1 **Examination of aerosol impacts on convective clouds and precipitation in two**
2 **metropolitan areas in East Asia; how varying depths of convective clouds between**
3 **the areas diversify those aerosol effects?**

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Abstract

This study examines the role played by aerosols which act as cloud condensation nuclei (CCN) in the development of clouds and precipitation in two metropolitan areas in East Asia that have experienced substantial increases in aerosol concentrations over the last decades. These two areas are the Seoul and Beijing areas and the examination has been done by performing simulations using the Advanced Research Weather Research and Forecasting model as a cloud-system resolving model. CCN are advected from the continent to the Seoul area and this increases aerosol concentrations in the Seoul area. These increased CCN concentrations induce the enhancement of condensation that in turn induces the enhancement of deposition and precipitation amount in a system of less deep convective clouds as compared to those in the Beijing area. In a system of deeper clouds in the Beijing area, increasing CCN concentrations also enhance condensation but reduce deposition. This leads to CCN-induced negligible changes in precipitation amount. Also, in the system, there is a competition for convective energy among clouds with different condensation and updrafts. This competition results in different responses to increasing CCN concentrations among different types of precipitation, which are light, medium and heavy precipitation in the Beijing area. CCN-induced changes in freezing play a negligible role in CCN-precipitation interactions as compared to the role played by CCN-induced changes in condensation and deposition in both of the areas.

82   **1. Introduction**

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84   With increasing aerosol loading or concentrations, cloud-particle sizes can be changed. In

85   general, with increasing droplet sizes, the efficiency of collision and collection among

86   droplets increases. Increasing aerosol loading is known to make the droplet size smaller

87   and thus make the efficiency of collision and collection among droplets lower. This leads

88   to less droplets or cloud liquid forming raindrops and there is more cloud liquid present in

89   the air to be evaporated or frozen. Studies have shown that increases in cloud-liquid mass

90   due to increasing aerosol loading can enhance the freezing of cloud liquid and parcel

91   buoyancy, which lead to the invigoration of convection (Rosenfeld et al., 2008; Fan et al.,

92   2009). Via the invigoration of convection, precipitation can be enhanced. The dependence

93   of aerosol-induced invigoration of convection and precipitation enhancement on aerosol-

94   induced increases in condensational heating in the warm sector of a cloud system has been

95   shown (e.g., van den Heever et al., 2006; Fan et al., 2009; Lee et al., 2018). Increasing

96   cloud-liquid mass induces increasing evaporation, which intensifies gust fronts. This in

97   turn strengthens convective clouds and increases the amount of precipitation (Khain et al.,

98   2005; Tao et al., 2007; Storer et al., 2010; Tao et al., 2012; Lee et al., 2017; Lee et al.,

99   2018). It is notable that aerosol-induced precipitation enhancement is strongly sensitive to

100   cloud types that can be defined by cloud characteristics such as cloud depth (e.g., Tao et

101   al., 2007; Lee et al., 2008; Fan et al., 2009).

102   Since East Asia was industrialized, there have been substantial increases in aerosol

103   concentrations over the last decades in East Asia (e.g., Lee et al., 2013; Lu et al., 2011; Oh

104   et al., 2015; Dong et al., 2019). These increases are far greater than those in other regions

105   such as North America and Europe (e.g., Lu et al., 2011; Dong et al., 2019). While those

106   increasing aerosols affect clouds, precipitation and hydrologic circulations in the

107   continental East Asia, the increase in the advected aerosols from the continent to the

108   Korean Peninsula affect clouds, precipitation and hydrologic circulations in the Korean

109   Peninsula (Kar et al., 2009). This study aims to examine effects of the increasing aerosols,

110   which particularly act as cloud condensation nuclei (CCN), and their advection on clouds

111   and precipitation in East Asia. This study focuses on aerosols which act as CCN, but not

112   ice-nucleating particles (INPs), to examine those effects, based on the fact that CCN

account for most of aerosol mass that affects clouds and precipitation, and CCN, but not INPs, are associated with above-described aerosol-induced invigoration of convection and intensification of gust fronts. Note that these aerosol-induced invigoration and intensification are two well-established major theories of aerosol-cloud interactions. As a first step to the examination, this study focuses on two metropolitan areas in East Asia which are the Beijing and Seoul areas. The population of each of the Beijing and Seoul areas is ~ 20 millions. Associated with this, these areas have lots of aerosol sources (e.g., traffic) and have made a substantial contribution to the increases in aerosol concentrations in East Asia. Hence, we believe that these two cities can represent overall situations related to increasing aerosol concentrations in East Asia.

As mentioned above, aerosol-cloud interactions (and their impacts on precipitation) are strongly dependent on cloud types and thus to gain a more general understanding of those interactions, we select cases from the Beijing and Seoul areas with different cloud types. A selected case from the Beijing area involves deep convective clouds that reach the tropopause, while a selected case from the Seoul area involves comparatively shallow (or less deep) convective clouds. Via comparisons between these two cases, we aim to identify mechanisms that control varying aerosol-cloud interactions with cloud types.

To examine impacts of aerosols, which act as CCN, on clouds and precipitation in the cases, numerical simulations are performed, as a way of fulfilling above-described aim. These simulations use a cloud-system resolving model (CSRM) that has reasonably high resolutions to resolve cloud-scale processes that are related to cloud microphysics and dynamics. Hence, these simulations are able to find process-level mechanisms in association with cloud-scale processes.

2. Case description

In the Seoul area, South Korea, there is an observed mesoscale convective system (MCS) for a period from 03:00 LST (local solar time) to 18:00 LST December 24th 2017. During this period, there is a recorded moderate amount of precipitation and its maximum precipitation rate reaches ~ 13 mm hr^{-1}. Here, precipitation in the Seoul area is measured by rain gauges in automatic weather stations (AWSs) (King, 2009). The measurement is

144   performed hourly with a spatial resolution that ranges from ~1 km to ~10 km. The Seoul

145   area is marked by an inner rectangle in Figure 1a and dots in the rectangle in Figure 1a

146   mark the selected locations of rain gauges. At 21:00 LST December 23rd 2017, synoptic-

147   scale features develop in favor of the formation and development of the selected MCS and

148   associated moderate rainfall. These features involve the southwesterly low-level jets that

149   transport warm and moist air to the Korean Peninsula. The southwesterly low-level jet

150   plays an important role in the formation and development of rainfall events in the Korean

151   Peninsula by fetching warm and moist air (Hwang and Lee 1993; Lee et al. 1998; Seo et

152   al. 2013; Oh et al. 2018).

153   There was another observed MCS case in the Beijing area, China for a period from

154   14:00 LST on July 27th to 00:00 LST July 28th 2015. There is a substantial recorded amount

155   of precipitation for this period and its maximum precipitation rate reaches ~ 45 mm hr^{-1}.

156   Here, similar to the situation in the Seoul area, precipitation in the Beijing area is measured

157   by rain gauges in AWSs hourly with a spatial resolution that ranges from ~1 km to ~10 km.

158   The Beijing area is marked by an inner rectangle in Figure 1b and dots in the rectangle in

159   Figure 1b mark the selected locations of rain gauges. At 09:00 LST July 27th 2015,

160   synoptic-scale features develop in favor of the formation and development of the selected

161   MCS. These features involve the southerly low-level jet that develops heavy rainfall events

162   in the Beijing area by transporting warm and moist air to the area. Synoptic features which

163   are described here are based on reanalysis data that are produced by the Met Office Unified

164   Model (Brown et al., 2012) every 6 hours with a 0.11° × 0.11° resolution.

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166   **3. CSRM and simulations**

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168   **3.1 CSRM**

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170   The Advanced Research Weather Research and Forecasting (ARW) model (version 3.3.1)

171   is used as a CSRM. The ARW model is a compressible model with a nonhydrostatic status.

172   A 5th-order monotonic advection scheme is used to advect microphysical variables (Wang

173   et al., 2009). The Rapid Radiation Transfer Model (RRTMG; Mlawer et al., 1997; Fouquart

174   and Bonnel, 1980) is adopted to parameterize shortwave and longwave radiation in

175   simulations. A microphysics scheme that is used in this study calculates the effective sizes

176   of hydrometeors that are fed into the RRTMG, and the RRTMG simulates how these

177   effective sizes affect radiation.

178   The CSRM adopts a bin scheme as a way of parameterizing microphysics. The

179   Hebrew University Cloud Model (HUCM) detailed in Khain et al. (2011) is the bin scheme.

180   A set of kinetic equations is solved by the bin scheme to represent a size distribution

181   function for each of seven classes of hydrometeors and aerosols acting as CCN. Hence,

182   there are seven size distribution functions for hydrometeors. The seven classes of

183   hydrometeors are water drops, three types of ice crystals, which are plates, columns and

184   dendrites, snow aggregates, graupel and hail. Drops whose radius is smaller (larger) than

185   40 µm are categorized to be droplets (raindrops). There are 33 bins for each size

186   distribution in a way that the mass of a particle m_j in the j bin is to be $m_j = 2m_{j-1}$.

187   A cloud-droplet nucleation parameterization based on Köhler theory represents cloud-

188   droplet nucleation. Arbitrary aerosol mixing states and aerosol size distributions can be fed

189   to this parameterization. To represent heterogeneous ice-crystal nucleation,

190   parameterizations by Lohmann and Diehl (2006) and Möhler et al. (2006) are used. In these

191   parameterizations, contact, immersion, condensation-freezing, and deposition nucleation

192   paths are all considered by taking into account the size distribution of INPs, temperature

193   and supersaturation. Homogeneous droplet freezing is

194   considered following the theory developed by Koop et al. (2000).

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196   **3.2 Control runs**

197

198   For a three-dimensional CSRM simulation of the observed case of convective clouds in the

199   Seoul (Beijing) area, i.e., the control-s (control-b) run, a domain just over the Seoul

200   (Beijing) area, which is shown in Figure 1a (1b), is used. This domain adopts a 300-m

201   resolution. The control-s run is for a period from 03:00 LST to18:00 LST December 24th

202   2017, while the control-b run is for a period from 14:00 LST on July 27th to 00:00 LST

203   July 28th 2015. The length of the domain is 170 (140) km in the east-west (north-south)

204   direction for the control-s run, and 280 (240) km for the control-b run. There are 100

205   vertical layers and these layers employ a sigma coordinate that follows the terrain. The top

206　pressure of the model is 50 hPa for both of the control-s and control-b runs. On average,

207　the vertical resolution is ~200 m.

208　　Reanalysis data, which are produced by the Met Office Unified Model (Brown et al.,

209　2012), represent the synoptic-scale features, provide initial and boundary conditions of

210　variables such as wind, potential temperature, and specific humidity for the simulations.

211　The simulations adopt an open lateral boundary condition. The Noah land surface model

212　(LSM; Chen and Dudhia, 2001) calculates surface heat fluxes.

213　　The current version of the ARW model is not able to consider the spatiotemporal

214　variation of aerosol properties. In order to take into account the spatiotemporal variation of

215　aerosol properties, which is typical in metropolitan areas, such as composition and number

216　concentration, an aerosol preprocessor, which is able to consider the variability of aerosol

217　properties, is developed and used in the simulations. This aerosol preprocessor interpolates

218　or extrapolates background aerosol properties in observation data such as aerosol mass

219　(e.g., $PM_{2.5}$ and PM_{10}) into grid points and time steps in the model. In this study, the inverse

220　distance weighting method is used for the extrapolation and interpolation of observation

221　data including aerosol mass into grid points and time steps in the model. PM stands for

222　particulate matter. The mass of aerosols with diameter smaller than 2.5 (10.0) μm per unit

223　volume of the air is $PM_{2.5}$ (PM_{10}).

224　　There are surface observation sites, which measure aerosol properties, in the domains

225　and these sites are classified into two types; the selected locations of these sites are marked

226　by dots in the inner rectangles in Figure 1. The distance between the observation sites

227　ranges from ~1 km to ~10 km and the time interval between observations is ~10 minutes.

228　More than 90% of the sites belong to the first type of the sites. These first-type sites are

229　managed by the government in South Korea or China, and measure $PM_{2.5}$ or PM_{10} but not

230　other aerosol properties such as aerosol composition and size distributions. Less than 10%

231　of the sites belong to the second type of the sites. These second-type sites are a part of

232　aerosol robotic network (AERONET; Holben et al., 2001) and measure aerosol

233　composition and size distributions. The production of aerosol data in these second-type or

234　AERONET sites is viable only in the presence of the sun. The first-type sites observe $PM_{2.5}$

235　or PM_{10} using the beta-ray attenuation method (Eun et al., 2016; Ha et al., 2019) and hence,

236　produce $PM_{2.5}$ or PM_{10} data whether the sun is present or not. $PM_{2.5}/PM_{10}$ data from the

237   first-type sites are used to represent the spatiotemporal variability of aerosols over the

238   domains and the simulation periods. To represent aerosol composition and size

239   distributions, data from the AERONET sites are employed.

240   The AERONET data are averaged over the AERONET sites at 02:00 LST December

241   24th 2017 (13:00 LST July 27th 2015), which is 1 hour before the observed MCS forms, for

242   the Seoul (Beijing) case. Based on the average data, it is assumed that aerosol particles are

243   internally mixed with 70 (80) % ammonium sulfate and 30 (20) % organic compound for

244   the Seoul (Beijing) case. This mixture is assumed to represent aerosol chemical

245   composition in the whole domain and during the entire simulation period. Since ammonium

246   sulfate and organic compound are representative components of CCN, it is assumed that

247    $PM_{2.5}$ and PM_{10}, which are from the first-type sites, represent the mass of aerosols that act

248   as CCN for the Seoul and Beijing areas, respectively. Aerosols reflect, scatter and absorb

249   shortwave and longwave radiation before they are activated. This type of aerosol-radiation

250   interactions is not taken into account in this study. This is mainly based on the fact that in

251   the mixture, there is insignificant amount of radiation absorbers; black carbon is a

252   representative radiation absorber. The average AERONET data indicate that the size

253   distribution of background aerosols acting as CCN follows the bi-modal log-normal

254   distribution for both of the Seoul and Beijing cases. Based on the average AERONET data,

255   it is assumed that for the whole domain and simulation period, the size distribution of

256   background aerosols acting as CCN follows a shape of distribution with specific size

257   distribution parameters (i.e., modal radius and standard deviation of each of accumulation

258   and coarse modes, and the partition of aerosol number among those modes) for each of the

259   cases. Modal radius of the shape of distribution is 0.110 (0.085) and 1.413 (1.523) μm,

260   while standard deviation of the shape of distribution is 1.54 (1.63) and 1.75 (1.73) for

261   accumulation and coarse modes, respectively, in the Seoul (Beijing) case. The partition of

262   aerosol number, which is normalized by the total aerosol number of the size distribution,

263   is 0.999 and 0.001 for accumulation and coarse modes, respectively, in both of the cases.

264   By using $PM_{2.5}$ or PM_{10}, which is not only from the first-type sites but also interpolated

265   and extrapolated to grid points immediately above the surface and time steps, and based on

266   the assumption of aerosol composition and size distribution above, which is in turn based

267   on data from the AERONET sites, the background number concentrations of aerosols

268 acting as CCN are obtained for the simulation for each of the cases. There is no variation
269 with height in background concentrations of aerosols acting as CCN from immediately
270 above the surface to the top of the planetary boundary layer (PBL). However, it is assumed
271 that they decrease exponentially with height from the PBL top upward. With this
272 exponential decrease, when the altitude reaches the tropopause, background concentrations
273 of aerosols acting as CCN reduce by a factor of ~10 as compared to those at the PBL top.
274 The size distribution and composition of aerosols acting as CCN do not vary with height.
275 Once background aerosol properties (i.e., aerosol number concentrations, size distribution
276 and composition) are put into each grid point and time step, those properties at each grid
277 point and time step do not change during the course of the simulations.

278   For the control-s and control-b runs, aerosol properties of INPs are not different from
279 those of CCN except for the fact that the concentration of background aerosols acting as
280 CCN is 100 times higher than the concentration of background aerosols acting as INPs at
281 each time step and grid point, following a general difference between CCN and INPs in
282 terms of their concentrations (Pruppacher and Klett, 1978).

283   Once clouds form and background aerosols start to be in clouds, those aerosols are
284 not background aerosols anymore and the size distribution and concentrations of those
285 aerosols begin to evolve through aerosol sinks and sources that include advection and
286 aerosol activation (Fan et al., 2009). For example, once aerosols are activated, they are
287 removed from the corresponding bins of the aerosol spectra. In clouds, after aerosol
288 activation, aerosol mass starts to be inside hydrometeors and via collision-collection, it
289 transfers to different types and sizes of hydrometeors. In the end, aerosol mass disappears
290 in the atmosphere when hydrometeors with aerosol mass touches the surface. In non-cloudy
291 areas, aerosol size and spatial distributions are designed to be identical to the size and
292 spatial distributions of background aerosols, respectively. In other words, for this study,
293 we use "the aerosol recovery method". In this method, at any grid points, immediately after
294 clouds disappear entirely, aerosol size distributions and number concentrations recover to
295 background properties that background aerosols at those points have before those points
296 are included in clouds. In this way, we can keep concentrations of background aerosols
297 outside clouds in the simulations at observed counterparts. This enables spatiotemporal
298 distributions of background aerosols in the simulations to mimic those distributions that

299 are observed and particularly associated with observed aerosol advection in reality. In the
300 aerosol recovery method, there is no time interval between the cloud disappearance and the
301 aerosol recovery. Here, when the sum of mass of all types of hydrometeors (i.e., water
302 drops, ice crystals, snow aggregates, graupel and hail) is not zero at a grid point, that grid
303 point is considered to be in clouds. When this sum becomes zero, clouds are considered to
304 disappear. Many studies using CSRM have employed this aerosol recovery method. They
305 have proven that with the recovery method, reasonable simulations of overall cloud and
306 precipitation properties are accomplished (e.g., Morrison and Grabowski, 2011; Lebo and
307 Morrison, 2014; Lee et al., 2016; Lee et al., 2018).

308

309 **3.3 Additional runs**

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311 We repeat the control-s run by getting rid of aerosol-advection induced increases in
312 concentrations of aerosols acting as CCN as a way of investigating how the aerosol
313 advection affects the cloud system in the Seoul area. This repeated run is named the low-
314 aerosol-s run. An aerosol layer, which is advected from East Asia or from the west of the
315 Seoul area to it, increases aerosol concentrations in the Seoul area. There are stations in
316 islands in the Yellow Sea that monitor the aerosol advection (Eun et al., 2016; Ha et al.,
317 2019). To monitor and identify the aerosol advection, $PM_{2.5}$ which is measured by a station
318 in Baekryongdo island in Yellow Sea are compared to those which are measured in stations
319 in and around the Seoul area. In Figure 1a, a dot outside the inner rectangle marks the
320 island. The time evolution of $PM_{2.5}$ measured by the station on the island and the average
321 $PM_{2.5}$ over stations in the Seoul area, between 07:00 LST on December 22nd and 21:00 LST
322 on December 24th in 2017 when there is the strong advection of aerosols from East Asia to
323 the Seoul area, is shown in Figure 2. At 09:00 LST on December 22nd, the advection of
324 aerosols from East Asia enables aerosol mass to start going up and attain its peak around
325 05:00 LST on December 23rd on the island. Following this, aerosol mass starts to increase
326 in the Seoul area around 01:00 LST on December 23rd, and the mass attains its peak at
327 15:00 LST on December 23rd in the Seoul area. This is because aerosols, which are
328 advected from East Asia, move through the island to reach the Seoul area.

329   In the low-aerosol-s run, as a way of getting rid of aerosol-advection induced increases

330   in concentrations of aerosols acting as CCN, it is assumed that $PM_{2.5}$, which is assumed to

331   represent the mass of aerosols acting as CCN, and the associated background concentration

332   of aerosols acting as CCN after 01:00 LST on December 23rd do not evolve with the aerosol

333   advection in the Seoul area. Hence, the background concentration of aerosols acting as

334   CCN is assumed to have that at 01:00 LST on December 23rd at each time step and grid

335   point at the beginning of the simulation period. However, to isolate CCN effects on clouds,

336   background aerosol concentration acting as INPs at each time step and grid point in the

337   low-aerosol-s run is not different from that in the control-s run during the simulation period.

338   In the observed PM data for the Seoul area, there is reduction in PM by a factor of ~10 on

339   average over a period between ~07:00 and ~14:00 LST on December 24th, since

340   precipitation scavenges aerosols (Figure 2). To emulate this scavenging and reflect it in

341   background aerosols acting as CCN for the low-aerosol-s run, $PM_{2.5}$ and corresponding

342   background concentrations of aerosols acting as CCN at each grid point is gradually

343   reduced for the period between 07:00 and 14:00 LST on December 24th. This reduction is

344   done in a way that background concentrations of aerosols acting as CCN at each grid point

345   at 14:00 LST on December 24th is 10 times lower than that at 07:00 LST on December 24th

346   in the low-aerosol-s run. Then, $PM_{2.5}$ and corresponding background concentrations of

347   aerosols acting as CCN at each grid point at 14:00 LST on December 24th maintains until

348   the end of the simulation period. This results in the evolution of the average $PM_{2.5}$ over the

349   Seoul area in the low-aerosol-s run as shown in Figure 2. Here, the concentration of

350   background aerosols acting as CCN, which is averaged over the whole domain and

351   simulation period, in the control-s run is 3.1 times higher than that in the low-aerosol-s run.

352   Via comparisons between the runs, how the increasing concentration of background

353   aerosols acting as CCN due to the aerosol advection has an impact on clouds can be

354   examined. The concentration of background aerosols acting as CCN is different among

355   grid points and time steps in the control-s run. Hence, the ratio of the concentration of

356   background aerosols acting as CCN between the runs is different among grid points and

357   time steps.

358   For the Beijing case, to examine how aerosols acting as CCN affect clouds and

359   precipitation, we repeat the control-b run with simply reduced concentrations of

360   background aerosols acting as CCN at each time step and grid point by a factor of 3.1. This

361   repeated run is named the low-aerosol-b run. The 3.1-fold increase in aerosol

362   concentrations from the low-aerosol-b run to the control-b is based on the 3.1-fold increase

363   in the average concentration of background aerosols acting as CCN from the low-aerosol-

364   s run to the control-s run. However, as in the control-s and low-aerosol-s runs, to isolate

365   CCN effects on clouds, background aerosol concentration acting as INPs at each time step

366   and grid point in the low-aerosol-b run is identical to that in the control-b run during the

367   simulation period. Hence, on average, a pair of the control-s and low-aerosol-s runs has the

368   same perturbation of aerosols acting as CCN as in a pair of the control-b and low-aerosol-

369   b runs. Here, we define aerosol perturbation as a relative increase in aerosol concentration

370   when compared to that before the increase occurs. The brief summary of all simulations in

371   this study is given in Table 1.

372

373   **4. Results**

374

375   **4.1 Cumulative precipitation**

376

377   We compare the observed precipitation to the simulated counterpart in the control-s run for

378   the Seoul case and in the control-b run for the Beijing case. For this comparison, the

379   observed and simulated precipitation rates at the surface are averaged over the domain for

380   each of the Seoul and Beijing cases (Figures 3a and 3b). Here, the simulated precipitation

381   rates are smoothed over 1 hour. The comparison shows that the evolution of the simulated

382   precipitation rate does not deviate from the observed counterpart significantly (Figures 3a

383   and 3b).

384   In the Seoul case, overall, the precipitation rate is higher in the control-s run than in

385   the low-aerosol-s run. As a result of this, the domain-averaged cumulative precipitation

386   amount at the last time step is 14.1 mm and 12.0 mm in the control-s run and the low-

387   aerosol-s run, respectively. The control-s run shows ~20 % higher cumulative precipitation

388   amount. In the Beijing case, the evolution of the mean precipitation rate in the control-b

389   run is not significantly different from that in the low-aerosol-b run. Due to this, the control-

390   b run shows only ~2 % higher cumulative precipitation amount, despite the fact that the

391   concentrations of background aerosols acting as CCN are ~3 times higher in the control-b

392   run than in the low-aerosol-b run. Note that in the Seoul case, the time- and domain-

393   averaged concentration of background aerosols acting as CCN is also ~3 times higher in

394   the control-s run than in the low-aerosol-s run. Despite this, the difference in the cumulative

395   precipitation amount between the runs with different concentrations of background

396   aerosols acting as CCN is greater in the Seoul case than in the Beijing case.

397

398   **4.2 Precipitation, and associated latent-heat and dynamic processes**

399

400   Figures 4a and 4b show the cumulative frequency distributions of precipitation rates at the

401   last time step in the simulations for the Seoul and Beijing cases, respectively. In each of

402   those figures, the observed frequency distribution is shown and compared to the simulated

403   distribution. The observed distribution is obtained by interpolating and extrapolating the

404   observed precipitation rates to grid points and time steps in each of the control-s and

405   control-b runs. The observed maximum precipitation rates are 13.0 and 44.5 mm hr^{-1} for

406   the Seoul and Beijing cases, respectively, and these maximum rates are similar to those in

407   the control-s and control-b runs, respectively. Overall, the observed and simulated

408   frequency distributions are in good agreement for each of the cases. This enables us to

409   assume that results in the control-s (control-b) run are benchmark results to which results

410   in the low-aerosol-s (low-aerosol-b) run can be compared to identify how aerosols acting

411   as CCN have an impact on clouds and precipitation for the Seoul (Beijing) case. Here, it is

412   notable that for the Beijing case, while differences in the cumulative precipitation amount

413   between the control-b and low-aerosol-b runs are not significant, features in the frequency

414   distribution of precipitation rates between those runs are substantially different (Figure 4b).

415

416   **1) Seoul case**

417

418   **a.   Precipitation Frequency distributions**

419

420   Regarding precipitation whose rates are higher than ~2 mm hr^{-1}, the cumulative

421   precipitation frequency at the last time step is higher in the control-s run as compared to

422   that in the low-aerosol-s run (Figure 4a). In particular, for the precipitation rate of 11.4 mm

423   h^{-1}, there is an increase in the cumulative frequency by a factor of as much as ~10 in the

424   control-s run. When it comes to precipitation rates above 11.5 mm hr^{-1}, precipitation is

425   present in the control-s run and precipitation is absent in the low-aerosol-s run. Regarding

426   precipitation whose rates are lower than ~2 mm hr^{-1}, differences in the cumulative

427   frequency between the runs are insignificant. Hence, we see that there are significant

428   increases in the frequency of relatively heavy precipitation whose rates are above ~2 mm

429   hr^{-1} in the control-s run when compared to that in the low-aerosol-s run. At the last time

430   step, this results in a larger amount of cumulative precipitation in the control-s run than in

431   the low-aerosol-s run.

432   The time evolution of the cumulative precipitation frequency is shown in Figure 5. At

433   06:00 LST December 24th 2017, which corresponds to the initial stage of the precipitation

434   development, the maximum precipitation rate reaches ~3 mm hr^{-1} and there is the greater

435   frequency over most of precipitation rates in the control-s run than in the low-aerosol-s run

436   (Figure 5a).   With the time progress from 06:00 LST to 10:00 LST, the maximum

437   precipitation rate increases to reach 12 mm hr^{-1} and the cumulative frequency is higher over

438   precipitation whose rates are higher than ~3 mm hr^{-1} in the control-s run, while for

439   precipitation whose rates are lower than ~3 mm hr^{-1}, differences in the cumulative

440   frequency between the runs are negligible (Figures 5a and 5b). When time reaches 12:00

441   LST, which is around time when the peak in the evolution of the area-averaged

442   precipitation rates occurs and thus the system is at its mature stage, the maximum

443   precipitation rate increases up to ~13 mm hr^{-1} (Figures 3a and 5c). The basic patterns of

444   differences in the cumulative precipitation frequency between the runs with the maximum

445   precipitation rate around 13 mm hr^{-1}, which are established at 12:00 LST, maintain until

446   the end of the simulation period (Figures 4a and 5c).

447

448   **b. Condensation, deposition, updrafts and associated variables**

449

450   Note that the source of precipitation is precipitable hydrometeors which are raindrops,

451   snow, graupel and hail particles. Droplets and ice crystals are the source of those

452   precipitable hydrometeors mostly via collision and coalescence processes. Droplets and ice

453   crystals gain their mass mostly via condensation and deposition. Based on this, to explain

454   the greater cumulative precipitation amount in the control-s run than in the low-aerosol-s

455   run, the evolutions of differences in condensation, deposition and associated updrafts

456   between the runs are analyzed. The vertical profiles of differences in the area-averaged

457   condensation, deposition and freezing rates, updraft mass fluxes and the associated mass

458   density of each class of hydrometeors between the runs at 03:20, 03:40, 06:00 and 12:00

459   LST are shown in Figure 6. In Figure 6, differences in freezing rates are added for a more

460   comprehensive understanding of processes that are related to differences in cumulative

461   precipitation amount between the runs. Freezing includes riming processes between liquid

462   and solid hydrometeors and these riming processes act as a source of precipitable

463   hydrometeors. Cloud fractions are 0.32 (0.30), 0.85 (0.82), 0.93 (0.92) and 1.00 (1.00) in

464   the control-s (low-aerosol-s) run at 03:20, 03:40, 06:00 and 12:00 LST, respectively. We

465   see that cloud fraction varies 0-~6% between the runs. Note that in all of figures, which

466   display snow and hail mass density and include Figure 6, snow mass density includes ice-

467   crystal mass density, while hail mass density includes graupel mass density for the sake of

468   the display brevity. In Figure 6, horizontal black lines represent the altitudes of freezing

469   and melting.

470   Condensation rates in the control-s run start to be larger than that in the low-aerosol-

471   s run at 03:20 LST (Figure 6a). Higher aerosol or CCN concentrations induce more

472   nucleation of droplets, higher cloud droplet number concentration (CDNC) and associated

473   greater integrated surface of droplets in the control-s run. CDNC, which is averaged over

474   grid points and time steps with non-zero CDNC, is 1050 and 352 cm^{-3} in the control-s and

475   low-aerosol-s runs, respectively. Hence, more droplet surface is provided for water vapor

476   to condense onto in the control-s run. This leads to more condensation in the control-s run.

477   This establishes stronger feedbacks between updrafts and condensation, leading to greater

478   droplet (or cloud-liquid) mass at 03:20 LST in the control-s run (Figure 6a). Then, these

479   stronger feedbacks, which involve stronger updrafts particularly above 2 km in altitude,

480   subsequently induce greater deposition and snow mass as time progresses from 03:20 LST

481   to 03:40 LST, while more condensation and greater droplet mass maintain in the control-s

482   run with the time progress to 03:40 LST (Figure 6b). These stronger updrafts enable clouds

483   to grow higher in the control-s run. This eventually leads to a situation where the maximum

484   cloud depth is ~7 km in the control-s run and this depth is ~5 % deeper than that in the low-

485   aerosol-s run for the whole simulation period.

486   Through aerosol-induced stronger feedbacks between condensation, deposition and

487   updrafts in the control-s run, while more condensation and more overall deposition

488   maintain in the control-s run, differences in condensation and deposition between the

489   control-s and low-aerosol-s runs increase as time progresses from 03:40 LST to 06:00 LST

490   (Figures 6b and 6c). Associated with this, the greater mass of raindrops and hail particles

491   appears up, while the greater mass of droplets and snow in the control-s run than in the

492   low-aerosol-s run maintains with the time progress from 03:40 LST to 06:00 LST (Figure

493   6c). At 06:00 LST, there is more freezing starting to occur in the control-s run than in the

494   low-aerosol-s run. However, differences in freezing are ~one and ~two orders of magnitude

495   smaller than those in deposition and condensation, respectively. After 06:00 LST until time

496   reaches 12:00 LST when the overall differences in the cumulative precipitation frequency

497   between the runs are established, differences in freezing become ~3 times smaller than

498   those in deposition and ~one order of magnitude smaller than those in condensation

499   (Figures 6c and 6d). The greater mass of hydrometeors in the control-s run also continues

500   after 06:00 LST until time reaches 12:00 LST (Figures 6c and 6d). At 12:00 LST,

501   condensation, deposition and freezing rates are still higher in the control-s run. Here, we

502   see that CCN-induced more cumulative precipitation amount and associated differences in

503   the precipitation frequency distribution between the control-s and low-aerosol-s runs are

504   primarily associated with CCN-induced more condensation, which induce CCN-induced

505   more deposition and higher mass density of hydrometeors as sources of precipitation, but

506   weakly connected to CCN-induced changes in freezing. This is supported by the fact that

507   the time- and domain-averaged differences in freezing rate are ~one to ~two order of

508   magnitude smaller than those in condensation and deposition rates.

509

510   **c. Condensation frequency distributions and horizontal distributions of**

511   **condensation and precipitation**

512

513   Based on the importance of condensation for CCN-induced changes in precipitation, the

514   horizontal distribution of the column-averaged condensation rates over the domain and the

515   cumulative frequency distribution of the column-averaged condensation rates at each time

516   step is obtained. To better visualize the role of condensation in precipitation, the horizontal

517   distribution of the column-averaged condensation rates is superimposed on that of

518   precipitation rates (Figure 7). At 03:40 LST, condensation mainly occurs around the

519   northern part of the domain as marked by a yellow rectangle. The synoptic wind condition

520   in the marked area favors the collision between northward and southward wind and the

521   associated convergence around the surface (Figures 7a and 7b). This convergence induces

522   updrafts and condensation in the marked area. In the marked area, more aerosols acting as

523   CCN induce more and more extensive condensation, which leads to the higher domain-

524   averaged condensation rates in the control-s run than in the low-aerosol-s run (Figures 6b,

525   7a and 7b). More droplets are formed on more aerosols acting as CCN and more droplets

526   provide more surface areas where condensation occurs and this enables more and more

527   extensive condensation in the control-s run than in the low-aerosol-s run (Figures 6b, 7a

528   and 7b).

529   At 06:20 LST, a precipitating system is advected into the domain via the western

530   boundary, and as seen in Figures 7c and 7d for 08:40 LST, as time progresses to 08:40 LST,

531   the advected precipitating system is further advected to the east and extended mostly over

532   areas in the northern part of the domain where condensation mainly occurs. This confirms

533   that condensation is the main source of cloud mass and precipitation. In the eastern part of

534   the domain, there are mountains and in particular, higher mountains are on the northeastern

535   part of the domain than in the other parts of the domain. These higher mountains induce

536   forced convection and associated condensation more effectively in the northeastern part

537   than in the other parts. This is in favor of the precipitating system that extends further to

538   the east in the northern part of the domain. Due to more aerosols acting as CCN,

539   condensation, which is induced by forced convection over mountains, is more and more

540   extensive in the control-s run (Figures 7c and 7d). In association with this, there is more

541   extension of the precipitating system in the control-s run than in the low-aerosol-s run. This

542   enables the system in the control-s run to reach the eastern boundary at 08:40 LST, which

543   is earlier than in the low-aerosol-s run (Figures 7c and 7d). The system in the low-aerosol-

544   s run reaches the eastern boundary at 09:00 LST. Here, we see that although aerosols acting

545   as CCN do not change overall locations of the precipitation system, they affect how fast

546   the system extends to the east by affecting the amount of condensation which is produced

547   by forced convection. Associated with this, as seen in Figure 8, the control-s run has the

548   much higher cumulative condensation frequency than the low-aerosol-s run over all of

549   condensation rates during the period between 07:20 and 09:00 LST. Contributed by this,

550   the higher precipitation frequency over most of precipitation rates occurs in the control-s

551   run during and after the period (Supplementary Figures 1a and 1b and Figures 5b and 5c).

552   At 10:00 LST, in the southern part of the domain, there is a precipitating area forming

553   as marked by a yellow rectangle (Figures 7e and 7f). The precipitation area in the southern

554   part of the domain extends and merges into the advecting main precipitating system in the

555   northern part of the domain. The merge leads to precipitation that occupies most of the

556   domain at 12:00 LST (Figures 7g and 7h). After 10:00 LST, associated with this merge,

557   the maximum precipitation rate increases to 13 mm hr^{-1} at 12:00 LST (Figures 5c). After

558   13:00 LST, the precipitation enters its dissipating stage and its area reduces and nearly

559   disappears. Even after the merge, CCN-induced more condensation maintains and this in

560   turn contributes to a situation where the control-s run has the greater precipitation

561   frequency over most of precipitation rates than in the low-aerosol-s run until the

562   simulations progress to their last time step (Figures 4a, 5c and 6d).

563

564   **2) Beijing case**

565

566   Stronger convection and deeper clouds develop in the Beijing case than in the Seoul case.

567   The maximum cloud depth is ~7 and ~12 km in the control-s and control-b runs,

568   respectively. In the Seoul case, clouds do not reach the tropopause, while they reach the

569   tropopause in the Beijing case. Deeper clouds in the Beijing case produce the maximum

570   precipitation rate of ~45 mm hr^{-1} in the control-b run. However, less deep clouds in the

571   Seoul case produce the maximum precipitation rate of ~13 mm hr^{-1} in the control-s run

572   (Figure 4).

573

574   **a. Precipitation frequency distributions**

575

576   When it comes to precipitation whose rates are higher than ~12 mm hr^{-1}, the control-b run

577   has the higher cumulative precipitation frequency at the last time step than the low-aerosol-

578   b run (Figure 4b). Particularly, for the precipitation rates of 28.1 and 30.0 mm hr^{-1}, the

579   cumulative frequency increases by a factor of as much as ~10. Moreover, regarding

580   precipitation rates higher than ~33 mm hr^{-1}, precipitation is present in the control-b run,

581   however, precipitation is absent in the low-aerosol-b run. Hence, we see that the frequency

582   of comparatively heavy precipitation whose rates are higher than ~12 mm hr^{-1} rises

583   significantly in the control-b run as compared to that in the low-aerosol-b run. Below ~2

584   mm hr^{-1}, there is also the greater precipitation frequency in the control-b run than in the

585   low-aerosol-b run. Unlike the situation for precipitation rates above ~12 mm hr^{-1} and below

586   ~2 mm hr^{-1}, for precipitation rates from ~2 mm hr^{-1} to ~12 mm hr^{-1}, the control-aerosol-b

587   run has the lower precipitation frequency than in the low-aerosol-b run. Here, we see that

588   the higher precipitation frequency above ~12 mm hr^{-1} and below ~2 mm hr^{-1} balances out

589   the lower precipitation frequency between ~2 and ~12 mm hr^{-1} in the control-b run. This

590   results in the similar cumulative precipitation amount between the runs.

591   Figure 9 shows the time evolution of the cumulative precipitation frequency. When

592   precipitation starts around 16:00 LST, the higher precipitation frequency occurs over most

593   of precipitation rates in the low-aerosol-run-b run than in the control-b run (Figure 9a). At

594   16:00 LST, the maximum precipitation rate is lower than 1.0 mm hr^{-1} for both of the runs.

595   As time progresses to 17:00 LST, the maximum precipitation rate increases to ~17 mm hr^{-}

596   ^{1} and the higher (lower) cumulative precipitation frequency over precipitation rates higher

597   than ~12 mm hr^{-1} (between ~2 and ~12 mm hr^{-1}) in the control-b run than in the low-

598   aerosol-b run, which is described above as shown in Figure 4b for the last time step, starts

599   to emerge (Figure 9b). At 17:20 LST, the higher frequency for precipitation rates below 2

600   mm hr^{-1} in the control-b run, which is also described above as shown in Figure 4b for the

601   last time step, starts to show up, while the higher (lower) frequency for precipitation rates

602   higher than ~12 mm hr^{-1} (between ~2 and ~12 mm hr^{-1}) in the control-b run, which is

603   established at 17:00 LST, maintains as time progresses from 17:00 LST to 17:20 LST

604   (Figure 9c). At 17:20 LST, the maximum precipitation rate increases to 42 (19) mm hr^{-1} in

605   the control-b (low-aerosol-b) run (Figure 9c). At 19:00 LST, the maximum precipitation

606   rate increases to ~45 (33) mm hr^{-1} for the control-b (low-aerosol-b) run, while the

607   qualitative nature of differences in the precipitation frequency distributions with the tipping

608   precipitation rates of ~2 and ~12 mm hr^{-1} between the runs does not vary much between

609   17:20 and 19:00 LST (Figures 9c and 9d). The qualitative nature of differences in the

610   cumulative precipitation frequency between the runs and the maximum precipitation rates

611   in each of the runs, which are established at 19:00 LST, do not vary significantly until the

612   end of the simulation period (Figures 4b and 9d).

613

614   **b.   Condensation, deposition, updrafts and associated variables**

615

616   As done for the Seoul case, as a way of better understanding differences in the cumulative

617   precipitation amount and frequency between the control-b and low-aerosol-b runs, the

618   evolutions of differences in the vertical distributions of the area-averaged condensation

619   rates, deposition rates, freezing rates, the mass density of each class of hydrometeors and

620   updrafts mass fluxes are obtained and shown in Figures 10. Cloud fractions are 0.12 (0.11),

621   0.25 (0.22), 0.36 (0.32), 0.43 (0.40) and 0.48 (0.47) in the control-b (low-aerosol-b) run at

622   14:20, 15:40, 16:00, 17:20 and 19:00 LST, respectively. Here, we see that cloud fraction

623   varies by ~2-12% between the runs. In Figure 10, horizontal black lines represent the

624   altitudes of freezing and melting. As seen in Figure 3b, precipitation starts around 16:00

625   LST but differences in condensation rates start at 14:20 LST with higher condensation rates

626   in the control-b run (Figure 10a). Similar to the situation in the Seoul case, higher

627   concentrations of aerosols acting as CCN induce more nucleation of droplets, higher

628   CDNC and associated greater integrated surface of droplets in the control-b run. CDNC,

629   which is averaged over grid points and time steps with non-zero CDNC, is 992 and 341

630   cm^{-3} in the control-b and low-aerosol-b runs, respectively. Hence, more droplet surface is

631   provided for water vapor to condense onto in the control-b run. This leads to more

632   condensation in the control-b run. Due to this, cloud-liquid or droplet mass becomes greater

633   in the control-b run at 14:20 LST (Figure 10a). Increased condensation rates induce

634   increased condensational heating and thus intensified updrafts (Figure 10a). These

635   updrafts enable the maximum cloud depth to be ~ 12 km in the control-b run and this depth

636   is just ~1 % deeper than that in the low-aerosol-b run for the whole simulation period. This

637   negligible difference in the maximum cloud depth between the runs is due to the fact that

638 clouds with the maximum depth reach the tropopause in both of runs and thus there is not

639 much wiggle room to make significant differences in cloud depth between the runs.

640   When time reaches 15:40 LST, deposition rates and snow mass start to show

641 differences between the runs, while higher condensation rates and droplet mass maintain

642 in the control-b run with the time progress from 14:20 LST to 15:40 LST. However, unlike

643 the situation in the Seoul case, higher concentrations of aerosols acting as CCN result in

644 lower deposition rates and snow mass in the control-b run (Figure 10b). When time

645 progresses from 15:40 LST to 16:00 LST, differences in freezing start to occur and freezing

646 rates are lower (higher) at altitudes between ~6 and ~8 km (~4 and ~6 km), while higher

647 condensation rates and droplet mass, and lower snow mass maintain in the control-b run

648 (Figure 10c). Due to stronger updrafts, which are mainly ascribed to more condensation,

649 deposition rates start to be higher at altitudes between ~7 and ~9 km and freezing rates are

650 higher at altitudes between ~4 and ~6 km in the control-b run with the time progress from

651 15:40 LST to 16:00 LST (Figure 10c). Differences in freezing rates are similar to those in

652 deposition and ~two orders of magnitude smaller than those in condensation at 16:00 LST

653 (Figure 10c). At 16:00 LST, differences in hail mass between the runs appear up and hail

654 mass is slightly lower in the control-b run (Figure 10c). At 17:20 LST, overall, freezing

655 rates are lower at altitudes between ~4 and ~8 km, while overall, snow and hail mass is still

656 lower, and droplet mass is still higher in the control-b run (Figure 10d). Differences in

657 freezing rates are ~2 times smaller than those in deposition and ~one order of magnitude

658 smaller than those in condensation at 17:20 LST (Figure 10d). Due to more condensation

659 and droplet mass, greater raindrop mass appears up in the control-b run at 17:20 LST

660 (Figure 10d). As the time progresses to 19:00 LST, deposition rates become lower at the

661 altitudes from ~7 km to ~12 km and overall freezing rates become higher at altitudes from

662 ~ 4 km to ~10 km in the control-b run (Figure 10e). Overall, lower snow and hail mass

663 maintains in the control-b run as time progresses from 17:20 LST to 19:00 LST. As time

664 progresses from 17:20 LST to 19:00 LST, overall higher condensation rates, droplet and

665 raindrop mass maintain in the control-b run (Figure 10e). Here, while the time- and

666 domain-averaged deposition (condensation and freezing) rates are lower (higher) in the

667 control-b run over the whole simulation period, the average differences in freezing rates

668 are ~one to ~two orders of magnitude smaller than those in deposition and condensation

669   rates between the runs. Hence, more condensation (but not deposition and freezing) is a

670   main cause of stronger updrafts in the control-b run. More condensation and more freezing

671   tend to induce increases in the mass of precipitable hydrometeors in the control-b run. Less

672   deposition tends to induce decreases in the mass of precipitable hydrometeors in the

673   control-b run. This competition between condensation, deposition and freezing leads to

674   negligible differences in the cumulative precipitation amount at the last time step between

675   the control-b and low-aerosol-b runs, although roles of freezing in this competition are

676   negligible as compared to those of condensation and deposition.

677

678   **c. Condensation frequency distributions, horizontal distributions of**

679   **condensation and precipitation, and condensation-precipitation**

680   **correlations**

681

682   Figure 11 shows the horizontal distribution of the column-averaged condensation rates over

683   the domain and Figure 12 shows the cumulative frequency distributions of column-

684   averaged condensation rates at selected times. As in the Seoul case, the horizontal

685   distribution of condensation rates is superimposed on that of precipitation rates and the

686   terrain in Figure 11. At 14:20 LST, condensation starts to occur in places with mountains,

687   which induce forced convection, and condensation is concentrated around the center of the

688   domain as marked by a yellow circle (Figures 11a and 11b). Note that condensation does

689   not occur in the plain area which is the south of the 100-m terrain-height contour line

690   (Figures 11a and 11b). Due to higher concentrations of aerosols acting as CCN, there is

691   more condensation around the center in the control-b run than in the low-aerosol-b run

692   (Figures 11a and 11b). This leads to a situation where the control-b run has the higher area-

693   averaged condensation rates than the low-aerosol-b run (Figure 10a). Then, as time

694   progresses to 17:20 LST, the condensation area extends to the eastern and western parts of

695   the domain mostly over mountain areas (Figures 11c and 11d). Hence, the main source of

696   condensation is considered to be forced convection over mountains. As seen in Figures 11c

697   and 11d, higher concentrations of aerosols acting as CCN induce the control-b run to have

698   much more condensation spots and thus much bigger areas with condensation than the low-

699   aerosol-b run at 17:20 LST. Associated with this, CCN-induced more condensation in the

700   control-b run maintains with the time progress to 17:20 LST (Figure 10d). At 17:20 LST,

701   precipitation mainly occurs in a spot which is in the western part of areas with relatively

702   high condensation rates (Figures 11c and 11d).

703   At 17:20 LST, as seen in the cumulative frequency of condensation rates, the control-

704   b run has the higher condensation frequency above condensation rate of ~10 ×10^{-3} g m^{-3} s^{-}

705   ^{1} and below that of ~3×10^{-3} g m^{-3} s^{-1} than the low-aerosol-b run (Figure 12a). This pattern

706   of differences in the condensation frequency distribution with the tipping condensation-

707   rate points at ~10 ×10^{-3} and ~3 ×10^{-3} g m^{-2} s^{-1} continues up to 19:00 LST (Figures 12b).

708   Figure 13 shows the mean precipitation rate over each of the column-averaged

709   condensation rates for the period up to 17:20 LST in the control-b run. A column-averaged

710   condensation rate in an air column with a precipitation rate at its surface is obtained and

711   these condensation and precipitation rates are paired at each column and time step. Then,

712   collected precipitation rates are classified and grouped based on the corresponding paired

713   column-averaged condensation rates. The classified precipitation rates corresponding to

714   each of the column-averaged condensation rates are averaged arithmetically to construct

715   Figure 13. There are only less than 10 % differences in the mean precipitation rate for each

716   of the column-averaged condensation rates between the control-b and low-aerosol-b runs

717   (not shown). Figure 13 shows that generally a higher condensation rate is related to a higher

718   mean precipitation rate. It is also roughly shown that, according to the mean precipitation

719   rate for each condensation rate, overall, condensation rates below ~3 × 10^{-3} g m^{-3} s^{-1} and

720   above ~10 × 10^{-3} g m^{-3} s^{-1} are correlated with precipitation rates below ~2 mm hr^{-1} and

721   above ~12 mm hr^{-1}, respectively, while condensation rates between ~3 and ~10 × 10^{-3} g

722   m^{-3} s^{-1} are correlated with precipitation rates between ~2 and ~12 mm hr^{-1} (Figure 13).

723   Hence, on average, the higher frequency of condensation with rates above ~10 × 10^{-3} g m^{-}

724   ^{3} s^{-1} and below ~ 3 × 10^{-3} g m^{-3} s^{-1} can be considered to lead to the higher frequency of

725   precipitation whose rates are higher than ~12 mm hr^{-1} and lower than ~ 2 mm hr^{-1} in the

726   control-b run, respectively. It can also be considered that the lower condensation frequency

727   between ~3 and ~10 × 10^{-3} g m^{-3} s^{-1} leads to the lower precipitation frequency between ~2

728   and ~12 mm hr^{-1} in the control-b run. It is found that this correspondence between

729   condensation and precipitation rates is valid whether analyses to construct Figure 13 are

730   repeated only for a time point at 16:30 LST or for a period between 16:30 and 17:00 LST.

731 These time point and period are related to analyses of the moist static energy as described

732 in Section e below.

733   At 17:20 LST, the larger precipitation frequency between ~2 and ~12 mm hr^{-1} in the

734 low-aerosol-b run nearly offsets the larger precipitation frequency in the other ranges of

735 precipitation rates in the control-b run (Figure 9c). This leads to the similar average

736 precipitation rate between the runs at 17:20 LST and contributes to the similar cumulative

737 precipitation at the last time step between the runs (Figure 3b).

738

739   **d. Evaporation and gust fronts**

740

741 As time progresses from 17:00 to 19:00 LST, the precipitation system moves northward

742 (Figure 14). At the core of the precipitation system, due to evaporation and downdrafts,

743 there is the horizontal outflow forming at 17:00 LST (Figures 14a and 14b). The core is

744 represented by the field of precipitation whose rates are higher than 1 mm hr^{-1} in Figure 14.

745 At the core, the northward outflow is magnified by the northward synoptic-scale wind,

746 while at the core, the outflow in the other directions is offset by the northward synoptic-

747 scale wind. Hence, the outflow is mainly northward from 17:00 LST onwards as marked

748 by yellow circles in Figures 14. This enables convergence or a gust front, which is produced

749 by the outflow from the core, to be mainly formed at the north of the core. Note that the

750 intensity of a gust front is proportional to that of outflow from a core of precipitation or

751 convective system (Weisman and Klemp, 1982; Houze, 1993). The strong gust front at the

752 north of the core generates strong updrafts, a significant amount of condensation and

753 precipitation. Then, a subsequent area with clouds and precipitation is formed at the north

754 of the core as time progresses, which means that the precipitation system extends or moves

755 to the north as seen in comparisons between sub-panels with different times in Figure 14.

756 This movement, which is induced by collaborative work between outflow, synoptic wind

757 and gust fronts, is typical in deep convective clouds.

758   As described above, the more droplet nucleation and greater integrated droplet surface

759 induce more condensation before 17:00 LST in the control-b run. This and lower efficiency

760 of collision and collection among droplets enable the control-b run to have a larger amount

761 of cloud liquid or droplets as a source of evaporation. This in turn enables more droplet

762   evaporation, more associated cooling and stronger downdrafts, although less rain

763   evaporation is in the control-b run particularly for the period from 17:00 LST to 19:00 LST.

764   The time- and domain-averaged droplet and rain evaporation rates are 0.72 (0.31) and 0.08

765   (0.13) g m^{-3} h^{-1}, respectively, while the time- and domain-averaged downdraft mass flux is

766   0.15 (0.10) kg m^{-2} s^{-1} over the period from 17:00 LST to 19:00 LST in the control-b (low-

767   aerosol-b) run. More evaporation of droplets and associated stronger downdrafts with

768   higher concentrations of aerosols acting as CCN have been shown by the numerous

769   previous studies (e.g., Tao et al., 2007; Tao et al., 2012; Khain et al., 2008; Lee et al., 2018).

770   During the period between 17:00 and 19:00 LST, with the development of convergence

771   or the gust front, as mentioned above, the maximum precipitation rate increases from ~ 17

772   (17) to ~ 45 (33) mm hr^{-1} in the control-b (low-aerosol-b) run (Figure 9). This indicates

773   that the gust-front development contributes to the overall intensification of the precipitation

774   system, while it moves northward. If there were only northward synoptic-scale wind with

775   no formation of the gust front, the system would move northward with less intensification.

776   Over the period from 17:00 LST to 19:00 LST, stronger downdrafts and associated stronger

777   outflow generate a stronger gust front and more subsequent condensation in the control-b

778   run. This enhances the small initial difference, which is at 17:00 LST, in the frequency of

779   precipitation with rates above ~12 mm hr^{-1} between the runs substantially as time

780   progresses from 17:00 LST to 19:00 LST (Figure 9). Associated with this, with the time

781   progress, the nearly identical maximum precipitation rate between the runs at 17:00 LST

782   turns into the significantly higher maximum precipitation rate in the control-b run than in

783   the low-aerosol-b run (Figure 9). Around 19:00 LST, the system enters its dissipating stage,

784   accompanying reduction in the precipitating area and the area-averaged precipitation rate

785   (Figures 3b).

786

787   **e. Moist static energy**

788

789   Condensation, which controls droplet mass and precipitation, is controlled by updrafts and

790   updrafts are in turn controlled by instability. One of important factors that maintain

791   instability is the moist static energy. Motivated by this, to better understand differences in

792   the precipitation frequency distribution in association with those in the condensation

793   frequency distribution between the control-b and low-aerosol-b runs, we calculate the flux

794   of the moist static energy and the flux is defined as follows:

795

796    $\vec{Fs} = S \times \rho \times \vec{V}$ (1),

797

798   where $\vec{Fs}$ represents the flux of the moist static energy, S the moist static energy, ρ the air

799   density and $\vec{V}$ the horizontal-wind vector. In Eq. (1), we see that the flux is in the vector

800   form and has two components, which are its magnitude and direction. The fluxes of the

801   moist static energy in the PBL are obtained over the domain at 16:30 LST, since in general,

802   the moist static energy in the PBL has much stronger effects on instability and updrafts

803   than that above the PBL. In particular, we focus on the PBL fluxes of the energy that cross

804   the boundary over a time step at 16:30 LST between areas with the column-averaged

805   condensation rate from 3×10^{-3} g m^{-3} s^{-1} to 10×10^{-3} g m^{-3} s^{-1}, which are referred to as "area

806   A", and those with the column-averaged condensation rate above 10×10^{-3} g m^{-3} s^{-1}, which

807   are referred to as "area B". This is because we are interested in the exchange of the moist

808   static energy between areas A and B and this exchange can be seen by looking at those

809   fluxes which cross the boundary between those areas.

810   We are interested in the exchange of the energy, since we hypothesized that the

811   exchange somehow alters instability in each of areas A and B in a way that there are

812   increases (decreases) in instability, the updraft intensity, condensation and precipitation

813   with increasing concentrations of aerosols acting as CCN in area B (A), leading to the

814   higher (lower) frequency of condensation whose rates are higher than 10×10^{-3} g m^{-3} s^{-1}

815   (between 3×10^{-3} and 10×10^{-3} g m^{-3} s^{-1}) and precipitation whose rates are higher than 12

816   mm hr^{-1} (between 2 and 12 mm hr^{-1}) in the control-b run than in the low-aerosol-b run.

817   When the PBL fluxes, which crosses the boundary over the time step at 16:30 LST, are

818   summed at 16:30 LST, there is the net flux from area A to area B. This means that there is

819   the net transportation of the moist static energy from areas with condensation rates between

820   3×10^{-3} and 10×10^{-3} g m^{-3} s^{-1} to those with condensation rates greater than 10×10^{-3} g m^{-3}

821   s^{-1} in the PBL at 16:30 LST as shown in Table 2. Table 2 shows the net summed flux of the

822   moist static energy which crosses the boundary between areas A and B in the control-b run

823   as well as the low-aerosol-b run. To calculate the net flux at 16:30 LST in Table 2, the

824   fluxes, which cross the boundary between areas A and B over the time step at 16:30 LST,

825   only at grid points in the PBL are summed. For the calculation, the flux from area A to area

826   B has a positive sign, while the flux from area B to area A has a negative sign. Since the

827   net flux is positive for both of the runs as shown in Table 2, there is the net flux from area

828   A to area B in the PBL. The above-described analysis for the fluxes crossing the boundary

829   between areas A and B is repeated for every time step between 16:30 and 17:00 LST and

830   based on this, the net summed flux over the period between 16:30 and 17:00 LST is

831   obtained. As shown in Table 2, the net flux for the period between 16:30 and 17:00 LST is

832   also positive as in the situation only for 16:30 LST. This means that there is the net

833   transportation of the moist static energy from area A to area B in the PBL during the period

834   between 16:30 and 17:00 LST.

835   At 16:30 LST, condensation with rates above 10×10^{-3} g m^{-3} s^{-1} starts to develop and

836   this forms area B. Area B has stronger updrafts via greater condensational heating than in

837   other areas, including area A, with lower condensation rates. Stronger updrafts in area B

838   induce the convergence of air and associated moist static energy from area A to area B.

839   Since the average condensation rate and updrafts at 16:30 LST over area B are higher and

840   stronger due to increasing concentrations of aerosols acting as CCN, respectively, the air

841   convergence and the associated transportation of the moist static energy in the PBL from

842   area A to area B are stronger and more, respectively, in the control-b run than in the low-

843   aerosol-b run (Table 2). Stated differently, area B steals the moist static energy from area

844   A, and this occurs more effectively in the control-b run. This increases instability and

845   further intensifies updrafts in area B, and decreases instability and weakens updrafts in area

846   A, while these increases and decreases (intensification and weakening) of instability

847   (updrafts) are greater in the control-b run for the period from 16:30 LST to 17:00 LST.

848   This increases condensation, cloud mass and precipitation whose rates are higher than 12

849   mm hr^{-1} in area B, and decreases condensation, cloud mass and precipitation whose rates

850   are from 2 mm hr^{-1} to 12 mm hr^{-1} in area A. These increases and decreases occur more

851   effectively for the control-b run than for the low-aerosol-b run during the period. This in

852   turn leads to the lower precipitation frequency for the precipitation rates from 2 mm hr^{-1} to

853   12 mm hr^{-1} and the higher frequency for the precipitation whose rates are higher than 12

854   mm hr^{-1} at 17:00 LST in the control-b run (Figure 9b). The weakened updrafts and reduced

855   condensation turn a portion of precipitation with rates between 2 and 12 mm hr^{-1} to

856   precipitation whose rates are below 2 mm hr^{-1}, and this takes place more efficiently in the

857   control-b run during the period between 16:30 and 17:00 LST. This eventually increases

858   the frequency of precipitation rates below 2 mm hr^{-1} and this increase is greater for the

859   control-b run, leading to the greater precipitation frequency for the precipitation rates

860   below 2 mm hr^{-1} in the control-b run at 17:20 LST (Figure 9c).

861

862   **5. Discussion**

863

864   **5.1 Comparison of the Seoul and Beijing cases**

865

866   In this section, we compare the Seoul case to the Beijing case. For the comparison,

867   remember that on average, a pair of the control-s and low-aerosol-s runs has the same

868   perturbation of aerosols acting as CCN as in a pair of the control-b and low-aerosol-b runs.

869   Associated with the fact that clouds in the Seoul case are less deep than those in the Beijing

870   case, overall, updrafts in the Seoul case are not as strong as those in the Beijing case. Hence,

871   unlike the situation in the Beijing case, stronger updrafts, which accompany higher

872   condensation rates, and associated convergence in the Seoul case are not strong enough to

873   steal the sufficient amount of the moist static energy from weaker updrafts which

874   accompany lower condensation rates. This makes the redistribution of the moist static

875   energy between areas with relatively higher condensation rates and those with relative

876   lower condensation rates, such as that between areas A and B for the Beijing case,

877   ineffective for the Seoul case. Due to this, the sign of CCN-induced changes in the

878   frequency of precipitation rates does not vary throughout all of the precipitation rates

879   except for the range of low precipitation rates where there are nearly no CCN-induced

880   changes in the frequency in the Seoul case as shown in Figure 4a. As seen in Figure 4a,

881   mainly due to increases in condensation and deposition, precipitation frequency increases

882   for most of precipitation rates, although the precipitation frequency does not show

883   significant changes as concentration of aerosols acting as CCN increases for relatively low

884   precipitation rates in the control-s run as compared to that in the low-aerosol-s run. This

885   means that there are no tipping precipitation rates where the sign of CCN-induced changes

886   in the frequency of precipitation rates changes in the Seoul case, contributing to the higher

887   cumulative precipitation amount in the simulation with higher concentrations of aerosols

888   acting as CCN for the Seoul case, which are different from the situation in the Beijing case.

889   In the Beijing case with deeper clouds as compared to those in the Seoul case, clouds

890   develop gust fronts via strong downdrafts and associated strong outflow. These gust fronts

891   play an important role in developing strong convection and associated high precipitation

892   rates. Unlike the situation in the Seoul case, there are strong clouds and associated updraft

893   entities that are able to steal heat and moisture (or the moist static energy) as sources of

894   instability from areas with relatively less strong clouds and updrafts with medium strength;

895   note that these strong clouds here involve stronger updrafts via greater condensational

896   heating as described in Section e above and this enables these clouds to be thicker and have

897   higher cloud mass than these less strong clouds. This further intensifies strong clouds and

898   weakens less strong clouds with medium strength. Due to this, the cumulative frequency

899   of heavy (medium) precipitation in association with strong clouds (less strong clouds with

900   medium strength) increases (decreases). Some of the weakened clouds eventually produce

901   light precipitation, which increase the cumulative frequency for light precipitation. The

902   intensification of strong clouds and the weakening of less strong clouds with medium

903   strength gets more effective with increasing concentration of aerosols acting as CCN.

904   Hence, in the Beijing case, for medium precipitation in association with less strong clouds,

905   the simulation with higher concentration of aerosols acting as CCN shows the lower

906   cumulative precipitation frequency at the last time step. However, for heavy precipitation,

907   which is associated with strong clouds, and light precipitation, the simulation with higher

908   concentrations of aerosols acting as CCN shows the higher cumulative precipitation

909   frequency at the last time step. These differential responses of precipitation to increasing

910   concentration of aerosols acting as CCN among different types of precipitation occur in the

911   circumstances of the similar cumulative precipitation amount between the simulations with

912   different concentration of aerosols acting as CCN. This similar precipitation amount is due

913   to above-mentioned competition between CCN-induced changes in condensation,

914   deposition and freezing.

915   In both of the Seoul and Beijing cases, CCN-induced changes in condensation plays an

916   important role in making differences in the precipitation amount and/or the precipitation

frequency distribution between the simulations with different concentration of aerosols
acting as CCN. It is notable that in less deep clouds in the Seoul case, in addition to
condensation, deposition plays a role in precipitation to induce CCN-induced increases in
the precipitation amount. CCN-induced increases in condensation initiate the differences
in cloud mass and precipitation and then CCN-induced increases in deposition follow to
further enhance those differences. In deep clouds in the Beijing case, condensation tends
to induce increases in cloud mass and precipitation, while deposition tends to induce
decreases in cloud mass and precipitation with increasing concentration of aerosols acting
as CCN. Hence, as clouds get shallower and thus ice processes become less active, the role
of deposition in CCN-induced changes in precipitation amount turns from CCN-induced
suppression of precipitation to enhancement of precipitation. Here, we find that contrary
to the traditional understanding, the role of variation of freezing, which is induced by the
varying concentration of aerosols acting as CCN but not INPs, in precipitation is negligible
as compared to that of condensation and deposition in both of the cases.

6. Summary and conclusions

This study examines impacts of aerosols, which act as CCN, on clouds and precipitation in
two metropolitan areas, which are the Seoul and Beijing areas, in East Asia that has
experienced substantial increases in aerosol concentrations over the last decades. The
examination is performed via simulations, which use a CSRM. These simulations are for
deep clouds which reach the tropopause in the Beijing case and for comparatively less deep
clouds which do not reach the tropopause yet grow above the level of freezing in the Seoul
case.

In both of the cases, CCN-induced changes in condensation plays a critical role in
CCN-induced variation of precipitation properties (e.g., the precipitation amount and the
precipitation frequency distribution). In the Seoul case, CCN-induced increases in
condensation and subsequent increases in deposition lead to CCN-induced increases in the
precipitation frequency over most of precipitation rates and thus in the precipitation amount.
However, in the Beijing case, while there are increases in condensation with increasing
CCN concentrations, there are decreases in deposition with increasing CCN concentrations.

948   This competition between increases in condensation and decreases in deposition leads to

949   negligible CCN-induced changes in cumulative precipitation amount in the Beijing case.

950   In both of the cases, CCN-induced changes in freezing are negligible as compared to those

951   in condensation and deposition. In the Beijing case, there is another competition for the

952   moist static energy among clouds with different updrafts and condensation. This

953   competition results in CCN-induced differential changes in the precipitation frequency

954   distributions. With clouds getting deeper from the Seoul case to the Beijing case, clouds

955   and associated updrafts, which are strong enough to steal the moist static energy from other

956   clouds and their updrafts, appear. This makes strong clouds stronger and clouds with

957   medium strength weaker. With higher CCN concentrations, strong clouds steal more

958   energy, and thus strong clouds become stronger and clouds with medium strength weaker

959   with a greater magnitude. As a result of this, there are more frequent heavy precipitation

960   (whose rates are higher than 12 mm hr^{-1}) and light precipitation (whose rates are lower than

961   2 mm hr^{-1}), and less frequent medium precipitation (with rates from 2 mm hr^{-1} to 12 mm

962   hr^{-1}) with increasing CCN concentrations in the Beijing case.

963   In both of the Seoul and Beijing cases, there are mountains and they play an important

964   role in how cloud and precipitation evolve with time and space. In both of the cases, the

965   precipitating system moves or expands over mountains which induce forced convection

966   and generate condensation. This important role of mountains and forced convection in the

967   formation and evolution of the precipitation system has not been examined much in the

968   previous studies of aerosol-cloud interactions, since many of those previous studies (e.g.,

969   Jiang et al., 2006; Khain et al., 2008; Li et al., 2011; Morrison et al., 2011) have dealt with

970   convective clouds that develop over plains and oceans. Hence, findings in this study, which

971   are related to mountain-forced convection and its interactions with aerosols, can be

972   complementary to those previous studies. Stated differently, this study can shed light on

973   our path to the understanding of aerosol-cloud interactions over more general domains not

974   only with no terrain but also with terrain.

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979 **Code/Data source and availability**

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981 Our private computer system stores the code/data which are private and used in this study.

982 Note that in particular, the stored PM data are provided by the Korea Environment

983 Cooperation in South Korea and State Key Laboratory of Severe Weather in China. Upon

984 approval from funding sources, the data will be opened to the public. Projects related to

985 this paper have not been finished, thus, the sources prevent the data from being open to the

986 public currently. However, if information on the data is needed, contact the corresponding

987 author Seoung Soo Lee (slee1247@umd.edu).

988

989 **Author contributions**

990 Essential initiative ideas are provided by SSL, KJH and KHS to start this work. Simulation

991 and observation data are analyzed by SSL, JC and GK. JU and YZ review the results and

992 contribute to their improvement. CHJ and JG perform additional simulations, which are

993 required by the review process, and their basic analyses. CHJ, SKS, YGL and NU provide

994 ideas to handle the reviewers' comments.

995

996 **Competing interests**

997 The authors declare that they have no conflict of interest.

998

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FIGURE CAPTIONS

Figure 1. Inner rectangles in (a) and (b) mark the Seoul area in the Korean Peninsula and the Beijing area in the East-Asia continent, respectively. A dot outside the inner rectangle in (a) marks Baekryongdo island. Dots in the inner rectangles in (a) and (b) mark the selected locations where precipitation and aerosol mass are measured. In (a) and (b), the light blue represents the ocean and the green the land area.

Figure 2. Time series of $PM_{2.5}$ observed at the ground station in Baekryongdo island (blue line) and of the average $PM_{2.5}$ over ground stations in the Seoul area (red line) between 07:00 LST on December 22nd and 21:00 LST on December 24th in 2017. Note that $PM_{2.5}$ observed at stations in the Seoul area is applied to the control-s run whose period is marked by the dashed rectangle. Time series of the average $PM_{2.5}$ over stations in the Seoul area in the low-aerosol-s run for the simulation period is also shown (black solid line).

Figure 3. Time series of precipitation rates at the surface, which are averaged over the domain and smoothed over 1 hour, (a) for the control-s and low-aerosol-s runs in the Seoul area and (b) for the control-b and low-aerosol-b runs in the Beijing area. In (a) and (b), the averaged and observed precipitation rates over the observation sites in the Seoul and Beijing areas, respectively, are also shown.

Figure 4. Observed and simulated cumulative frequency distributions of precipitation rates at the surface for (a) the Seoul case, which are collected over the Seoul area, and (b) the Beijing case, which are collected over the Beijing area, at the last time step. Simulated distributions are in the control-s and low-aerosol-s runs for the Seoul case and in the control-b and low-aerosol-b runs for the Beijing case. The observed distribution is obtained by interpolating and extrapolating the observed precipitation rates to grid points and time steps in the control-s and control-b runs for the Seoul and Beijing cases, respectively.

1199 Figure 5. Cumulative frequency distributions of the precipitation rates at the surface in the

1200 control-s and low-aerosol-s runs for the Seoul case at (a) 06:00, (b) 10:00 and (c) 12:00

1201 LST.

1202

1203 Figure 6. Vertical distributions of differences in the area-averaged condensation,

1204 deposition and freezing rates, and cloud-liquid, raindrop, snow and hail mass density, and

1205 updraft mass fluxes between the control-s and low-aerosol-s runs at (a) 03:20, (b) 03:40,

1206 (c) 06:00 and (d) 12:00 LST. The horizontal black line in each panel represents the altitude

1207 of freezing or melting. Here, for the sake of the display brevity, snow mass density includes

1208 ice-mass density, while hail mass density includes graupel mass density.

1209

1210 Figure 7. Spatial distributions of terrain heights, column-averaged condensation rates,

1211 surface wind vectors and precipitation rates at (a) and (b) 03:40, (c) and (d) 08:40, (e) and

1212 (f) 10:00, and (g) and (h)12:00 LST. The distributions in the control-s run are shown in (a),

1213 (c), (e) and (g), and the distributions in the low-aerosol-s run are shown in (b), (d), (f) and

1214 (h). Condensation rates are shaded. Dark-yellow and dark-red contours represent

1215 precipitation rates at 0.5 and 3.0 mm hr^{-1}, respectively, while beige, light brown and brown

1216 contours represent terrain heights at 100, 300 and 600 m, respectively. See text for yellow

1217 rectangles in (a), (b), (e) and (f).

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1219 Figure 8. Cumulative frequency distributions of the column-averaged condensation rates

1220 in the control-s and low-aerosol-s runs for the Seoul case at (a) 07:20 and (b) 09:00 LST.

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1222 Figure 9. Cumulative frequency distributions of the precipitation rates at the surface in the

1223 control-b and low-aerosol-b runs for the Beijing case at (a) 16:00, (b) 17:00, (c) 17:20, and

1224 (d) 19:00 LST.

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1226 Figure 10. Same as Figure 6 but for differences between the control-b and low-aerosol-b

1227 runs at (a) 14:20, (b) 15:40, (c) 16:00, (d) 17:20 and (e) 19:00 LST.

1228

1229   Figure 11. Spatial distributions of terrain heights, column-averaged condensation rates,

1230   surface wind vectors and precipitation rates at (a) and (b) 14:20, and (c) and (d) 17:20 LST.

1231   (a) and (c) are for the control-b run and (b) and (d) are for the low-aerosol-b run.

1232   Condensation rates are shaded. Dark-yellow and dark-red contours represent precipitation

1233   rates at 1.0 and 2.0 mm hr^{-1}, respectively, while beige, light brown, brown and dark brown

1234   contours represent terrain heights at 100, 500, 1000 and 1500 m, respectively. See text for

1235   yellow circles in (a) and (b).

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1237   Figure 12. Cumulative frequency distributions of the column-averaged condensation rates

1238   in the control-b and low-aerosol-b runs for the Beijing case at (a) 17:20 and (b) 19:00 LST.

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1240   Figure 13. Mean precipitation rates corresponding to each column-averaged condensation

1241   rate for the period between 14:00 and 17:20 LST in the control-b run. One standard

1242   deviation of precipitation rates is represented by a vertical bar at each condensation rate.

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1244   Figure 14. Spatial distributions of precipitation rates (shaded) and wind vectors (arrows)

1245   for the Beijing case at (a) and (b) 17:00, and (c) and (d) 19:00. The distributions in the

1246   control-b run are in (a) and (c). The distributions in the low-aerosol-b run are in (b) and (d).

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Simulations	Site	Concentrations of background aerosols acting as CCN
Control-s run	Seoul area	Observed and affected by the aerosol advection
Low-aerosol-s run	Seoul area	Same as those in the control-s run but unaffected by the aerosol advection
Control-b run	Beijing area	Observed
Low-aerosol-b run	Beijing area	Reduced by a factor of 3.1 as compared to those observed

1259

1260   Table 1. Summary of simulations

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Simulations	The net flux of the moist static energy which crosses the boundary between areas A and B ($J\ m^{-2}\ s^{-1}$)	
	At 16:30 LST	16:30 to 17:00 LST
Control-b run	1.57×10^{12}	1.07×10^{15}
Low-aerosol-b run	1.15×10^{12}	7.55×10^{14}

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1264   Table 2. The net flux of the moist static energy which crosses the boundary between areas

1265   A and B at 16:30 LST and for a period from 16:30 to 17:00 LST.

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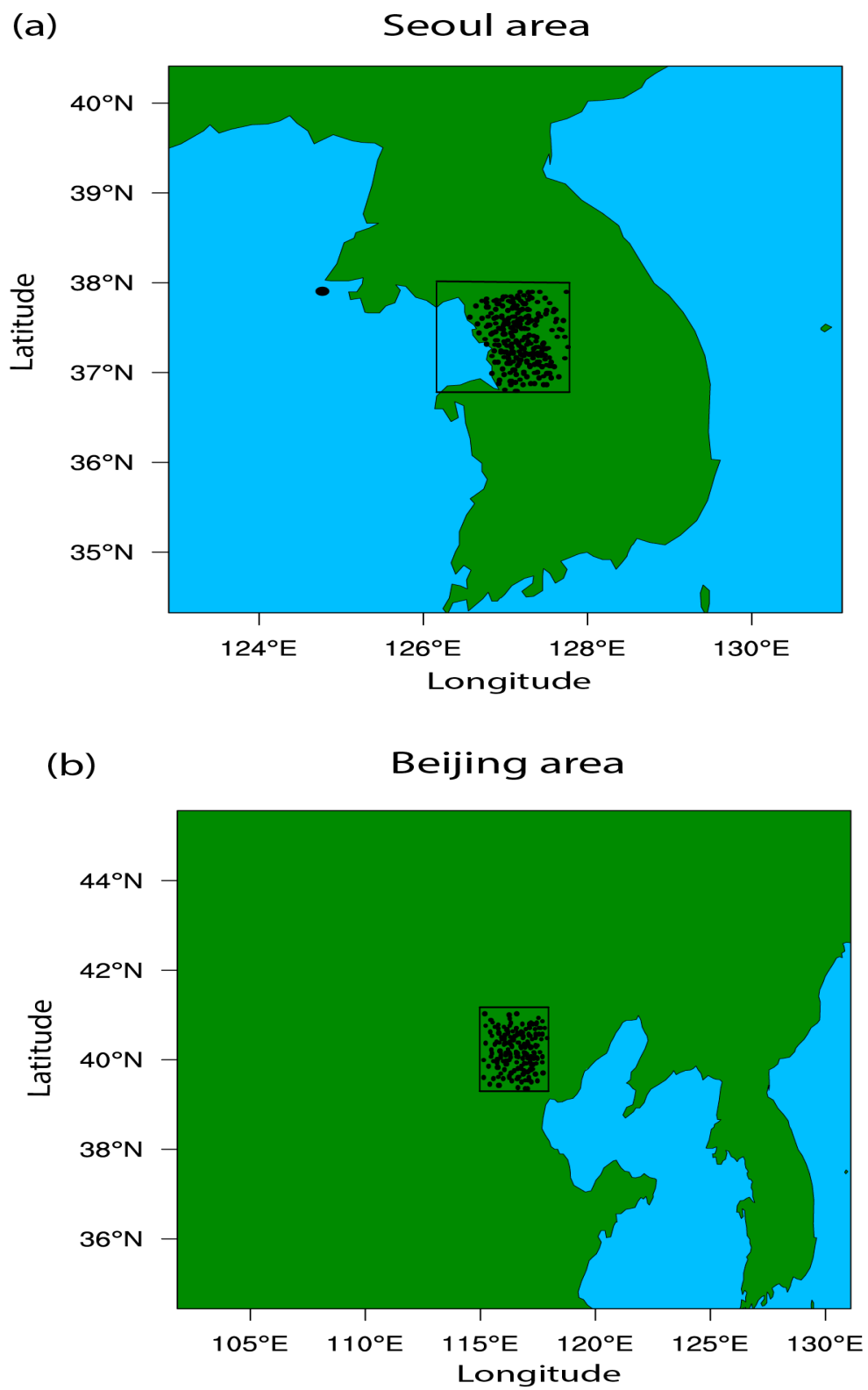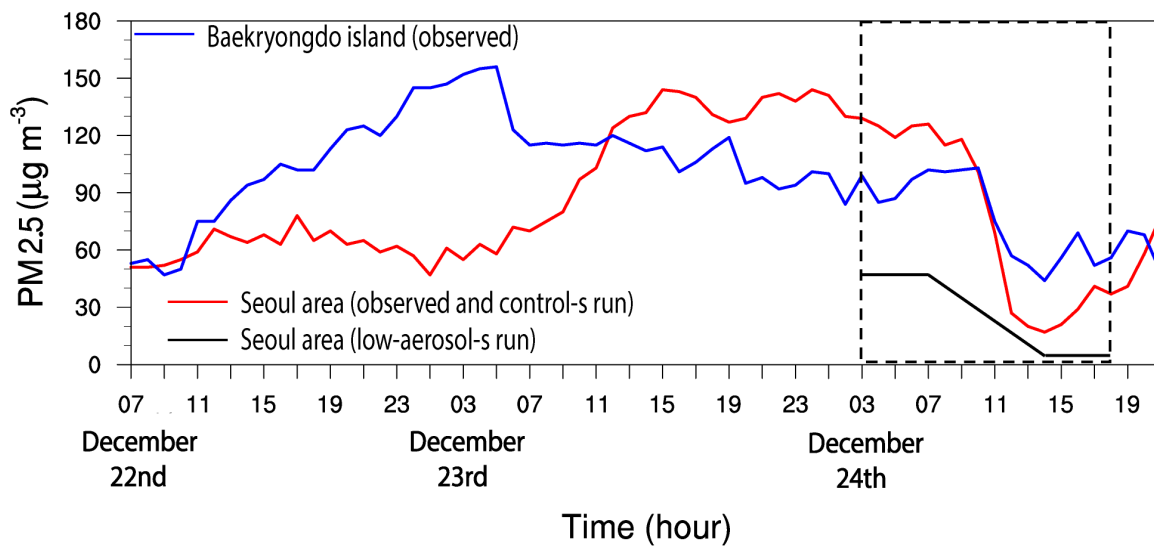
(a)

Seoul area

(b)

Beijing area

1272

Figure 1

1273

Figure 2

Seoul case

Beijing case

Figure 3

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(a)

Seoul case

(b)
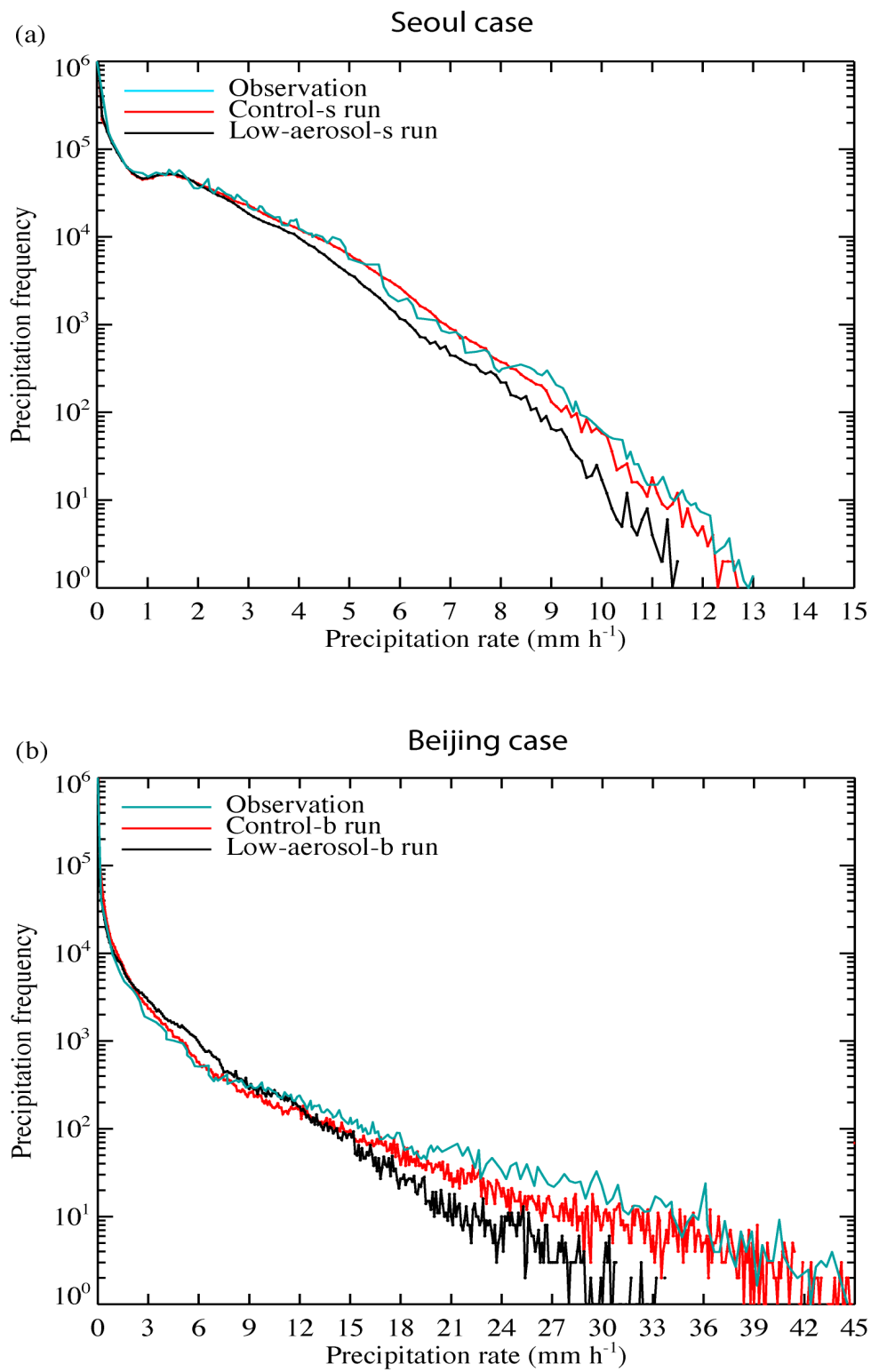
Beijing case

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1289   **Figure 4**

Seoul case

(a)

06:00 LST

(b)

10:00 LST

(c)

12:00 LST
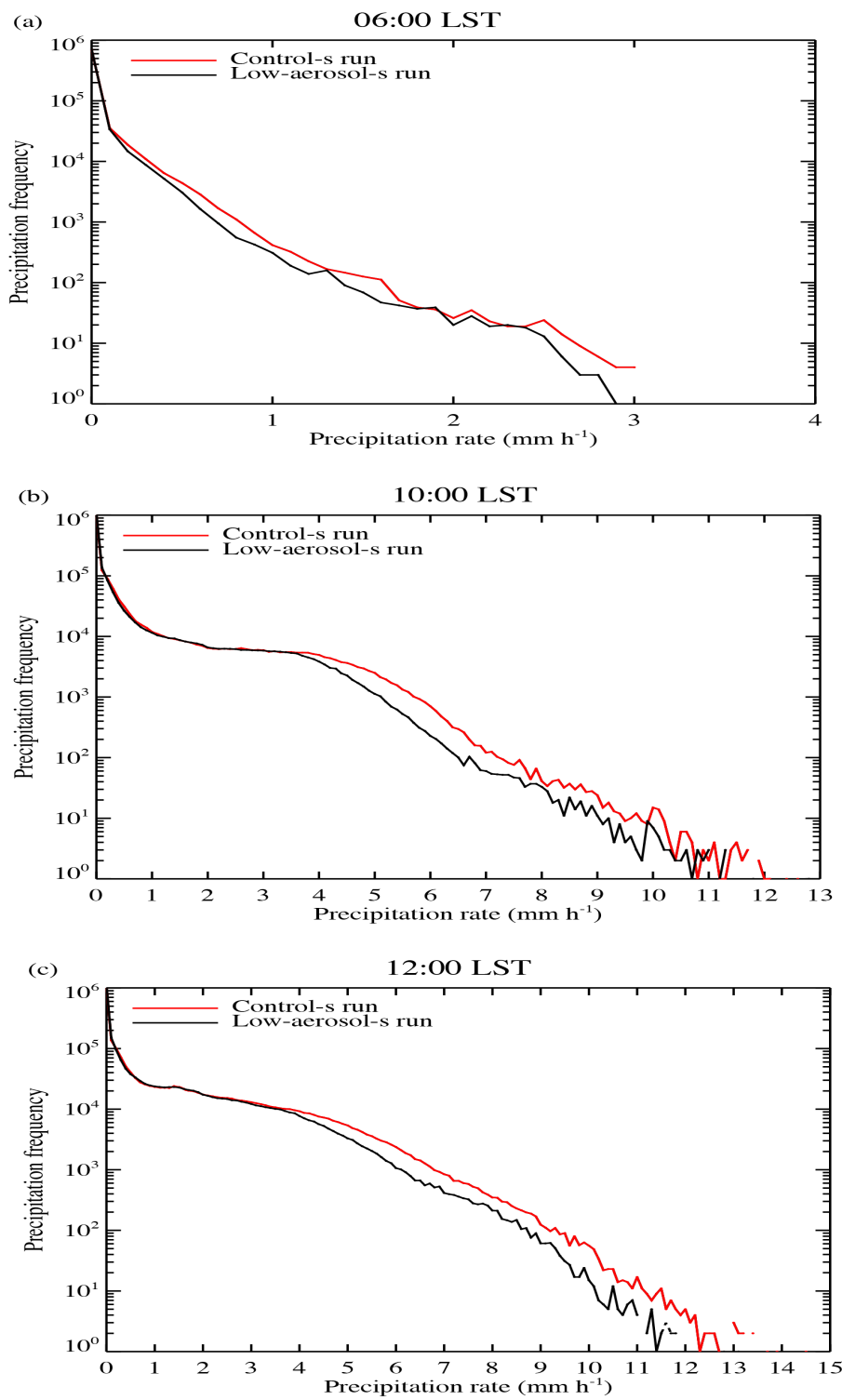
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Figure 5

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Seoul case
(control-s run minus low-aerosol-s run)

(a)

03:20 LST

Condensation rate $(\times 10^{-4} \text{g m}^{-3} \text{s}^{-1})$
Deposition rate $(\times 10^{-5} \text{g m}^{-3} \text{s}^{-1})$
Freezing rate $(\times 10^{5} \text{g m}^{-3} \text{s}^{-1})$
Updraft mass flux $(\times 10^{-2} \text{ kg m}^{-2} \text{s}^{-1})$
Cloud-liquid mass density $(\times 10^{-2} \text{g m}^{-3})$
Snow mass mass density $(\times 10^{-2} \text{g m}^{-3})$
Rain mass density $(\times 10^{-2} \text{g m}^{-3})$
Hail mass density $(\times 10^{-3} \text{g m}^{-3})$

(b)

03:40 LST

Condensation rate $(\times 10^{-4} \text{g m}^{-3} \text{s}^{-1})$
Deposition rate $(\times 10^{5} \text{g m}^{-3} \text{s}^{-1})$
Freezing rate $(\times 10^{5} \text{g m}^{-3} \text{s}^{-1})$
Updraft mass flux $(\times 10^{-2} \text{ kg m}^{-2} \text{s}^{-1})$
Cloud-liquid mass density $(\times 10^{-2} \text{g m}^{-3})$
Snow mass mass density $(\times 10^{-2} \text{g m}^{-3})$
Rain mass density $(\times 10^{-2} \text{g m}^{-3})$
Hail mass density $(\times 10^{-3} \text{g m}^{-3})$

1292

1293
Figure 6

(c)

06:00 LST

(d)

12:00 LST

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1295   **Figure 6**

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Figure 7

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1300

Figure 7
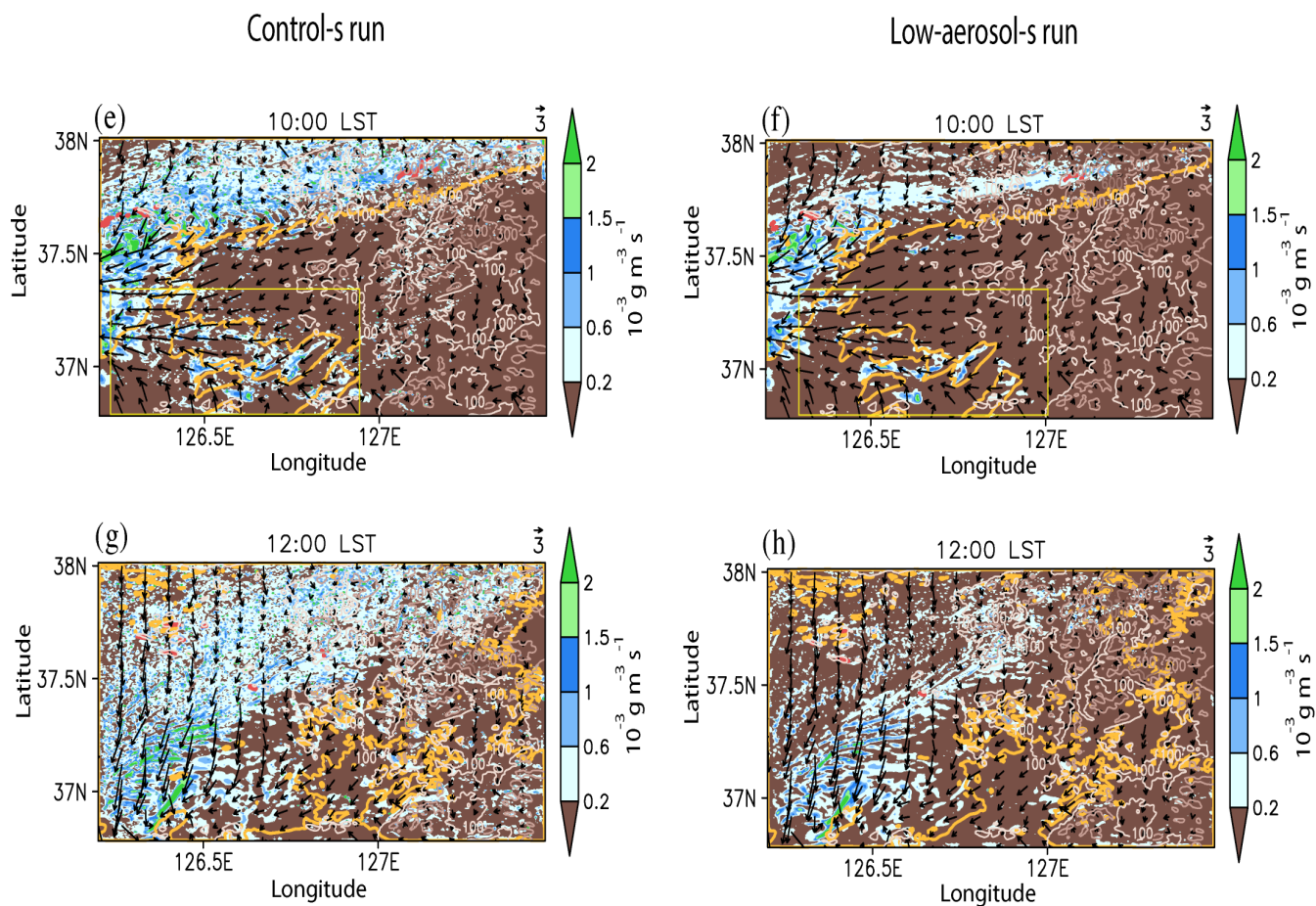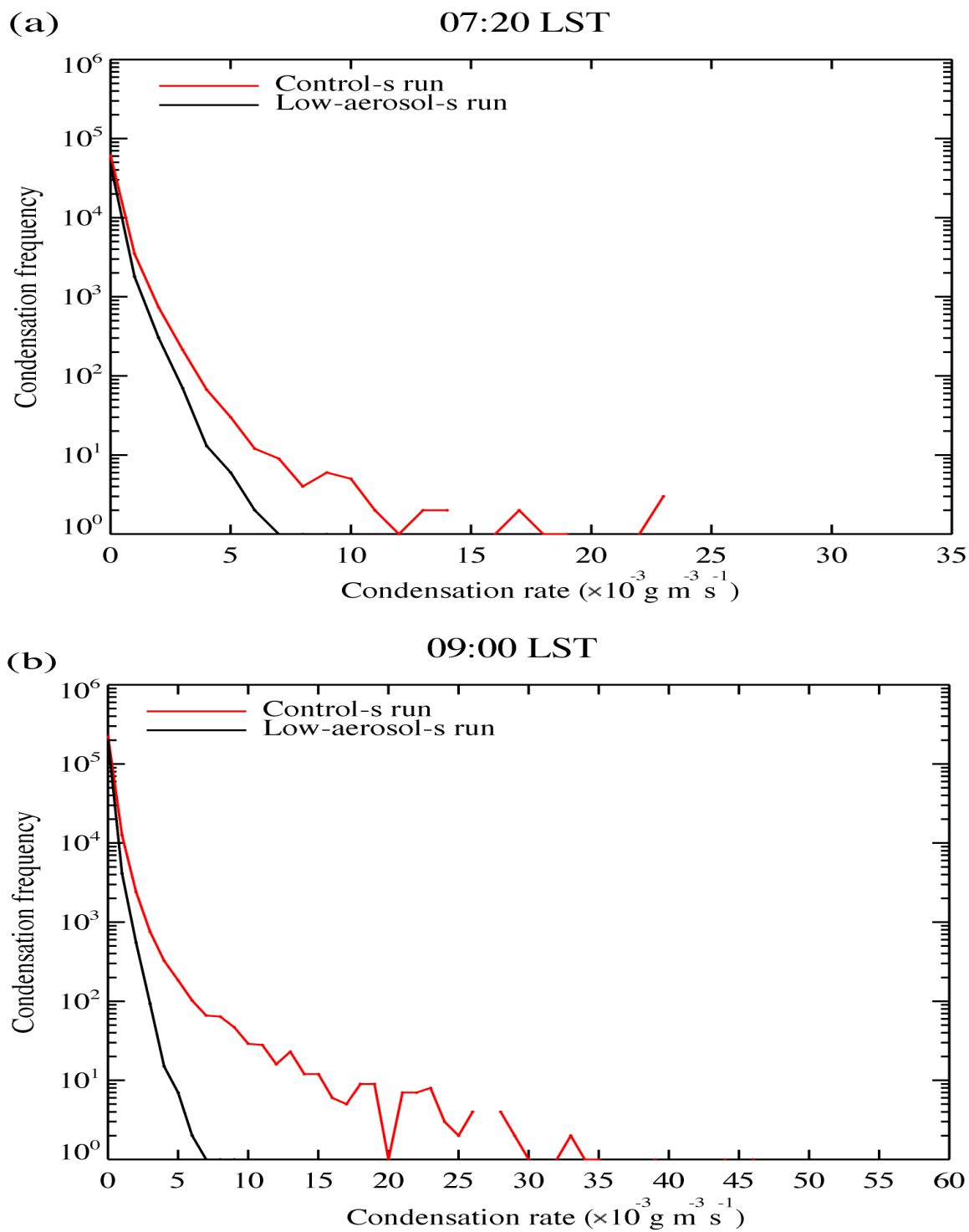
Seoul case

(a)

07:20 LST

(b)

09:00 LST

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1306 **Figure 8**

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Beijing case

16:00 LST

(a)

17:00 LST

(b)

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Figure 9

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(c)

17:20 LST

(d)

19:00 LST
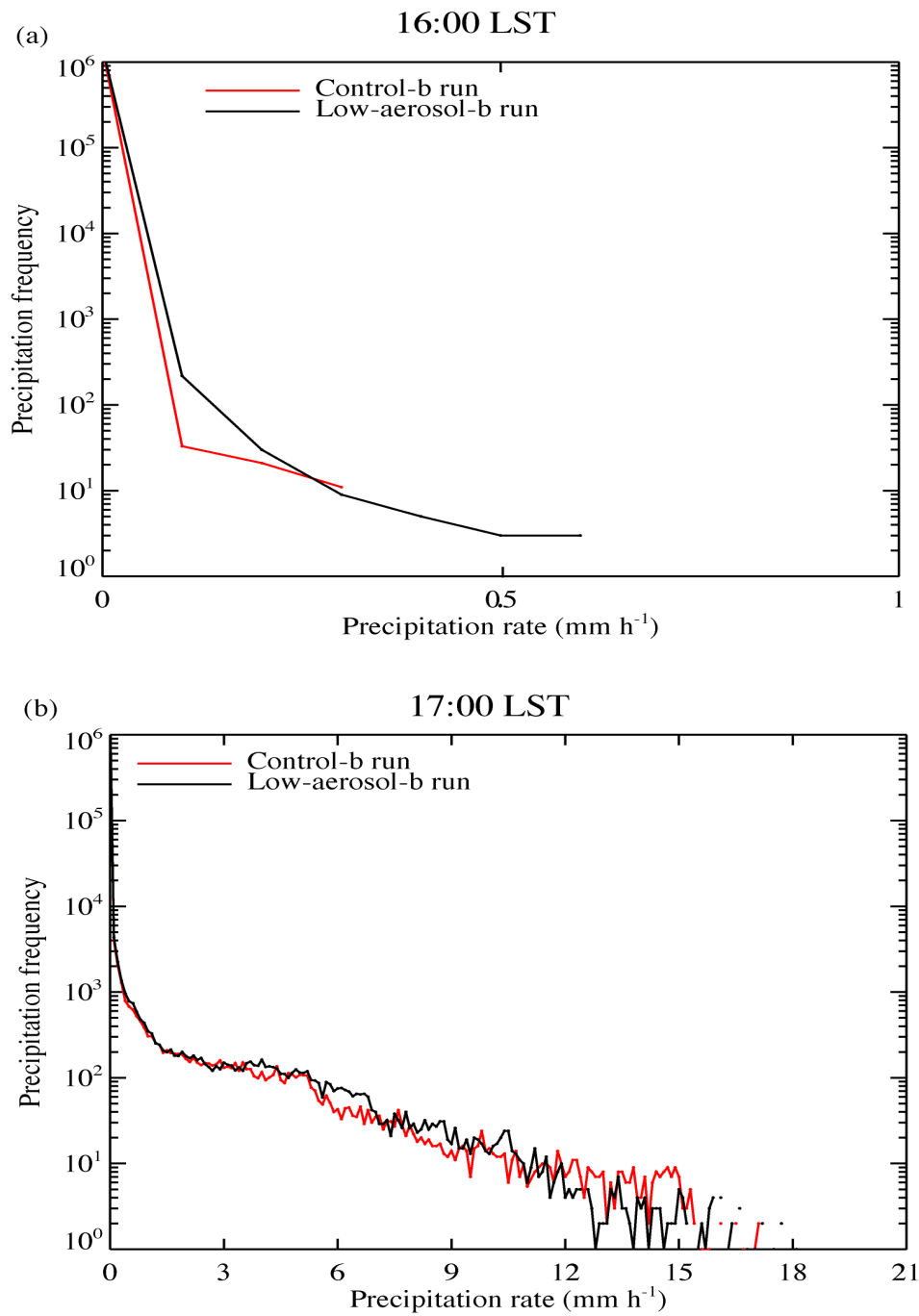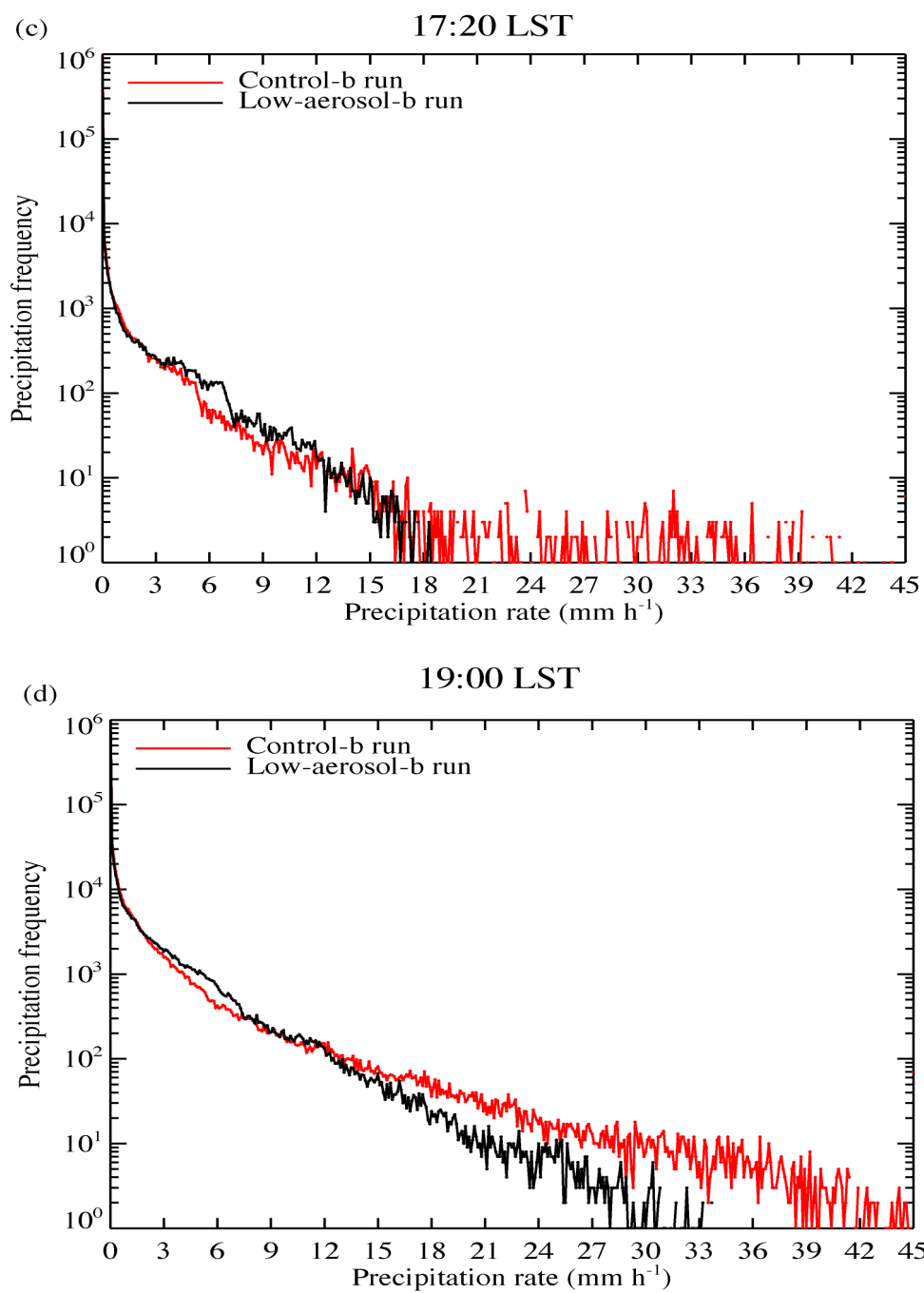
1311
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1313
Figure 9
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Beijing case
(control-b run minus low-aerosol-b run)

(a) 14:20 LST

(b) 15:40 LST
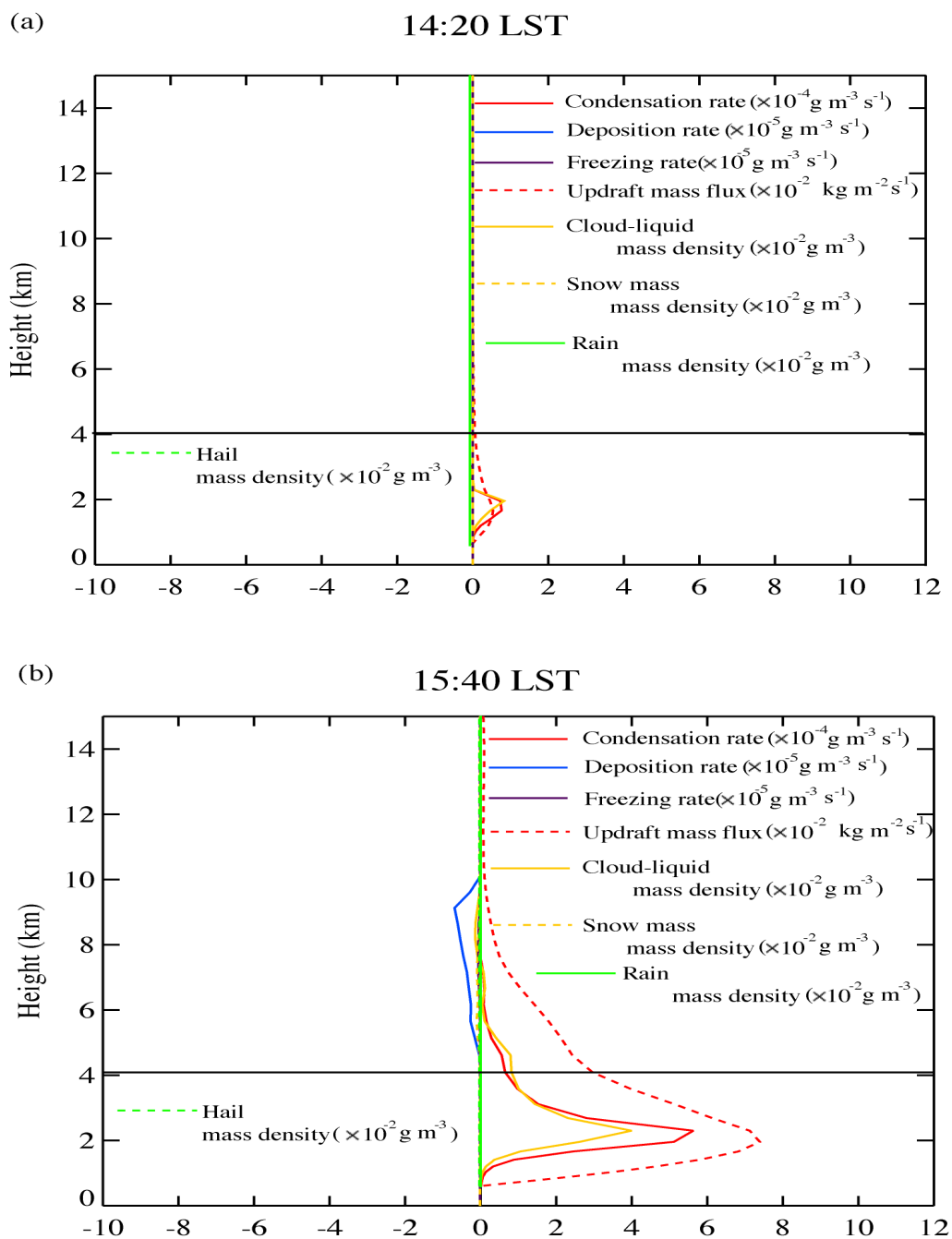
1317

1318 **Figure 10**

(c) 16:00 LST

(d) 17:20 LST

(e) 19:00 LST
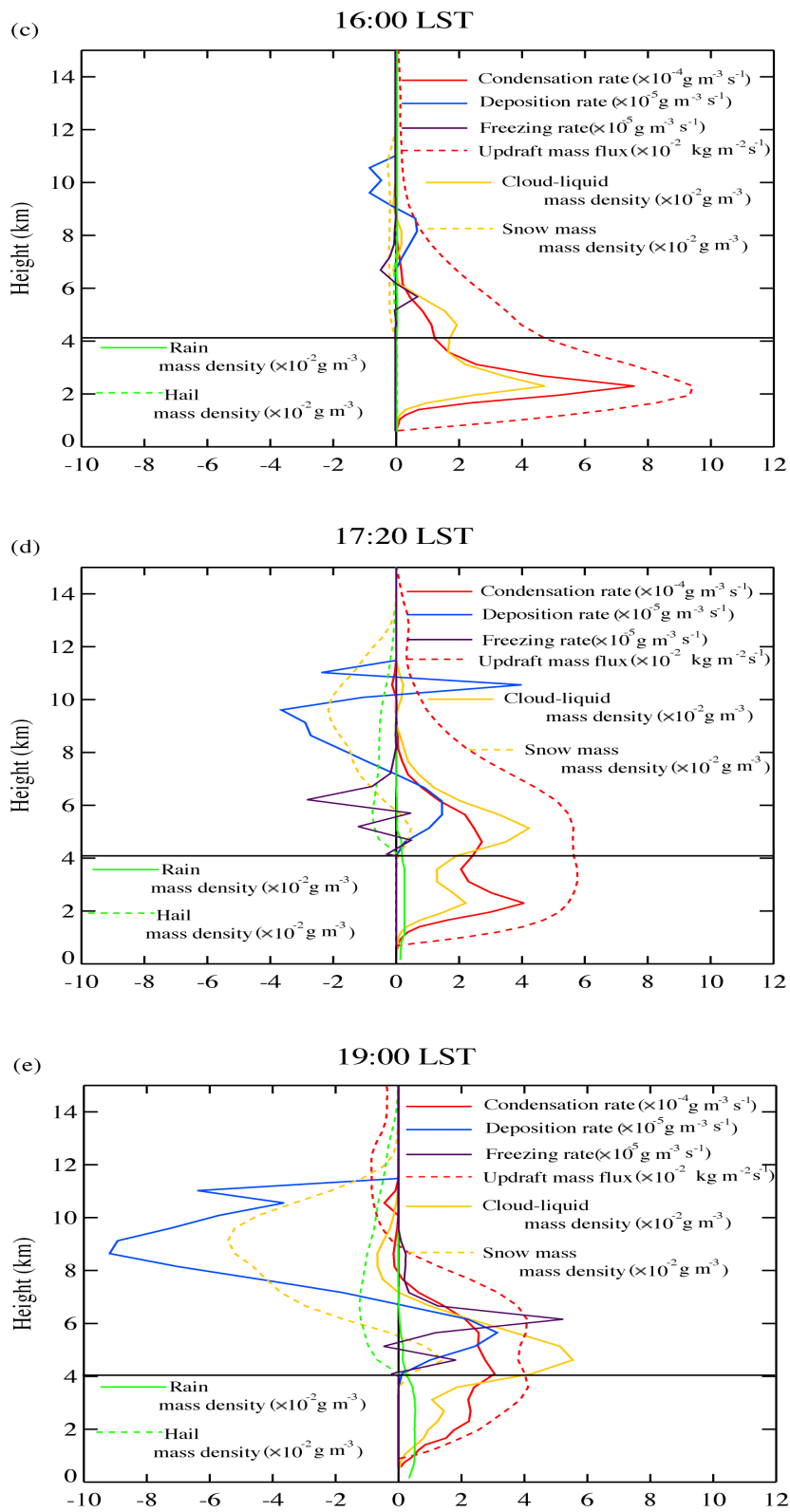
Figure 10

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1320

Figure 11

Beijing case

(a)

17:20 LST

(b)

19:00 LST

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1327 **Figure 12**

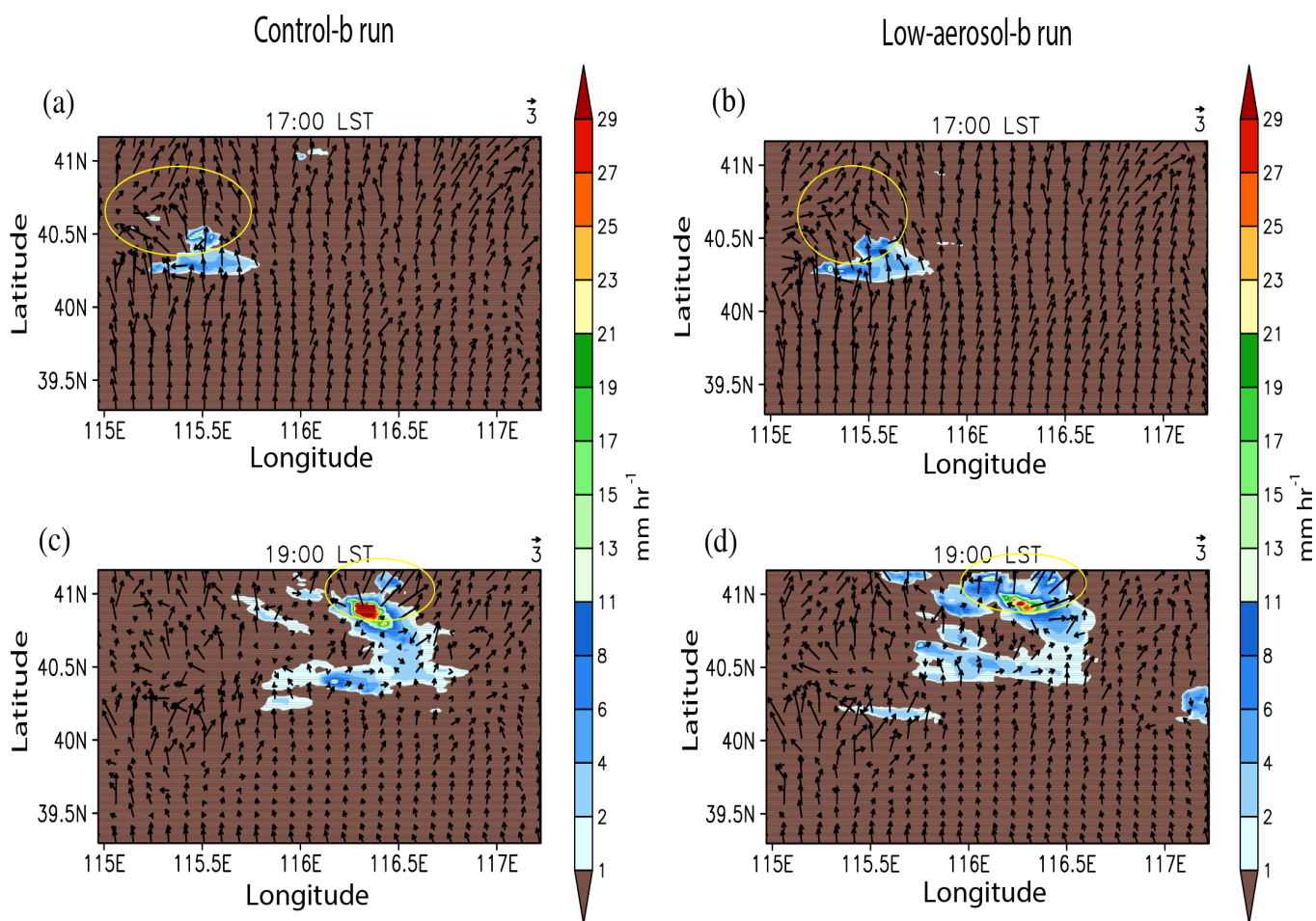Condensation rates vs. Precipitation rates
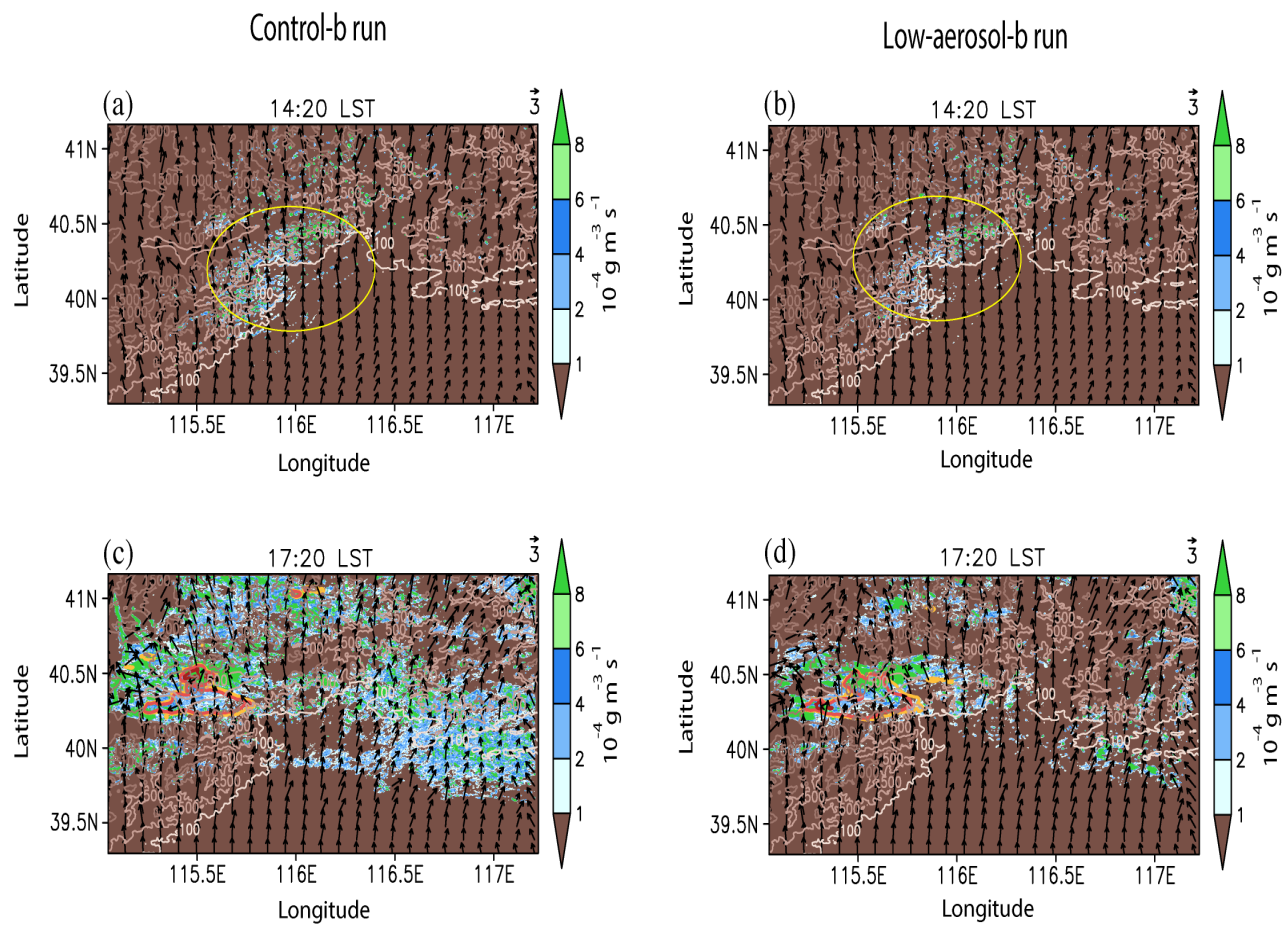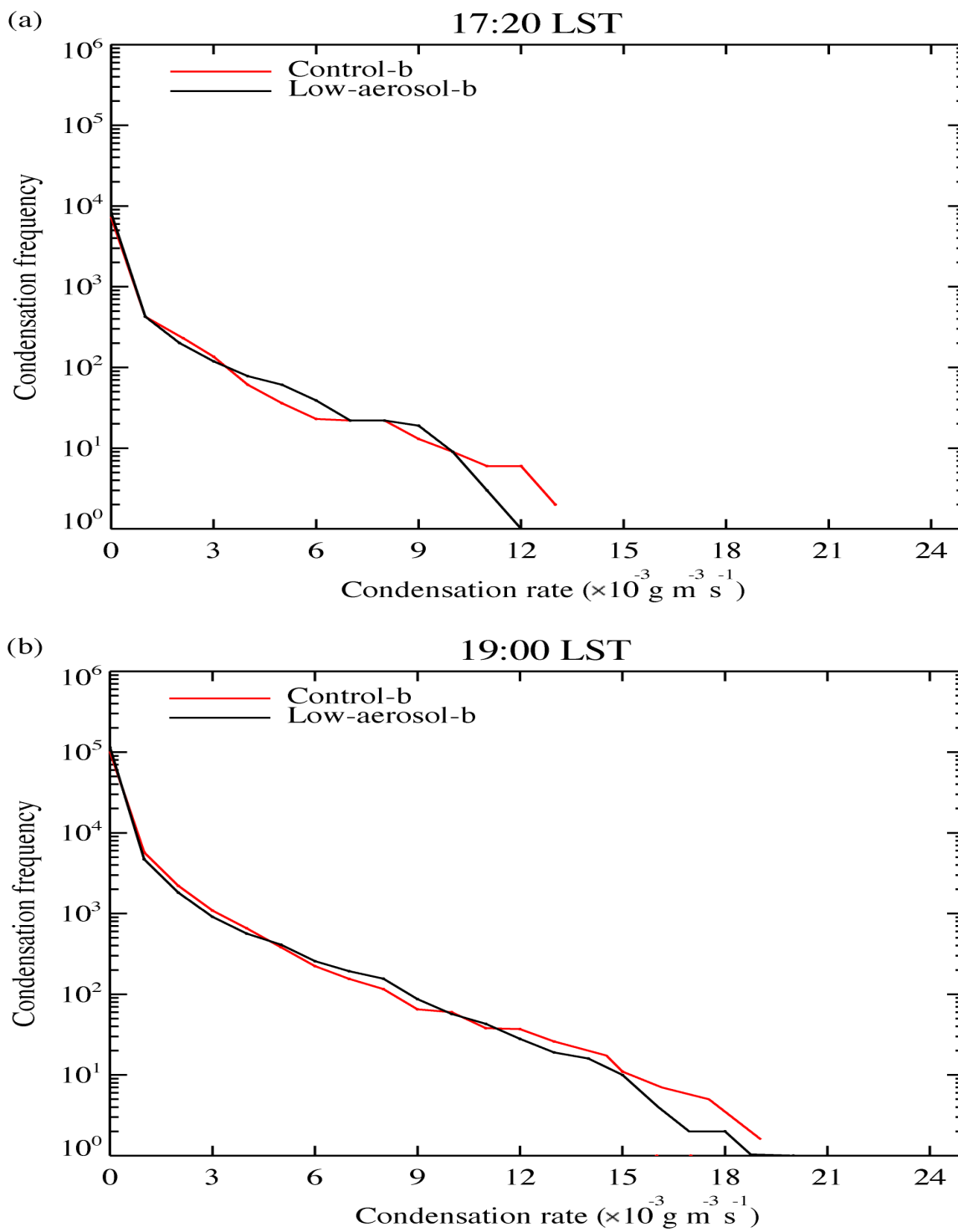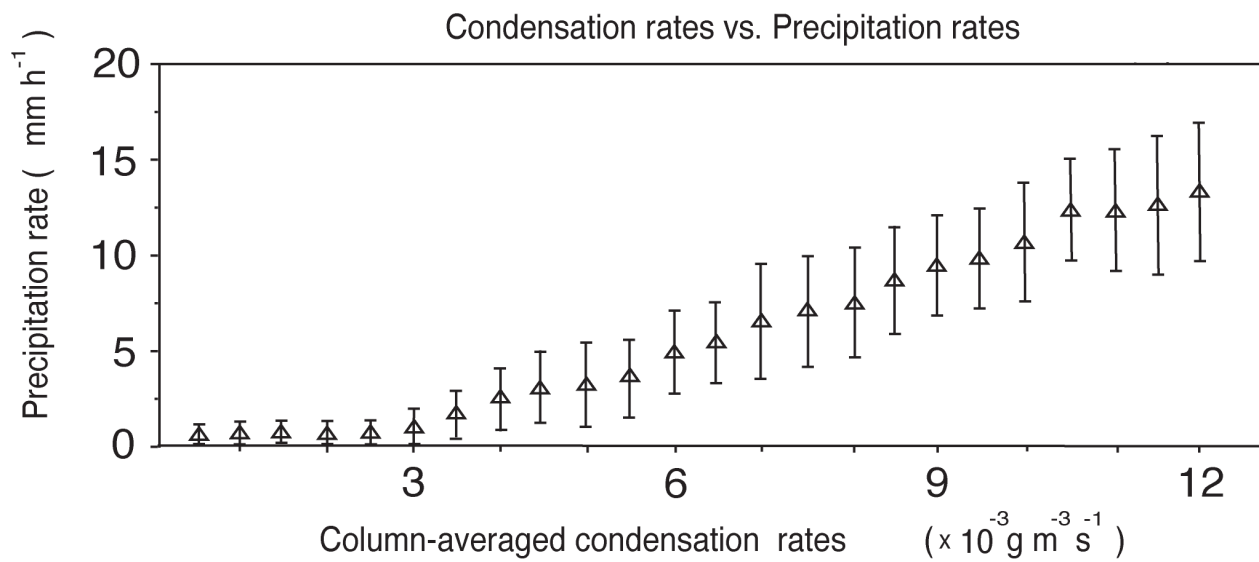
Figure 13

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Figure 14

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