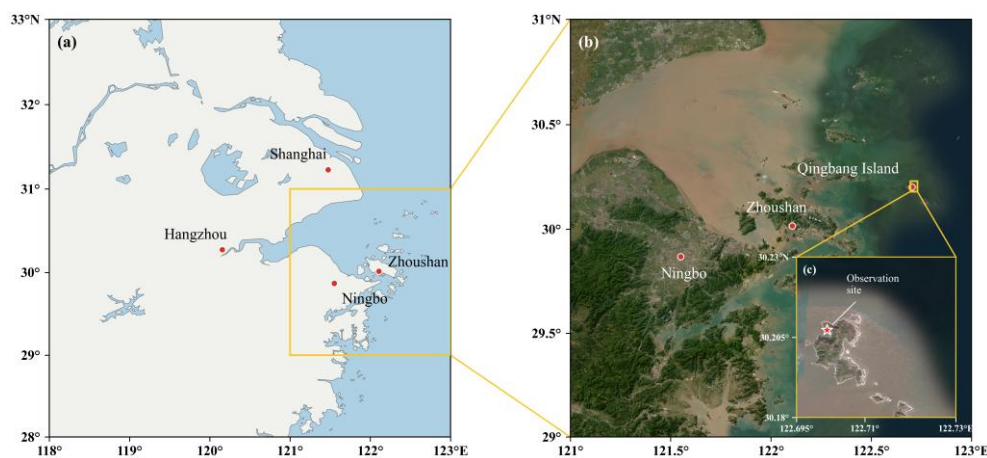


Figure 1: Location of Observation Stations. Satellite imagery in panels (b) and (c) is from Esri World Imagery (sources: Esri, Vantor, Earthstar Geographics, and the GIS User Community | Powered by Esri).

The routine meteorological elements are observed by the AWS1900 automatic weather station with a sampling interval of 10 seconds, and it can simultaneously provide temperature (T), relative humidity (RH), wind speed (WS), and wind direction (WD). The mass concentration of particulate matter PM and horizontal visibility are measured using an environmental particulate matter monitor (PMS100) and a forward-scatter visibility meter (VM3000), respectively. The particulate matter measurement module adopts a heated inlet sampling method to ensure that the RH of the measured particulate matter drops to 40%, allowing the particles to be in a dry state during measurement. The characteristic parameters of fog droplets are measured using an FM-120 fog droplet spectrometer manufactured by Droplet Measurement Technologies (DMT). The instrument is based on the single-particle forward-scattering principle and measures the fog droplet number concentration and size distribution over a diameter range of 2–50 μm in 30 size bins at a sampling frequency of 1 Hz. The FM-120 provides measurement parameters of fog droplet number concentration (N_d), liquid water content



(LWC), and fog droplet diameter (D). To achieve the collaborative analysis of multi-source data, in this study, the data from FM-120, AWS1900, VM3000, and PMS100 were uniformly averaged to a common time resolution of 1 minute and strictly matched by time.

Table 1. Instruments used in sea fog analysis and their measured parameters.

Instrument Name	Equipment model	Measurement parameters	Temporal resolution
Fog Monitor	FM-120	N_d , LWC, D	1s
Automatic Weather Station	AWS1900	T, RH, WS, WD	10s
Forward Scatter Visibility Sensor	VM3000	Vis	10s
Particulate Matter Monitor	PMS100	$PM_{2.5}$	10s

2.2 Sea fog event identification and life-cycle stage classification

Sea fog events are identified using a combined criterion of horizontal visibility and relative humidity (WMO/GAW, 2003; Met Office, 1994). Specifically, sea fog is identified when horizontal visibility remains below 1 km and relative humidity exceeds 90%. The stages of each sea fog event are classified according to the following method. The sea fog process (with relative humidity all greater than 90%) can mainly be divided into 4 stages. The formation stage (visibility decreases from 1500 m to 1000 m), development stage (between the formation and mature stages of sea fog), mature stage (visibility shows only small variations after decreasing to its minimum value, with a continuous duration of more than 30 min), and dissipation stage (visibility begins to increase from its minimum value to 1000 m). Sea fog events are also often accompanied by two additional stages, the explosive enhancement stage (visibility decreases by 500 m within 30 min) and the oscillation stage (visibility is below 1000 m while showing frequent oscillations).

2.3 Methods

We investigate the microphysical characteristics of five sea fog events in detail using observations from the FM-120. From these observations, we calculate fog droplet number concentration (N_d), liquid water content (LWC), and fog droplet effective diameter (ED) as follows.

$$N_d = \int n(D) dD \quad (1)$$

$$LWC = \int \frac{\pi}{6} \rho_w n(D) D^3 dD \quad (2)$$

$$ED = \frac{\int D^3 n(D) dD}{\int D^2 n(D) dD} \quad (3)$$

where $n(D)$ is the number of fog droplets within the unit volume and unit droplet size, with the unit being $\text{cm}^{-3} \cdot \mu\text{m}^{-1}$. ρ_w is the density of liquid water, which is $1.0 \text{ g} \cdot \text{cm}^{-3}$.



The aerosol mass extinction efficiency (MEE) is defined as the ratio of the aerosol extinction coefficient (σ_{ext}) to the aerosol mass concentration $\rho(\text{PM}_{2.5})$ (Saide et al., 2022). Under sea fog conditions, MEE includes the extinction contributions from both fog droplets and aerosols. The unit of σ_{ext} is Mm^{-1} , the unit of $\rho(\text{PM}_{2.5})$ is $\mu\text{g}\cdot\text{m}^{-3}$, and the unit of MEE is $\text{m}^2\cdot\text{g}^{-1}$. The value 3.912 is the Koschmieder constant (Lee and Shang, 2016).

180
$$\text{MEE} = \sigma_{\text{ext}} / \rho(\text{PM}_{2.5}) \quad (4)$$

$$\sigma_{\text{ext}} = 3912 / \text{Vis} \quad (5)$$

To quantitatively assess the effects of aerosols on sea fog microphysical characteristics from observations, this study uses the aerosol–cloud interaction index (ACI). The index was first proposed by Kaufman and Fraser (1997). It uses changes in cloud droplet effective radius caused by changes in aerosol optical depth to quantify aerosol-related microphysical responses. ACI can also be calculated from effective radius, optical depth, or cloud droplet number concentration. ACI calculated from effective radius or optical depth usually needs to be compared under similar LWC conditions. Otherwise, its variation may reflect the evolution of sea fog itself rather than aerosol effects. McComiskey et al. (2009) pointed out that cloud droplet number concentration directly corresponds to the activation of aerosol particles into cloud droplets and does not require a fixed LWC condition, and therefore has an advantage. $\text{PM}_{2.5}$ mass concentration is selected as the proxy for aerosol loading. Lin et al. (2026) verified that the ACI calculated using $\text{PM}_{2.5}$ mass concentration is similar to that calculated using aerosol number concentration (N_a). Therefore, in this study, ACI is calculated using $\text{PM}_{2.5}$ mass concentration and fog droplet number concentration (N_d).

185
$$\text{ACI}_{N_d, \text{PM}_{2.5}} = \frac{1}{3} \left(\frac{\partial \ln N_d}{\partial \ln \text{PM}_{2.5}} \right) \quad (6)$$

Theoretically, when all newly added aerosol particles are activated and grow into fog droplets, $\partial \ln N_d / \partial \ln \text{PM}_{2.5}$ equals 1. Therefore, $0 < \text{ACI} \leq 0.33$ is an ideal reference range for the microphysical process in which aerosol particles are activated and grow into fog droplets. However, variations in actual atmospheric conditions and the complexity of aerosol composition can cause ACI to exceed this range or become negative. A positive ACI indicates a positive logarithmic sensitivity between $\text{PM}_{2.5}$ and N_d , which is consistent in direction with the Twomey response. A negative ACI indicates a negative logarithmic sensitivity between $\text{PM}_{2.5}$ and N_d can be classified as a non-Twomey response. A negative ACI does not mean that it has no physical significance. The mechanisms responsible for this phenomenon are complex. They may involve not only the semi-direct effect caused by absorbing aerosols, but also the covariation of macroscopic meteorological variables and the extreme competition for water vapor caused by aerosol overloading during sea fog.

To better quantify the effects of aerosols on sea fog microphysical characteristics without constraining LWC, the relative relationship between liquid water content and aerosol loading needs to be considered. Therefore, this study uses the cloud–aerosol mass mixing ratio (MCR) proposed by Lin et al. (2026). MCR is defined as the liquid water content per unit aerosol mass and is used to quantify the abundance of liquid water relative to aerosol loading in sea fog.



$$M_{Cr} = \log \left(\frac{LWC \times 10^6}{PM_{2.5}} \right) \quad (7)$$

where LWC is expressed in $g \cdot m^{-3}$, $PM_{2.5}$ is expressed in $\mu g \cdot m^{-3}$, and 10^6 is the unit conversion factor. M_{Cr} is a dimensionless ratio of cloud water mass to aerosol mass.

210 The fog droplet size can be altered by aerosols, and the resulting changes can feed back through the collision-coalescence process, thereby affecting the maintenance and dissipation of sea fog. The fog droplet diameter must exceed a critical diameter, D_c , for the collision-coalescence process to become effective. Liu and Daum (2005) and Liu et al. (2006) proposed the autoconversion threshold function (T) to evaluate the strength of the collision-coalescence process during fog. The value of T ranges from 0 to 1, and a larger value indicates a stronger collision-coalescence process.

$$215 \quad T = \frac{P}{P_0} = \left[\frac{\int_{D_c}^{\infty} D^6 n(D) dD}{\int_0^{\infty} D^6 n(D) dD} \right] \left[\frac{\int_{D_c}^{\infty} D^3 n(D) dD}{\int_0^{\infty} D^3 n(D) dD} \right] \quad (8)$$

$$D_c = 8.18 \times 10^{-4} \beta_{con}^{1/6} \frac{N_d^{1/6}}{LWC^{1/3}} \quad (9)$$

where D_c represents the automatic conversion critical diameter, and $\beta_{con}=1.15 \times 10^{23}$ is the empirical coefficient (Liu et al., 2004).

2.4 Removal of edge effects and determination of the characteristic time window W

220 During the evolution of sea fog, rapid changes in dynamical and meteorological conditions are often accompanied by sudden changes in air entrainment and advection. To reduce the influence of rapid changes during the initial and final dissipation periods of sea fog on the reasonable determination of the characteristic time window, the first 10 min and the final 10 min of all sea fog events were removed for basic quality control. The data from each instrument were uniformly averaged to 1 minute and matched by time. Subsequently, a unified characteristic time window (W) was determined using the changes
 225 with window size in the normalized mean standard errors of $PM_{2.5}$, N_d , and LWC and in their first-order differences (SEM Differential) for the five sea fog events (Fig. 2). The use of this unified window ensures a consistent basis for ACI calculations across different events. For a given sliding window size n (min), the mean standard error $SEM_{v,j}(n)$ of the physical parameters $v(PM_{2.5}, N_d, LWC)$ in event j is defined as follows.

$$SEM_{v,j}(n) = \frac{\sigma_{v,j}(n)}{\sqrt{n}} \quad (10)$$

230 where $\sigma_{v,j}(n)$ represents the mean sample standard deviation of parameter v within sliding windows of length n for event j. To achieve equal-weighted horizontal comparison of different background concentration events, the SEM is normalized based on the minimum test window ($n_{min} = 2min$) to obtain $SEM_{norm,v,j}$.

$$SEM_{norm,v,j}(n) = \frac{SEM_{v,j}(n)}{SEM_{v,j}(n_{min})} \quad (11)$$

To further characterize the sensitivity of normalized SEM to changes in window length, the first-order forward difference with respect to window length n was calculated.

$$D_{v,j}(n) = \frac{SEM_{norm,v,j}(n + \Delta n) - SEM_{norm,v,j}(n)}{\Delta n} \quad (12)$$

where Δn represents the interval between two adjacent test windows. The smaller the absolute value of $D_{v,j}(n)$, the lower the sensitivity of the normalized SEM to the further increase in window length.

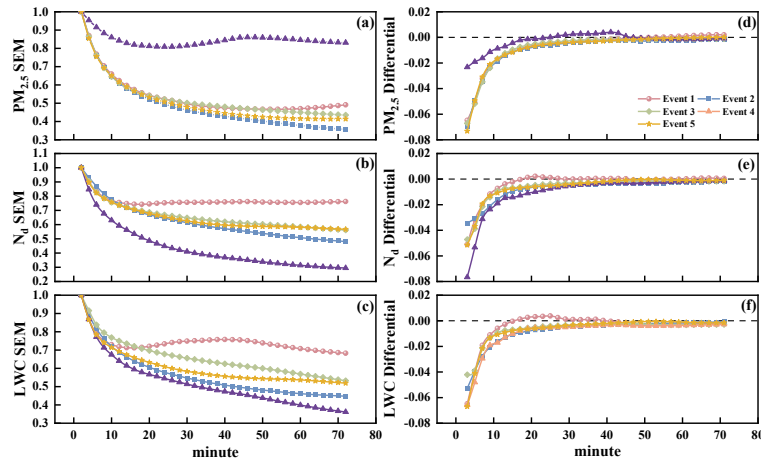


Figure 2: Normalized SEM and Differential of PM_{2.5}, N_d, and LWC for the five sea fog events under different window lengths.

Fig. 2(a), (b), and (c) show that the normalized SEM of PM_{2.5}, N_d, and LWC decrease rapidly when W is approximately 2-20 min. This indicates that the statistical results are relatively sensitive to high frequency variations at shorter window lengths, which may increase the sensitivity of ACI calculations to short-term fluctuations and reduce their stability. Fig. 2(d), (e), and (f) show that, as W increases, the normalized SEM curves of all parameters gradually become flatter, and their first-order differences also gradually approach zero. When W is between 38 and 40 min, the first-order differences of PM_{2.5}, N_d, and LWC for all events generally fall within a low amplitude range close to zero. This indicates that the sensitivity of the normalized SEM to further increases in window length has markedly decreased. Considering both the reduction in the influence of high-frequency variations and the preservation of information on sea fog evolution, a characteristic time window of 40 min was selected and applied uniformly to this dataset. All ACI calculations were calculated at this timescale. This time window represents an empirical characteristic timescale for the present dataset and should not be regarded as a universally optimal value for other sea fog datasets.

3 Results and discussion

3.1 Fog droplet and aerosol characteristics during the five sea fog events

This study is based on spring observations over the East China Sea from March to May 2025. During the observation period, five sea fog events were recorded with complete simultaneous measurements of aerosols and fog droplet. The basic meteorological information for these events, including event duration, mean relative humidity (RH), mean wind speed (WS), mean wind direction (WD), and mean visibility (Vis), is shown in Table 2. During all five sea fog events, RH remained at 100%, the mean near-surface wind speed ranged from 3.44 to 4.56 m·s⁻¹, and the mean wind direction ranged from 170° to 181°, with all five events dominated by southerly flow.

Table 2. Basic information of 5 sea fog events.

Event	Duration(h)	RH(%)	WS(m/s)	WD(°)	Vis(km)
Event 1	2.4	100	3.45	181	0.44
Event 2	6.1	100	3.44	177	0.59
Event 3	7.6	100	3.71	170	0.55
Event 4	3.0	100	3.57	175	0.66
Event 5	4.1	100	4.56	170	0.58

Under similar meteorological and dynamical conditions, the five events exhibited clear differences in aerosol loading and sea fog microphysical characteristics (Fig. 3).

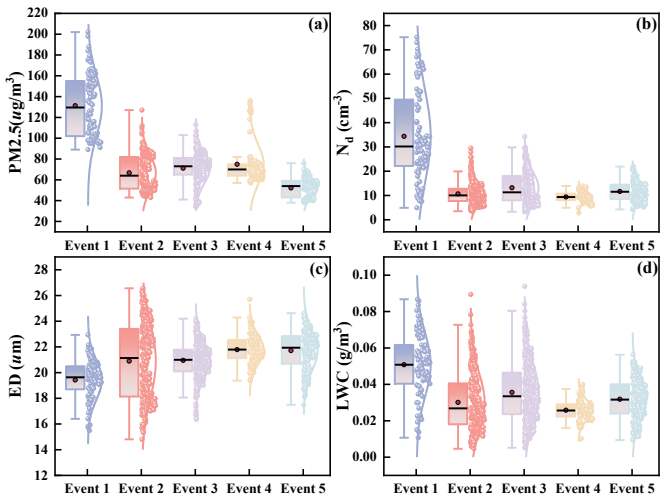


Figure 3: Microphysical characteristic parameters of aerosols and droplets.

Fig. 3(a) shows that the mean PM_{2.5} value of Event 1 was 131 μg·m⁻³, markedly higher than those of the other four sea fog events. The mean PM_{2.5} values of Events 2, 3, and 4 were at intermediate levels, ranging from 67 to 75 μg·m⁻³. The mean PM_{2.5} value of Event 5 was the lowest, at 52 μg·m⁻³. Fig. 3(b) shows that the mean N_d of Event 1 was 34 cm⁻³, approximately three times those of the other events, and it also showed the largest dispersion. The mean N_d values of Events



2-5 were concentrated within the range of $9\text{--}13\text{ cm}^{-3}$. Event 1 was characterized not only by high aerosol loading but also by sufficient LWC (Fig. 3(d)). It also had the smallest mean ED, at $19.5\text{ }\mu\text{m}$ (Fig. 3(c)). These results are highly consistent with the expected characteristics of the classical Twomey effect. The mean $\text{PM}_{2.5}$ value of Event 4 was $75\text{ }\mu\text{g}\cdot\text{m}^{-3}$, the second highest among the five events, whereas its LWC and N_d were both the lowest, at $0.026\text{ g}\cdot\text{m}^{-3}$ and 9.4 cm^{-3} , respectively. This combination of limited water vapor and relatively high aerosol loading may lead to suppressed activation.

The overall statistics for each event can often obscure the transient formation and dissipation processes within sea fog. Therefore, data from the automatic weather station, particulate matter monitor, forward-scatter visibility meter, and fog droplet spectrometer were used to calculate and present the temporal evolution of aerosol and fog-droplet microphysical parameters for the five sea fog events (Fig. 4). The parameters include visibility (Vis), mass extinction efficiency (MEE), fog droplet number concentration (N_d), $\text{PM}_{2.5}$ mass concentration, liquid water content (LWC), and air temperature (T_a).

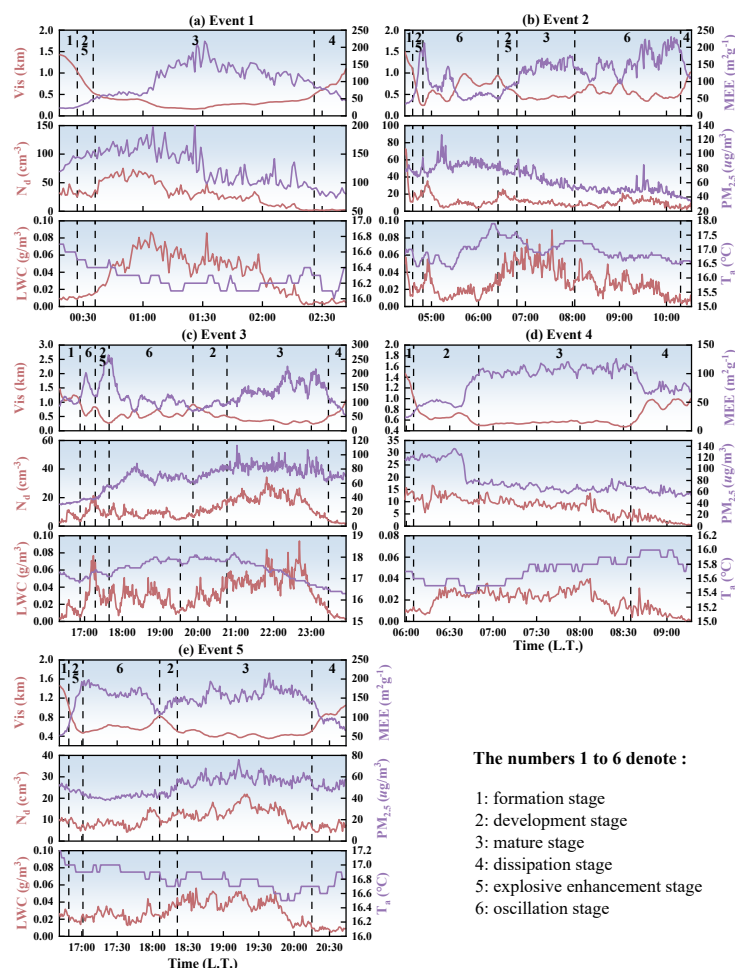


Figure 4: Temporal evolution of microphysical parameters and aerosols during the five sea fog events. The numbers in the figure indicate the different stages of sea fog. The numbers 1 to 6 denote the formation stage, development stage, mature stage, dissipation stage, explosive enhancement stage, and oscillation stage, respectively.



During the formation stage, Vis in all five events decreased from 1.43-1.51 km to 1.0 km, while MEE began to increase. An explosive enhancement stage occurred in four of the five events, and all of these stages occurred during the development stage. In the four events with explosive enhancement, the development stage was accompanied by a decrease in T_a . The explosive enhancement mainly occurred at night, in the early morning, or in the evening, suggesting that near-surface cooling may have provided favorable conditions for the rapid enhancement of sea fog. During the explosive enhancement stage, Vis decreased by at least 500 m within 30 min, while MEE increased markedly. In Event 3, MEE increased from 125 to 265 $\text{m}^2 \cdot \text{g}^{-1}$, whereas N_d decreased from 21 to 9 cm^{-3} . During this stage in Event 5, MEE increased by 135%, while N_d showed only a slight change. In Events 3 and 5, the increase in MEE during the explosive enhancement stage was accompanied by a decrease or little change in N_d . This indicates that the rapid deterioration in Vis was not accompanied by a simultaneous increase in N_d and may be related to the combined changes in LWC, fog droplet size distribution, and air-mass properties. The mature stage is the most complex stage of aerosol- sea fog interactions, during which Vis generally remained at a low level. During this stage, LWC reached relatively high values in all events, whereas N_d showed different variations among the events. Among them, Event 1 generally shows higher LWC, N_d , and aerosol loading, with peak values of 0.086 $\text{g} \cdot \text{m}^{-3}$, 75 cm^{-3} , and 202 $\mu\text{g} \cdot \text{m}^{-3}$, respectively. Event 2 showed the coexistence of high LWC and relatively low N_d . In Event 3, the peak N_d

was 34 cm^{-3} under relatively high LWC conditions. In Event 5, the peak N_d was 22 cm^{-3} under moderate LWC conditions. In Event 4, although $\text{PM}_{2.5}$ remained at a relatively high level of 70-80 $\mu\text{g} \cdot \text{m}^{-3}$ under relatively low-LWC conditions, N_d remained at 3-9 cm^{-3} and showed a decreasing trend. These differences among the events indicate that there is no fixed relationship between aerosol loading and N_d . The microphysical response may be regulated by environmental conditions such as LWC. The oscillation stage occurred only in Events 2, 3, and 5, which had relatively long sea fog lifetimes. During this stage, Vis oscillated between 0.2 and 0.9 km, while MEE and LWC also showed oscillatory variations. These repeated changes in multiple parameters indicate that the microphysical state of sea fog underwent multiple adjustments during its maintenance. This may reflect the intermittent maintenance of sea fog under complex boundary-layer conditions.

3.2 Temporal evolution of ACI and its relationship with MCr during the sea fog events

The analysis of aerosol and fog-droplet microphysical characteristics for the five sea fog events in Sect. 3.1 preliminarily indicates that there are nonlinear characteristics between aerosol $\text{PM}_{2.5}$ and LWC and N_d . To further quantify the effects of aerosols on sea fog, the ACI index of aerosol-cloud interaction as a quantitative measurement standard. The temporal evolution of ACI for the five sea fog events was calculated using the unified 40 min characteristic window in this paper (Fig. 5). Positive ACI indicates a positive logarithmic sensitivity relationship between $\text{PM}_{2.5}$ and N_d , and the direction is consistent with the Twomey response. Negative ACI indicates a negative logarithmic sensitivity relationship between $\text{PM}_{2.5}$ and N_d , and it represents a non-Twomey response. The specific mechanism needs to be determined by combining LWC, effective diameter, fog droplet size spectrum, and meteorological changes.

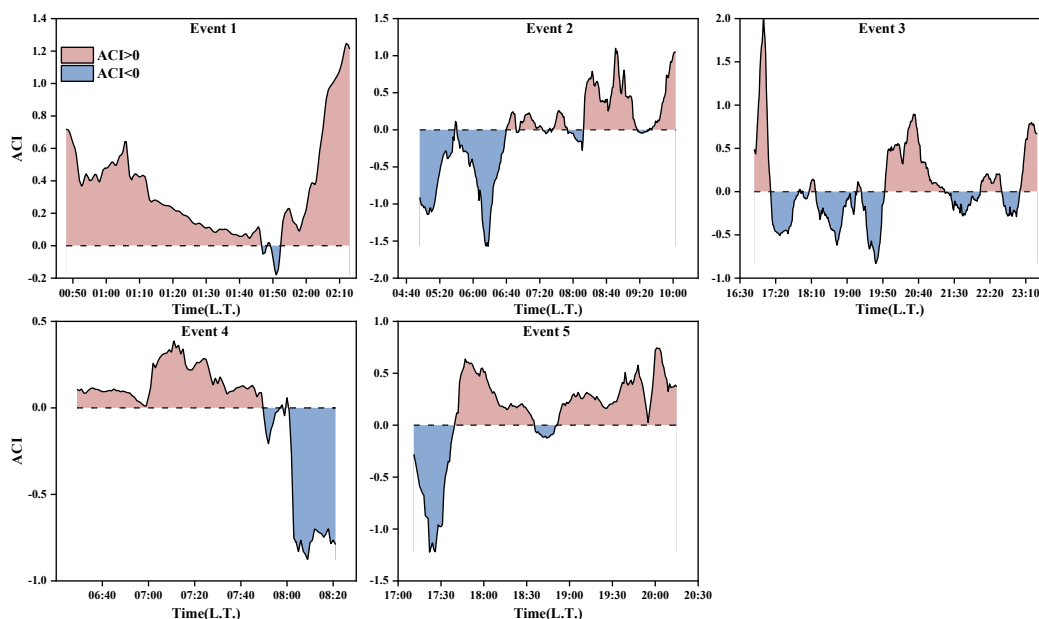
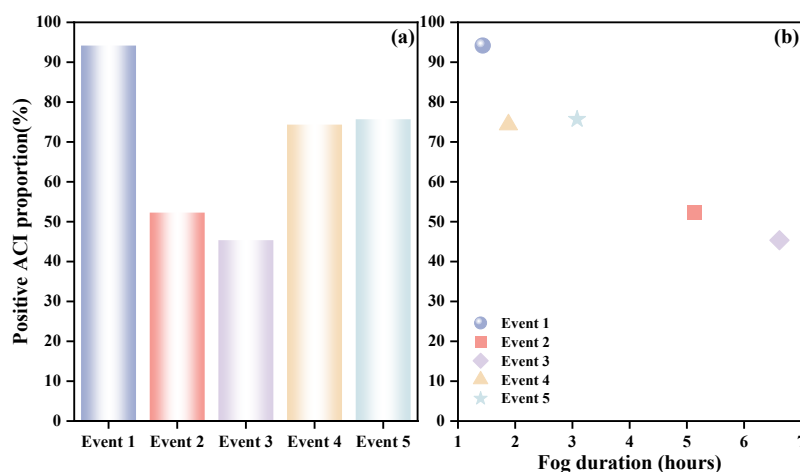


Figure 5: Temporal evolution of ACI during the five sea fog events.

In Event 1, which had both high LWC and high aerosol loading, ACI remained positive for most of the time. During the early mature stage (00:48-01:47), ACI remained positive and reached 0.7. During the same period, N_d increased from 37 to 75 cm^{-3} , and LWC increased from 0.016 to 0.086 $\text{g}\cdot\text{m}^{-3}$. This stage showed a positive response consistent in direction with the Twomey effect. It should be noted that the ACI value of 0.7 exceeded the ideal reference range of $0 < \text{ACI} \leq 0.33$ under the assumption of one-to-one aerosol activation. Because $\text{PM}_{2.5}$ mass concentration was used as the aerosol proxy in this study, this range cannot be regarded as a strict upper limit for the ACI. The relatively high ACI may also be affected by the rapid evolution of LWC, air mass transport, and the common time trends within the window. During the oscillation stages of the long-lived sea fog Events 2 and 3, ACI showed continuous negative ACI responses for more than 1 hour. Some periods of negative ACI were accompanied by an increase in T_a and continuous accumulation of $\text{PM}_{2.5}$, while N_d did not increase correspondingly. This phenomenon is similar to the semi-direct effect reported by Lin et al. (2026) in free-tropospheric warm clouds (T_a increased and N_d was insensitive to aerosols). In Event 4, ACI remained within the range of 0.1-0.4 during the development stage and the early mature stage (06:29-07:50), but became negative during the middle and late mature stage (07:50-08:21), showing a non-Twomey response. In addition, Event 5 showed a negative ACI response during the oscillation stage (17:11-17:40). Overall, ACI in the five events showed clear alternations between positive and negative values over time, indicating that the response relationship between $\text{PM}_{2.5}$ and N_d would change with the evolution state of sea fog, rather than being solely determined by the aerosol load.



335 Figure 6: The proportion of positive ACI time for 5 sea fog events and its relationship with the duration of the events.

The proportion of time with positive ACI ranged from 45% to 94% among the different sea fog events (Fig. 6(a)). The proportion was 94% in Event 1, 74% and 75% in Events 4 and 5, respectively, and 52% and 45% in Events 2 and 3, respectively. Fig. 6(b) shows that, among the five events, the proportion of time with positive ACI generally decreased as the sea fog duration increased. Because of the limited number of events, this trend is not sufficient to establish a universal statistical relationship. Physically, this trend may be related to the relatively lower availability of water relative to aerosol loading in longer-duration sea fog events. Enhanced competition for water vapor may weaken the positive response of N_d to increasing aerosol loading and limit the further growth of fog droplets, thereby reducing the proportion of time with positive ACI responses that are consistent with the Twomey effect. If this water-limited state also suppresses the formation of larger fog droplets as well as collision-coalescence and sedimentation, it may favor the sustained maintenance of sea fog.

345 To identify the empirical transition range of the mean MCr-ACI response in sea fog and establish a unified grouping basis for the subsequent comparison of aerosol-sea fog microphysical characteristics, the ACI results obtained using the 40 min window for the five sea fog events and their corresponding MCr results were used in this section (Fig. 7). The kernel density distribution of MCr-ACI was shown. The binned medians with a bin width of 0.02 and a quadratic polynomial fitting curve were also superimposed to jointly identify the range over which the MCr-ACI response changes.

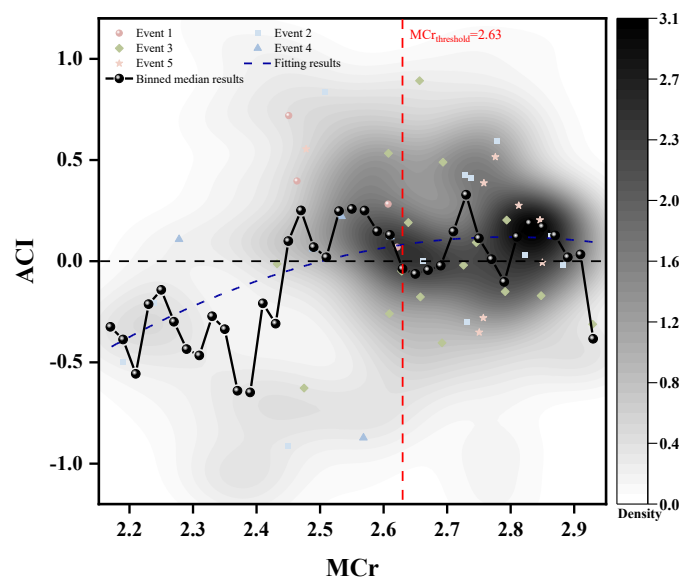


Figure 7: Scatter plot and kernel density distribution of MCr-ACI for the five sea fog events. The kernel density distribution represents the distribution calculated from all data, while the scatter points with different shapes and colors represent the distributions of the individual sea fog events. For the scatter plot, one data point is displayed for every 25 calculations.

The MCr values of the five sea fog events were distributed within the range of 2.17-2.93, which is lower than the MCr threshold value of 4.5 reported by Lin et al. (2026) for warm clouds. When calculated using the same definition, this result indicates that the relative state between LWC and $PM_{2.5}$ aerosol loading in the sea fog samples of this study generally corresponds to a lower MCr range. Due to differences in cloud type, LWC, aerosol composition, and observation system, these two values should not be interpreted as directly interchangeable universal thresholds. For the binned median curve, when $MCr < 2.44$, the binned median ACI ranged from -0.65 to -0.14, with a non-Twomey response dominant. When $MCr \in [2.44, 2.62]$, the binned median ACI began to shift from negative values toward near zero and positive values. This range lies outside the high-value region of the kernel density distribution, where the sample size is relatively limited and the binned medians show large fluctuations. Therefore, this range alone is insufficient to serve as the basis for grouping. When $MCr \in [2.62, 2.93]$, the data were located within the kernel density region, and the binned median ACI crossed zero at $MCr = 2.63$. The quadratic polynomial fitting curve crossed zero at $MCr = 2.51$. However, the amount of data on the left side of this value was relatively small and not representative. The fitting curve reached a local maximum the MCr range of 2.78-2.85. Based on these results, the range of 2.50-2.63 was identified as the empirical response transition range, and $MCr_{threshold} = 2.63$ was selected as the empirical grouping threshold for the subsequent grouping analysis of aerosol-sea fog microphysical characteristics. This value is used only for the samples in this study and does not represent a universal physical critical point.

The scatter points from different events still showed clear differences in ACI within the same MCr range. Most data points from Event 1 were distributed in the region of $MCr \in [2.45, 2.62]$ and $ACI > 0$. Although most samples from this event had MCr values below the empirical grouping threshold, positive responses were dominant, indicating that $MCr < 2.63$

does not necessarily correspond to a non-Twomey response. Most data points from Event 4 were distributed within $MCr \in [2.27, 2.61]$, with ACI reaching a relatively low negative value of -0.87, corresponding to activation suppression under water-vapor-deficient conditions. Most data points from Event 5 were concentrated within $MCr \in [2.7, 2.9]$, but ACI changed from negative to positive. Negative ACI was also observed in Event 3 when $MCr > 2.7$, and the ACI variation may be related to the rise in T_a and boundary-layer disturbances during the oscillation stage. The above results collectively indicate that MCr being higher or lower than 2.63 is not sufficient to independently determine the response direction. LWC and thermal conditions can still further change the response within the same MCr range.

To evaluate the stability of the candidate grouping threshold and the selection of nearby thresholds, a scanning threshold sensitivity test was used in this study. Within $MCr \in [2.25, 2.85]$, $MCr_{\text{threshold}}$ was set sequentially at an interval of 0.05. For each selected threshold, all data were divided into a low- MCr group ($MCr \leq MCr_{\text{threshold}}$) and a high- MCr group ($MCr > MCr_{\text{threshold}}$). The mean ACI, the correlation coefficient (r) between $\ln(PM_{2.5})$ and $\ln(N_d)$, and the root mean square error (RMSE) were then calculated separately for the two groups (Fig. 8).

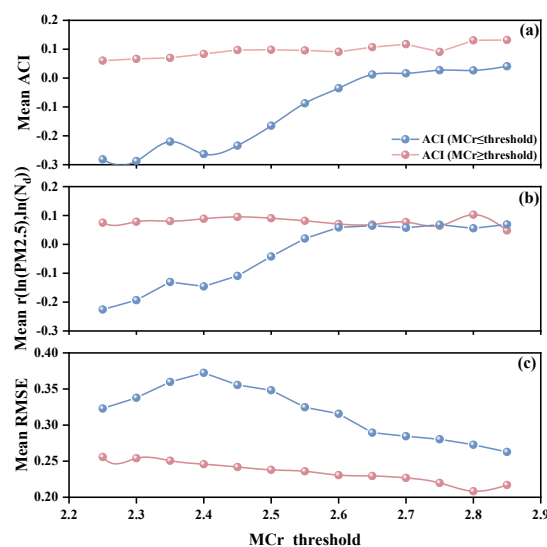


Figure 8: (a) Mean ACI, (b) mean r , and (c) mean RMSE.

The mean ACI of the high- MCr group remained within the range of 0.06-0.13 throughout the entire scanning range (Fig. 8(a)). The mean r remained stable within the positive range of 0.07-0.10 (Fig. 8(b)). The mean RMSE ranged from 0.21 to 0.26 (Fig. 8(c)). The three metrics changed only slightly with the grouping threshold, indicating that the mean response direction of the high- MCr group was relatively robust to the threshold selection. It should be noted that the r values were generally small. Therefore, the robustness here mainly refers to the direction of the mean ACI response and does not indicate a strong linear correlation between $PM_{2.5}$ and N_d .

In the low- MCr group, when $MCr_{\text{threshold}} \leq 2.40$, the mean ACI ranged from -0.27 to -0.22, and the mean r ranged from -0.23 to -0.10. The mean RMSE reached a peak of 0.37 when $MCr_{\text{threshold}}$ was 2.40. As the selected threshold



increased, the metrics of the low-MCr group gradually shifted toward those of the high-MCr group. The mean r changed
 395 from negative to positive when $MCr_threshold$ was approximately 2.50. The mean ACI remained negative at when
 $MCr_threshold$ was 2.60, but became positive when $MCr_threshold$ was 2.65, indicating that the zero crossing occurred
 within the range of 2.60-2.65. $MCr = 2.63$ falls within this range and is also within the dense region of the MCr -ACI kernel
 density distribution (Fig. 7). This also indicates that the response on the low-MCr side did not complete its transition at a
 single MCr value, but underwent a gradual transition from a change in the direction of the mean r to a positive mean ACI.
 400 Based on Fig. 7 and Fig. 8, the range of approximately 2.50-2.63 was finally identified as the empirical transition range of
 the mean MCr -ACI response, and $MCr = 2.63$ was adopted as the empirical grouping threshold for the subsequent
 comparison of microphysical characteristics.

3.3 Difference in the effective diameter of fog droplets

Aerosols regulate the surface energy budget through radiative effects (Huang et al., 2014; Bellouin et al., 2020; Liu et al.,
 405 2021). However, the influence of surface radiation processes on the thermodynamic structure of the boundary layer
 (Deaconu et al., 2019; Graaf et al., 2020) directly affects the formation, maintenance, and dissipation of sea fog. The
 temporal evolution results of radiation parameters and the evolution results of boundary layer temperature with time and
 space during the observation period (Fig. 8 and Fig. 9) were used to analyse how radiation affects the stable structure of the
 boundary layer and the evolution characteristics of the sea fog life cycle.

410 To analyze the relationship between aerosol load and the effective diameter (ED) of droplets under different MCr and
 LWC conditions, this study divided the samples into high- and low- MCr intervals based on $MCr = 2.63$. The high- MCr
 range was $MCr \in [2.63, 2.93]$, and the low- MCr range was $MCr \in [2.17, 2.63]$. Meanwhile, the full sample was divided into
 low-, medium-, and high-LWC groups according to the 33.33rd and 66.67th percentiles of LWC. The low-LWC range was
 $LWC \in [0, 0.025] \text{ g} \cdot \text{m}^{-3}$, the medium-LWC range was $LWC \in [0.025, 0.039] \text{ g} \cdot \text{m}^{-3}$, and the high-LWC range was $LWC \in$
 415 $[0.039, 0.094] \text{ g} \cdot \text{m}^{-3}$. The relationship between $\ln(PM_{2.5})$ and ED was investigated for the full sample and within the different
 MCr and LWC ranges (Fig. 9). The aerosol-cloud interaction index based on effective radius, $ACI_{re} = -\partial \ln(re) / \partial \ln(PM_{2.5})$,
 and the corresponding correlation coefficient $r(ACI)$ were also calculated under constant liquid water conditions. Because
 $ED = 2re$, $\ln(ED)$ and $\ln(re)$ differ only by a constant. Therefore, the corresponding logarithmic sensitivity slopes are
 equivalent.

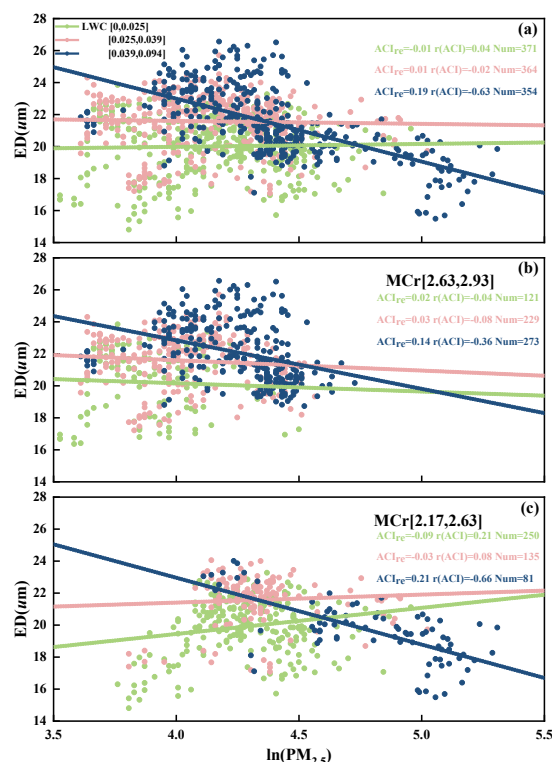


Figure 9: The relationship between aerosols and ED under different MCr and LWC conditions.

The results for the full sample (Fig. 9(a)) show that, without constraining MCr, only the high-LWC group exhibited a clear decrease in ED with increasing aerosol load, and the direction of the droplet size change was consistent with the Twomey effect. In contrast, the ACI_{re} values of the medium- and low-LWC groups were -0.01 and 0.01 , respectively, and their regression lines were nearly horizontal. These results indicate that the response between aerosol load and fog droplet size has a clear dependence on LWC conditions. These results also show clear differences in the relationship between $PM_{2.5}$ and ED among different LWC groups, and the overall analysis without LWC grouping may mask these differences.

In the high-MCr range (Fig. 9(b)), the ACI_{re} values of the three LWC groups increased successively as LWC increased. The ACI_{re} values of the low-, medium-, and high-LWC groups were 0.02 , 0.03 , and 0.14 , respectively. The absolute values of $r(ACI)$ also increased as LWC increased. The $r(ACI)$ values of the low-, medium-, and high-LWC groups were -0.04 , -0.08 , and -0.36 , respectively. The relationship between $PM_{2.5}$ and ED was weak in the low- and medium-LWC groups, whereas a relatively clear negative correlation was observed in the high-LWC group. The direction of the droplet size change was consistent with the Twomey effect. This indicates that even within the high-MCr range, there are still differences in the association between $PM_{2.5}$ and ED among the LWC groups, and high-MCr does not necessarily correspond to a stronger response of droplet size.

In the low-MCr range (Fig. 9(c)), the ACI_{re} values of the three LWC groups changed from negative to positive as LWC increased. The ACI_{re} values of the low-, medium-, and high-LWC groups were -0.09 , -0.03 , and 0.21 , respectively. The



correlation coefficient $r(\text{ACI})$ showed a trend from positive to negative. The $r(\text{ACI})$ values of the low-, medium-, and high-LWC groups were 0.21, 0.08, and -0.66, respectively. In the low- and medium-LWC groups, ED showed a weak non-Twomey change with increasing $\text{PM}_{2.5}$. In the high-LWC group ($\text{LWC} \geq 0.039 \text{ g} \cdot \text{m}^{-3}$), ED decreased markedly with increasing $\text{PM}_{2.5}$, and the direction of the droplet size change shifted to one consistent with the Twomey effect. This high-LWC group mainly originated from the high aerosol load and high LWC state during the early mature stage of Event 1. Because this group was clearly dominated by a single event, its representativeness across events still requires further examination using more independent observations. These results indicate that, under low-MCr conditions, the response direction between $\text{PM}_{2.5}$ and ED still shows a clear dependence on LWC conditions. MCr below 2.63 does not necessarily correspond to a non-Twomey response. In the high-LWC group of this dataset, the droplet size response under low-MCr conditions can shift from a non-Twomey type to a direction consistent with the Twomey effect. Overall, the relationship between $\text{PM}_{2.5}$ and ED shows a clear dependence on the combined MCr-LWC state.

3.4 Differences in droplet size

In this section, within the high- and low-MCr subranges, the data were divided into high and low aerosol loading groups, based on the median of $\text{PM}_{2.5}$ in each subinterval. Under this classification, the normalized frequency distributions of LWC, ED, and N_d were investigated (Fig. 10).

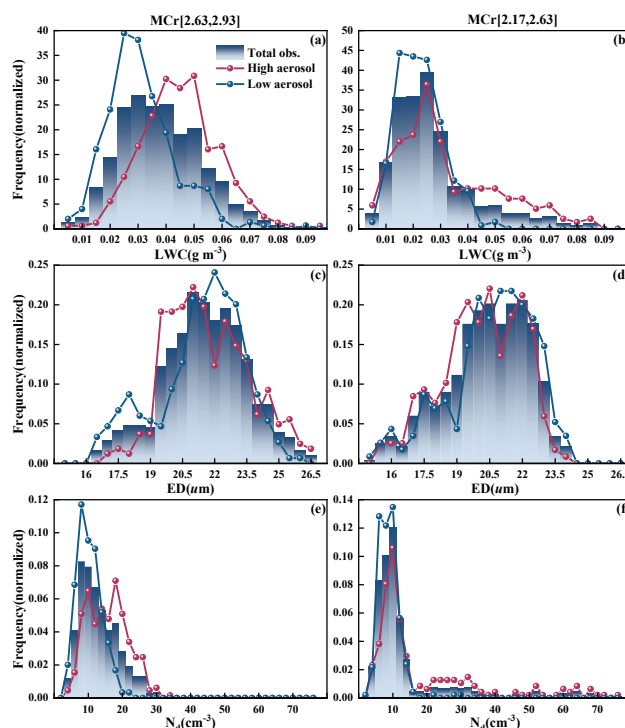


Figure 10: Normalized frequency distributions of LWC, ED, and N_d for the high and low aerosol loading groups within different MCr ranges. The left columns (a), (c), and (e), shows the normalized frequency distributions of LWC, ED, and N_d , respectively, within the high-



M_{Cr} range. The right columns (b), (d), and (f), shows the normalized frequency distributions of LWC, ED, and N_d , respectively, within the low-M_{Cr} range.

In the high-M_{Cr} range in Fig. 10, the high aerosol loading group showed higher N_d and smaller ED, exhibiting the typical characteristics of the Twomey effect. For the N_d distribution (Fig. 10(e)), the low aerosol loading group showed a strong unimodal distribution, with the mode concentrated at 8-10 cm^{-3} . The high aerosol loading group showed a bimodal structure, with the main mode shifting to 18-20 cm^{-3} . Meanwhile, the frequency of $N_d > 20 \text{ cm}^{-3}$ was higher than that in the low aerosol loading group. For the ED distribution (Fig. 10(c)), the mode of the low aerosol loading group was concentrated at 21.5-22.5 μm , whereas the mode of the high aerosol loading group shifted to a smaller ED of approximately 21 μm . With the increase in aerosol load, the synergistic change of N_d increase and ED decrease is consistent with the Twomey effect. Fig. 10(a) further shows that the LWC distribution of the high aerosol loading group shifted overall toward higher values, with the mode at approximately 0.04-0.05 $\text{g} \cdot \text{m}^{-3}$, compared with approximately 0.025 $\text{g} \cdot \text{m}^{-3}$ for the low aerosol loading group.

In the low-M_{Cr} range in Fig. 10, the main distributions of the three microphysical parameters were relatively similar between the high and low aerosol loading groups. For the LWC distribution (Fig. 10(b)), the modes of both the high and low aerosol loading groups were concentrated within the relatively low-LWC range of 0.015-0.025 $\text{g} \cdot \text{m}^{-3}$, while the distribution of the high aerosol loading group extended toward higher LWC. The N_d and ED distributions of the high and low aerosol loading groups (Fig. 10(f, d)) showed a high degree of overlap within their main distribution ranges and did not exhibit the clear combined characteristics of increased N_d and decreased ED observed in the high-M_{Cr} range. The main peaks of N_d were both concentrated at 7-10 cm^{-3} . The high aerosol loading group showed a weak tailing distribution over 20-75 cm^{-3} . These samples mainly corresponded to the high aerosol loading and high LWC state during the early mature stage of Event 1 in the low-M_{Cr} and high-LWC subgroup shown in Fig. 9(c). Overall, after empirical grouping using $M_{Cr} = 2.63$, clear differences were observed in the fog droplet microphysical distributions associated with changes in aerosol load between the high- and low-M_{Cr} ranges, indicating that this empirical grouping threshold can effectively characterize different sea fog microphysical states.

To further compare the differences in fog droplet size distributions between the high and low aerosol loading groups within different M_{Cr} ranges, the mean fog droplet spectra of the high and low aerosol loading groups and the observational mean of the full sample were presented (Fig. 11).

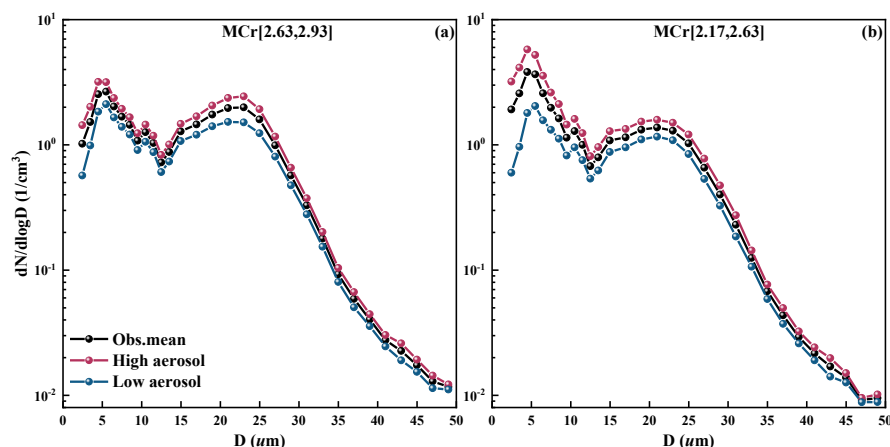


Figure 11: The mean fog droplet spectra of the high and low aerosol loading groups within the two MCr ranges, together with the observational mean of the full sample.

485 The mean fog droplet spectra in the two MCr ranges both showed similar bimodal structures. The small droplet modal peak occurred at 4-6 μm , the spectral valley occurred at 12-13 μm , and the large droplet modal peak occurred at 20-23 μm . The fog droplet number size distribution decreased rapidly at diameters larger than 25 μm . In the high-MCr range (Fig. 11(a)), fog droplets in both the small and large droplet size ranges increased markedly under high aerosol loading conditions. The small droplet modal peaks of the high and low aerosol loading groups were 3.0 and 2.0 cm^{-3} , respectively, with a ratio of approximately 1.5. The large droplet modal peaks of the high and low aerosol loading groups were 2.4 and 1.5 cm^{-3} , respectively, with a ratio of approximately 1.6. The peak ratios of the two modes in the high and low aerosol load groups were basically consistent. This result showed simultaneous enhancement of the small and large droplet modes. This indicates that, while small droplets associated with activation increased, larger fog droplets also increased accordingly. Overall, this spectral feature is consistent with continued growth of fog droplets toward larger sizes after activation.

495 In the low-MCr range (Fig. 11(b)), the fog droplet spectrum showed response characteristics that were clearly different from those in the high-MCr range. The main feature was a marked enhancement of the small droplet mode, while the enhancement of the large droplet mode was relatively limited. The small droplet modal peaks of the high and low aerosol loading groups were 5.8 and 1.7 cm^{-3} , respectively, with a ratio of approximately 3.4. Meanwhile, in the low-MCr range, the small droplet modal peak of the high aerosol loading group was approximately twice that of the high-MCr range. This indicates a clear accumulation characteristic of small droplets under low-MCr conditions. However, the large droplet modal peaks of the high and low aerosol loading groups were 1.45 and 1.05 cm^{-3} , respectively, with a ratio of approximately 1.38. The large droplet mode also increased under high aerosol loading conditions, but its increase was clearly smaller than that of the small droplet mode. Compared with the high-MCr range, the large droplet modal peak of the high aerosol loading group was also relatively lower in the low-MCr range. This spectral feature indicates that, under low-MCr conditions, increasing aerosol load mainly corresponded to a marked enhancement of the small droplet mode, while the enhancement of the large droplet mode was relatively limited. This reflects that when more aerosols participate in droplet activation, due to the

relatively limited water content conditions, the ability of droplets to further develop towards larger sizes is weakened, thereby forming a clear small droplet accumulation spectral feature. From the spectral feature, this reflects that the activation-growth pathway of fog droplets was somewhat restricted under low-MCr conditions, and fog droplet growth remained relatively more concentrated at the small droplet end.

Based on the results in Fig. 11, the LWC constraint was further introduced to compare the differences in fog droplet spectra corresponding to changes in aerosol load under different MCr-LWC combined states (Fig. 12).

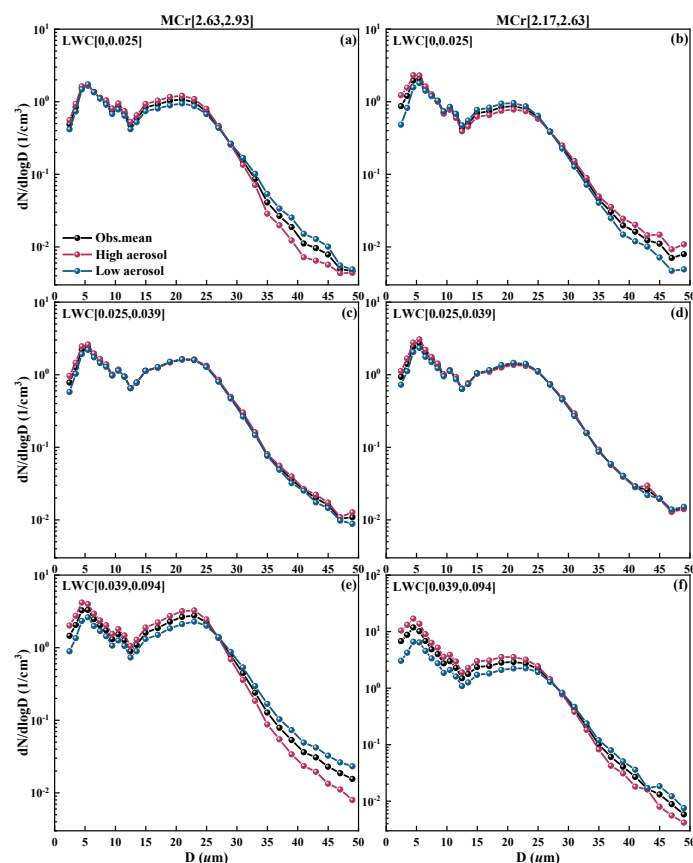


Figure 12: The mean fog droplet spectra of the high and low aerosol loading groups within the two MCr ranges, together with the observational mean of the full sample.

In the high-MCr range, under low-LWC conditions (Fig. 12(a)), the peak ratios between the high and low aerosol loading groups for the small and large droplet modes were approximately 1.13 and 1.41, respectively. Under medium-LWC conditions (Fig. 12(c)), the peak ratio between the high and low aerosol loading groups for the small droplet mode was 1.24, while the large droplet modal peaks of the two groups completely overlapped. Under high-LWC conditions (Fig. 12(e)), the peak ratios between the high and low aerosol loading groups for the small and large droplet modes further increased to 1.59 and 1.43, respectively. In the decay region at the large droplet end ($D \geq 30 \mu\text{m}$), the spectral values of the fog droplets $dN/d\log D$ in the low aerosol load group were generally higher than those in the high aerosol load group, and the difference



between the two groups was more obvious at the larger droplet sizes end. As LWC increased, the peak ratio between the high and low aerosol loading groups for the small droplet mode continuously increased from 1.13 to 1.24 and then to 1.59. Because the small droplet mode mainly represents fog droplets formed by aerosol activation, this result indicates that, under high-MCr conditions, the fog droplet activation response corresponding to an increase in aerosol load gradually strengthened as LWC increased. The direction of this change was consistent with the Twomey effect. This result and the finding in Fig. 9(b) mutually corroborate each other. Under high-MCr conditions, ACI_{re} increased with LWC (0.02, 0.03, and 0.14), reflecting a strengthening of the Twomey effect with increasing LWC.

In the low-MCr range, under low-LWC conditions (Fig. 12(b)), the small droplet modal peaks of the high and low aerosol loading groups were 2.3 and 1.8 cm^{-3} , respectively, with a ratio of 1.28. The large droplet modal peaks of the high and low aerosol loading groups were 0.78 and 0.95 cm^{-3} , respectively, with a ratio of 0.82. A similar spectral feature was observed under medium-LWC conditions (Fig. 12(d)). Because the large droplet mode mainly reflects the further growth of fog droplets toward larger sizes after activation, this spectral feature of an enhanced small droplet mode and a relatively weakened large droplet mode is consistent with a state in which subsequent fog droplet growth is restricted under relatively limited water content conditions. Under high-LWC conditions (Fig. 12(f)), the spectral values of the fog droplets $dN/d\log D$ increased overall across all droplet size ranges. The small droplet modal peak of the high aerosol loading group (16.9 cm^{-3}) was much higher than that of the low aerosol loading subgroup (6.5 cm^{-3}). The peak ratios between the high and low aerosol loading groups for the small and large droplet modes were 2.6 and 1.6, respectively.

Under the high-LWC constraint in Fig. 12(e) and (f), the small and large droplet modal peaks of the high aerosol loading group were both higher than those of the low aerosol loading group. In both MCr ranges, the small droplet mode was enhanced as aerosol load increased, indicating that the small droplet response associated with aerosol activation was strengthened in both ranges. The main differences between the two MCr ranges were reflected in the magnitude of the spectral values of the fog droplets $dN/d\log D$ and in the relative changes of the small and large droplet modes. The small droplet modal peak of the high aerosol loading group in Fig. 12(f) was four times that of the corresponding peak in Fig. 12(e). Meanwhile, the spectral feature in Fig. 12(f) showed a more pronounced small droplet accumulation characteristic. These differences in the magnitude of the spectral values and in the relative changes of the small and large droplet modes further indicate that empirical grouping using $MCr = 2.63$ can distinguish two different microphysical states. One state occurs under high-MCr and high-LWC conditions, where both the small and large droplet modes are enhanced as aerosol load increases, showing a relatively complete activation-growth pathway. The other state occurs under low-MCr and high-LWC conditions, where the enhancement of the small droplet mode is clearly greater than that of the large droplet mode. This state shows a pronounced small droplet accumulation characteristic and reflects a restricted activation-growth pathway.

3.5 Differences in the autoconversion threshold function (T)

Aerosols can affect the T by changing the fog droplet spectral shape, thereby influencing the collision-coalescence processes of fog droplets. Therefore, the normalized frequency distributions of the T in the high- and low-MCr ranges were calculated

in this section. The value of T ranges from 0 to 1, and a larger T indicates a stronger collision–coalescence process of fog droplets. Meanwhile, the relationship between T and $\ln(\text{PM}_{2.5})$ was analyzed under different LWC constraint (Fig. 13).

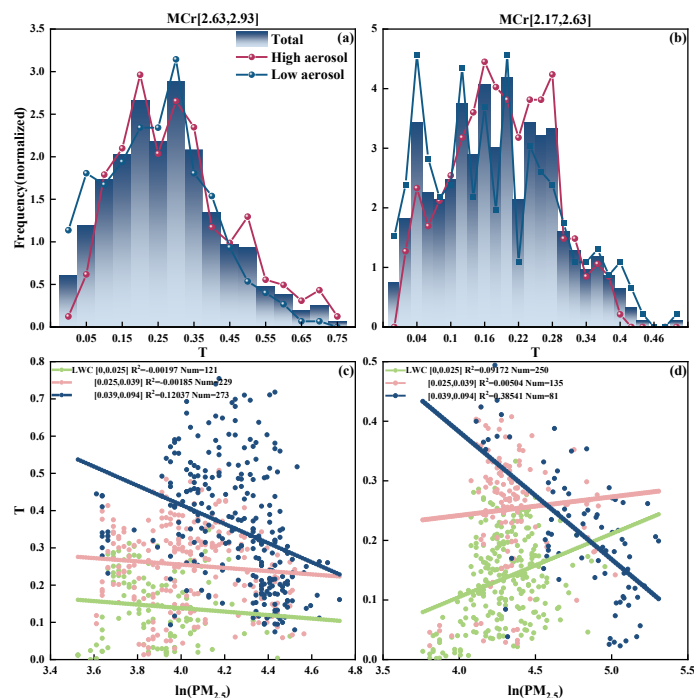


Figure 13: Normalized frequency distributions of the T in different MCr ranges (a, b), and the relationships between T and $\ln(\text{PM}_{2.5})$ under different LWC constraints (c, d).

In the overall distribution of T (Fig. 13(a) and (b)), the T distribution range in the high-MCr range (0-0.75) was higher than that in the low-MCr range (0-0.5). This indicates that higher T states occurred more readily in the high-MCr range, whereas the collision–coalescence process was relatively weaker overall in the low-MCr range. In the high-MCr range (Fig. 13(c)), the regression lines between T and $\ln(\text{PM}_{2.5})$ were nearly horizontal under low- and medium-LWC conditions, indicating that the effect of aerosol load on T was weak when LWC was relatively low. Under high-LWC conditions, T was negatively correlated with $\ln(\text{PM}_{2.5})$, and T decreased from 0.53 to 0.23 as $\ln(\text{PM}_{2.5})$ increased. This result is consistent with the Twomey effect in the direction of droplet size reduction under high-MCr and high-LWC conditions shown in Fig. 9(b). This indicates that, as aerosol load increased, fog droplet size decreased and T also decreased markedly, corresponding to a weakened collision-coalescence process. In the low-MCr range (Fig. 13(d)), T was positively correlated with $\ln(\text{PM}_{2.5})$ under low-LWC conditions, and T increased from 0.07 to 0.24 as $\ln(\text{PM}_{2.5})$ increased. This change is consistent with the non-Twomey response of ED increasing with aerosol load in the low-MCr and low-LWC condition in Fig. 9(c). This indicates that an increase in aerosol load was accompanied by increases in fog droplet size and T , corresponding to an enhanced collision-coalescence process. Under high-LWC conditions, T showed a strong negative correlation with $\ln(\text{PM}_{2.5})$. This phenomenon was mainly contributed by the high aerosol loading and high-LWC state during the early mature stage of Event 1. This result is consistent with the decrease in ED with increasing aerosol load under low-MCr and high-LWC conditions



shown in Fig. 9(c). This indicates that, as aerosol load increased, the decrease in fog droplet size and the decrease in T occurred simultaneously, corresponding to a tendency toward a weakened collision-coalescence process. Therefore, within the low-MCr range, the response direction of T to aerosol load changed with LWC conditions, further indicating a clear dependence of the collision-coalescence process on the combined MCr-LWC state.

580 4 Summary and conclusions

This study was based on high temporal resolution observations from the ground-based observation platform on Qingbang Island in the East China Sea. Five advection sea fog events observed during spring from March to May 2025 were investigated in detail over their entire life cycles. Based on the convergence characteristics of the normalized standard errors of $PM_{2.5}$, N_d , and LWC and their first-order differences, a characteristic time window of 40 min was determined for the ACI calculation. This window was used to suppress high-frequency fluctuations while retaining the main evolution characteristics of the sea fog processes. The MCr framework was extended from warm clouds to sea fog. An empirical transition range of the MCr-ACI response was identified, and an empirical MCr threshold was selected for the subsequent grouping analysis. On this basis, the temporal evolution of ACI and its relationship with MCr were further analyzed. The differences in sea fog microphysical responses corresponding to aerosol load were then analyzed by combining ED, fog droplet spectral characteristics, and the T under different MCr and LWC conditions. The results show that the microphysical responses of aerosol-sea fog interactions exhibit a clear dependence on the combined MCr-LWC state (Fig. 14). The main conclusions are as follows.

Conceptual Diagram of Aerosol–Sea Fog Microphysical Responses under Different MCr–LWC States

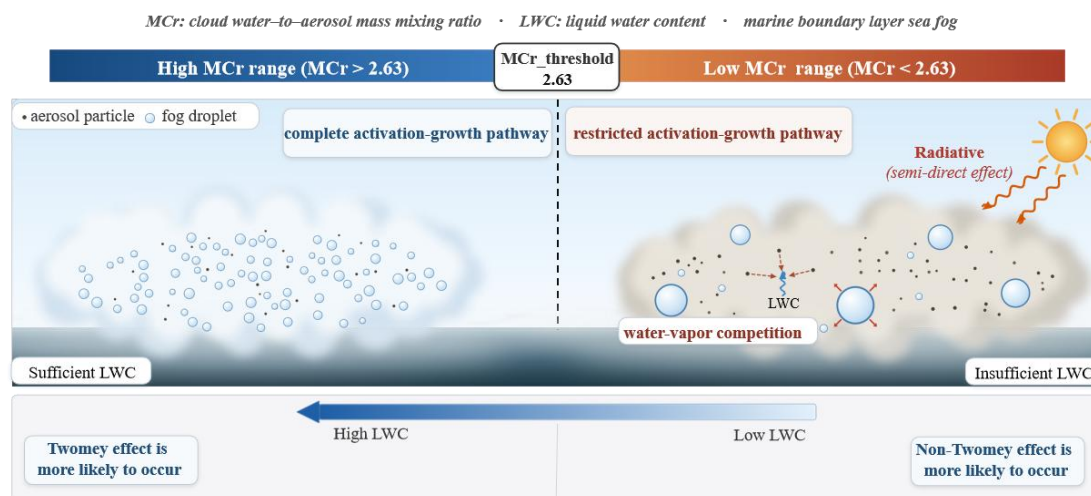


Figure 14: Conceptual diagram of the microphysical responses of aerosol-sea fog interactions under the constraints of MCr-LWC.

595 The five sea fog events show that aerosol-sea fog microphysical responses varied markedly among events and life-cycle stages rather than being determined by aerosol loading alone. Event 1 showed the highest N_d and the smallest ED under high



aerosol loading and sufficient LWC, which was consistent with the classical Twomey effect. In contrast, low LWC and N_d were observed in Event 4 despite relatively high aerosol loading. The six-stage analysis further showed that explosive enhancement occurred during the development stage, whereas oscillation was observed only in long-lived events. Consistently, ACI alternated between positive and negative values across events and stages, and the proportion of time with positive ACI generally decreased with increasing sea fog duration. These results indicate that the $PM_{2.5}$ - N_d response changed with the evolving sea fog state and the relative availability of water. These differences in microphysical response states were further characterized using MCr. MCr ranged from 2.17 to 2.93, and the MCr-ACI response showed a gradual transition from negative to positive within approximately 2.50-2.63. MCr = 2.63 was therefore selected as the empirical grouping threshold for subsequent microphysical analysis. This value is substantially lower than the MCr threshold of 4.5 reported by Lin et al. (2026) for free-tropospheric warm clouds. The threshold sensitivity analysis showed a relatively stable mean response on the high-MCr side and greater sensitivity on the low-MCr side. Thus, MCr boundaries derived for different cloud and fog regimes cannot be directly transferred between systems (Figs. 3-8).

Within the empirical MCr grouping, both the direction and strength of the aerosol-related microphysical responses were further regulated by LWC. In the high-MCr range, ACI_{re} increased from 0.02 to 0.14 with increasing LWC, indicating a progressively stronger droplet-size reduction response consistent with the Twomey effect. In the low-MCr range, ACI_{re} shifted from -0.09 and -0.03 under low- and medium-LWC conditions to 0.21 under high-LWC conditions, demonstrating that low MCr does not necessarily correspond to a non-Twomey response. Fog droplet spectra and T provided mutually consistent microphysical evidence for this state dependence. Under high-MCr and high-LWC conditions, both the small and large droplet modes were enhanced with increasing aerosol load, indicating a relatively complete activation-growth pathway. Under low-MCr and high-LWC conditions, more pronounced small droplet accumulation and relatively restricted subsequent growth were observed. The small droplet modal peak under high aerosol loading was approximately four times that in the corresponding high-MCr group. The response direction of T to aerosol loading likewise changed with the combined MCr-LWC state. Taken together, the ACI_{re} , ED, fog droplet spectra, and T results indicate that MCr distinguishes different relative aerosol-liquid water states, while LWC further regulates fog droplet activation, subsequent growth, and collision-coalescence. Aerosol effects on sea fog microphysics should therefore be interpreted in terms of the combined MCr-LWC state rather than aerosol concentration alone (Figs. 9-13).

This study shows that aerosol-sea fog interactions exhibit a clear dependence on the combined MCr-LWC state. The single MCr threshold identified for warm clouds cannot be directly extended to sea fog, and aerosol responses in sea fog should not be parameterized as a monotonic function of aerosol concentration. Aerosol concentration alone is insufficient to determine the direction and strength of aerosol effects on sea fog microphysics. Combining MCr-LWC can more effectively distinguish fog droplet activation, droplet-size growth, and collision-coalescence responses under different aerosol-liquid water combined states. This helps deepen the understanding of the microphysical processes of aerosol-sea fog interaction. At the same time, it provides observational evidence for the analysis and prediction of the evolution of sea fog under high pollution conditions. This study uses $PM_{2.5}$ mass concentration as a proxy for cloud condensation nucleus number



concentration, and does not directly characterize the size distribution and chemical composition of cloud condensation nuclei. The supersaturation under near-saturated conditions is also difficult to observe directly, and the number of observed events and the sample sizes of some subranges remain relatively limited. In the future, more sea fog events can be observed, and combined with direct measurements of cloud condensation nuclei, the applicability of the MCr-LWC framework in different regions and types of sea fog can be further tested.

Data availability

The relevant data of the observation platforms at the observation sites in this study is available at <https://doi.org/10.5281/zenodo.22940490> (Ma, 2026).

Author contributions

Xirui Ma: Writing-original draft, Software, Data curation, Funding acquisition. Yun Zhang: Review & editing, Funding acquisition. Yanqiong Xie: Review, Validation. Panyan Ge, Lingbing Bu and Yan Yin: Supervision.

Competing interests

The authors in this article declare that they have no conflict of interest with others.

Acknowledgements

We thank the anonymous reviewers and editors for their instructive suggestions and comments.

Financial support

This study was supported by the National Natural Science Foundation of China (NSFC) (grants nos. 42530611, U2542209, and 42505094), the Hunan Provincial Key Research and Development Program (grants no. 2025AQ2012) and the Postgraduate Research and Practice Innovation Program of Jiangsu Province (grant no. 26CXJH2422).

References

Albrecht, B. A.: Aerosols, cloud microphysics, and fractional cloudiness, *Science*, 245, 1227–1230, <https://doi.org/10.1126/science.245.4923.1227>, 1989.



- Andreae, M. O. and Rosenfeld, D.: Aerosol–cloud–precipitation interactions. Part 1. The nature and sources of cloud-active aerosols, *Earth-Sci. Rev.*, 89, 13–41, <https://doi.org/10.1016/j.earscirev.2008.03.001>, 2008.
- 655 Arun, N., Sarangi, C., and Qian, Y.: Particulate pollution increases fog duration over the Indo-Gangetic Plain, *Geophys. Res. Lett.*, 52, e2024GL114095, <https://doi.org/10.1029/2024GL114095>, 2025.
- Arun, N., Sarangi, C., Kanawade, V. P., Singh, M. K., and Gautam, R.: Widespread fog invigoration due to aerosols over Northern India, *Sci. Adv.*, 12, eadx1662, <https://doi.org/10.1126/sciadv.adx1662>, 2026.
- Bellouin, N., Quaas, J., Gryspeerdt, E., Kinne, S., Stier, P., Watson-Parris, D., N., Boucher, O., Carslaw, K. S., Christensen, M., Daniau, A.-L., Dufresne, J.-L., Feingold, G., Fiedler, S., Forster, P., Gettelman, A., Haywood, J. M., Lohmann, U., Malavelle, F., Mauritsen, T., McCoy, D. T., Myhre, G., Mülmenstädt, J., Neubauer, D., Possner, A., Rugenstein, M., Sato, Y., Schulz, M., Schwartz, S. E., Sourdeval, O., Storelvmo, T., Toll, V., Winker, D., and Stevens, B.: Bounding global aerosol radiative forcing of climate change, *Rev. Geophys.*, 58, e2019RG000660, <https://doi.org/10.1029/2019RG000660>, 2020.
- 665 Boucher, O., Randall, D., Artaxo, P., Bretherton, C., Feingold, G., Forster, P., et al.: Clouds and aerosols, in: *Climate Change 2013: The Physical Science Basis*, Cambridge University Press, Cambridge, 571–658, <https://doi.org/10.1017/CBO9781107415324.016>, 2013.
- Boutle, I., Price, J., Kudsztsa, I., Kokkola, H., and Romakkaniemi, S.: Aerosol–fog interaction and the transition to well-mixed radiation fog, *Atmos. Chem. Phys.*, 18, 7827–7840, <https://doi.org/10.5194/acp-18-7827-2018>, 2018.
- 670 Carslaw, K. S., Lee, L. A., Reddington, C. L., Pringle, K. J., Rap, A., Forster, P. M., Mann, G. W., Spracklen, D. V., Woodhouse, M. T., Regayre, L. A., and Pierce, J. R.: Large contribution of natural aerosols to uncertainty in indirect forcing, *Nature*, 503, 67–71, <https://doi.org/10.1038/nature12674>, 2013.
- Chen, Y.-C., Wang, S.-H., Min, Q., Lu, S., Lin, P.-L., Lin, N.-H., Chung, K.-S., and Joseph, E.: Aerosol impacts on warm-cloud microphysics and drizzle in a moderately polluted environment, *Atmos. Chem. Phys.*, 21, 4487–4502, <https://doi.org/10.5194/acp-21-4487-2021>, 2021.
- 675 Dorman, C. E., Mejia, J., Koraćin, D., and McEvoy, D.: World marine fog analysis based on 58-years of ship observations, *Int. J. Climatol.*, 40, 145–168, <https://doi.org/10.1002/joc.6200>, 2020.
- Dusek, U., Frank, G. P., Hildebrandt, L., Curtius, J., Schneider, J., Walter, S., Chand, D., Drewnick, F., Hings, S., Jung, D., Borrmann, S., Andreae, M. O.: Size matters more than chemistry for cloud-nucleating ability of aerosol particles, *Science*, 312, 1375–1378, <https://doi.org/10.1126/science.1125261>, 2006.
- 680 Fan, J., Wang, Y., Rosenfeld, D., and Liu, X.: Review of aerosol–cloud interactions: mechanisms, significance, and challenges, *J. Atmos. Sci.*, 73, 4221–4252, <https://doi.org/10.1175/JAS-D-16-0037.1>, 2016.
- Feingold, G., Remer, L. A., Ramaprasad, J., and Kaufman, Y. J.: Analysis of smoke impact on clouds in Brazilian biomass burning regions: an extension of Twomey's approach, *J. Geophys. Res.-Atmos.*, 106, 22907–22922, <https://doi.org/10.1029/2001JD000732>, 2001.
- 685



- Fernando, H. J. S., Gultepe, I., Dorman, C., Pardyjak, E., Wang, Q., Hoch, S. W., et al.: C-FOG: life of coastal fog, *Bull. Am. Meteorol. Soc.*, 102, E244–E272, <https://doi.org/10.1175/BAMS-D-19-0070.1>, 2021.
- Gao, S., Lin, H., Shen, B., and Fu, G.: A heavy sea fog event over the Yellow Sea in March 2005: analysis and numerical modeling, *Adv. Atmos. Sci.*, 24, 65–81, <https://doi.org/10.1007/s00376-007-0065-2>, 2007.
- 690 Gryspeerdt, E., Quaas, J., and Bellouin, N.: Constraining the aerosol influence on cloud fraction, *J. Geophys. Res.-Atmos.*, 121, 3566–3583, <https://doi.org/10.1002/2015JD023744>, 2016.
- Gultepe, I., Heymsfield, A. J., Fernando, H. J. S., Pardyjak, E., Dorman, C. E., Wang, Q., Creegan, E., Hoch, S. W., Flagg, D. D., Yamaguchi, R., Krishnamurthy, R., Gaberšek, S., Perrie, W., Perelet, A., Singh, D. K., Chang, R., Nagare, B., Wagh, S., and Wang, S.: A review of coastal fog microphysics during C-FOG, *Bound.-Layer Meteorol.*, 181, 227–265, 695 <https://doi.org/10.1007/s10546-021-00659-5>, 2021.
- Gultepe, I., Müller, M. D., and Boybeyi, Z.: A new visibility parameterization for warm-fog applications in numerical weather prediction models, *J. Appl. Meteorol. Climatol.*, 45, 1469–1480, <https://doi.org/10.1175/JAM2423.1>, 2006.
- Gultepe, I., Tardif, R., Michaelides, S.C., Cermak, J., Bott, A., Bendix, J., Müller, M.D., Pagowski, M., Hansen, B., Ellrod, G., Jacobs, W., Toth, G., and Cober, S.G.: Fog research: a review of past achievements and future perspectives, *Pure Appl. Geophys.*, 164, 1121–1159, <https://doi.org/10.1007/s00024-007-0211-x>, 2007. 700
- Haeffelin, M., Bergot, T., Elias, T., Tardif, R., Carrer, D., Chazette, P., Colomb, M., Drobinski, P., Dupont, E., Dupont, J.-C., Gomes, L., Musson-Genon, L., Pietras, C., Plana-Fattori, A., Protat, A., Rangognio, J., Raut, J.-C., Rémy, S., Richard, D., Sciare, J., and Zhang, X.: PARISFOG: shedding new light on fog physical processes, *Bull. Am. Meteorol. Soc.*, 91, 767–783, <https://doi.org/10.1175/2009BAMS2671.1>, 2010.
- 705 He, J., Ren, X., Wang, H., Shi, Z., Zhang, F., Hu, L., Zeng, Q., and Jin, X.: Analysis of the microphysical structure and evolution characteristics of a typical sea fog weather event in the eastern sea of China, *Remote Sens.*, 14, 5604, <https://doi.org/10.3390/rs14215604>, 2022.
- Huang, H., Liu, H., Huang, J., Mao, W., and Bi, X.: Atmospheric boundary layer structure and turbulence during sea fog on the southern China coast, *Mon. Weather Rev.*, 143, 1907–1923, <https://doi.org/10.1175/MWR-D-14-00207.1>, 2015.
- 710 Jia, X., Quan, J., Zheng, Z., Liu, X., Liu, Q., He, H., and Liu, Y.: Impacts of anthropogenic aerosols on fog in North China Plain, *J. Geophys. Res.-Atmos.*, 124, 252–265, <https://doi.org/10.1029/2018JD029437>, 2019.
- Kaufman, Y. J. and Fraser, R. S.: The effect of smoke particles on clouds and climate forcing, *Science*, 277, 1636–1639, <https://doi.org/10.1126/science.277.5332.1636>, 1997.
- Kaufman, Y. J. and Fraser, R. S.: The Effect of Smoke Particles on Clouds and Climate Forcing, *Science*, 277, 1636–1639, 715 <https://doi.org/10.1126/science.277.5332.1636>, 1997.
- Köhler, H.: The nucleus in and the growth of hygroscopic droplets, *Trans. Faraday Soc.*, 32, 1152–1161, <https://doi.org/10.1039/TF9363201152>, 1936.
- Koračin, D., Dorman, C. E., Lewis, J. M., Hudson, J. G., Wilcox, E. M., and Torregrosa, A.: Marine fog: a review, *Atmos. Res.*, 143, 142–175, <https://doi.org/10.1016/j.atmosres.2013.12.012>, 2014.



- 720 Koraćin, D., Lewis, J., Thompson, W. T., Dorman, C. E., and Businger, J. A.: Transition of stratus into fog along the California coast: observations and modeling, *J. Atmos. Sci.*, 58, 1714–1731, [https://doi.org/10.1175/1520-0469\(2001\)058<1714:TOSIFA>2.0.CO;2](https://doi.org/10.1175/1520-0469(2001)058<1714:TOSIFA>2.0.CO;2), 2001.
 Lee, Z., and Shang, S.: Visibility: How Applicable is the Century-Old Koschmieder Model?, *J. Atmos. Sci.*, 73, 4573–4581, <https://doi.org/10.1175/JAS-D-16-0102.1>, 2016.
- 725 Lewis, J. M., Koraćin, D., Rabin, R., and Businger, J.: Sea fog off the California coast: viewed in the context of transient weather systems, *J. Geophys. Res.-Atmos.*, 108, 4457, <https://doi.org/10.1029/2002JD002833>, 2003.
 Li, X., Wang, F., Zhang, X., Chan, P. W., Yang, Y., Lu, C., and Gao, M.: Anomalous increase in sea fog frequency along coastal China in the early 21st century and the aerosol influence, *npj Clim. Atmos. Sci.*, 8, 365, <https://doi.org/10.1038/s41612-025-01247-x>, 2025.
- 730 Liang, J. and Griswold, J. D. S.: Characteristics of Yellow Sea fog under the influence of Eastern China aerosol plumes, *Remote Sens.*, 16, 2262, <https://doi.org/10.3390/rs16132262>, 2024.
 Lin, P.-H., Wang, S.-H., Klemm, O., and Lin, N.-H.: Separating the Twomey effect and the semi-direct effect in absorbing aerosol environments through the cloud-aerosol mixing ratio, *Atmos. Chem. Phys.*, 26, 2083–2102, <https://doi.org/10.5194/acp-26-2083-2026>, 2026.
- 735 Lin, P.-H., Wang, S.-H., Klemm, O., and Lin, N.-H.: Separating the Twomey effect and the semi-direct effect in absorbing aerosol environments through the cloud-aerosol mixing ratio, *Atmos. Chem. Phys.*, 26, 2083–2102, <https://doi.org/10.5194/acp-26-2083-2026>, 2026.
 Liu, J.-W., Sun, Y., and Yang, L.: Interannual variability in summertime sea fog over the northern Yellow Sea and its association with the local sea surface temperature, *J. Geophys. Res.-Atmos.*, 126, e2020JD034439, <https://doi.org/10.1029/2020JD034439>, 2021.
- 740 Liu, Q., Shen, X., Sun, J., Zhang, Y., Qi, B., Ma, Q., Han, L., Xu, H., Hu, X., Lu, J., Liu, S., Yu, A., Liang, L., Gao, Q., Wang, H., Che, H., and Zhang, X.: Characterization of fog microphysics and their relationships with visibility at a mountain site in China, *Atmos. Chem. Phys.*, 25, 3253–3267, <https://doi.org/10.5194/acp-25-3253-2025>, 2025.
 Liu, Y., Daum, P. H., and McGraw, R. L.: Size truncation effect, threshold behavior, and a new type of autoconversion parameterization, *Geophys. Res. Lett.*, 32, L11811, <https://doi.org/10.1029/2005GL022636>, 2005.
- 745 Liu, Y., Daum, P. H., and McGraw, R. L.: Size truncation effect, threshold behavior, and a new type of autoconversion parameterization, *Geophys. Res. Lett.*, 32, L11811, <https://doi.org/10.1029/2005GL022636>, 2005.
 Liu, Y., Daum, P. H., and McGraw, R.: An analytical expression for predicting the critical radius in the autoconversion parameterization, *Geophys. Res. Lett.*, 31, L06121, <https://doi.org/10.1029/2003GL019117>, 2004.
- 750 Liu, Y., Daum, P. H., and McGraw, R.: An analytical expression for predicting the critical radius in the autoconversion parameterization, *Geophys. Res. Lett.*, 31, L06121, <https://doi.org/10.1029/2003GL019117>, 2004.



- Liu, Y., Daum, P. H., McGraw, R., and Miller, M.: Generalized threshold function accounting for effect of relative dispersion on threshold behavior of autoconversion process, *Geophys. Res. Lett.*, 33, L11804, <https://doi.org/10.1029/2005GL025500>, 2006.
- 755 Liu, Y., Daum, P. H., McGraw, R., and Miller, M.: Generalized threshold function accounting for effect of relative dispersion on threshold behavior of autoconversion process, *Geophys. Res. Lett.*, 33, L11804, <https://doi.org/10.1029/2005GL025500>, 2006.
- Lohmann, U. and Feichter, J.: Global indirect aerosol effects: a review, *Atmos. Chem. Phys.*, 5, 715–737, <https://doi.org/10.5194/acp-5-715-2005>, 2005.
- 760 Ma, X.: Aerosol-sea fog interaction data, Zenodo [Dataset], <https://doi.org/10.5281/zenodo.22940490>, 2026.
- Maalick, Z., Kühn, T., Korhonen, H., Kokkola, H., Laaksonen, A., and Romakkaniemi, S.: Effect of aerosol concentration and absorbing aerosol on the radiation fog life cycle, *Atmos. Environ.*, 133, 26–33, <https://doi.org/10.1016/j.atmosenv.2016.03.018>, 2016.
- Mazoyer, M., Burnet, F., and Denjean, C.: Experimental study on the evolution of droplet size distribution during the fog life cycle, *Atmos. Chem. Phys.*, 22, 11305–11321, <https://doi.org/10.5194/acp-22-11305-2022>, 2022.
- 765 Mazoyer, M., Burnet, F., Denjean, C., Roberts, G. C., Haeffelin, M., Dupont, J.-C., and Elias, T.: Experimental study of the aerosol impact on fog microphysics, *Atmos. Chem. Phys.*, 19, 4323–4344, <https://doi.org/10.5194/acp-19-4323-2019>, 2019.
- McComiskey, A., Feingold, G., Frisch, A. S., Turner, D. D., Miller, M. A., Chiu, J. C., Min, Q., and Ogren, J. A.: An assessment of aerosol-cloud interactions in marine stratus clouds based on surface remote sensing, *J. Geophys. Res.-Atmos.*, 114, D09203, <https://doi.org/10.1029/2008JD011006>, 2009.
- 770 McComiskey, A., Feingold, G., Frisch, A. S., Turner, D. D., Miller, M. A., Chiu, J. C., Min, Q., and Ogren, J. A.: An assessment of aerosol-cloud interactions in marine stratus clouds based on surface remote sensing, *J. Geophys. Res.-Atmos.*, 114, D09203, <https://doi.org/10.1029/2008JD011006>, 2009.
- McFiggans, G., Artaxo, P., Baltensperger, U., Coe, H., Facchini, M. C., Feingold, G., Fuzzi, S., Gysel, M., Laaksonen, A., Lohmann, U., Mentel, T. F., Murphy, D. M., O'Dowd, C. D., Snider, J. R., and Weingartner, E.: The effect of physical and chemical aerosol properties on warm cloud droplet activation, *Atmos. Chem. Phys.*, 6, 2593–2649, <https://doi.org/10.5194/acp-6-2593-2006>, 2006.
- Met Office: Handbook of Aviation Meteorology, 3rd edn., HMSO, London, ISBN 0-11-400365-3, 1994, <https://library.metoffice.gov.uk/Portal/Default/en-GB/RecordView/Index/209477>.
- 780 Niu, F., Li, Z., Li, C., Lee, K.-H., and Wang, M.: Increase of wintertime fog in China: potential impacts of weakening of the Eastern Asian monsoon circulation and increasing aerosol loading, *J. Geophys. Res.-Atmos.*, 115, D00K20, <https://doi.org/10.1029/2009JD013484>, 2010.
- Niu, S., Lu, C., Liu, Y., Zhao, L., Lü, J., and Yang, J.: Analysis of the microphysical structure of heavy fog using a droplet spectrometer: a case study, *Adv. Atmos. Sci.*, 27, 1259–1275, <https://doi.org/10.1007/s00376-010-8192-6>, 2010b.



- 785 Oliphant, A. J., Baguskas, S. A., and Fernandez, D. M.: Impacts of low cloud and fog on surface radiation fluxes for ecosystems in coastal California, *Theor. Appl. Climatol.*, 144, 239–252, <https://doi.org/10.1007/s00704-021-03518-y>, 2021.
- Petters, M. D. and Kreidenweis, S. M.: A single parameter representation of hygroscopic growth and cloud condensation nucleus activity, *Atmos. Chem. Phys.*, 7, 1961–1971, <https://doi.org/10.5194/acp-7-1961-2007>, 2007.
- Pilié, R. J., Mack, E. J., Rogers, C. W., Katz, U., and Kocmond, W. C.: The formation of marine fog and the development of fog-stratus systems along the California coast, *J. Appl. Meteorol.*, 18, 1275–1286, [https://doi.org/10.1175/1520-0450\(1979\)018<1275:TFOMFA>2.0.CO;2](https://doi.org/10.1175/1520-0450(1979)018<1275:TFOMFA>2.0.CO;2), 1979.
- 790 Poku, C., Ross, A. N., Hill, A. A., Blyth, A. M., and Shipway, B.: Is a more physical representation of aerosol activation needed for simulations of fog?, *Atmos. Chem. Phys.*, 21, 7271–7292, <https://doi.org/10.5194/acp-21-7271-2021>, 2021.
- Pope, N. H. and Igel, A. L.: Counteracting influences of gravitational settling modulate aerosol impacts on cloud-base-lowering fog characteristics, *Atmos. Chem. Phys.*, 25, 5433–5444, <https://doi.org/10.5194/acp-25-5433-2025>, 2025.
- 795 Quaas, J., Ming, Y., Menon, S., Takemura, T., Wang, M., Penner, J. E., Gettelman, A., Lohmann, U., Bellouin, N., Boucher, O., Sayer, A. M., Thomas, G. E., McComiskey, A., Feingold, G., Hoose, C., Kristjánsson, J. E., Liu, X., Balkanski, Y., Donner, L. J., Ginoux, P. A., Stier, P., Grandey, B., Feichter, J., Sednev, I., Bauer, S. E., Koch, D., Grainger, R. G., Kirkevåg, A., Iversen, T., Seland, Ø., Easter, R., Ghan, S. J., Rasch, P. J., Morrison, H., Lamarque, J.-F., Iacono, M. J., Kinne, S., and
- 800 Schulz, M.: Aerosol indirect effects – general circulation model intercomparison and evaluation with satellite data, *Atmos. Chem. Phys.*, 9, 8697–8717, <https://doi.org/10.5194/acp-9-8697-2009>, 2009.
- Quan, J., Liu, Y., Jia, X., Liu, L., Dou, Y., Xin, J., and Seinfeld, J. H.: Anthropogenic aerosols prolong fog lifetime in China, *Environ. Res. Lett.*, 16, 044048, <https://doi.org/10.1088/1748-9326/abef32>, 2021.
- Quan, J., Zhang, Q., He, H., Liu, J., Huang, M., and Jin, H.: Analysis of the formation of fog and haze in North China Plain (NCP), *Atmos. Chem. Phys.*, 11, 8205–8214, <https://doi.org/10.5194/acp-11-8205-2011>, 2011.
- 805 Ramanathan, V., Crutzen, P. J., Kiehl, J. T., and Rosenfeld, D.: Aerosols, climate, and the hydrological cycle, *Science*, 294, 2119–2124, <https://doi.org/10.1126/science.1064034>, 2001.
- Reutter, P., Su, H., Trentmann, J., Simmel, M., Rose, D., Gunthe, S. S., Wernli, H., Andreae, M. O., and Pöschl, U.: Aerosol- and updraft-limited regimes of cloud droplet formation: influence of particle number, size and hygroscopicity on the activation of cloud condensation nuclei (CCN), *Atmos. Chem. Phys.*, 9, 7067–7080, <https://doi.org/10.5194/acp-9-7067-2009>, 2009.
- 810 Rosenfeld, D., Andreae, M. O., Asmi, A., Chin, M., de Leeuw, G., Donovan, D. P., Kahn, R., Kinne, S., Kivekäs, N., Kulmala, M., Lau, W., Schmidt, K. S., Suni, T., Wagner, T., Wild, M., and Quaas, J.: Global observations of aerosol-cloud-precipitation-climate interactions, *Rev. Geophys.*, 52, 750–808, <https://doi.org/10.1002/2013RG000441>, 2014.
- 815 Rosenfeld, D., Lohmann, U., Raga, G. B., O'Dowd, C. D., Kulmala, M., Fuzzi, S., Reissell, A., and Andreae, M. O.: Flood or drought: how do aerosols affect precipitation?, *Science*, 321, 1309–1313, <https://doi.org/10.1126/science.1160606>, 2008.
- Rosenfeld, D., Zhu, Y., Wang, M., Zheng, Y., Goren, T., and Yu, S.: Aerosol-driven droplet concentrations dominate coverage and water of oceanic low-level clouds, *Science*, 363, eaav0566, <https://doi.org/10.1126/science.aav0566>, 2019.



- Saide, P. E., Thapa, L. H., Ye, X., Pagonis, D., Campuzano-Jost, P., Guo, H., Schuneman, M. L., Jimenez, J. L., Moore, R.,
820 Wiggins, E., Winstead, E., Robinson, C., Thornhill, L., Sanchez, K., Wagner, N. L., Ahern, A., Katich, J. M., Perring, A. E.,
Schwarz, J. P., Lyu, M., Holmes, C. D., Hair, J. W., Fenn, M. A., and Shingler, T. J.: Understanding the Evolution of Smoke
Mass Extinction Efficiency Using Field Campaign Measurements, *Geophys. Res. Lett.*, 49(18), e2022GL099175,
<https://doi.org/10.1029/2022GL099175>, 2022.
- Seinfeld, J. H., Bretherton, C., Carslaw, K. S., Coe, H., DeMott, P. J., Dunlea, E. J., Feingold, G., Ghan, S., Guenther, A. B.,
825 Kahn, R., Kraucunas, I., Kreidenweis, S. M., Molina, M. J., Nenes, A., Penner, J. E., Prather, K. A., Ramanathan, V.,
Ramaswamy, V., Rasch, P. J., Ravishankara, A. R., Rosenfeld, D., Stephens, G., and Wood, R.: Improving our fundamental
understanding of the role of aerosol–cloud interactions in the climate system, *Proc. Natl. Acad. Sci. USA*, 113, 5781–5790,
<https://doi.org/10.1073/pnas.1514043113>, 2016.
- Stolaki, S., Haeffelin, M., Lac, C., Dupont, J.-C., Elias, T., and Masson, V.: Influence of aerosols on the life cycle of a
830 radiation fog event. A numerical and observational study, *Atmos. Res.*, 151, 146–161,
<https://doi.org/10.1016/j.atmosres.2014.04.013>, 2015.
- Tardif, R. and Rasmussen, R. M.: Event-based climatology and typology of fog in the New York City region, *J. Appl.
Meteorol. Climatol.*, 46, 1141–1168, <https://doi.org/10.1175/JAM2516.1>, 2007.
- Twomey, S.: The influence of pollution on the shortwave albedo of clouds, *J. Atmos. Sci.*, 34, 1149–1152,
835 [https://doi.org/10.1175/1520-0469\(1977\)034<1149:TIOPOT>2.0.CO;2](https://doi.org/10.1175/1520-0469(1977)034<1149:TIOPOT>2.0.CO;2), 1977.
- Twomey, S.: The nuclei of natural cloud formation part II: the supersaturation in natural clouds and the variation of cloud
droplet concentration, *Geofis. Pura Appl.*, 43, 243–249, <https://doi.org/10.1007/BF01993560>, 1959.
- Wainwright, C., Chang, R. Y.-W., and Richter, D.: Aerosol activation in radiation fog at the Atmospheric Radiation Program
Southern Great Plains site, *J. Geophys. Res.-Atmos.*, 126, e2021JD035358, <https://doi.org/10.1029/2021JD035358>, 2021.
- 840 WMO/GAW: Aerosol Measurement Procedures, Guidelines and Recommendations, WMO TD No. 1178, GAW Report No.
153, World Meteorological Organization, Geneva, Switzerland,
https://gml.noaa.gov/aero/docs/GAW_153_Aerosol_Measurement_Procedures.pdf, 2003.
- Yan, S., Zhu, B., and Kang, H.: Long-term fog variation and its impact factors over polluted regions of East China, *J.
Geophys. Res.-Atmos.*, 124, 1741–1754, <https://doi.org/10.1029/2018JD029389>, 2019.
- 845 Yan, S., Zhu, B., Huang, Y., Zhu, J., Kang, H., Lu, C., and Zhu, T.: To what extents do urbanization and air pollution affect
fog?, *Atmos. Chem. Phys.*, 20, 5559–5572, <https://doi.org/10.5194/acp-20-5559-2020>, 2020.
- Yan, S., Zhu, B., Zhu, T., Shi, C., Liu, D., Kang, H., Lu, W., and Lu, C.: The effect of aerosols on fog lifetime: observational
evidence and model simulations, *Geophys. Res. Lett.*, 48, e2020GL091156, <https://doi.org/10.1029/2020GL091156>, 2021.
- Yang, L., Ding, S., Liu, J.-W., and Zhang, S.-P.: Effects of radiative cooling on advection fog over the northwest Pacific
850 Ocean: observations and large-eddy simulations, *Atmos. Chem. Phys.*, 24, 6809–6824, [https://doi.org/10.5194/acp-24-6809-](https://doi.org/10.5194/acp-24-6809-2024)
2024, 2024.



Yang, L., Liu, J.-W., Ren, Z.-P., Xie, S.-P., Zhang, S.-P., and Gao, S.-H.: Atmospheric conditions for advection-radiation fog over the western Yellow Sea, *J. Geophys. Res.-Atmos.*, 123, 5455–5468, <https://doi.org/10.1029/2017JD028088>, 2018.

855 Zhang, S.-P., Xie, S.-P., Liu, Q.-Y., Yang, Y.-Q., Wang, X.-G., and Ren, Z.-P.: Seasonal variations of Yellow Sea fog: observations and mechanisms, *J. Climate*, 22, 6758–6772, <https://doi.org/10.1175/2009JCLI2806.1>, 2009.

Zhou, S., Xu, Z., Chen, Y., Zhao, M., Li, Y., and Yan, K.: Anthropogenic air pollutants strongly interact with natural aerosols over the eastern China seas: key processes, size distributions, and seasonalities, *Atmos. Chem. Phys.*, 25, 13597–13619, <https://doi.org/10.5194/acp-25-13597-2025>, 2025.