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3 Numerical Case Study of the Aerosol-Cloud-Interactions in Warm
4 Boundary Layer Clouds over the Eastern North Atlantic with an
5 Interactive Chemistry Module
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Abstract

The presence of warm boundary layer stratiform clouds over the Eastern North Atlantic (ENA) region is commonly influenced by the Azores High, especially during the summer season. To investigate comprehensive aerosol-cloud interactions, this study employs the Weather Research and Forecast model coupled with a chemistry component (WRF-Chem), incorporating aerosol chemical components that are relevant with formation of cloud condensation nuclei (CCN) and accounting for aerosol spatiotemporal variation. This study focuses on aerosol indirect effects, particularly long-range transport aerosols, in the ENA region under three different weather regimes: ridge with surface high-pressure system, post-trough with surface high-pressure system, and weak trough. The WRF-Chem simulations conducted at a near **large-eddy** scale offer valuable insights into the model's performance, especially regarding its high spatial resolution in **capturing** the **mesoscale** cloud **features** across various weather regimes. Our result shows that introducing five times more aerosols to either non-precipitating or precipitating clouds significantly increases ambient CCN numbers, resulting in varying degrees of higher LWP. The substantial aerosol-cloud interaction especially occurs in the precipitating clouds and demonstrates the LWP susceptibility to changes in CCN under different regimes. Conversely, **thin**, non-rain clouds at the edges of a cloud system are prone to evaporation, exhibiting an aerosol drying effect. The aerosols released during this process transition back to the accumulation mode, facilitating future activation. This dynamic behavior is not adequately represented in prescribed-aerosol simulations.

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68 1 Introduction

69 Low-level stratiform clouds are predominantly generated over oceanic regions and are
70 categorized into three main types: warm boundary layer stratiform clouds located on the eastern
71 side of oceanic subtropical highs, stratocumulus clouds that develop over warm western boundary
72 currents during winter cold outbreaks, and Arctic stratus (Klein and Hartmann, 1993). Warm
73 boundary layer stratocumulus clouds, on average, blanket around 20% of the Earth's surface
74 annually (Wood, 2012; Warren et al., 1988). Their influence on the Earth's energy balance is
75 substantial, primarily through their ability to reflect incoming solar radiation, resulting in
76 significant shortwave cloud radiative effects leading to a pronounced negative net radiative effect
77 (Chen et al., 2000; Stephens and Greenwald, 1991; Hartmann et al., 1992).

78 Research on aerosol-cloud interactions in warm boundary layer clouds has been ongoing
79 since the 1970s. Twomey (1974) proposed that aerosols play an important role in influencing the
80 Earth's energy budget by serving as cloud condensation nuclei (CCN). These CCN are crucial for
81 cloud formation. A higher concentration of CCN results in the formation of clouds with a greater
82 number of smaller-sized cloud droplets (Twomey, 1991). These smaller droplets enhance cloud
83 albedo, known as the first indirect effect, and inhibit precipitation formation while prolonging
84 cloud lifetime, known as the second indirect effect (Albrecht, 1989). In addition to these indirect
85 effects, aerosol particles have direct, semi-direct, and indirect impacts on the atmosphere's energy
86 budgets and surface, leading to changes in atmospheric stability (Lee et al., 2008). Until now, our
87 understanding of aerosol-cloud interactions remains incomplete. In a recent review paper,
88 Feingold et al. (2024) highlighted that the response of cloud amount (including liquid water
89 content, spatial coverage, and cloud persistence) to aerosol perturbations is still unclear. Both
90 positive and negative adjustments in liquid water path (LWP) and cloud fraction (CF) have been

91 observed. Increases in cloud amount (positive adjustments) are linked to rain suppression, whereas
92 enhanced evaporation of smaller droplets and entrainment feedback tend to decrease cloud amount
93 (negative adjustments).

94 This study focuses on warm boundary layer stratiform clouds located on the eastern side
95 of oceanic subtropical highs, specifically targeting the area over the Eastern North Atlantic (ENA)
96 region, where the U.S. Department of Energy (DOE) Atmospheric Radiation Measurement (ARM)
97 program developed a ground-based user facility in the Azores archipelago (Mather and Voyles,
98 2013). Long-term ground-based observations at the ARM ENA site, aircraft field campaigns near
99 the Azores islands, and satellite retrievals over the ENA region provide comprehensive datasets
100 for observational studies on aerosol-cloud interactions (Zheng et al., 2022; Zheng et al., 2023;
101 Ghate et al., 2023; Qiu et al., 2024).

102 The presence of stratocumulus clouds over the ENA region is commonly influenced by the
103 Azores High, also known as the Bermuda-Azores High (Rémillard and Tselioudis, 2015). This
104 semi-permanent high-pressure system typically develops over the subtropical North Atlantic
105 Ocean. The Azores High often brings stable and relatively dry conditions to the region, which can
106 contribute to the formation and maintenance of stratocumulus clouds. During the summer season,
107 the Azores High tends to strengthen and expand, leading to more persistent high-pressure
108 conditions and often warmer, drier weather in its vicinity. Although synoptic intrusions from high
109 latitudes are less frequent in the summer compared to the winter season (Wood et al., 2015), the
110 ENA region still experiences synoptic variability from weak troughs during the summer months
111 (Mechem et al., 2018; Zheng et al., 2024).

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113 Leveraging the marine boundary layer cloud observations from the ARM ENA
114 observatory, this study aims to study aerosol indirect effects (AIE), especially long-range transport
115 aerosols, in the warm boundary layer clouds over the ENA region under three different synoptic
116 regimes: ridge with surface high-pressure system, post-trough with surface high-pressure system,
117 and weak trough (Mechem et al., 2018; Zheng et al., 2024). These regimes are chosen because the
118 ARM site experiences northerly wind conditions, during the passage of troughs. This minimizes
119 the influence of the island effect on the observations (Ghate and Cadetdu, 2019; Zheng and Miller,
120 2022).

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121 Only a few numerical studies examined aerosol-cloud interactions in marine boundary
122 layer clouds over this region (Zhang et al., 2021; Wang et al., 2020; Kazemirad and Miller, 2020;
123 Christensen et al., 2024). Wang et al. (2020), for example, used the Weather Research and Forecast
124 (WRF) model with prescribed CCN profiles to simulate perturbed long-range transport aerosol
125 concentration for two different cases of marine boundary layer (MBL) clouds. They concluded
126 that when long-range transport aerosol plumes penetrate down into the drizzling cloud deck, the
127 simulations show an increase in marine cloud fractions with larger water content, supporting a
128 positive cloud amount adjustment to CCN perturbations. Christensen et al. (2024) utilized an
129 advanced WRF configuration integrated with a Lagrangian framework to assess the effects of
130 aerosols on developing cloud fields across 10 case study days during the ENA field campaign and
131 got the same conclusion. However, a limitation of these studies is that they do not account for
132 aerosol composition acting as CCN or the changes in aerosol populations following the cloud
133 evaporation process, even though aerosol wet removal is included in their simulations.

134 To further investigate the impacts of realistic aerosol chemical components and aerosol
135 spatiotemporal variation on the AIE, this study adopts the WRF model coupled with a chemistry

138 component (WRF-Chem) to examine the AIE in the ENA region across different synoptic regimes.
139 A brief description of observational data and the WRF-Chem model, as well as the configuration
140 and numerical experiments, are given in Section 2. Simulated results are discussed in Section 3,
141 including model evaluation, model sensitivity tests, and cloud susceptibilities. The discussion and
142 summary are provided in Section 4.

143 2 Methodology

144 2.1 Observational data

145 2.1.1 MERRA-2

146 The Modern-Era Retrospective analysis for Research and Applications, Version 2
147 (MERRA-2), represents the latest advancement in global atmospheric reanalysis during the
148 satellite era. Produced by NASA's Global Modeling and Assimilation Office (GMAO), it utilizes
149 the Goddard Earth Observing System Model (GEOS) version 5.12.4 (Molod et al., 2015). The
150 aerosol species are from the dataset, inst3_3d_aer_Nv, which is an instantaneous 3-dimensional 3-
151 hourly data collection in MERRA-2 (Modeling and Office, 2015). The dataset comprises
152 assimilations of aerosol mixing ratio parameters at a native resolution of 0.5° latitude x 0.625°
153 longitude across 72 model layers, encompassing dust, sea salt, sulfur dioxide (SO_2), sulfate (SO_4),
154 black carbon (BC), and organic carbon (OC). The data is provided every three hours, beginning
155 at 00:00 UTC. Based on Wang et al. (2020), we also adopt MERRA-2 to drive the WRF-Chem
156 initial and boundary conditions for this study (see [Section 2.2.2](#) for details).

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157 2.1.2 Geostationary satellite retrievals (Meteosat)

158 Cloud properties are derived from the Spinning Enhanced Visible and Infrared Imager
159 (SEVIRI) on Meteosat-10 and Meteosat-11, which offer a spatial resolution of 3 km at nadir and

161 a half-hourly temporal resolution over the ENA region. These SEVIRI cloud products are
162 generated using the Satellite CLOud and Radiation Property retrieval System (SatCORPS)
163 algorithms (Painemal et al., 2021). These methods, developed by the Clouds and Earth's Radiant
164 Energy System (CERES) project, are specifically tailored to support ARM ground-based
165 observation sites (Minnis et al., 2011; Minnis et al., 2021). Specifically, this study adopts cloud
166 fraction for all clouds as the observational reference over the ENA region. The adopted data have
167 been specifically processed (e.g., solar zenith angle, cloud optical thickness, and cloud labels) and
168 averaged to $25 \text{ km} \times 25 \text{ km}$ (Qiu et al., 2024).

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169 2.1.3 Aircraft observation

170 The U.S. DOE ARM Aerosol and Cloud Experiments in the Eastern North Atlantic (ACE-
171 ENA) aircraft field campaign near the Azores islands provided extensive observations of the
172 vertical distributions of aerosol and cloud properties (Wang et al., 2022). Intensive operational
173 periods (IOPs) of the ACE-ENA took place in late June and July 2017, as well as January to
174 February 2018. During the 2017 summer IOP, the ARM Aerial Facility's (AAF) Gulfstream-159
175 (G-1) aircraft delivered precise measurements of aerosol size distribution, total aerosol number
176 concentration, and chemical constituents both below and above cloud layers. SO_4 and OC mass
177 concentrations were measured using the Aerodyne High-Resolution Time-of-Flight Aerosol Mass
178 Spectrometer (HR-ToF-AMS), while refractory BC was measured by the Single Particle Soot
179 Photometer (SP2). Detailed information about each instrument is available on the ARM website
180 (<https://www.arm.gov/research/campaigns/aaf2017ace-ena>). In this study, aircraft measurements
181 of SO_4 , OC, and BC from 19 July 2017, are utilized to assess the simulated aerosol vertical profile.
182 However, uncertainties arising from the measurements and spatiotemporal sampling strategies
183 may hinder direct comparisons of absolute values between the observations and modeled results.

185 2.1.4 ARM ground-based observations

186 The DOE ARM ground-based instruments deployed on Graciosa Island in the Azores
187 archipelago provide comprehensive measurement of aerosols, clouds, radiation, atmospheric
188 boundary layer, and other atmospheric properties. In this study, LWP is retrieved from the
189 brightness temperature measured by the microwave radiometer (MWR) at 23.8 and 31.4 GHz
190 (Liljegren et al., 2001) and used for model evaluation. The temperature and moisture profiles are
191 from the interpolated sonde data, derived from the radiosonde measurement.

192 2.2 The model

193 2.2.1 WRF-Chem

194 The Weather Research and Forecasting (WRF) model version 4.4.2 (Skamarock et al.,
195 2021) coupled with a chemistry component (WRF-Chem) (Grell et al., 2005) is used in this study.
196 The standard WRF-Chem permits the simulation of the combined direct, indirect, and semi-direct
197 effects of aerosols (Grell et al., 2005; Fast et al., 2006; Chapman et al., 2009). WRF-Chem version
198 4.4.2 has sophisticated packages to represent chemistry processes (i.e., gas-phase reaction, gas-to-
199 particle conversion, coagulation, etc.) and aerosol size and composition (Binkowski and Shankar,
200 1995). In this study, the Regional Acid Deposition Model version 2 (RADM2) photochemical
201 mechanism (Stockwell et al., 1997) is integrated alongside the Modal Aerosol Dynamics Model
202 for Europe (MADE) and the Secondary Organic Aerosol Model (SORGAM) (Ackermann et al.,
203 1998; Schell et al., 2001) to simulate atmospheric chemistry and the evolution of anthropogenic
204 aerosols. MADE/SORGAM adopts a modal approach to represent the aerosol size distribution,
205 predicting mass and number concentrations across three aerosol modes (Aiken, accumulation, and
206 coarse). MADE/SORGAM has inorganic, organic, and secondary organic aerosols and contain
207 aerosol formation processes including nucleation, condensation, and coagulation. WRF-Chem

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210 tracks the number of particles and the mass of chemical compounds (e.g., SO_4^{2-} , NH_4^+ , NO_3^- , Na^+ ,
211 Cl^- etc.) in each aerosol mode, including both interstitial aerosols and aerosols present in liquid
212 water (the sum of cloud and rain), as prognostic variables.

213 The size, composition, and mixing state of aerosols significantly influence their capability
214 to activate as CCN (Zaveri et al., 2010). A physically based aerosol activation parameterization
215 scheme has been developed for climate models to simulate CCN concentration accurately and
216 efficiently (Abdul-Razzak and Ghan, 2000). This aerosol activation parameterization was initially
217 designed for a single aerosol type with a lognormal size distribution. Then, they expanded this
218 parameterization to accommodate multiple externally mixed lognormal modes, with each mode
219 consisting of both soluble and insoluble materials internally mixed. However, WRF-Chem
220 (MADE/SORGAM) chemistry package adopts this global internal mixing assumption, where all
221 particles within a log-normal mode within the same grid cell are instantly combined, resulting in
222 the same chemical composition. This instantaneous internal mixing assumption modifies the
223 optical and chemical characteristics of particles in WRF-Chem simulations, potentially impacting
224 aerosol-cloud interactions, such as aerosol activation as CCN (Zhang et al., 2014).

225 2.2.2 The configuration

226 Our focus in this study is to examine aerosol-cloud interactions close to the scale of large-
227 eddy simulation (LES) over the ARM ENA site. We use WRF-Chem with a full chemistry
228 package involving sophisticated gaseous and aqueous chemical processing calculations and dry
229 and wet depositions. The numerical simulations are employed with 4 domains consisting of 4
230 horizontal resolutions of 5 km, 1.67 km, 0.56 km, and 0.19 km, respectively (Fig. 1), with one-
231 way nesting. There are 550×530 grids for D1, 451×430 grids for D2, 553×532 grids for D3,
232 and 553×532 grids for D4. The domain size of domain 4 is about 1 degree which is similar to

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the spacing resolution of global climate models. Seventy-five vertically staggered layers are stretched to have a higher resolution near the surface based on a terrain-following pressure coordinate system. With this setup, the model has roughly 24 model layers in the boundary layer (~2000 m). The time step is 30 and 10 seconds for advection and physics calculation for domains 1 and 2, respectively. The nesting inner domains 3 and 4 have the time step of 3 seconds and 1 second, respectively. The physics schemes adopted in the simulations are listed in Table 1. The initial and boundary meteorological conditions are taken from ERA5, developed by the Copernicus Climate Change Service (C3S) at ECMWF (European Centre for Medium-Range Weather), stands as the fifth generation of ECMWF atmospheric reanalysis, spanning from January 1940 to the present day (Hersbach et al., 2023). This comprehensive dataset offers hourly estimates of numerous atmospheric, land, and oceanic climate variables, covering the entirety of Earth on a 31 km grid. The atmospheric component is resolved using 137 levels, spanning from the surface up to 80 km in height.

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The computational expense of conducting a 4-domain WRF-Chem simulation, particularly with LES resolution, is exceedingly high. To mitigate this, we execute WRF solely for the two outer domains (d01 and d02), leveraging the WRF downscaling module (ndown) (Skamarock et al., 2008) to generate meteorological initial and boundary conditions for domain 3. As a result, we only need to perform WRF-Chem simulations for the two inner domains (d03 and d04), leading to an almost 50% reduction in total computational costs (compared to the original 4-domain run, which had a throughput of 4 hours per day using 1080 cores). It is important to note that a high temporal frequency for domain 3 boundary conditions is essential due to its fine horizontal resolution (0.56 km). In this context, we update the boundary condition every 5 minutes for domain 3.

259 To enhance the realism of aerosol mass simulation in remote marine regions, such as the
260 ENA site, we account for major aerosol species (BC, OC, and SO₄), as well as SO₂, from MERRA-
261 2 into the boundary conditions of domain 3. Aerosols in the initial condition are introduced into
262 the restart file (wrfirst) following a one-hour initial run, rather than in the initial condition file
263 (wrfinput), to address certain numerical challenges. According to the emission setup for
264 MADE/SORGAM, we assume that the Aiken mode and the accumulation mode account for 20%
265 and 80% of the aerosol mass (BC and OC), respectively (Tuccella et al., 2012). Conversely, for
266 SO₄, 80% is allocated to the Aiken mode and 20% to the accumulation mode, reflecting the faster
267 growth rate of SO₄ and a longer duration of growth from the domain 3 boundary. Because
268 MERRA-2 only provides aerosol mass, the aerosol number concentrations for different aerosol
269 species are estimated with the density assumption of BC (1.7 g cm⁻³), OC (1.0 g cm⁻³) and SO₄
270 (1.77 g cm⁻³) based on Liu et al. (2012). Aerosol optical depth retrieved from satellite remote
271 sensing can offer valuable information for comparison; however, it may be subject to high bias in
272 our study cases due to cloud cover.

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273 It is common to consider that the ENA region is an unpolluted area because it is far away
274 from the anthropogenic pollution sources. Besides long-range transport aerosols, two local aerosol
275 sources, dimethyl sulfide (DMS) and sea salts, are also important for the aerosol budget. Kazil et
276 al. (2011) pointed that the observed DMS flux from the ocean in the VOCALS-REx field campaign
277 over the Southeast Pacific can support a nucleation source of aerosol. DMS oxidation by nitrate
278 (NO₃) produces SO₂ and then increases SO₄ concentration (Toon et al., 1987). Since we adapted
279 SO₂ and SO₄ concentration from MERRA-2 in the initial and boundary conditions, we did not
280 double count DMS emissions in our simulations. As a result, chemical species emissions, except
281 for sea salt, are excluded from the simulations. The emission of sea salt particles is parameterized

283 using the method outlined by Clarke et al. (2006) in WRF-Chem. ~~Sea salt emissions are driven by~~
 284 ~~surface wind speed. The simulated surface wind speed aligns closely with ERA-5 data; however,~~
 285 ~~the sea salt concentration is only one-third of the value found in the MERRA-2 analysis. To~~
 286 ~~improve the alignment with the sea salt aerosol concentration observed in the MERRA-2~~
 287 ~~reanalysis, we~~ adjust the parameter factor for sea salt ~~emissions~~ to three times the original estimate
 288 ~~(further comparison can be found in Section 3).~~

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289 2.3 Study cases and numerical experiment design

290 We select three specific study cases to assess the impact of long-range transport aerosols on
 291 warm boundary layer clouds, with each case representing a typical meteorological regime observed
 292 over the ENA site. The first case, dated 1 July 2016, exhibits the formation of overcast
 293 stratocumulus clouds (Fig. 2a) within a meteorological regime characterized by a ridge system in
 294 the free troposphere and a high-pressure system near the surface (Fig. 2d). Predominant
 295 northwesterly and northerly winds in the area of the ARM ENA site coincide with the presence of
 296 long-range transport aerosols, commonly found along the periphery of the high surface pressure
 297 system (Logan et al., 2014; Gallo et al., 2023).

298 The second case on 19 July 2017 is a stratocumulus cloud case (Fig. 2b) within a post-
 299 trough regime featuring a high surface pressure under the influence of a trough system (Fig. 2e).
 300 Following the trough passage, robust northwesterly winds facilitated the influx of long-range
 301 transport aerosols into the region, which then shifted to northerly winds as the trough moved away.
 302 Because the ACE-ENA aircraft field campaign ~~intensive operations period (IOP) was during this~~
 303 time, ~~aircraft~~ aerosol observational data can be used to evaluate the model performance for this
 304 case.

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313 Finally, the third case, dated 23 August 2019, occurred during a period of weak trough
314 activity (Fig. 2f). Here, we noted the presence of broken, thicker stratocumulus clouds, often
315 accompanied by deeper cloud formations (Fig. 2c). Long-range transport aerosols were again
316 observed, primarily carried by northwesterly and northerly winds, albeit with weaker surface wind
317 speeds compared to the preceding two cases.

318 All simulations start at 12 UTC on the preceding day of the study case, spanning a duration
319 of 36 hours, with the initial 12 hours dedicated to spin-up. Again, aerosols in the initial condition
320 are introduced into the restart file after one-hour initial run (i.e., 13 UTC). The three
321 aforementioned cases, labeled as control cases (20160701_control, 20170719_control, and
322 20190823_control), are utilized to examine the behavior of warm boundary layer clouds under
323 diverse meteorological conditions. Additionally, we formulated three perturbed cases
324 (20160701_perturbed, 20170719_perturbed, and 20190823_perturbed) by amplifying aerosol
325 concentrations in both initial and boundary conditions, as well as sea salt emissions, by a factor of
326 five relative to each control case. These control cases represent clean conditions, with near-surface
327 CCN concentrations below 100 cm^{-3} at the ARM ENA site. A comparison between the control
328 and perturbed cases elucidates the sensitivity of warm boundary layer clouds to aerosol
329 enhancements under varying meteorological conditions, thereby contributing to a deeper
330 understanding of cloud microphysics processes under varying atmospheric dynamics.

331 3 Results

332 3.1 Model evaluation

333 3.1.1 Meteorological conditions

334 Figures 2g, 2h, and 2i display the model-simulated LWP in the control runs over domain 3
335 and 4. The modeled LWP is calculated in-cloud LWP only. The simulations with fine spatial

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339 resolution effectively capture synoptic frontal systems and cloud features, particularly when
 340 compared to the cloud images from the Meteosat satellite (Figs. 2a, 2b, and 2c). Thin, uniform
 341 stratocumulus clouds on 1 July 2016 are simulated in 20160701_control, while the solid
 342 stratocumulus and frontal system on 19 July 2017 are also well captured in 20170719_control.
 343 Broken stratocumulus clouds on 23 August 2019 are reproduced in the simulation of
 344 20190823_control. Figures 2j, 2k, and 2l display the model-simulated LWP in the perturbed runs.
 345 The overcast stratocumulus clouds are simulated better in 20160701_perturbed (Fig. 2j), which
 346 more accurately reflects reality.

347 The control runs serve as a basis for comparing the boundary layer structure against the
 348 interpolated soundings obtained from the ARM ENA site. Figure 3 depicts the comparison,
 349 showing the simulated air temperature aligning closely with the observed values. However, on 1
 350 July 2016, the model (20160701_control) displays a warm bias in capturing the temperature
 351 inversion (Figs. 3a and 3b), with the simulated inversion layer situated approximately 200-300 m
 352 lower than observed, during the whole study period. Potential temperature and relative humidity
 353 have consistent performance as illustrated in Figs. S1b and S2b, respectively. While the model
 354 indicates high relative humidity (> 90%) within 1000 m, observations show this extending up to
 355 ~1200 m.

356 For the 19 July 2017 case, the model (20170719_control) successfully represents the
 357 diurnal cycle of temperature vertical gradient within 1000 m height. However, compared to
 358 observations, the model does not catch the inversion at 1500 m height near the noon time and
 359 shows a warm bias in the model's simulated temperature (Figs. 3c and 3d). The model simulation
 360 also tends to depict drier conditions in the evening compared to the observation (Figs. S3c and
 361 S3d).

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372 On 23 August 2019, characterized by a weak trough regime and higher boundary layer
 373 height, the simulation of 20190823_control accurately captures warm and moist air advection in
 374 the morning but ~~has difficulty in maintaining~~ fidelity in the late afternoon. Notably, the lower
 375 troposphere becomes excessively warm and dry after ~~17 Z local time~~ compared to observations
 376 (Figs. 3e, 3f, ~~S2e, S2f, S3e, and S3f~~).

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377 In general, all simulations effectively capture large-scale conditions and cloud features
 378 (Fig. 2) across different synoptic regimes but do not accurately represent temperature inversions
 379 and air advection patterns. Discrepancies are noted in the simulated boundary layer height, which
 380 is lower, and the inversion is weaker than actual observations. Furthermore, the discrepancies tend
 381 to increase in the later stage of simulation.

382 3.1.2 Aerosol evolution

383 As mentioned in Section 2.2.2, we incorporate major aerosol species (BC, OC, and SO₄),
 384 from MERRA-2 into the domain 3 initial (in the restart file at 13 UTC) and boundary conditions
 385 to enhance the realism of aerosol simulation. Figure 4 shows time-series SO₄ vertical profiles
 386 from both MERRA-2 and WRF-Chem for three study cases. Here, we demonstrate the time
 387 evolution of SO₄. ~~This species~~ is the main aerosol component among the three introduced aerosol
 388 species, about 60~80% of total aerosol mass, in the initial conditions.

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389 Compared to the MERRA-2 data, 20160701_control well captures the long-range transport
 390 SO₄ between 1000 m and 2000 m, which is above the cloud deck, on 1 July 2016 (Figs. 4a and 4b).
 391 The observed high BC and OC are also concentrated in this layer (Figs. ~~S4a and S5a~~), as well as
 392 simulated ones (Figs. ~~S4b and S5b~~). ~~Figures~~ 4c and 4e show two MERRA-2 time-series vertical
 393 distributions of SO₄ on 19 July 2017 and 23 August 2019, both showing low-altitude (below 1500
 394 m) aerosol plumes. On 19 July 2017, the concentrations of BC and OC showed two peaks – one

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404 near the surface and another above 1500 m in the free troposphere (Figs. [S4c](#) and [S5c](#)). This
 405 pattern indicates the presence of a biomass-burning signature in the plume on that day (Wang et
 406 al., 2020). While the simulation of 20170719_control did not capture the near surface BC, OC,
 407 and SO₄ concentration after [11 Z local time](#) on 19 July 2017 (Figs. [S4d](#), [S5d](#), and 4d). It is because
 408 in the case of the post-tough regime, the wind direction changes from northwesterly to northerly
 409 when the trough moved away, [and](#) the aerosol plume in domain 3 did not propagate into domain 4
 410 when the wind direction [changed](#) (figure not shown). However, the simulation of
 411 20170719_control still captures the BC and OC plumes in the free troposphere (above 2000 m
 412 height) (Figs. [S4d](#) and [S5d](#)).

413 Aircraft observations during [ACE-ENA](#) provide more accurate depictions of aerosol
 414 vertical distribution and aerosol layer heights, with differentiation of aerosol type. Figure 5a shows
 415 the vertical distribution of aerosol mass concentrations averaged over the flights on 19 July 2017.
 416 BC, OC and SO₄ all increase with height above clouds (~1000 m), indicating downward
 417 propagation of aerosol plumes and possible interaction with MBL clouds (600 – 1000 m). Here,
 418 we also see that high SO₄ in the free troposphere, [similar to](#) MERRA-2, but the model
 419 underestimates the OC concentration in the free troposphere. On the other hand, within the MBL,
 420 there is a much higher concentration of SO₄ in the MBL than those of BC and OC in the
 421 observations. This phenomenon is also captured by the WRF-Chem simulation (Fig. 5b), but the
 422 model did not capture the magnitude of [the](#) SO₄ concentration.

423 Similarly, for the case of [20190823](#), within the low boundary layer, there is a much higher
 424 concentration of SO₄ in the low boundary layer (Fig. 4e). After [the noon time](#) on 23 August 2019,
 425 BC and OC show both high-altitude plumes and low-altitude plumes approaching into the domain,
 426 which indicate potentially two different aerosol sources (Figs. [S4e](#) and [S5e](#)). Again, while the

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443 simulation of 20190823_control well captures the time evolution of aerosol plume, the boundary
 444 of high-altitude plumes and low-altitude plumes appears 300 m lower in the simulations (~600 m
 445 in altitude; Figs. S4f and S5f) compared to the observations (~900 m in altitude).

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446 Sea salt particles serve as an important source of CCN over the ocean, particularly in
 447 unpolluted conditions. However, due to their larger particle size, sea salt particles tend to
 448 accumulate near the ocean surface and are swiftly removed by dry deposition and sedimentation
 449 processes (Chin et al., 2002). The simulation of 20160701_control (Figs. S6a and S6b) accurately
 450 reproduces sea salt concentrations, both in magnitude and vertical distribution, consistent with
 451 observations, same as the case of 20170719 (Figs. S6c and S6d). Nevertheless, the model
 452 encounters difficulties in simulating sea salt concentrations for the case of 23 August 2019 (Figs.

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453 S5e and S5f), corresponding to a weak-trough system (Fig. 2c). Although the simulated surface
 454 wind speed matches well with ERA-5 (Fig. S7), the underestimation of sea salt concentrations may
 455 be attributed to limitations in the emission parameterization, which is overly reliant on surface
 456 wind speed (Gong, 2003).

Deleted: As discussed in Section 2.2, we adjusted the parameter factor to three times its original value to better align with the MERRA-2 dataset.

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457 3.1.3 Cloud properties

458 In Fig. 6, we observe a comparison between the simulated results and observations of LWP
 459 and CF at different spatial scales (4 km- and domain-average, respectively) to leverage the
 460 spatiotemporal advantages offered by both sets of observations. The ARM ground-based
 461 instrument recorded an LWP of over 400 g m⁻² during the nighttime with drizzle droplets reaching
 462 to the surface on 1 July 2016 (Fig. 6a). As the sunrise (around 6 Z local time), the LWP decreases
 463 to a range of about 100 g m⁻², and then increases again to 600 g m⁻² after 21 Z local time.

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485 To compare with the ARM ground-based observations, the WRF-Chem simulated result is
 486 averaged over 20×20 grids centered on the Azores, which corresponds to an approximate
 487 resolution of 4 km (Fig. 6a). Overall, the control run generates a thin cloud layer with an
 488 underestimated LWP during the nighttime, capturing only 10-20% of the observed LWP. The
 489 simulated clouds are more consistent with the observations during the daytime, especially in the
 490 perturbed run. However, it is important to note that the LWP retrieved by MWR experiences
 491 significant uncertainties during drizzling or precipitating conditions. This is primarily due to the
 492 scattering effects of large raindrops and raindrops accumulating on the instrument's radome, which
 493 can result in an overestimation of LWP (Tian et al., 2019; Cadeddu et al., 2020).

494 Figure 6b depicts the comparison of CF between observations and WRF-Chem. The CF
 495 values obtained from Meteosat are close to 1, indicating a solid cloud field. In contrast, the CF
 496 simulated by 20160701_control range between 0.5 and 0.9 on a domain-averaged scale. Similar
 497 to the LWP results, the simulated CFs from 20160701_control exhibit a diurnal cycle, with higher
 498 values during the nighttime and lower values during the daytime. Due to the thinner clouds
 499 simulated in 20160701_control based on LWP, the modeled CF is 40-60% lower than the
 500 observation in the afternoon, indicating that clouds dissipate more quickly in the control run.
 501 Conversely, the 20160701_perturbed scenario demonstrates improved performance in both LWP
 502 and CF. This indicates that the 20160701 case is sensitive to aerosol variations, with the CCN
 503 number being too low in the control run.

504 Compared to a ridge system like the case of 20160701, the WRF-Chem model has
 505 difficulty in capturing the warm boundary layer clouds under a regime characterized by a post-
 506 trough system (20170719) or a weak trough system (20190823). Compared to the observations,
 507 the simulated LWP in 20170719_control is about 30% of the observed value (Fig. 6c. In contrast,

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517 the simulated CF performs better, reaching about 75% of the observed value (Fig. 6d). The
 518 discrepancy between the modeled results and observations may arise from delayed moisture
 519 transfer from the outer domain or insufficient vertical resolution. In this instance, the cloud
 520 systems move quickly under the post-trough weather regime. A 5-minute moisture input from the
 521 boundary condition using WRF downscaling (ndown) may not be sufficient to transport moisture
 522 into the inner domain, making it difficult for the model to develop thicker marine stratocumulus
 523 clouds, especially for such high spacing resolution. On the other hand, in another ongoing project,
 524 we increased the vertical layers to 99, which is to double vertical layers below 2km. However, the
 525 test with 99 levels only slightly improves the cloud cover and LWP. The estimated LWP
 526 susceptibility shows little to no change. The insignificant improvement indicates even higher
 527 resolution is needed to fully simulate the cloud processes near the sharp boundary layer inversion
 528 for these solid stratocumulus cloud layer. Another possible reason is that the 6th Order Horizontal
 529 Diffusion used in the study (diff_6th_opt = 2) rapidly dissipates marine stratiform clouds,
 530 especially in the high spacing resolution (Knievel et al., 2007). It is worth noting that Christensen
 531 et al. (2024) conducted sensitivity tests using various shallow cumulus and microphysics schemes,
 532 and the different combinations of these schemes had a substantial impact on the simulated cloud
 533 amount as well.

534 Moving to the case of 20190823, overall, compared to the observations, the control run
 535 captures LWP and CF slightly better, especially in the domain-averaged scale (Figs. 6e and 6f).
 536 Based on the LWP observed from ARM, there are two systems passing in the area, one between 5
 537 Z to 13 Z local time and the other between 17 Z to 22 Z local time on 23 August 2019, (high ARM
 538 LWP in Fig. 6e). The simulation of 20190823_control captures the first system, but slightly
 539 underestimates LWP; however, the model misses the second system. The model simulated CFs

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552 also match well with Meteosat (Fig. 6f). Only after 17 Z, the model misses catching the second
553 system. The CFs drops 50 – 70 % compared to the observations.

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554 The underestimation of LWP and CF in model simulations leads to insufficient longwave
555 cooling at the cloud top. This reduced cooling weakens cloud-top entrainment, resulting in a less
556 pronounced boundary layer inversion and a shallower boundary layer height identified in Section
557 3.1.1. This creates a negative feedback loop, where the initial inaccuracies in cloud properties
558 affect boundary layer dynamics (Zheng et al., 2021). On the other hand, in the perturbed runs, the
559 results show an adjustment in the amounts of LWP and CF, aligning more closely with the
560 observations. This suggests that the CCN number is underestimated in the control runs (more
561 discussion in Section 3.3). The model's response to aerosol changes highlights its capability for
562 studying aerosol-cloud interactions.

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563 3.2 Aerosol composition and activation

564 The advantage of utilizing WRF-Chem to investigate aerosol-cloud interactions stems from
565 its capability to simulate the spatiotemporal distribution of CCN. This modeling is based on
566 various aerosol components and their sizes, as well as their dynamic responses to wet removal
567 processes associated with clouds and precipitation. In traditional simulations that rely on fixed or
568 prescribed aerosol distributions, accurately representing these factors can be particularly
569 challenging. WRF-Chem, however, allows for a more nuanced understanding by dynamically
570 modeling how aerosol populations evolve over time, especially after cloud evaporation processes.
571 During the evaporation process, the reduction in cloud water can lead to a re-entrainment of
572 aerosols back into the atmosphere, altering their concentration and properties. This change can
573 affect subsequent cloud formation and precipitation patterns, highlighting the importance of
574 capturing these interactions for reliable predictions.

583 In this section, we concentrate on aerosol activation, considering its size and chemical
584 composition across three different cases. The following section will discuss the aerosol indirect
585 effect and how changes in cloud properties feedback into the aerosol population and its activation
586 capability.

587 In Fig. 7a, the blue solid line and blue dashed line represent the vertical profiles of total
588 aerosol number concentration (including Aiken mode and accumulation mode) and aerosol number
589 concentration of the Aiken mode, respectively. These profiles are averaged over domain 4 on 1
590 July 2016. The environment shown in the figure is characterized by its cleanliness, with a total
591 aerosol number concentration below the cloud top (approximately 1000 m in height) measuring
592 less than 300 cm^{-3} . In the 20160701_control simulation, the total aerosol number is low, and
593 approximately 70% of the total aerosol numbers belong to the Aiken mode. According to the study
594 conducted by McCoy et al. (2024), which utilized aerosol number concentration measurements
595 from ARM airborne observations on 15 July 2017, it was found that the ratio of the Aiken mode
596 to the total aerosol number was approximately 50-60% within an altitude of 1000 m. Compared
597 to this observational analysis, our simulations generate an overabundance of small-sized aerosols,
598 which result in a low concentration of CCN. This discrepancy arises from the assumptions made
599 when constructing the aerosol initial and boundary conditions (80% for Aiken mode and 20% for
600 accumulation mode SO_4).

601 The CCN calculation presented in Fig. 7a is based on the Köhler theory, which considers
602 both the aerosol size (curvature effect) and the chemical composition (solution effect) to estimate
603 the theoretical CCN number concentration at different supersaturations. Below 1.0%
604 supersaturation, the CCN number concentration is found to be 42% of the total number of aerosol
605 number (could be estimated from 100% of accumulation mode and 16% of Aiken mode) (Fig.

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610 ~~S8a~~). In the simulation of 20160701_control, the CCN number concentration ~~below~~ 0.2% (0.5%)
 611 supersaturation is only 11% (25%) of the total aerosol number, which is lower than the
 612 observations reported in Wang et al. (2020), where the observed CCN number concentration under
 613 0.35% supersaturation was approximately 25% of the total aerosol number. Even though SO₄ is
 614 the dominant chemical component, accounting for nearly 50% (as shown in Figs. ~~8b~~ and ~~8c~~), the
 615 presence of an excessive number of Aiken mode aerosols may be the primary reason for the low
 616 activation rate. The curvature effect caused by these Aiken mode aerosols hinders their ability to
 617 act as CCN effectively.

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618 In the simulation of 20170719_control, the most aerosols ~~are~~ within a height of 1000 m,
 619 which is also the cloud layer (Fig. 7d). The average aerosol number concentration across the entire
 620 domain is measured to be 1286 cm⁻³ within a height of 2000 m and the Aiken mode is 80% of the
 621 total aerosol number in this case. The chemical composition of aerosols in the 20170719_control
 622 mainly is SO₄, ~~with the other species exhibiting lower concentrations (Figs. 7e and 7f)~~. This
 623 variation in vertical distribution leads to more aerosols being activated under the cloud top at a
 624 height of 1500 m. This is attributed to the presence of a peak value of accumulation mode aerosols
 625 and SO₄ at this height.

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626 Because of ~~the high number concentration of simulated~~ Aiken mode aerosols, overall, the
 627 activation rate is low. ~~Below 1.0% supersaturation, the CCN number concentration is estimated~~
 628 to be 25% of the total aerosol number. This could be a result of 100% of the accumulation mode
 629 aerosols and 7% of the Aiken mode aerosols contributing to the CCN population (Fig. ~~S8c~~). The
 630 CCN number concentration ~~below~~ 0.2% (0.5%) supersaturation is only 4% (12%) of the total
 631 aerosol number.

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641 Among the three cases studied, the case of 20190823 stands out as the most polluted case,
 642 but the aerosol component and vertical distribution are close to the case of 20170719 (Figs. 7c and
 643 7i). The average aerosol number concentration across the entire domain is measured to be 1850
 644 cm^{-3} within a height of 2000 m. The Aiken mode aerosols are also high and contribute to more
 645 than 75% of the total aerosol number in this case (Figs. 7g and S8e). The large SO_4 species
 646 concentration also leads to more aerosols being activated under the cloud top at a height of 2000
 647 m. The CCN number concentration below 0.2% (0.5%) supersaturation is only 6% (17%) of the
 648 total aerosol number, slightly better than the case of 20170719.

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649 The cloud droplet numbers observed in the three cases fall within the range of CCN
 650 numbers below 0.1% and 0.2% supersaturation. Therefore, in the subsequent sections, we utilize
 651 the CCN number concentration below 0.2% supersaturation as a representative of the CCN
 652 activation rate.

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653 3.3 Cloud responses to aerosol perturbations

654 Figure 8 illustrates the comparison of time series profiles of cloud water content (CWC)
 655 and CCN number concentration below 0.2% supersaturation between the control runs and
 656 perturbed runs. This figure also demonstrates the CCN spatiotemporal variation in our
 657 simulations. Specifically, for the case of 20160701, it is evident that the CWC in
 658 20160701_perturbed exhibits a positive response to increased CCN compared to the CWC in
 659 20160701_control. This result aligns with most WRF studies that use fixed or prescribed CCN
 660 numbers to investigate aerosol-cloud interactions (Wang et al., 2020; Christensen et al., 2024).

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661 Echoing the insufficient longwave cooling at the cloud top due to the underestimation of
 662 LWP and CF in model simulations discussed in Section 3.1.3, Terai et al. (2014) also observed

671 weak cloud-top entrainment in their study of five pockets of open cells, using aircraft data from
 672 the VOCALS Regional Experiment (VOCALS-REx). Their research indicated that clouds tend to
 673 break up easily in the absence of aerosols, characteristic of a pristine environment. Consistent
 674 with their findings, our study demonstrates that in-cloud collision-coalescence processes
 675 effectively remove aerosols, particularly because of the wider range of cloud droplet sizes present
 676 in a clean environment (i.e., the control runs) (Table 2). Even in the control run for 20160701, our
 677 results indicate that broken open cell clouds in a pristine environment struggle to develop into
 678 closed clouds.

679 Figure 9a depicts the time series of domain-averaged LWP, encompassing both cloud and
 680 rain, and CCN number concentration below 0.2% supersaturation for both the 20160701_control
 681 and 20160701_perturbed cases. This visualization provides a quantitative representation of the
 682 change in CCN number concentration, which increases from a mean value of 32.52 cm^{-3} in the
 683 control run to 127.68 cm^{-3} in the perturbed run, approximately four times higher than the control
 684 run. Because we want to avoid counting high CCN number concentration above cloud top which
 685 do not readily become cloud droplets, the CCN number concentration is averaged from the surface
 686 up to the height of 1000 m (Wang et al., 2020).

687 The LWP in the 20160701_control case exhibits a domain mean value of 64.88 g m^{-2} ,
 688 which subsequently increases to 123.27 g m^{-2} in the 20160701_perturbed case. As mentioned in
 689 Section 3.1, the LWP for the 20160701 case follows a diurnal cycle, with higher values during
 690 nighttime and lower values during daytime. This diurnal cycle is also observed in the perturbed
 691 simulation, with the larger differences in CCN and LWP between the control run and perturbed
 692 run during nighttime (Fig. 9a).

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698 After increasing ~~the aerosol concentration~~, the cloud droplet number in the
699 20160701_perturbed run demonstrates similar responses. In the 20160701_control case, the
700 domain mean value of ~~the~~ cloud droplet number is 14.03 cm^{-3} , which subsequently increases to
701 45.52 cm^{-3} in the 20160701_perturbed case. As the cloud droplet number increases, the cloud
702 radius decreases from $12.23 \text{ }\mu\text{m}$ in the control run to $10.08 \text{ }\mu\text{m}$ in the perturbed case.

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703 The case of 20170719 represents a post-trough weather regime, and Fig. 8c illustrates the
704 passage of a frontal system in the area after ~~8 Z local time~~ on that day. In the 20170719_perturbed
705 simulation, the CWC increases following the system's passage (Fig. 8d) compared to the CWC in
706 the 20170719_control run. Additionally, the ambient CCN number in the perturbed run is also
707 higher. The time variation of CCN concentration in Fig. 9c shows elevated CCN numbers before
708 and after the system enters the domain. In the 20170719_control case, the domain mean value of
709 CCN number concentration is 60.51 cm^{-3} , which subsequently increases to 253.51 cm^{-3} in the
710 20170719_perturbed case. The domain-averaged LWP also exhibits an increase, rising from 59.31
711 g m^{-2} in the 20170719_control run to 74.07 g m^{-2} in the 20170719_perturbed case. Notably, this
712 change primarily occurs after the passage of the frontal system.

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713 The cloud droplet number consistently shows higher values in the perturbed case (Fig. 9d),
714 and this pattern is similar to the difference in CCN between the two runs of 20170719 (Fig. 9c).
715 In the 20170719_control case, the domain mean value of cloud droplet number is 20.70 cm^{-3} , while
716 the value is 56.09 cm^{-3} in the perturbed case. When the cloud droplet number increases in the
717 perturbed run, the cloud radius decreases from $9.90 \text{ }\mu\text{m}$ in the control run to $7.49 \text{ }\mu\text{m}$ in the
718 perturbed case. This reduction in cloud radius is even smaller than the cloud radius observed in
719 the case of 20160701.

722 The case of 20190823 is similar to the case of 20170719, but it represents a weak trough
 723 weather regime. Figure 8e also illustrates the passage of a cloud system in the area between 5 Z
 724 to 17 Z local time on 23 August 2019, and the CWC in the perturbed run increases during this
 725 period. Quantitatively, in the 20190823_control case, the domain mean value of CCN number
 726 concentration is 124.32 cm^{-3} , which subsequently increases to 475.37 cm^{-3} in the
 727 20190823_perturbed case, which is also about three times higher. The domain-averaged LWP
 728 also exhibits an increase, rising from 48.92 g m^{-2} in the 20190823_control run to 58.53 g m^{-2} in
 729 the 20190823_perturbed case.

730 Differing from the case of 20170719, the frontal system moved away from the study
 731 domain after the noontime, the differences of CCN number or cloud droplet number between the
 732 control and perturbed runs becomes even more pronounced after the system (Figs. 9e and 9f). It
 733 is because aerosols are transported to the area following the frontal system (Fig. 4f) and then more
 734 aerosols activated as CCN. In the 20190823_control case, the domain mean value of cloud droplet
 735 number is 33.94 cm^{-3} , while the value is 79.97 cm^{-3} in the perturbed case. When the cloud droplet
 736 number increases in the perturbed run, the cloud radius decreases from $8.51 \text{ }\mu\text{m}$ in the control run
 737 to $6.45 \text{ }\mu\text{m}$ in the perturbed case. This reduction in cloud radius is similar to the cloud radius
 738 observed in the case of 20170719.

739 We observe that large aerosol-induced LWP occurs during the periods of rainfall (Fig. S9).
 740 To accurately quantify the differences, we calculate the average LWP over approximately 25 km
 741 of domain 4. This results in 16 averaged grids per output file, with each file generated every 10
 742 minutes. This averaging process is based on Arola et al. (2022) and Zhou and Feingold (2023) to
 743 avoid the impact of heterogeneity and co-variability on the results. Specifically, we aggregate the

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750 simulation grids with a spacing resolution of approximately 190 m to form a larger grid of around
751 25 km for each 10-minute simulation output, as shown in Fig. S10a.

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752 Table 2 presents the 10-minute mean and standard deviation of several variables, including
753 CCN, LWP, cloud droplet number (N_c), cloud radius (R_e), and rainfall intensity (RI), across three
754 study cases. The classification of "rain" and "non-rain" is based on the RI (unit: mm hr^{-1}) on the
755 averaged grid. Specifically, a grid is considered as "rain" if the RI is greater than zero. In the
756 control cases, the averaged CCN number is 73.07 cm^{-3} , and the corresponding LWP is 53.17 g m^{-2} .
757 However, in the perturbed cases, the CCN number increases approximately threefold, reaching
758 218.21 cm^{-3} , and the LWP increases by 49% to 79.25 g m^{-2} . The introduction of additional aerosols
759 in the perturbed cases also leads to a significant increase in the N_c number, from 22.68 cm^{-3} in the
760 control cases to 59.74 cm^{-3} in the perturbed cases. Consequently, the R_e decreases by 21% from
761 $9.97 \mu\text{m}$ to $7.83 \mu\text{m}$, and the RI decreases by 11% from 0.009 mm hr^{-1} to 0.008 mm hr^{-1} .

762 To investigate the interaction between aerosols and clouds, we analyze the results
763 separately for rain and non-rain grids. In both the control and perturbed cases, we observe that the
764 CCN number within 1000 m is lower in the rain grids compared to the non-rain grids, primarily
765 due to the washout effect caused by rainfall. Additionally, the LWP over the rain grids is generally
766 higher than that over the non-rain grids. Furthermore, when comparing the control and perturbed
767 cases, we find that the LWP over the rain grids increases by 57% from 58.57 g m^{-2} to 91.81 g m^{-2} .
768 In contrast, the LWP over the non-rain grids only increases by 28% (Table 2). This difference can
769 be attributed to the conversion of cloud droplets to raindrops through processes like autoconversion
770 and collection, which occurs more prominently over the rain grids. We also observe that in the
771 non-rain grids, especially at the cloud edges (or low LWP), the perturbed cases reveal an increased
772 presence of small cloud droplets. This abundance of smaller droplets facilitates evaporation,

775 resulting in a reduced LWP (e.g., clouds in the bottom right corner of Figs. S10a and S10b).
776 Consequently, N_c over the rain grids is lower compared to N_c over the non-rain grids. Moreover,
777 when introducing aerosols in the perturbed runs, the results over the rain grids exhibit larger cloud
778 drops and a wider radius spectrum compared to the results over the non-rain grids. This suggests
779 that the presence of aerosols has a more pronounced effect on cloud properties within the rain
780 grids.

781 Zheng et al. (2022) conducted a study on the aerosol-cloud interaction using ground-based
782 measurements from the ARM program, focusing on the influence of environmental variables.
783 Their findings revealed that when there is ample water vapor and low CCN loading, the active
784 coalescence process leads to a broader size distribution of cloud droplets, resulting in an increase
785 in cloud droplet radius. On the other hand, when there is enhanced activation of CCN and
786 condensational growth of cloud droplets due to higher CCN loading below the cloud, the cloud
787 droplet radius decreases. This combined effect signifies an intensified aerosol-cloud interaction,
788 leading to a broad range of cloud droplet radii. The simulated results in our study, specifically
789 over the rain grids where a sufficient water vapor environment is considered, demonstrate a
790 significant aerosol-cloud interaction, where increased CCN introduces more newly converted
791 droplets, resulting in a broad range of cloud droplet radii.

792 Since we utilize a comprehensive aerosol module in WRF-Chem to examine aerosol-cloud
793 interactions, we are able to explore how changes in cloud properties, driven by increased CCN,
794 affect aerosol concentrations. For example, in the post-trough regime (20170917 case) and the
795 weak trough regime (20190823 case), we observe that the cloud structure exhibits more open-cell
796 stratocumulus clouds (Figs. 2h and 2i). As mentioned above, the increased number of smaller
797 cloud droplets at the cloud edge facilitate evaporation and results in a lower LWP (Fig. S10). The

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806 larger aerosols from the evaporated clouds return to the accumulation mode, making them more
807 likely to activate as CCN again.

808 To demonstrate how robust this process on the ACI, we calculate the time series of the
809 ratio of the ~~number concentration of CCN at a supersaturation of 0.2% in the perturbed runs to that~~
810 ~~in the control runs, normalized by the corresponding accumulation mode aerosol concentration,~~
811 defined as $(CCN_{0.2\%}/Accu.aerosols)_{perturbed} / (CCN_{0.2\%}/Accu.aerosols)_{control}$, shown in
812 Fig. 10. A ratio greater than 1 suggests that accumulation mode aerosols in the perturbed cases
813 are more readily activated as CCN at a supersaturation of 0.2%, especially in the cases of 20170719
814 and 20190823. Conversely, a ratio less than 1 is observed during the first half of the day in the
815 ~~20160701~~ case, which is attributed to the very low levels of accumulation mode aerosols in the
816 model.

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817 3.4 Cloud liquid water path (LWP) susceptibilities

818 In this study, the susceptibility of LWP to changes in CCN concentration is quantified
819 using the logarithmic slope between LWP and CCN, denoted as $d\ln(LWP)/d\ln(CCN)$
820 (Gryspeerd et al., 2019). This slope represents the sensitivity of ~~the LWP in~~ warm stratocumulus
821 ~~clouds~~ to variations in CCN concentration, like shown in Fig. ~~S10c~~. As presented in Table 2, we
822 aggregate the simulation grids with a spacing resolution of approximately 190 m to form a larger
823 grid of around 25 km for each 10-minute simulation output. This averaging process helps to reduce
824 the impact of heterogeneity and co-variability on the results.

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825 Figure 11 illustrates the averaged cloud susceptibilities for various LWP and CCN or Nc
826 bins across three study periods. The logarithmic slope between LWP and CCN is calculated at
827 each output time (every 10 minutes) using data from 16 aggregated grid points from the control

run and 16 aggregated grid points from the perturbed run. Our study reveals that when the CCN concentration is below 100 cm^{-3} , the susceptibility for different LWP and CCN values is positive and the values are large, indicating that changes in LWP are sensitive to variations in CCN number. Here also demonstrates the AIE is large when an increase in CCN can have a large impact on LWP enhancement. However, when the mean CCN concentration exceeds 100 cm^{-3} , the relationship between LWP and CCN becomes more complex, with both positive and negative susceptibilities observed. This suggests that the change in LWP is influenced by other factors, such as environmental conditions and cloud precipitation status. (as shown in Fig. 11a). It is important to note that the CCN number used in our study is averaged from the surface up to the height of 1000 m, which may introduce uncertainty to the absolute values of susceptibility by including aerosols that are not directly involved in the aerosol-cloud interaction (Wang et al., 2020).

Additionally, our simulations indicate that the N_c in this study is generally low, with a mean value typically below 80 cm^{-3} . For different LWP and N_c values, the susceptibility is mostly positive, indicating that changes in LWP are sensitive to variations in N_c number (as shown in Fig. 11c).

When we investigate the variation of LWP susceptibility over time, we observe that positive susceptibilities for different LWP and CCN (N_c) typically occur during periods of no rain or light rain (Figs. 12 and S9). On 1 July 2016, the time series of LWP susceptibility for different CCN or N_c shows a diurnal cycle, with large positive values during the nighttime and small positive values in the afternoon. During heavy rain events, such as from 8 to 15 Z local time on 19 July 2017 and from 23 Z to 14 Z on 23 August 2019 (Fig. S9), the LWP susceptibilities are negative or close to zero (Figs. 12c and 12e). In the perturbed cases, during the heavy rainfall periods, some aggregated grids show very low LWP (Fig. S10c). This reduction in LWP is caused

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874 by the evaporation from small cloud droplets at non-rain grids on the cloud edge. Those low LWP
 875 grids in the perturbed runs result in a negative or near-zero logarithmic slope between LWP and
 876 CCN (Figs. [S10c](#), 12c and 12e), although the domain mean LWP is higher in the perturbed case
 877 than in the control case (Figs. 9c and 9e).

878 To further illustrate the reduction in LWP due to evaporation at the cloud edge, Fig. 13
 879 presents the relative change in $\ln(LWP)$ between the perturbed and control cases across different
 880 LWP percentile ranges in the control case during the periods of negative LWP susceptibility, as
 881 shown in Fig. 12. The results indicate a decrease in LWP in the perturbed cases compared to the
 882 control cases for pixels with the lowest LWP percentile range (0-25%). We assume ~~that this range~~
 883 ~~corresponds to thin clouds~~ at the ~~edges of the cloud cover~~.

884 Figures 11b and 11d display the mean Re susceptibilities for different CCN ~~number~~ and
 885 Nc, respectively. The results consistently show that as CCN ~~number~~ or Nc increases, the radius
 886 of the cloud droplets decreases. Additionally, the change in Re is more pronounced when Nc (or
 887 CCN ~~number~~) is higher.

888 In this study, the logarithmic slope between LWP and CCN in the LWP susceptibility
 889 calculation is based on a linear assumption. Hoffmann et al. (2024) introduced a heuristic model
 890 that represents a significant advancement in understanding the process-level adjustments of cloud
 891 water in stratocumulus clouds, suggesting that the relationship may resemble a reversed "V" shape.
 892 Figure 13 indicates that the decrease in LWP (negative susceptibilities) in the perturbed cases
 893 occurs only in low LWP clouds (thin and non-rain clouds). Conversely, LWP increases (positive
 894 susceptibilities) in thicker, precipitating clouds under the perturbed scenarios, which is consistent
 895 with the findings of Hoffmann et al. (2024). Since our study aggregates grids to a 25 km resolution,
 896 we are able to capture such spatial heterogeneity and retrieve a negative susceptibility due to the

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evaporation of thin clouds at cloud edge while at 100-km scale (domain-averaged shown in Figs. S11 and S12), the signal is dominated by the increase of LWP at cloud core. However, based on the current study cases, we may not have sufficient data samples to illustrate a relationship beyond the linear assumption.

4 Discussion and summary

This study focuses on aerosol indirect effects (AIE), particularly involving long-range transport aerosols, in the Eastern North Atlantic (ENA) region. It specifically examines these effects on warm boundary layer stratiform clouds located on the eastern side of oceanic subtropical highs under three different weather regimes: a ridge with a surface high-pressure system, a post-trough with a surface high-pressure system, and a weak trough. We select three specific study cases (i.e., 20160701, 20170719, and 20190823) to assess the impact of long-range transport aerosols on warm boundary layer clouds, with each case representing a typical meteorological regime observed over the ENA site.

To investigate aerosol-cloud interactions more realistically, incorporating aerosol chemistry components that activate to cloud condensation nuclei (CCN) and accounting for aerosol spatiotemporal variation, this study employs the Weather Research and Forecast model coupled with a chemistry component (WRF-Chem). This approach provides a detailed examination of AIE in the ENA region under the three specified weather regimes. We employ a downscaling technique to conduct WRF-Chem simulations for the two inner domains (with the outer domains utilizing WRF). This approach results in nearly a 50% reduction in total computational costs, achieving a throughput of 8 hours per day using 1,080 cores.

We incorporate major aerosol species (BC, OC, and SO₄), as well as SO₂, from MERRA-2 to provide aerosol initial and boundary conditions, labeled as control cases. Additionally, we

925 formulate three perturbed cases by amplifying aerosol concentrations in both initial and boundary
926 conditions, as well as sea salt emissions, by a factor of five relative to each control case. Since
927 aerosol features are primarily determined by aerosol initial and boundary conditions, a higher
928 Aiken mode assumption in the major aerosol component (i.e., SO₄) regarding the aerosol mode
929 ratio (80% for Aiken mode and 20% for accumulation mode) results in fewer aerosols activating
930 as CCN due to the curvature effect in our simulations.

931 The WRF-Chem model captures the cloud structure in the case of 20160701, ~~but~~
932 ~~underestimates the cloud amount.~~ It simulates the formation of thin, uniform stratocumulus clouds
933 within a meteorological regime characterized by a ridge system in the free troposphere and a high-
934 pressure system near the surface. However, the cases of 20170719 and 20190823 exhibit the
935 development of ~~thicker but broken~~ solid stratocumulus clouds within a post-trough regime and a
936 weak trough, respectively. With the fast-moving cloud systems and strong surface wind, the WRF-
937 Chem model ~~struggles~~ to capture the development and movement of these cloud systems due to
938 delayed moisture transport from outer boundary condition and potential insufficient vertical
939 resolution.

940 In all cases, compared to the observations, the WRF-Chem model underestimates the liquid
941 water path (LWP) and cloud fraction due to warmer and lower simulated boundary layer. In the
942 perturbed cases, we find 57% higher aerosol-induced LWP, especially ~~for the rain grids.~~ We also
943 note that the perturbed cases exhibit lower rainfall intensity, indicating a rainfall suppression effect
944 attributed to high CCN concentrations as concluded in previous studies (Wang et al., 2020;
945 Christensen et al., 2024). In contrast, the LWP over the non-rain grids only increases by 28%.
946 Moreover, when introducing aerosols in the perturbed runs, the results over the rain grids exhibit
947 larger cloud drops and a wider radius spectrum compared to the results over the non-rain grids.

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953 This suggests that the presence of aerosols has a more pronounced effect on cloud properties within
954 the rain grids. The non-rain grids over the cloud edge can have lower LWP because smaller cloud
955 droplets are easy to evaporate.

956 Our study further elucidates the intricate feedback mechanisms governing aerosol-cloud
957 interactions and aerosol properties. In both the post-trough and weak trough regimes, we observe
958 a pronounced tendency for the cloud structure to develop more open-cell stratocumulus clouds.
959 At the peripheries of these clouds, the perturbed cases demonstrate a significant increase in the
960 presence of small cloud droplets. This heightened abundance of smaller droplets promotes
961 evaporation, thereby leading to a marked reduction in LWP.

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962 As these clouds evaporate, the aerosols that are released return to the accumulation mode.
963 This transition enhances their likelihood of reactivating as CCN. Consequently, this cycle
964 underscores the dynamic interplay between aerosol properties and cloud formation, highlighting
965 how changes in aerosol concentrations can influence cloud microphysics and, ultimately,
966 precipitation processes.

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967 Additionally, the susceptibility of LWP to changes in CCN concentration is quantified
968 using the logarithmic slope between LWP and CCN. Our result shows when the CCN
969 concentration is low, LWP is sensitive to variations in CCN number, with higher CCN number
970 concentration leading to higher LWP. However, when the mean CCN concentration is relatively
971 high, LWP is not as sensitive to changes in CCN, the LWP susceptibilities are small in magnitude,
972 with both positive and negative values. Those negative values are caused by the evaporation from
973 small cloud droplets at non-rain grids on the cloud edge.

974 In Wang et al. (2020), the LWC susceptibility for a light precipitation case on 18 July 2017
975 also shows positive values based on three sensitivity runs with CCN concentrations of 10, 100,

979 and 1000 cm⁻³. The cloud properties in their study are averaged over all cloud points in the
 980 innermost domain. We adopt the same method as Wang et al. (2020) to estimate the LWP
 981 susceptibility using the domain mean values, defined as
 982 $\Delta \ln(LWP_{perturbed} - LWP_{control}) / \Delta \ln(CCN_{perturbed} - CCN_{control})$. This approach
 983 predominantly yields positive values for LWP susceptibility across the three study cases (see Figs.

984 S11 and S12). This suggests that the 25-km resolution is able to capture such spatial heterogeneity
 985 and retrieve a negative susceptibility due to the evaporation of thin clouds at cloud edge while at
 986 100-km scale (domain-averaged), the signal is dominated by the increase of LWP at cloud core.

987 Conversely, the LWP susceptibilities associated with varying cloud droplet numbers
 988 reported in Qiu et al. (2024) reveal significant negative values in LWP susceptibility in response
 989 to high cloud droplet numbers, a trend that is partially reflected in our study. Further investigation
 990 is required to reconcile the difference in LWP responses between observational data and model
 991 simulations. Additionally, a more accurate estimation of LWP susceptibility to changes in CCN
 992 concentration is necessary.

993 Moreover, future research will focus on addressing the identified issues within the model,
 994 such as improving the representation of transported aerosol size distributions, resolving the overly
 995 fragmented stratocumulus cloud layers with sharp boundary layer inversions, and comparing the
 996 modeled cloud susceptibilities with observational data through simulations of ship tracks and local
 997 aerosol perturbations.

998

999 **Code and data availability:**

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- Deleted: critical for accurately estimating LWP
- Deleted: . A resolution that is too coarse (i.e., using
- Deleted: mean) may fail to capture finer details, such as
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1008 The WRF-Chem code (v4.4.2) used in this study has been released on GitHub
1009 (<https://github.com/wrf-model/WRF/releases/download/v4.4.2/v4.4.2.tar.gz>). The observational
1010 data used in this study are available at <https://doi.org/10.5281/zenodo.13356995>. Other WRF-
1011 Chem simulated outputs for the plots in this paper are available at
1012 <https://doi.org/10.5281/zenodo.13357040>.

1013

1014 **Author contributions:**

1015 H.-H. Lee and X. Zheng provided ideas and designed the experiments in this study. H.-H. Lee
1016 conducted all the simulations and analyses. H.-H. Lee leads and coordinates the manuscript with
1017 inputs from coauthors.

1018

1019 **Competing interests.**

1020 At least one of the (co-)authors is a member of the editorial board of Atmospheric Chemistry and
1021 Physics.

1022

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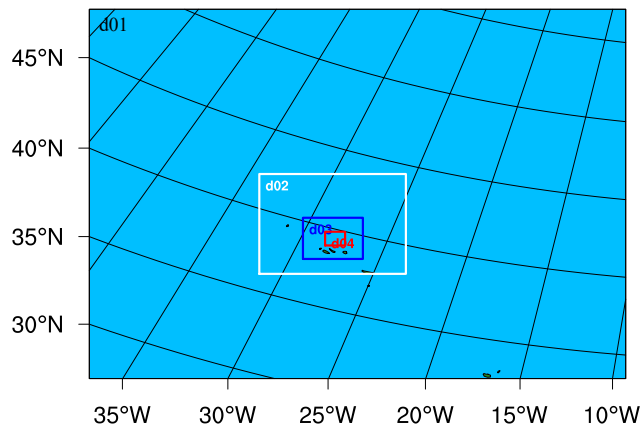
Table 1. WRF physics scheme configuration

Physics Processes	Scheme	Reference
Microphysics	Morrison (2 moments) scheme	Morrison et al. (2009)
Longwave radiation	RRTMG scheme	Mlawer et al. (1997)
Shortwave radiation	RRTMG scheme	Iacono et al. (2008)
Surface-layer	Monin-Obukhov surface layer	Monin and Obukhov (1954)
Land surface	Unified Noah land-surface model	Chen and Dudhia (2001)
Planetary boundary layer	MYJ (Eta) TKE scheme (d01 and d02 only)	Mellor and Yamada (1982) Janjić (1994)
Shallow cumulus parameterization	GRIMS scheme (d01 and d02 only)	Hong and Jang (2018)

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Table 2. Ten-minute mean and standard deviation of cloud condensation nuclei (CCN), liquid water path (LWP), cloud droplet number (Nc), cloud radius (Re), and rainfall intensity (RI) over three study cases. Data are averaged over ~25 km of the domain 4 and total 16 averaged grids are in the domain 4. Rain and non-rain are averaged the grids when the RI on the grid is larger than and equal to zero, respectively. Only CCN are averaged within 1000 m height over the domain 4, other variables are averaged within 2000 m height.

Area	Case	CCN (cm ⁻³)	LWP (g m ⁻²)	Nc (cm ⁻³)	Re (μm)	RI (mm hr ⁻¹)
Domain	Control	73.07 ± 48.77	53.17 ± 32.65	22.68 ± 11.59	9.97 ± 2.31	0.009 ± 0.033
	Perturbed	286.88 ± 183.69 (+293%)	79.25 ± 56.62 (+49%)	59.74 ± 27.29 (+163%)	7.83 ± 2.02 (-21%)	0.008 ± 0.033 (-11%)
Rain	Control	68.15 ± 48.05	58.57 ± 31.69	20.17 ± 9.33	10.47 ± 2.07	0.011 ± 0.035
	Perturbed	250.14 ± 153.23 (+267%)	91.81 ± 55.06 (+57%)	53.01 ± 20.39 (+163%)	8.35 ± 1.83 (-20%)	0.009 ± 0.036 (-18%)
Non-Rain	Control	103.73 ± 41.52	18.91 ± 9.81	38.57 ± 11.93	6.81 ± 0.76	0 ± 0
	Perturbed	444.47 ± 217.08 (+328%)	24.22 ± 15.80 (+28%)	89.24 ± 33.42 (+131%)	5.54 ± 0.90 (-19%)	0 ± 0



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Figure 1. Model domains are designed for simulations. The 4 domains ~~denote~~ 4 horizontal ~~resolutions~~ of 5 km (d01), 1.67 km (d02), 0.56 km (d03), and 0.19 km (d04), respectively.

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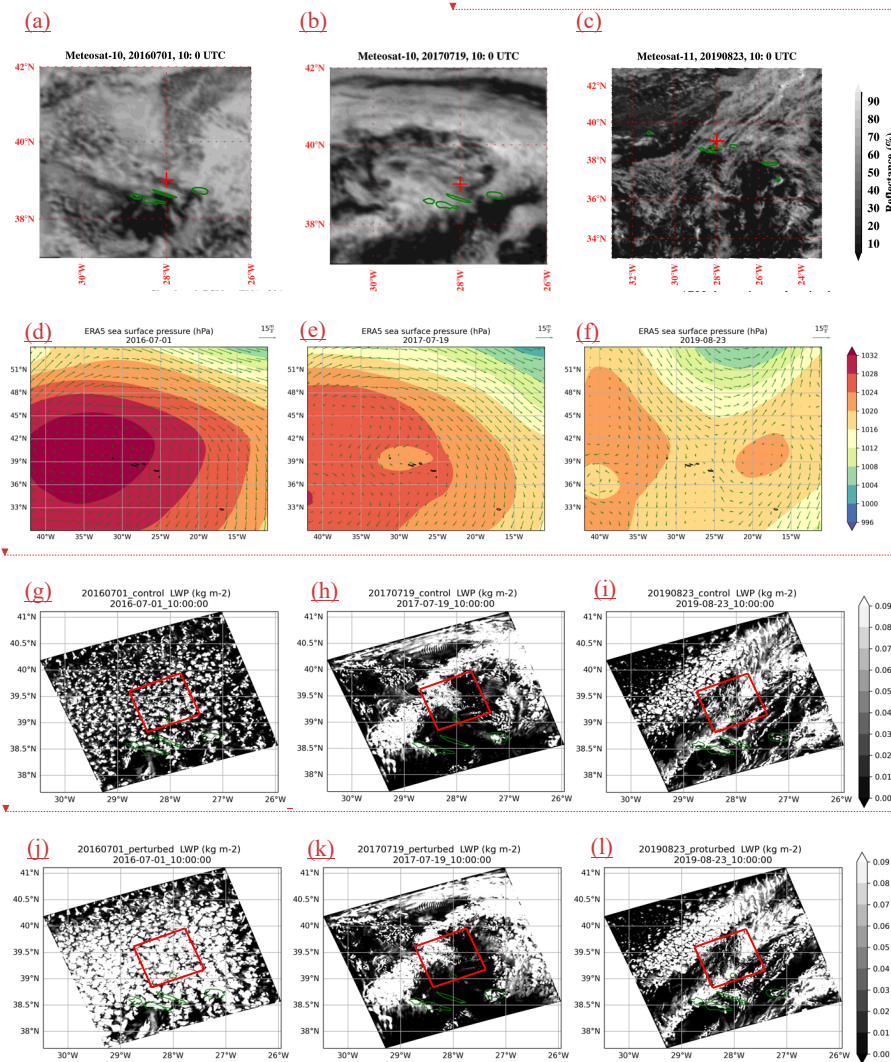


Figure 2. Spinning Enhanced Visible Infra-Red Imager (SEVIRI) images from the Meteosat satellite at 10:00 UTC on (a) 1 July 2016, (b) 19 July 2017, and (c) 23 August 2019 over the ENA. (d), (e), and (f) are on the same day of (a), (b), and (c), respectively, but from the ERA5 mean sea surface pressure (contour; units: hPa) and 10-meter surface wind (arrow; units: m s^{-1}).

1333 (g), (h), and (i) are on the same day of (a), (b), and (c), respectively, but ~~from the~~ WRF-Chem
1334 simulated liquid water path (LWP; units: kg m⁻²) in the control runs, ~~while (j), (k), and (l) are in~~
1335 ~~the perturbed runs~~. The red boxes in the figures indicate the result from domain 4.

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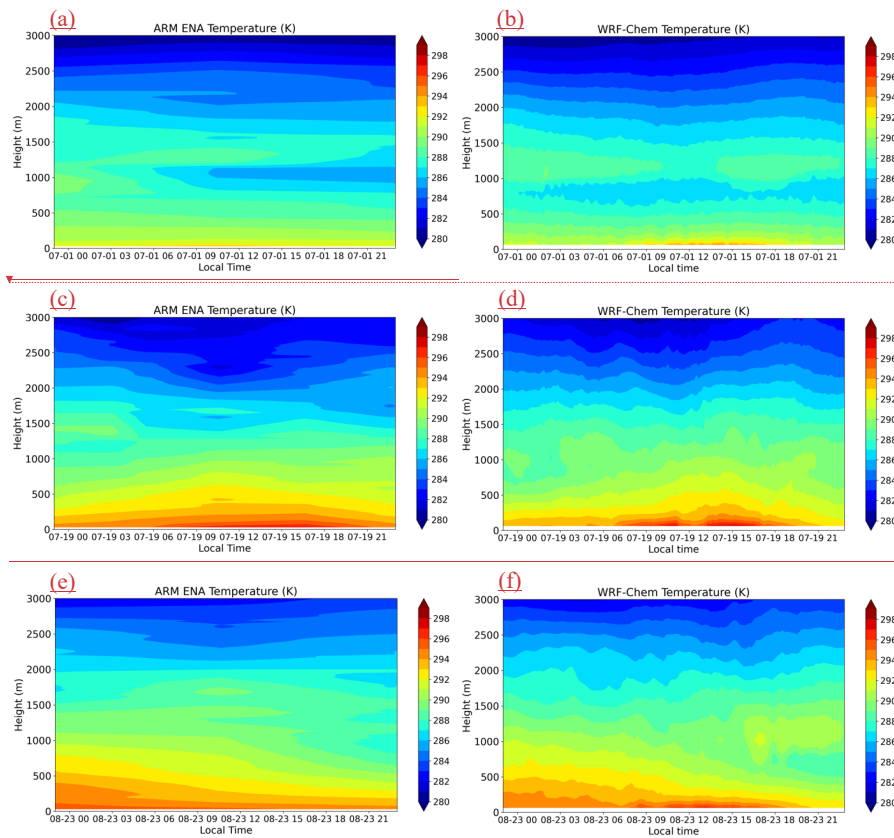
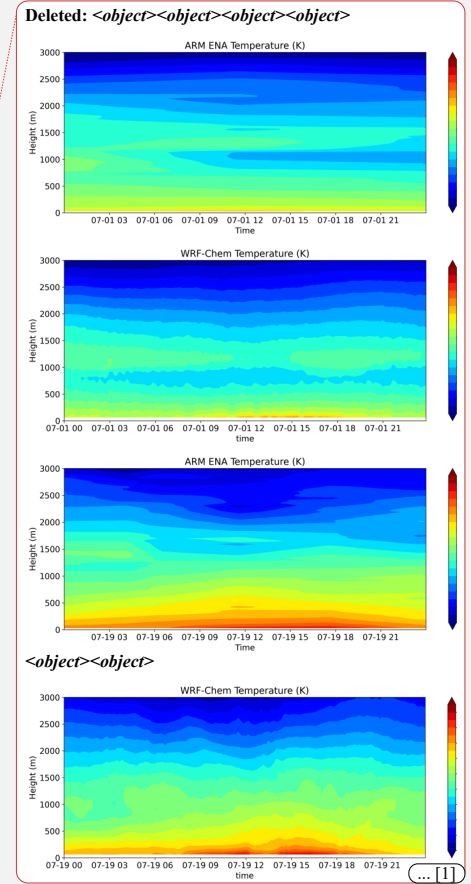


Figure 3. The time series (local time; UTC -1 hour) of temperature profiles (units: K) from ARM interpolated soundings at the Azores (39.09°N, -28.02°W) on (a) 1 July 2016, (c) 19 July 2017, and (e) 23 August 2019. Panels (b), (d), and (f) depict the same dates as (a), (c), and (e), respectively, but show the average temperature from WRF-Chem simulated results over 20×20 grids centered on the Azores (approximately 4 km resolution).



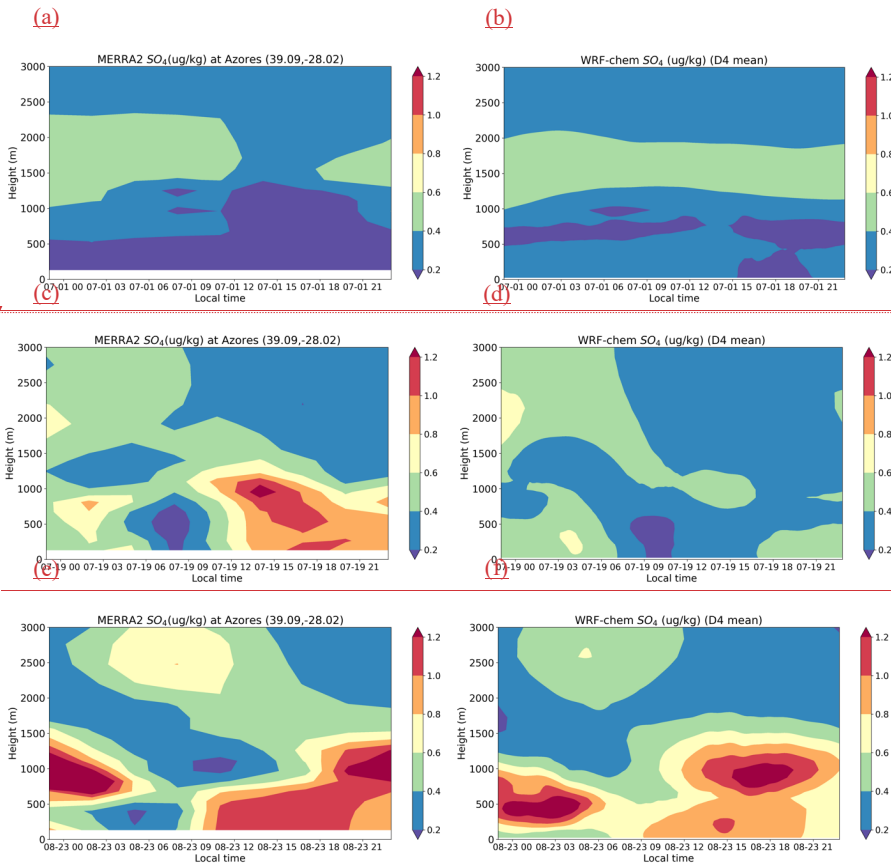
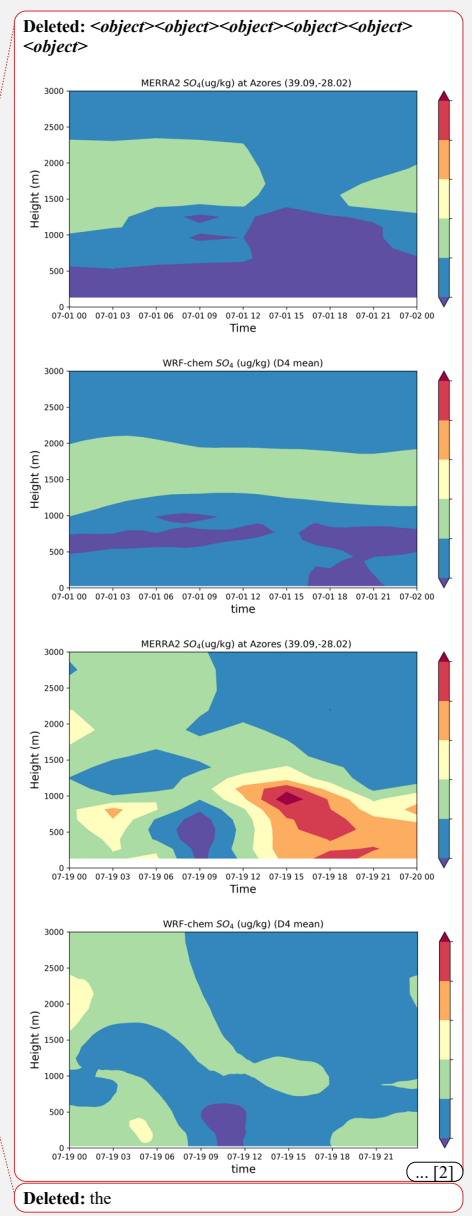
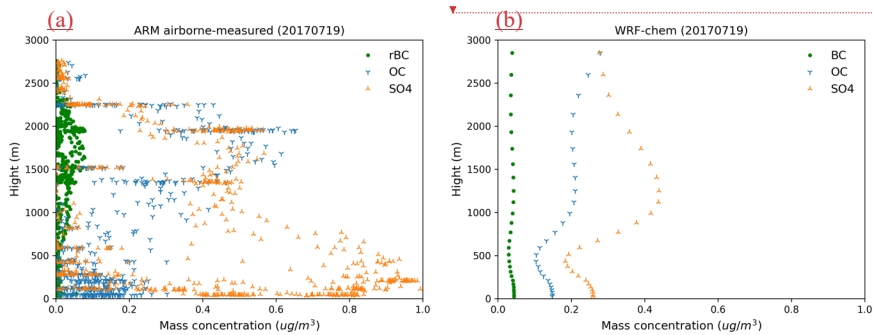


Figure 4. The time series (local time; UTC – 1 hour) of SO_4 profiles (units: $\mu\text{g kg}^{-1}$) from MERRA-2 at the Azores (39.09°N, -28.02°W) on (a) 1 July 2016, (c) 19 July 2017, and (e) 23 August 2019. Panels (b), (d), and (f) depict the same dates as (a), (c), and (e), respectively, but show the average aerosol concentration from WRF-Chem simulated data over domain 4.



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Figure 5. (a) ARM airborne-measured vertical profiles of SO₄, OC and refractory BC (rBC) mass concentration (units: $\mu\text{g cm}^{-3}$) averaged over multiple flights on 19 July 2017. **Note** that the highly uncertain and noisy aerosol observations between 600 – 1000 m height due to cloud contamination. (b) WRF-Chem simulated vertical profile of SO₄, OC, and BC mass concentration (units: $\mu\text{g cm}^{-3}$) averaged over domain 4 during the flight time from 8:40 to 11:50 UTC.

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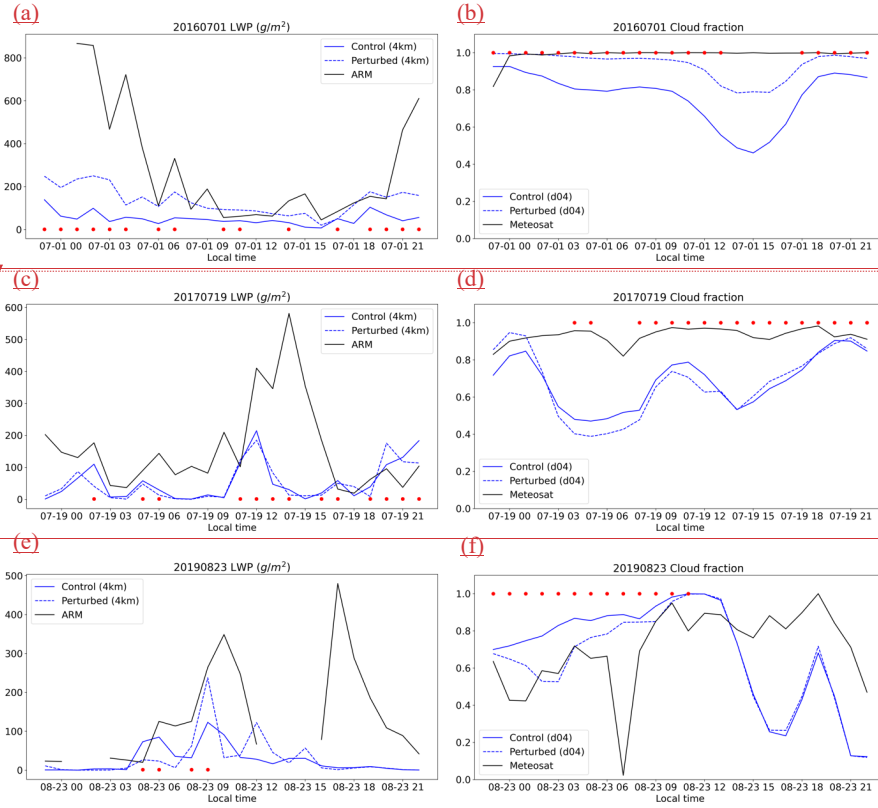
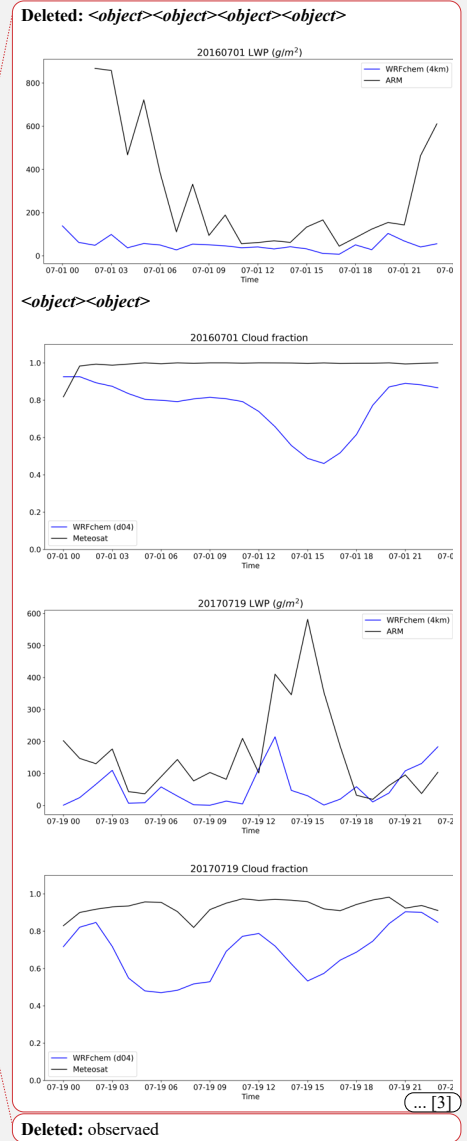


Figure 6. (a), (c), and (e) are the hourly time series (local time; UTC - 1 hour) of 4 km-averaged (4km) liquid water path (units: g m^{-2}) simulated from WRF-Chem (blue solid line) and observed from ARM (black solid line) on 1 July 2016, 19 July 2017, and 23 August 2019, respectively. (b), (d), and (f) are the hourly time series of domain-averaged (d04) cloud fraction simulated from WRF-Chem (blue solid line) and observed from Meteosat (black solid line) on 1 July 2016, 19 July 2017, and 23 August 2019, respectively. The 4 km-averaged data are averaged from the model simulated results over 20×20 grids centered on the Azores (approximately 4 km resolution). The red dots indicate when rainfall intensity is higher than 0.001 (0.01) mm hr^{-1} in 4-km averaged area (domain 4).



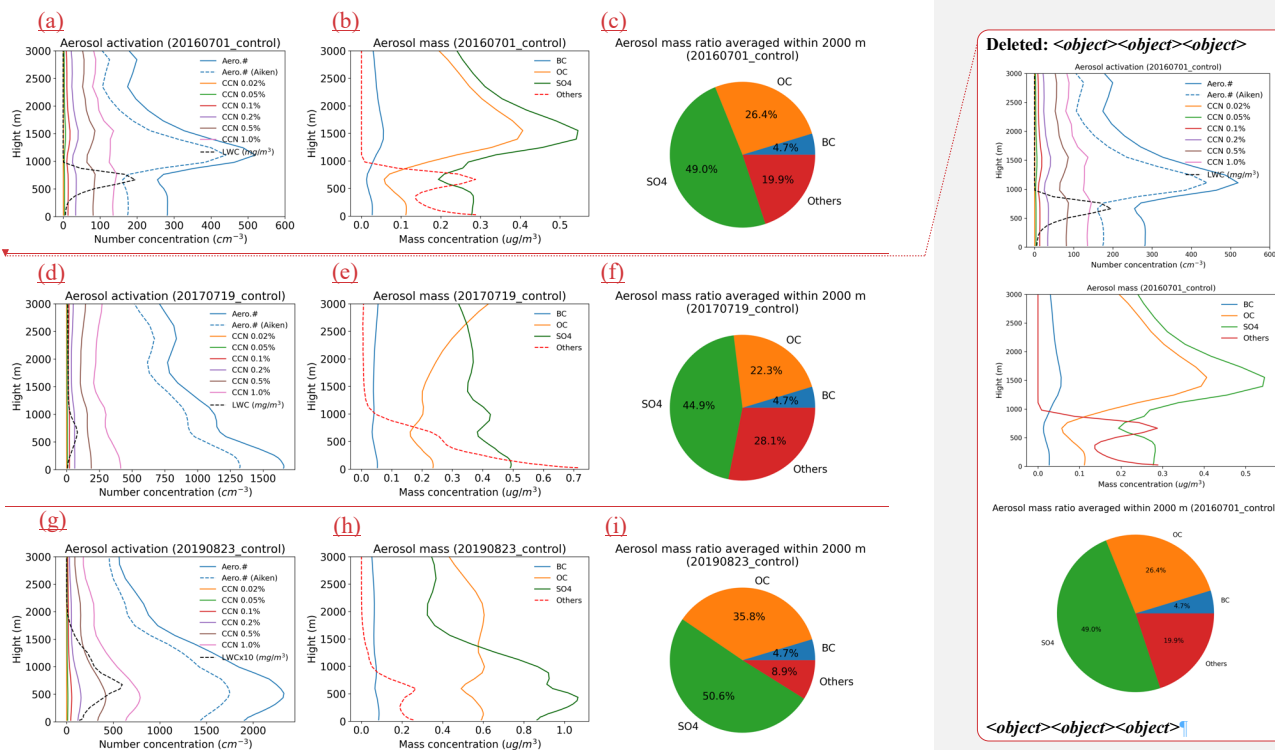


Figure 7. (a), (d), and (g) WRF-Chem vertical profiles of aerosol number concentration (Aiken mode and accumulation mode; units: cm^{-3}), aerosol number concentration (Aiken mode only; units: cm^{-3}), CCN number concentration below different supersaturations (units: cm^{-3}), and liquid water content (cloud and rain; units: mg m^{-3}) averaged over the domain 4 on 1 July 2016, 19 July 2017, and 23 August 2019, respectively, in the control runs. (b), (e), and (h) WRF-Chem vertical profiles of BC, OC, SO₄, and other species (like sea salt) (units: $\mu\text{g cm}^{-3}$) averaged over the domain 4 on 1 July 2016, 19 July 2017, and 23 August 2019, respectively, in the control runs. (c), (f), and (i) Pie chart of aerosol mass of different species averaged within 2000 m height on 1 July 2016, 19 July 2017, and 23 August 2019, respectively, in the control runs. Note that LWC is adjusted to fit the scale of x-axis for each case.

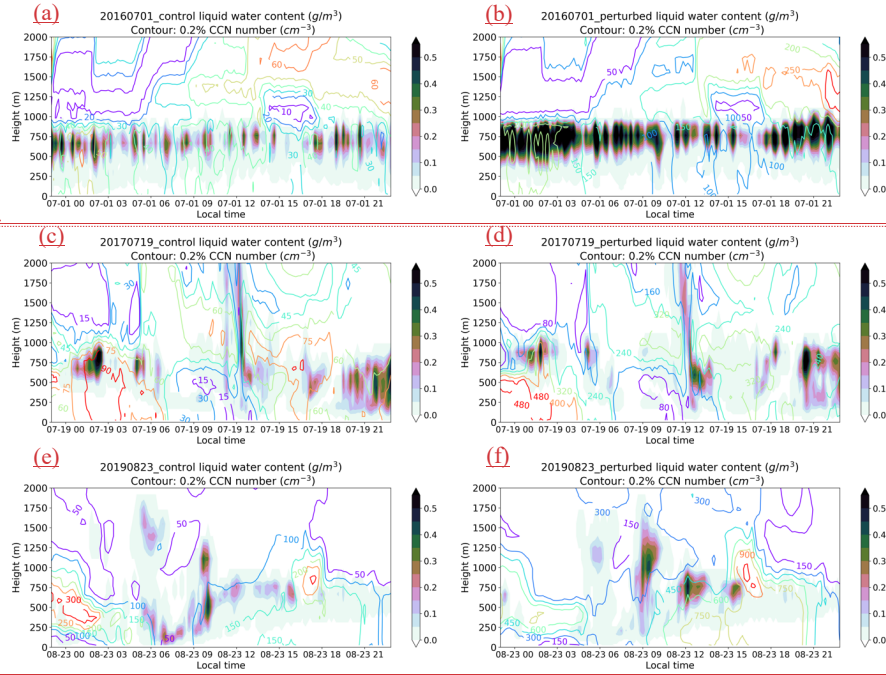
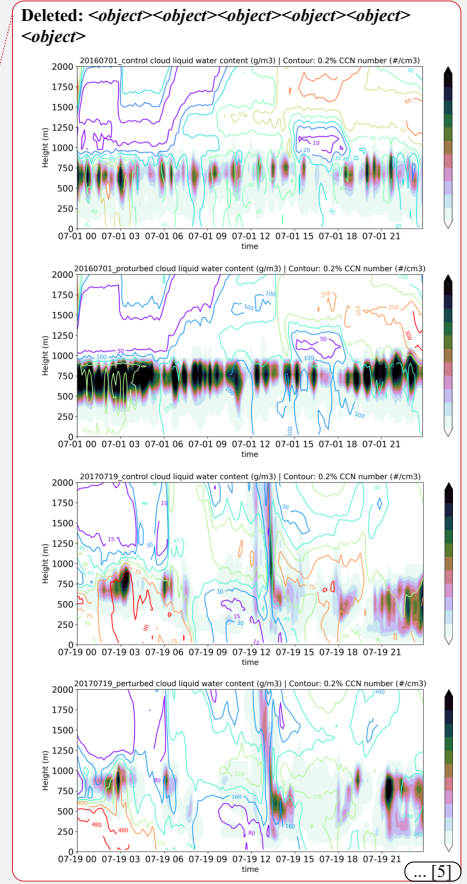


Figure 8. (a), (c), and (e) are the time series (local time: UTC – 1 hour) of 4 km-averaged cloud liquid water content profile (shade; units: g cm^{-3}) and CCN (0.2% supersaturation) number concentration profile (contour; units: $\# \text{ cm}^{-3}$) on 1 July 2016, 19 July 2017, and 23 August 2019, respectively, in the control runs. (b), (d), and (f) are the same as (a), (c), and (e), respectively, but in the perturbed runs. The data are averaged from the model simulated results over 20×20 grids centered on the Azores (approximately 4 km resolution).



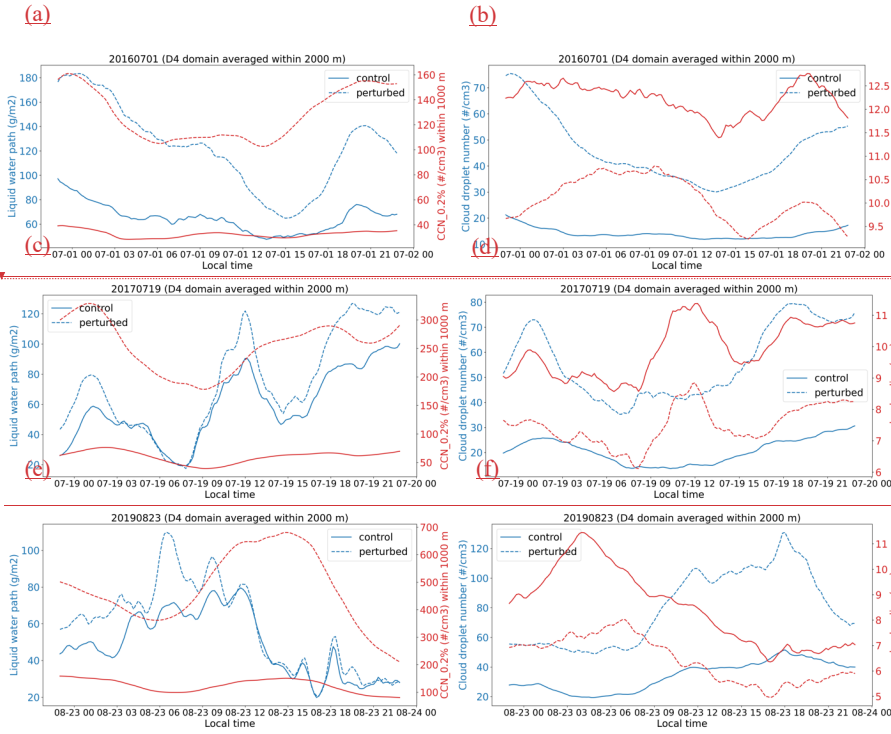
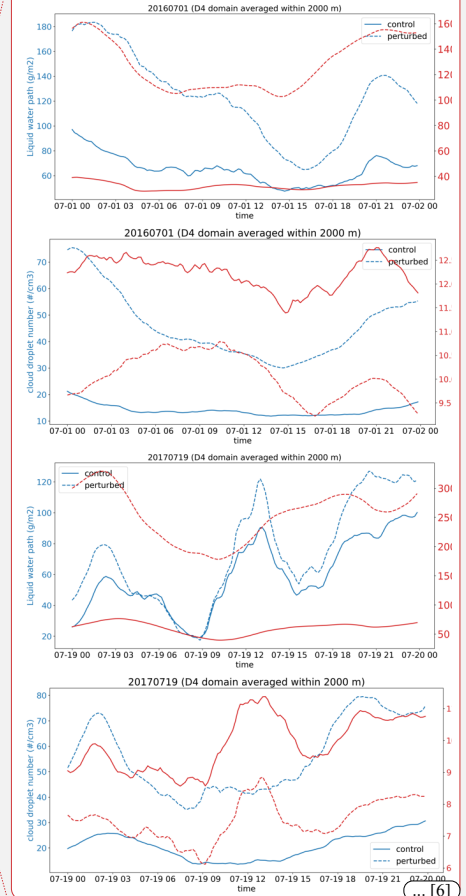


Figure 9. (a), (c), and (e) are the time series of domain-averaged liquid water path (blue lines; units: g m^{-2}) and CCN number concentration below 0.2% supersaturation (red lines; units: $\# \text{ cm}^{-3}$) for the control case (solid lines) and the perturbed case (dashed lines) on 1 July 2016, 19 July 2017, and 23 August 2019, respectively. (b), (d), and (f) are the time series of domain-averaged cloud droplet number (blue lines; units: $\# \text{ cm}^{-3}$) and cloud radius (red lines; units: μm) for the control (solid lines) and perturbed (dashed lines) on 1 July 2016, 19 July 2017, and 23 August 2019, respectively. Only CCN data are averaged within 1000 m height over domain 4, other variables are averaged within 2000 m height.

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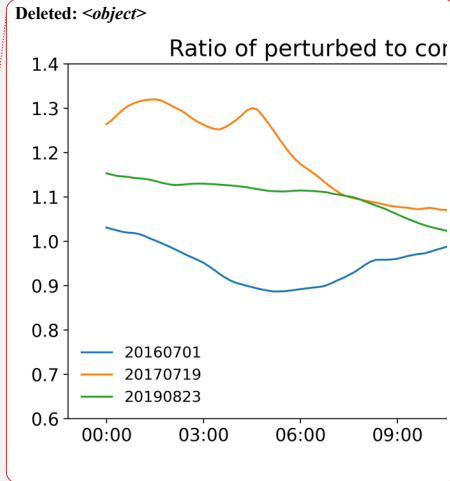
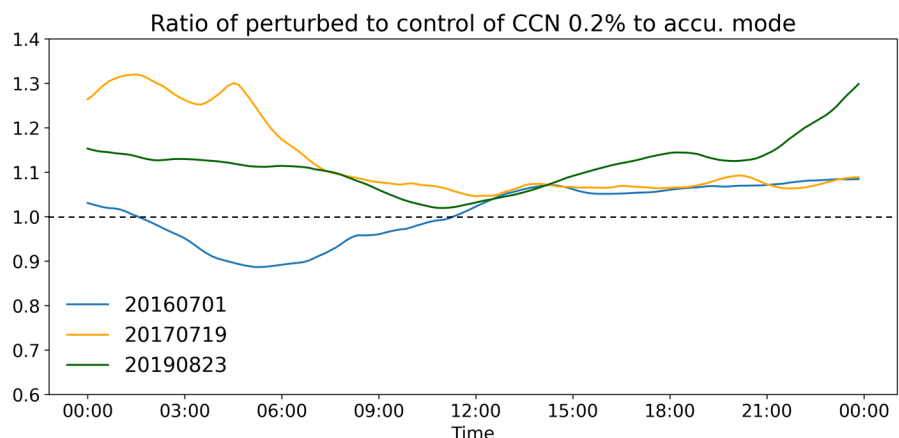
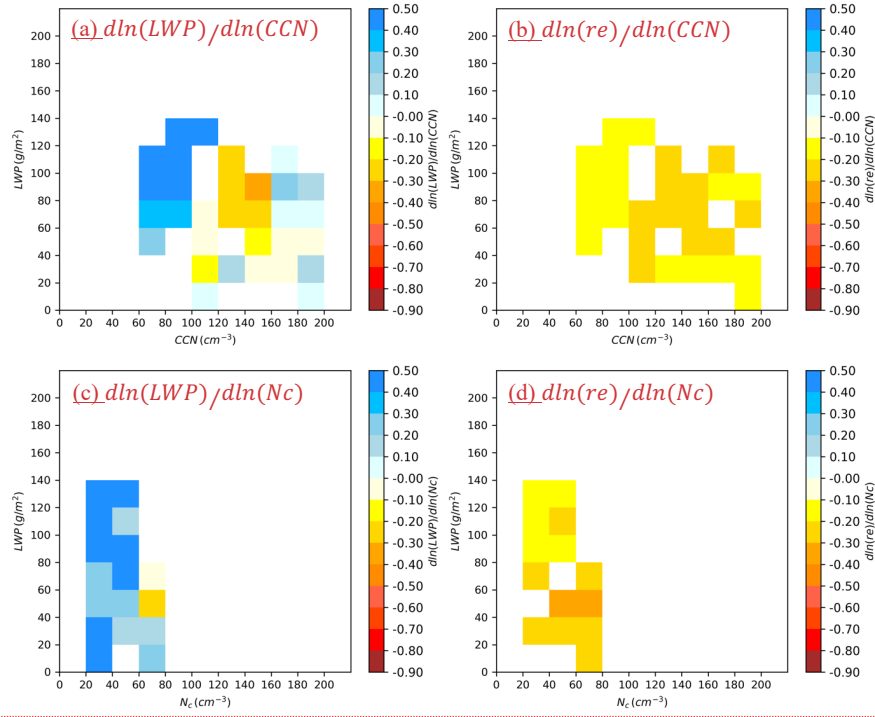


Figure 10. The time series of the ratio of the number concentration of CCN at a supersaturation of 0.2% in the perturbed runs to that in the control runs, normalized by the corresponding accumulation mode aerosol concentration, defined as $(CCN_{0.2\%}/Accu.aerosols)_{perturbed} / (CCN_{0.2\%}/Accu.aerosols)_{control}$. The black dashed line indicates the value of unity.

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1526 Figure 11. (a) and (b) are the mean liquid water path (LWP) and cloud radius (Re)
 1527 susceptibilities for different cloud condensation nuclei (CCN) and LWP bins for three study
 1528 cases, respectively. (c) and (d) are the same as (a) and (b), respectively, but for different cloud
 1529 droplet number (Nc) and LWP bins. The logarithmic slope between LWP and CCN, denoted as
 1530 $(d\ln(LWP)/d\ln(CCN))$, is calculated at each output time (every 10 minutes) using data from 16
 1531 aggregate grid points (~25 km for each grid point) from the control run and 16 aggregated grid
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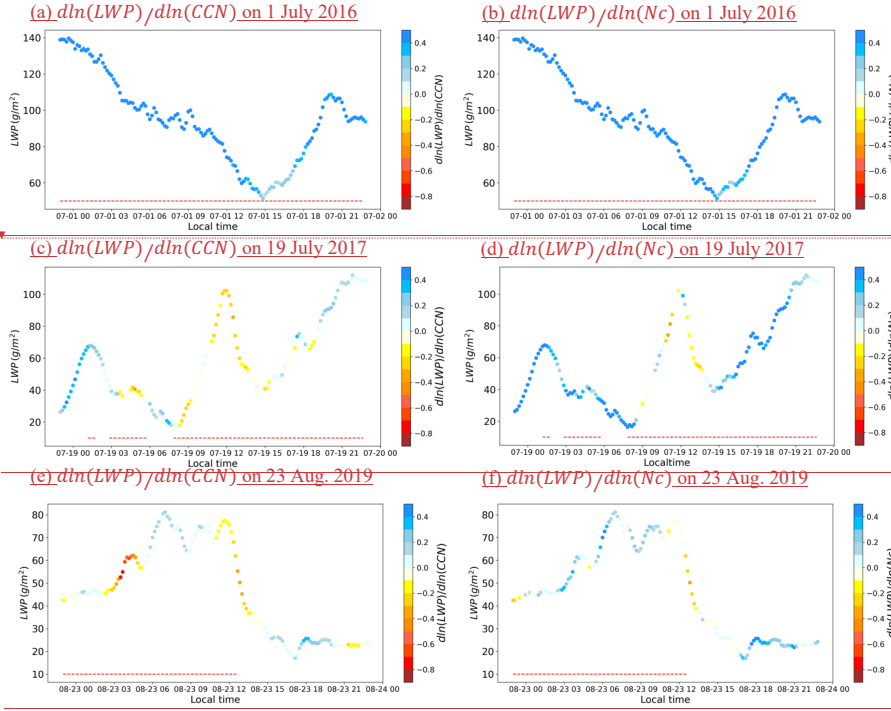
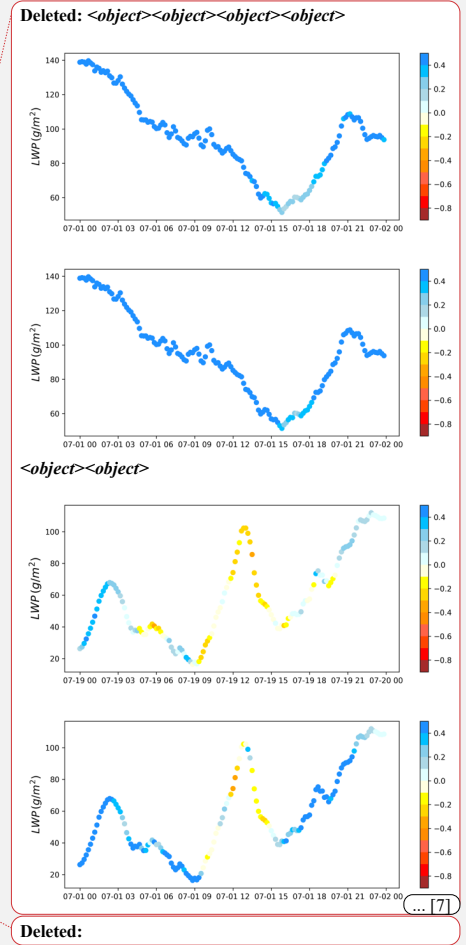
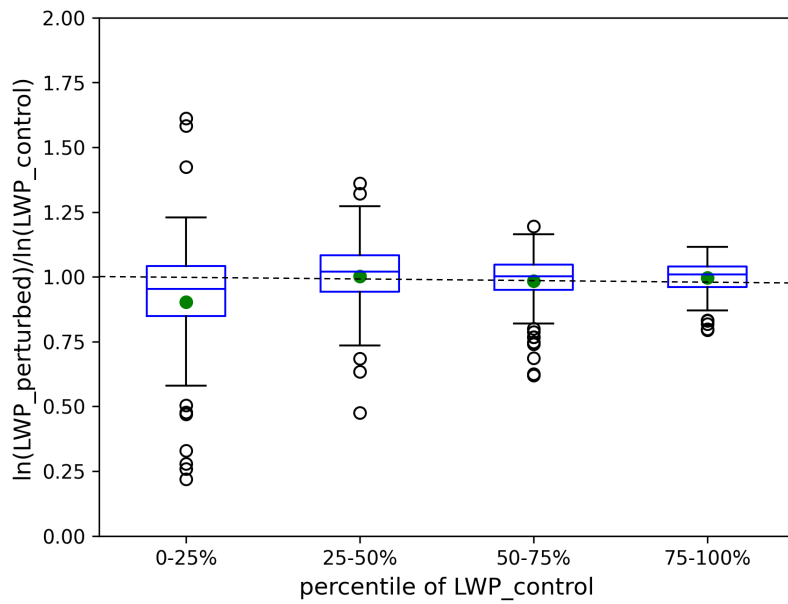


Figure 12. (a), (c) and (e) are the time variable of LWP susceptibility for different CCN concentration, denoted as $(d\ln(LWP)/d\ln(CCN))$, on 1 July 2016, 19 July 2017, and 23 August 2019, respectively. (b), (d) and (f) are the time variable of LWP susceptibility for different Nc concentration, denoted as $(d\ln(LWP)/d\ln(Nc))$, on 1 July 2016, 19 July 2017, and 23 August 2019, respectively. The logarithmic slope between LWP and CCN is calculated at each output time (every 10 minutes) using data from 16 aggregate grid points (~25 km for each grid point) from the control run and 16 aggregated grid points from the perturbed run. The red dashed lines indicate when rainfall intensity is higher than $0.001 \text{ mm } 10\text{-min}^{-1}$ in domain 4.





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Figure 13. **Boxplot** of the relative change in $\ln(LWP)$ between the perturbed and control cases across different LWP **Percentile** ranges in the control case during the negative susceptibility for LWP shown in Fig. 12. The box extends from the first quartile to the third quartile of the data, with a line at the median. The whiskers extend from the box to the farthest data point lying within 1.5x the inter-quartile range from the box. **Outlier values** are those past the end of the whiskers. Green dots are the mean value, and the back dashed line indicates the value of 1.0.

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