1 **Examination of effects of aerosols on a pyroCb and their dependence on fire**
2 **intensity and aerosol perturbation**
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Abstract

Through a modeling framework, this study investigates how a pyrocumulonimbus (pyroCb) event influences water vapor concentrations and cirrus cloud properties near the tropopause, specifically focusing on how fire-produced aerosols affect this role. Results from a case study show that when observed fire intensity is high, there is an insignificant impact of fire-produced aerosols on the development of the pyroCb and associated changes in water vapor and cirrus clouds near the tropopause. However, as fire intensity weakens, effects of those aerosols on microphysical variables and processes such as droplet size and autoconversion increase. Due to this, aerosol-induced invigoration of convection is significant for pyroCb with weak-intensity fires and associated weak surface heat fluxes. This leads to a situation where there is a greater aerosol effect on the transport of water vapor to the upper troposphere and the production of cirrus clouds with weak-intensity fires, whereas this effect is muted with strong-intensity fires.

56   **1. Introduction**

57

58   Recent studies (e.g., Pumphrey et al., 2011; Kablick et al., 2018) showed that

59   pyrocumulonimbus (pyroCbs) may transport significant amounts of water vapor to the

60   upper troposphere and the lower stratosphere (UTLS) and can possibly have an impact on

61   seasonal UTLS water vapor budgets. Any change in water vapor in the UTLS has an

62   exceptionally strong influence on the global radiation budget and thus Earth's climate

63   (Solomon et al., 2010). PyroCbs develop cirrus clouds associated with overshooting

64   convective tops that reach the UTLS. Changes in cirrus clouds in the UTLS are known to

65   have a strong influence on the global radiation budget (Solomon et al., 2010). The level of

66   our understanding of impacts of pyroCbs on water vapor and cirrus clouds in the UTLS on

67   a global scale is very low and studies have been conducted to improve this understanding

68   (Fromm et al., 2010). However, this paper does not focus on these pyroCb impacts at the

69   global scale. Instead, this paper aims to gain a process-level understanding of mechanisms

70   that control local impacts of individual pyroCbs on water vapor and cirrus clouds in the

71   UTLS. The examination of these mechanisms can provide useful information to

72   parameterize interactions among pyroCbs, water vapor and cirrus clouds in climate models.

73   PyroCbs initiate over a fire, and the large surface energy release mainly through fire-

74   induced latent- and sensible-heat fluxes at and near the surface affects the dynamic,

75   thermodynamic and microphysical development of pyroCbs (Fromm et al., 2010; Peterson

76   et al., 2017). However, questions remain about what role the large concentration of cloud

77   condensation nuclei (CCN) contained in smoke has on the vertical development and

78   microphysical properties of pyroCbs. Studies (e.g., Koren et al., 2008; Rosenfeld et al.,

79   2008; Storer et al., 2010; Tao et al., 2012) showed that aerosols affected cumulonimbus

80   clouds, thus raising the possibility that fire-generated aerosols may affect pyroCb

81   development. As an example of aerosol impacts on cumulonimbus clouds, these studies

82   have demonstrated that increases in aerosol loading can decrease the size of droplets (i.e.,

83   cloud-liquid particles). Individual aerosol particles act as seeds for the formation of

84   droplets, so increasing aerosol loading leads to more droplets formed. More droplets mean

85   more competition among them for available water vapor needed for their condensational

86   growth, decreasing the size of individual droplets (Twomey, 1977; Albrecht, 1989).

87 Aerosol-induced smaller droplet sizes reduce the efficiency of the growth of cloud-liquid

88 particles to raindrops via autoconversion, a collection process among cloud-liquid particles

89 in which they become raindrops (Pruppacher and Klett, 1978; Rogers and Yau, 1991). This

90 reduced efficiency leads to less cloud liquid converted to rain and thus more cloud liquid

91 available for transport by updrafts to altitudes above the freezing level. This eventually

92 induces more freezing of cloud liquid, enhanced parcel buoyancy and the invigoration of

93 updrafts and associated convection (Rosenfeld, 2008).

94   The role of fire-generated aerosols in the development of pyroCbs and their effects

95 on water vapor and cirrus clouds in the UTLS lacks a firm scientific understanding, and

96 hence this paper focuses on that role of those aerosols. To examine the role, this study

97 extends the previous modeling work by Kablick et al. (2018). The modeling work therein

98 showed that the effects of fire-generated aerosols on the development of a specific pyroCb

99 and its impacts on the UTLS water vapor and cirrus clouds were negligible compared to

100 the effects of fire-generated heat fluxes. However, aerosol effects on cloud development

101 vary with cloud properties such as typical updraft speeds (e.g., Khain et al., 2008; Lee et

102 al., 2008; Tao et al., 2012). For simplicity herein, an "updraft" refers to the general upward

103 motion of convective air, or to the actual updraft speed representing the updraft or

104 convective intensity, depending on the context. Typical updrafts are determined by

105 environmental instability as represented by convective available potential energy (CAPE).

106 Lee et al. (2008) demonstrated that different clouds with different typical updrafts showed

107 different sensitivity of cloud microphysical and thermodynamic development to aerosol

108 concentration. Hence, it is hypothesized that aerosol effects on pyroCb development and

109 its impacts on the UTLS water vapor and cirrus clouds vary depending on the pyroCb

110 typical updrafts.

111   To examine the potential variation of aerosol effects on pyroCb development and its

112 impacts on the UTLS water vapor and cirrus clouds with typical updrafts of pyroCbs,

113 numerical simulations are performed. Simulated is the pyroCb case examined by Kablick

114 et al. (2018) using a cloud-system resolving model (CSRM). The CSRM is capable of

115 resolving cloud-scale dynamic and thermodynamic processes. The basic modeling

116 methodology in this study is similar to that used by Kablick et al. (2018). However, this

study uses a more sophisticated microphysical scheme, i.e., a bin scheme, rather than the two-moment bulk scheme used by Kablick et al. (2018).

Note that Kablick et al. (2018) examined aerosol effects on the convective development of a specific pyroCb case with a typical updraft framework. The present study expands upon that work by performing sensitivity simulations in which typical updrafts in the pyroCb vary, enabling us to ascertain the dependence of aerosol effects on typical updrafts. Note that CAPE, which determines typical updrafts in convective clouds, is strongly dependent on surface latent and sensible heat fluxes (e.g., Houze, 1993), and in the case of pyroCbs, these fluxes are strongly controlled by fire intensity. Therefore, the present sensitivity simulations enable us to study the dependence of those aerosol effects on fire intensity. Since fire intensity is the dominant driver of the pyroCb typical updrafts, these typical updrafts are henceforth referred to as fire-driven updrafts.

Effects of fire-induced increases in aerosol concentration on pyroCbs are likely dependent on how much aerosol concentration increases (aerosol perturbation) (e.g., Rosenfeld et al., 2008; Koren et al., 2012). This study examines this dependence, not studied by Kablick et al. (2018).

2. Modeling framework

We use the Advanced Research Weather Research and Forecasting (ARW) model, a nonhydrostatic compressible model, as the CSRM. Prognostic microphysical variables are transported with a fifth-order monotonic advection scheme (Wang et al., 2009). Shortwave and longwave radiation is parameterized by the Rapid Radiation Transfer Model (RRTM; Mlawer et al., 1997; Fouquart and Bonnel, 1980).

To represent microphysical processes, the CSRM adopts a bin scheme based on the Hebrew University Cloud Model described by Khain et al. (2009). The bin scheme solves a system of kinetic equations for the size distribution functions of water drops, ice crystals (plate, columnar and branch types), snow aggregates, graupel and hail, as well as cloud condensation nuclei (CCN) and ice nuclei (IN). Each size distribution is represented by 33 mass doubling bins, i.e., the mass of a particle m_k in the kth bin is determined as $m_k = 2m_{k-1}$.

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

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

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

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

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

153   paths are all considered by taking into account the size distribution of IN, temperature and

154   supersaturation. Homogeneous aerosol (or haze particle) and droplet freezing is also

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

156

157   **3. Case description and simulations**

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159   **3.1 Control run**

160

161   The control run for an observed pyroCb case involved a forested site in the Canadian

162   Northwest Territories (60.03° N, 115.45° W). Kablick et al. (2018) give details about the

163   site and pyroCb case. The control run is identical to the Full Simulation in Kablick et al.

164   (2018) except for the different microphysical schemes. The period covered by the control

165   run is from 12:00 GMT on August 5th to 12:00 GMT on August 6th in 2014 and captures

166   the initial, mature, and decaying stages of the pyroCb. As described by Kablick et al. (2018),

167   balloon soundings of winds, temperature and dew-point temperature were obtained every

168   6 hours from the Ft. Smith observation station near the forested site. Sounding data at 12:00

169   GMT on August 5th prescribe the initial atmospheric conditions. Temperature and humidity

170   tendencies at each altitude from sequential soundings are obtained and applied to the

171   control run every time step by interpolation in a horizontally homogeneous manner. These

172   tendencies represent the impacts of synoptic- or large-scale motion on temperature and

173   humidity (Grabowski et al., 1996; Krueger et al., 1999; and Lee et al., 2018). The control

174   run is performed in a three dimensional domain with horizontal and vertical extends of 300

175   km and 20 km, respectively. The simulation horizontal and vertical resolutions are 500 m

176   and 200 m, respectively, to resolve cloud dynamic and thermodynamic processes.

177   Figure 1 shows a satellite image of the observed pyroCb and the fire spot (spatial length

178   is ~40 km) when the cloud is about to advance into its mature stage. To emulate this in the

179   simulation, a fire spot with a diameter of 40 km is placed at the center of the simulation

180   domain (Figure 2). In the fire spot, the surface latent and sensible heat fluxes are set at

181   1800 and 15000 W m^{-2}, respectively. In areas outside of the fire spot, the surface latent and

182   sensible heat fluxes are set at 310 and 150 W m^{-2}, respectively. These surface heat-flux

183   values follow previous studies which adopt boreal forest emissions (Trentmann et al., 2006;

184   Luderer et al., 2006). Following Kablick et al. (2018), the surface heat-flux values are

185   prescribed with no temporal variation and no consideration of interactions between heat

186   fluxes and the atmosphere in the control run. Thus, the setup for the surface heat fluxes is

187   idealized, enabling a better isolation of aerosol effects on pyroCb development and its

188   impacts on the UTLS water vapor and cirrus clouds by excluding effects of interactions

189   between the surface heat fluxes and atmosphere on this development and its impacts.

190   For the selected pyroCb case, aerosol chemical composition, size distribution and

191   concentration are unknown. Hence, in (outside) the fire spot at the first time step, the

192   concentration of aerosols acting as CCN is prescribed to be 15000 (150) cm^{-3}. This

193   prescription is for the planetary boundary layer (PBL) and the concentration decreases

194   exponentially with height above the PBL top. These prescribed aerosol concentrations are

195   typically observed in fire spots and their background (Pruppacher and Klett, 1997; Seinfeld

196   and Pandis, 1998; Reid et al., 1999; Andreae et al., 2004; Reid et al., 2005; Luderer et al.,

197   2009).

198   Reid et al. (2005) showed that aerosol mass produced by forest fires was generally

199   composed of (by mass) ~50-70% organic-carbon (OC) compounds, ~5-10% black-carbon

200   (BC) material, and ~20-45% inorganic species. The approximate median value of each of

201   these chemical component percentage ranges determines the aerosol particle composition

202   in the control run, i.e., 60% OC, 8% BC, and 32% inorganic species. OC is assumed to be

203   water soluble and composed of 18 % levoglucosan ($C_6H_{10}O_5$, density = 1600 kg m^{-3}, van't

204   Hoff factor = 1), 41 % succinic acid ($C_4H_6O_4$, density = 1572 kg m^{-3}, van't Hoff factor =

205   3), and 41 % fulvic acid ($C_{33}H_{32}O_{19}$, density = 1500 kg m^{-3}, van't Hoff factor = 5) based

206   on typically observed chemical composition of OC compounds over fire sites (Reid et al.,

207   2005). In the control run, the inorganic species are assumed to be ammonium sulfate, a

208   representative inorganic species associated with fires (Reid et al., 2005). This chemical

209 composition taken for aerosol particles is assumed to be spatiotemporally unvarying in the
210 control run.

211 The control run adopts the unimodal lognormal distribution as an initial aerosol size
212 distribution, a reasonable assumption in fire sites (Reid et al., 2005; Knobelspiessel et al.,
213 2011; Lee et al., 2014). Those studies reported that in general, median and standard
214 deviation aerosol diameters range from ~0.01 μm to ~0.03 μm and from ~2.0 μm to ~2.2
215 μm, respectively, for aerosols as CCN. The approximate median values of these ranges
216 determine median and standard deviation diameters of aerosols as CCN in the control run,
217 i.e., 0.02 μm and 2.1 μm, respectively. Assumed identical in the control run are IN and
218 CCN aerosol properties except that at the first time step, (1) their median and standard
219 deviation aerosol diameters differ, and (2) the IN concentration is 100 times lower than the
220 CCN concentration (Pruppacher and Klett, 1978; Fan et al., 2014 and 2017). Following
221 Seinfeld and Pandis (2006) and Phillips et al. (2007), for aerosols as IN, median and
222 standard deviation aerosol diameters are assumed to be 0.1 μm and 1.6 μm, respectively,
223 which are typical values in the continent.

224 Airflow in clouds diffuses and advects aerosols. After activation or capture by
225 precipitating hydrometeors, aerosols are transported within hydrometeors and removed
226 from the atmosphere once these hydrometeors reach the surface. Once clouds disappear
227 completely at any grid point, aerosol size distribution and number concentration recover to
228 the background values at the first time step. This assumption simulates overall aerosol
229 properties and their impacts on clouds and precipitation reasonably well (Morrison and
230 Grabowski, 2011; Lebo and Morrison, 2014; Lee et al., 2016). This assumption means that
231 fire continuously produces aerosols to maintain the initial background aerosol
232 concentration.

233 Located to the northeast of the fire spot is the observed cirrus cloud at the top of the
234 pyroCb, since winds advect the cloud northeastward (Figure 1). The extent of the observed
235 cirrus cloud is ~100 km. Figure 2 shows the simulated field of cloud-ice mass density at a
236 time that corresponds to the satellite image in Figure 1, representing the simulated cirrus
237 cloud in the control run. Located to the northeast of the fire spot is the simulated cirrus
238 cloud, and the extent of this cloud is ~100 km. The morphologies of the observed and
239 simulated cirrus clouds agree well.

240   The average liquid-water path (LWP) over areas with non-zero LWP in the control run

241   is 960 g m^{-2}, and the average ice-water path (IWP) over areas with non-zero IWP in the

242   control run is 202 g m^{-2}. These simulated LWP and IWP are ~10 % different from the

243   satellite-retrieved values. In this study, droplet mass but not raindrop mass is used to obtain

244   liquid-water content (LWC) and LWP, and the mass of ice crystals but not the mass of

245   snow aggregates, graupel and hail is used to obtain ice-water content (IWC) and IWP.

246   Drops with radii smaller (greater) than 20 μm are classified as droplets (raindrops). The

247   average cloud-top and cloud-base heights over the life span of the pyroCb are 10.3 km and

248   3.6 km in the control run, respectively, and these simulated heights are ~7% different from

249   the satellite-retrieved values. Overall, cloud macro-physical structures, as represented by

250   LWP, IWP, cloud-top and cloud-base heights, are simulated reasonably well compared to

251   the observation.

252   Compared are the control-run and observed reflectivity fields. Kablick et al. (2018)

253   provide details about the reflectivity field observed by CloudSat. Observations and the

254   control run both show increasing reflectivity up to ~7 km, decreasing reflectivity between

255   ~7 km and ~11 km and insignificant changes in reflectivity above ~11 km in altitude

256   (Figure 3). The average reflectivity over altitudes along the Cloudsat path in the control

257   run is -8.1 dBZe, ~15% different from the observed value. Hence, the reflectivity field is

258   simulated fairly well compared to the observation. These favorable comparisons between

259   the observed and simulated cirrus clouds, cloud macro-physical and reflectivity fields

260   demonstrate that the pyroCb-case simulation reasonably reproduces the event.

261

262   **3.2 Low-aerosol run**

263

264   To see the role played by fire-generated aerosols in the development of the pyroCb and its

265   effects on the UTLS water vapor and cirrus clouds, we repeat the control run by reducing

266   aerosol concentration in the fire spot from 15000 cm^{-3} to the background aerosol

267   concentration (i.e., 150 cm^{-3}). This reduction removes fire-generated aerosols in the fire

268   spot. Hence, comparisons between the control run and this repeated run, referred to as the

269   low-aerosol run, will identify the role played by fire-generated aerosols. The low-aerosol

270 run is identical to the Low Aerosol Simulation in Kablick et al. (2018) except for the

271 different microphysical schemes between them.

272

273 **3.3 Additional runs**

274

275 For the examination of the potential variation of effects of fire-generated aerosols on

276 pyroCb development and its impacts on the UTLS water vapor and cirrus clouds with fire

277 intensity and associated fire-driven updrafts, we repeat the control run by varying fire

278 intensity. Remember that fire intensity controls surface latent and sensible heat fluxes on

279 which fire-driven updrafts are strongly dependent. Therefore, variations in fire-induced

280 surface latent and sensible heat fluxes can represent variations in fire intensity. As a first

281 step, the control run is repeated by reducing fire-induced surface latent and sensible heat

282 fluxes by factors of 2 and 4, respectively. The first repeated run represents a case with

283 medium fire intensity, referred to as "the medium run". The second repeated run represents

284 a case with weak fire intensity, referred to as "the weak run". Relative to these repeated

285 runs, the control run represents a case with strong fire intensity. Then, to see effects of fire-

286 generated aerosols on pyroCb development for each of those different fire intensities, the

287 medium and weak runs are repeated with the identical initial aerosol concentration to that

288 in the low-aerosol run. The repeated medium and weak runs are referred to as "the medium-

289 low run" and "the weak-low run", respectively. The control, medium, and weak runs are

290 the polluted-scenario runs, and the low-aerosol, medium-low, and weak-low runs are the

291 clean-scenario runs.

292 Effects of fire-generated aerosols on pyroCb development and its impacts on the UTLS

293 water vapor and cirrus clouds can also depend on the magnitude of fire-induced increases

294 in aerosol concentration in a fire spot. To test this dependence, for each fire intensity, we

295 repeat the polluted-scenario run by increasing and decreasing the magnitude by a factor of

296 2 in the fire spot but not outside of the fire spot. The simulations with the increased

297 magnitude have an aerosol concentration of 30000 cm^{-3} at the first time step over the fire

298 spot in the PBL, referred to as the control-30000, medium-30000, and weak-30000 runs

299 for strong, medium, and weak fire intensities, respectively. The simulations with the

300 decreased magnitude have an aerosol concentration of 7500 cm^{-3} at the first time step over

301   the fire spot in the PBL, referred to as the control-7500, medium-7500, and weak-7500

302   runs for strong, medium, and weak fire intensities, respectively. Motivated by the analysis

303   described in Section 4.3, we additionally repeat the medium and weak runs with aerosol

304   concentrations of 2000 and 1000 cm^{-3}, respectively, at the first time step over the fire spot

305   in the PBL. The repeated medium (weak) run is referred to as the medium-2000 (weak-

306   1000) run. Table 1 summarizes the simulations.

307   The aerosol concentration of 30000 cm^{-3} (7500, 2000 and 1000 cm^{-3}) over the fire spot

308   corresponds to a situation where fire produces a larger (lower) concentration of aerosols

309   than what is typically observed, i.e., 10000 to 20000 cm^{-3} (Reid et al, 1999; Andreae et al,

310   2004; Reid et al, 2005; Luderer et al., 2009).

311

312   **4. Results**

313

314   **4.1 The control and low-aerosol runs**

315

316   The updraft mass flux was averaged over the simulation period between 17:00 GMT on

317   August 5th, approximately when the pyroCb started to form, and 12:00 GMT on August 6th

318   (Figure 4). The updraft mass flux is one of the most indicative variables of the upward air

319   motion and magnitude of convective invigoration. Since the updraft mass flux is updraft

320   speed that is multiplied by air density, and air density at each altitude varies negligibly,

321   differences in updraft mass fluxes are mostly explained by those in updraft speeds among

322   the simulations. Hence, with good confidence, differences in updraft mass fluxes mean

323   those in updraft speeds.

324   The upper troposphere is defined here to be between ~9 km in altitude and the

325   tropopause which is ~13 km in altitude. The equilibrium level where the buoyancy of a

326   rising air parcel becomes zero above the level of free convection is the tropopause

327   (Emanuel, 1994). The lower stratosphere is defined here to be between the tropopause and

328   an altitude 10 km above the tropopause. The UTLS is thus between ~9 km and ~23 km in

329   this study. The defined upper troposphere and lower stratosphere occupy around a quarter

330   of the total vertical extent of the troposphere and stratosphere, respectively.

331   Updraft mass fluxes in the control run are only ~3% greater than those in the low-

332   aerosol run (Figure 4 and Table 2). Given the hundredfold difference in aerosol loading

333   over the fire spot between the runs, this 3% difference in updraft fluxes is negligible. The

334   comparison between water-vapor mass density over the cloudy and non-cloudy columns in

335   the control run demonstrates that there is a substantial 5-fold increase in the amount of

336   water vapor in the part of the UTLS at and above the tropopause due to the pyroCb (Figure

337   5 and Table 2). Henceforth, the UTLS water vapor means water vapor in the part of the

338   UTLS at and above the tropopause.

339   For the simulation period between 17:00 GMT on August 5th and 12:00 GMT on

340   August 6th, the average water-vapor mass fluxes at the tropopause over cloudy and non-

341   cloudy grid columns are 8.30×10^{-6} and 0.57×10^{-6} kg m^{-2} s^{-1}, respectively. Due to the

342   presence of the pyroCb and associated updrafts in cloudy grid columns, there are

343   substantial increases in fluxes at the tropopause over cloudy grid columns compared to

344   fluxes in the background over non-cloudy grid columns. This explains the larger amount

345   of the UTLS water vapor over the pyroCb than in the background in the control run. The

346   vertical extent of water vapor reaches further up to ~16 km by the pyroCb beyond ~14 km

347   in the background (Figure 5). This means that air parcels that include water vapor overshoot

348   the tropopause by ~3 km in the pyroCb, while those parcels in the background do so by ~

349   1 km. This implies that air parcels and associated updrafts in the pyroCb are stronger,

350   reaching higher altitudes. Those stronger air parcels enable water-vapor layers to be

351   deepened in the lower stratosphere. These deepened layers and their greater water-vapor

352   mass contribute to more interception of longwave radiation by water vapor in the UTLS

353   over the pyroCb.

354   Similar to the situation with updraft mass fluxes, there is only a small (~2%) increase

355   in the average mass of the UTLS water vapor in the control run versus the low-aerosol run

356   for strong fire intensity (Figure 5 and Table 2). The small variation in updraft mass fluxes

357   between the runs results in a small variation in the average water-vapor fluxes at the

358   tropopause from 8.30×10^{-6} kg m^{-2} s^{-1} in the control run to 8.21×10^{-6} kg m^{-2} s^{-1} in the low-

359   aerosol run over cloudy columns for the simulation period between 17:00 GMT on August

360   5th and 12:00 GMT on August 6th.

361  The altitude of homogeneous freezing is 9 km, so cirrus clouds composed of ice crystals

362 (or cloud ice) are between 9 km and 13 km in the control run (Figure 6). The amount of

363 cirrus clouds in the control run, represented by the average cloud-ice mass density, ranges

364 from 0.028 to 0.037 g m^{-3} between 9 km and 13 km (Figure 6). The average cloud-ice

365 number concentration and size, represented by ice-crystal volume mean radius, between 9

366 km and 13 km range from 6 to 20 cm^{-3}, and from 10 to 20 μm, respectively. Henceforth,

367 "the UTLS cirrus clouds" refer to clouds in the upper troposphere.

368  Updrafts produce supersaturation, which leads to the primary source of cloud-ice mass

369 and associated cirrus clouds via deposition. Due to the negligible variation of updraft mass

370 fluxes, there are negligible variations of supersaturation and deposition (Figure 7). So, there

371 is only a negligible increase (~4%) in UTLS cirrus cloud mass in the control run compared

372 to that in the low-aerosol run (Figure 6 and Table 2). However, mainly due to the larger

373 aerosol concentrations, and associated greater homogeneous aerosol and droplet freezing,

374 there is a significant ~20-fold increase in cloud-ice number concentration and associated

375 with this, there is a significant ~2-fold decrease in cloud-ice size in the control run between

376 9 km and 13 km.

377  In summary, the pyroCb and associated updrafts cause a substantial enhancement of

378 the transport of water vapor to the UTLS at and above the tropopause. Wang (2019) also

379 reported this enhancement. Using modeling work and satellite observation, Wang (2019)

380 indicated that the upward transport of water vapor in deep convective storms was possibly

381 a major pathway through which water substance entered the stratosphere. Wang (2019)

382 showed that the upward transport of water vapor was aided by gravity wave and its

383 breaking in overshooting convective parcels. The pyroCb and its updrafts also produce

384 cirrus clouds. The effects of fire-generated aerosols on the pyroCb updrafts, cirrus cloud

385 mass and the enhancement of the water vapor transport are insignificant when fire intensity

386 is strong.

387

388 **4.2 Dependence of aerosol effects on fire intensity**

389

390 When fire-generated surface heat fluxes and fire intensity increase, in-cloud latent heat is

391 also likely to increase because a major source of in-cloud latent heating is surface heat flux.

392   Therefore, aerosol-induced perturbations of latent heating may be relatively small

393   compared with large in-cloud latent heat contributed by surface fluxes with very intense

394   burning. Thus, aerosol-induced increases in parcel buoyancy, updrafts and their impacts on

395   water vapor and the amount of cirrus clouds are relatively small compared with the large

396   buoyancy, strong fire-driven updrafts produced by strong fire intensity and their associated

397   impacts on water vapor and the amount of cirrus clouds.

398   When fire intensity and fire-generated surface heat fluxes decrease, in-cloud latent heat

399   is also likely to be smaller. Here, we are interested in how the magnitude of an aerosol-

400   induced perturbation of latent heating for a pyroCb with weak fire intensity compares to

401   that with strong fire intensity. This is to evaluate the possibility that with background in-

402   cloud latent heat varying with fire intensity, the relative magnitudes of aerosol-induced

403   perturbations of latent heat and surface flux-dominated latent heat may vary.

404

405   **4.2.1 Effects of Updrafts on the UTLS water vapor and cirrus clouds**

406

407   The average updraft mass fluxes in the low-aerosol, medium-low and weak-low runs

408   represent fire-driven updrafts for strong, medium and weak fire intensities, respectively

409   (Figure 4). Due to different fire intensity and associated CAPE, fire-driven updrafts vary

410   between these runs. All weak, medium and strong fire intensity cases show aerosol-induced

411   increases in updraft mass fluxes (Figure 4 and Table 2). Of interest is that the greatest

412   percentage increase in updraft mass flux is in the case of weak fire (weak-low to weak

413   runs), smallest in the case of strong fire (low-aerosol to control runs), and intermediate in

414   the case of medium fire (medium-low to medium runs) (Figure 4 and Table 2). Here, the

415   percentage difference, including both the percentage increase and decrease, is the relative

416   difference in the value of variables between the polluted-scenario and clean-scenario runs

417   for each fire intensity. The following equation determines this percentage difference for

418   the strong fire intensity case:

419

420     $\dfrac{Control\ run\ minus\ low-aerosol\ run}{Low-aerosol\ run} \times 100\ (\%)$ (1)

421

422   Replacing the control run with the medium (weak) run, and the low-aerosol run with the

423   medium-low (weak-low) run in Equation (1) determines the percentage difference for the

424   medium (weak) fire intensity case. Associated with the greater increases in updraft mass

425   fluxes, the percentage increases in the UTLS water vapor and cloud-ice mass (Equation 1)

426   are greater in the case of weaker fire (Figures 5 and 6 and Table 2).

427   In this section, we see that although fire-produced aerosols invigorate updrafts in all

428   three types of fire intensity, the invigoration-induced increases in the UTLS water-vapor

429   and cloud-ice mass increase as fire intensity weakens.

430

431   **4.2.2 Volume mean radius of droplets (R_v)**

432

433   **a.   Cloud droplet number concentration (N_d) and LWC**

434

435   The simulation period is divided into four sub-periods for this next analysis: period 1

436   between 17:00 and 19:00 GMT on August 5th (initial formation of the pyroCb), period 2

437   between 19:00 and 21:00 GMT on August 5th, and period 3 between 21:00 GMT and 23:00

438   GMT on August 5th (initial stages of cloud development), and period 4 between 23:00 GMT

439   on August 5th and 12:00 GMT on August 6th (mature and decaying stages). The average N_d

440   over period 1 decreases as the fire intensity and updrafts decrease (Figure 8). The polluted-

441   scenario run has higher aerosol concentrations over the fire spot (Table 1), leading to the

442   much higher average N_d in the polluted-scenario run than in the clean-scenario run for each

443   fire intensity. Increasing N_d enhances competition among droplets for a given amount of

444   water vapor. Enhanced competition eventually curbs the condensational growth and

445   reduces droplet size (R_v). This explains why the average R_v over period 1 is smaller in the

446   polluted-scenario run (Figure 8). Of interest is that as fire intensity weakens, although the

447   average N_d decreases, the average R_v decreases not only among the polluted-scenario runs

448   but also among the clean-scenario runs over the fire spot (Figure 8). This is because R_v is

449   proportional to $(\frac{LWC}{N_d})^{\frac{1}{3}}$ based on the following Equation (2):

450

451                $$R_v = (\frac{3}{4\,\pi\rho_w})^{\frac{1}{3}} (\frac{LWC}{N_d})^{\frac{1}{3}} \qquad\qquad\qquad (2)$$

452

453 Here, ρ_w represents water density at a constant value of 1000 kg m^{-3}, so $(\frac{3}{4\pi\rho_w})^{\frac{1}{3}}$ has a

454 constant value. LWC represents the given amount of water available for the condensational

455 growth of droplets. This proportionality means that for a given N_d, a decrease in LWC

456 decreases R_v. The average LWC over period 1 also decreases with weakening fire intensity

457 not only among the polluted-scenario runs but also among the clean-scenario runs (Figure

458 8). Effects of LWC on R_v outweigh those of N_d, leading to the decrease in the average R_v

459 with weakening fire intensity (Figure 8).

460   Using the average LWC and N_d from Figure 8, $(\frac{LWC}{N_d})^{\frac{1}{3}}$ decreases by 1.50×10^{-5} kg

461 from 3.50×10^{-5} kg in the control run to 2.00×10^{-5} kg in the weak run, while it decreases by

462 9.80×10^{-6} kg from 1.03×10^{-4} kg in the low-aerosol run to 9.32×10^{-5} kg in the weak-low run.

463 Associated with this, the average R_v shows a 47 % reduction from the control to weak runs,

464 while it shows a 10 % reduction from the low-aerosol to weak-low runs during period 1

465 (Figure 8).

466   In summary, the simulated LWC, N_d, their variations with fire intensity, and the

467 functional relation between LWC, N_d and R_v lead to a situation where R_v decreases much

468 more among the polluted-scenario runs than among the clean-scenario runs during the

469 period when the pyroCb initially forms.

470

471   **b. Equilibrium supersaturation**

472

473 During period 1, as fire intensity weakens and updraft speed decreases, parcel equilibrium

474 supersaturation decreases and thus, the minimum size of activated aerosol particles

475 increases not only among the clean-scenario runs but also among the polluted-scenario runs.

476 When the production of parcel supersaturation by updrafts and the consumption of

477 supersaturation by droplets balance out, parcel supersaturation reaches parcel equilibrium

478 supersaturation (Rogers and Yau, 1991). Mostly due to greater aerosol concentrations,

479 associated average equilibrium supersaturation and minimum size of activated aerosol

480 particles over areas with positive updraft speeds and period 1 are lower and larger,

481 respectively, in the polluted-scenario run than in the clean-scenario run for each fire
482 intensity (Rogers and Yau, 1991).

483   The average equilibrium supersaturation decreases from 0.21% in the control run to
484 0.10% in the weak run. Associated with this, the average minimum diameter increases from
485 0.09 μm in the control run to 0.12 μm in the weak run over period 1. The average
486 equilibrium supersaturation decreases from 0.55% in the low-aerosol run to 0.31% in the
487 weak-low run, and the average minimum size increases from 0.04 μm in the low-aerosol
488 run to 0.07 μm in the weak-low run over period 1.

489   The increase in the minimum-activation size with weakening fire intensity occurs
490 closer to the right tail of the assumed unimodal aerosol size distribution among the
491 polluted-scenario runs than among the clean-scenario runs. A smaller portion of the total
492 aerosol concentration is in the size range closer to the right tail of the distribution as long
493 as the range is on the right-hand side of the distribution peak where most of aerosol
494 activation occurs. So, a similar increase in the average minimum-activation size for a
495 weakened fire results in a smaller percentage reduction in the total activated aerosol
496 concentration and thus N_d among the polluted-scenario runs during period 1. The average
497 N_d over period 1 decreases by 8% from the control to weak runs. The average N_d decreases
498 by 76% from the low-aerosol to weak-low runs (Figure 8). This contributes to a greater
499 reduction in R_v as fire intensity weakens among the polluted-scenario runs during period 1.
500 This is for a similar LWC between the polluted-scenario and clean-scenario runs for each
501 fire intensity (Figure 8).

502   In association with larger aerosol concentration and the assumed aerosol size
503 distribution, a smaller percentage variation of the number of activated aerosols and N_d with
504 fire intensity is simulated in the polluted-scenario runs than in the clean-scenario runs. This
505 smaller variation of N_d aids the greater reduction in R_v among the polluted-scenario runs.

506

507   **4.2.3 Autoconversion, freezing, deposition and condensation**

508

509 Autoconversion is proportional to the size of cloud droplets (Pruppacher and Klett, 1978;
510 Rogers and Yau, 1991; Khairoutdinov and Kogan, 2000; Liu and Daum, 2004; Lee and
511 Baik, 2017). Due to the larger R_v during period 1, the subsequent average autoconversion

512   rates over period 2 are higher in the clean-scenario run than in the polluted-scenario run

513   for each fire intensity (Figure 9a). Due to the larger absolute and percentage reduction in

514   R_v among the polluted-scenario runs than among the clean-scenario runs with weakening

515   fire intensity during period 2, the average autoconversion rates decrease by 74% (14%)

516   from the control (low-aerosol) to weak (weak-low) runs (Figure 9a). Associated with this,

517   differences in the average autoconversion rates between the polluted-scenario and clean-

518   scenario runs increase as fire intensity weakens during period 2 (Figure 9a).

519   The increasing differences in autoconversion rates between the polluted-scenario

520   and clean-scenario runs increase those differences in the amount of cloud liquid available

521   for freezing with weakening fire intensity (Figure 9a). Thus, differences in the average rate

522   of cloud-liquid freezing and freezing-related latent heat over the period 2 between the runs

523   increase with weakening fire intensity (Figure 9a). Enhanced freezing-related latent heat

524   strengthens updrafts in places where freezing occurs and this, in turn, enhances deposition

525   and deposition-related latent heat (Lee et al., 2017). Although the average deposition over

526   period 2 is slightly lower, those strengthened updrafts enable the average deposition and

527   deposition-related latent heat to be greater in the polluted-scenario run than in the clean-

528   scenario run for each fire intensity during period 3 (Figures 9a and 9b). Differences in the

529   average freezing rate (and thus the average freezing-related latent heating) between the

530   runs do not change much up to ~20:30 GMT (Figure 10). However, after ~20:30 GMT,

531   these differences start to increase as time goes by for each fire intensity. This is because as

532   convection intensifies, the transport of cloud liquid to places above the freezing level starts

533   to be effective around 20:30 GMT. The greater freezing and thus freezing-related latent

534   heat eventually cause updrafts to be stronger in the polluted-scenario run starting at ~21:00

535   GMT (Figure 10). Then, the stronger updrafts induce deposition to be greater in the

536   polluted-scenario run around 21:10 GMT (Figure 10). Note that deposition-related latent

537   heat is about one order of magnitude greater than freezing-related latent heat for a unit

538   mass of hydrometeors involved in phase-transition processes. This contributes to much

539   greater differences in deposition-related latent heat during period 3 than those in freezing-

540   related latent heat between the runs during period 2 or 3 (Figures 9a and 9b).

541   To satisfy mass conservation, the freezing- and deposition-enhanced updrafts above

542   the freezing level induce more updraft mass fluxes below the freezing level in the polluted-

543   scenario run than in the clean-scenario run for each fire intensity. This leads to more

544   convergence around and below cloud base in the polluted-scenario run. The higher mass

545   fluxes and convergence below the freezing level, in turn, increase condensation starting

546   around 22:30 GMT in the polluted-scenario run (Figure 10). This induces the greater

547   average condensation and condensation-related latent heat in the polluted-scenario run

548   during period 4 (Figure 9c). Enhanced condensation in turn enhances updrafts, establishing

549   a positive feedback between freezing, deposition, condensation, and updrafts, thus further

550   enhancing freezing, deposition, condensation, and updrafts. This enhancement, due to the

551   feedback, eventually determines the overall differences in the pyroCb properties and their

552   impacts on the UTLS water vapor and cloud ice between the runs.

553   Due to the increasing differences in freezing-related latent heat between the polluted-

554   scenario and clean-scenario runs with weakening fire intensity during period 2, those

555   differences in the average freezing-affected updrafts and subsequently in deposition-

556   related latent heat over period 3 increase with weakening fire intensity (Figures 9a, 9b and

557   10). Those differences, calculated using Equation (1), in deposition-related latent heat are

558   16%, 181%, and 417 % for strong, medium, and weak fire intensities, respectively (Figures

559   9b and 10). Since percentage increases in deposition-related latent heat increase, the

560   subsequent percentage increases in updrafts in the polluted-scenario run increase with

561   weakening fire intensity, particularly during period 3 (Figure 10). During period 4, due to

562   these greater increases in updrafts in the polluted-scenario run with weaker fire intensity,

563   the percentage increases in condensation in the polluted-scenario run increase with

564   weakening fire intensity (Figures 9c and 10). The greater increases in condensation cause

565   the greater further enhancement of the increases in updrafts in the polluted-scenario run

566   with weaker fire intensity. This leads to the overall greater effects of fire-produced aerosols

567   on the UTLS water vapor and ice with weaker fire intensity.

568   This section shows that the smaller R_v leads to lower autoconversion rates and a

569   larger amount of cloud liquid as a source of freezing, which in turn induces higher freezing

570   rates and stronger feedbacks between freezing, deposition, condensation and updrafts in

571   the polluted-scenario run than in the clean-scenario run for each fire intensity. This results

572   in stronger updrafts and their impacts on the UTLS water vapor and ice in the polluted-

573   scenario run. The greater R_v reduction among the polluted-scenario runs than among the

574 clean-scenario runs with weakening fire intensity increases the differences in
575 autoconversion, freezing and the feedbacks between the polluted-scenario and clean-
576 scenario runs as fire intensity weakens. This results in the greater impacts of aerosol-
577 induced stronger updrafts on the UTLS water vapor and ice with weaker fire intensity.
578

4.3 Dependence of aerosol effects on the magnitude of aerosol perturbation

580

581 Table 3 shows that for each of the strong-, medium-, and weak-fire cases, there are
582 increases in the UTLS water-vapor and cirrus-cloud mass in the run with fire-induced
583 aerosol perturbations of 30000 or 7500 cm^{-3}. These increases are relative to the mass in the
584 low-aerosol run for the strong-fire case, in the medium-low run for the medium-fire case,
585 and in the weak-low run for the weak-fire case. Note that for each of the three types of fire-
586 induced aerosol perturbations of 30000, 15000 and 7500 cm^{-3}, aerosol-perturbation-
587 induced percentage increases in the UTLS water-vapor and cirrus-cloud mass increase as
588 fire intensity weakens (Tables 2 and 3). The qualitative nature of results regarding the
589 dependence of the percentage increases in the UTLS water-vapor and cirrus-cloud mass on
590 fire intensity thus does not depend on the magnitude of the fire-induced aerosol
591 perturbation.

592   Until now, we have taken interest in the sensitivity to fire intensity of an aerosol
593 perturbation on pyroCb development, the UTLS water vapor and cirrus clouds. To isolate
594 the sensitivity, we have shown comparisons among sensitivity simulations by varying only
595 fire intensity while maintaining a constant aerosol perturbation. While working well for
596 the isolation aspect, this strategy does not reflect reality well. It may be that weaker fire
597 intensity produces a lower aerosol concentration. This possibility is not that unrealistic,
598 since stronger fires likely involve more material burnt and higher aerosol emissions.

599   With this in mind, we make comparisons among three pairs of simulations: the low-
600 aerosol and control-30000 runs for strong fire vs. the medium-low and medium runs for
601 medium fire vs. the weak-low and weak-7500 runs for weak fire. Among these three pairs,
602 the magnitude of the fire-induced aerosol perturbation decreases with weakening fire,
603 emulating the possibility that weaker fire intensity involves a lower amount of aerosols.
604 The perturbation-related aerosol concentration is 30000 cm^{-3} for strong fire, 15000 cm^{-3} for

605   medium fire, and 7500 cm^{-3} for weak fire. Comparisons among these three pairs show that

606   relative importance of aerosol effects on pyroCb development and its impacts on the UTLS

607   water vapor and cirrus clouds increases for weaker fires, and that it does not matter if the

608   aerosol perturbation decreases or stays constant with weakening fire intensity (Tables 2

609   and 3). In these comparisons, it is also possible that when the fire-induced aerosol

610   perturbation is very low for medium or weak fire intensity, the latent heat perturbation by

611   aerosol perturbation can be very low. This very low latent heat is not large enough to

612   increase the relative importance of those aerosol effects with weakening fire intensity.

613   Based on this, the medium run is repeated with a fire-induced aerosol perturbation of 2000

614   cm^{-3} down from 15000 cm^{-3} (the medium-2000 run). The weak run is repeated with a fire-

615   induced aerosol perturbation of 1000 cm^{-3} down from 7500 cm^{-3} (the weak-1000 run). The

616   percentage increases in the UTLS water-vapor and cirrus-cloud mass from the medium-

617   low to medium-2000 runs or from the weak-low to weak-1000 runs are smaller than the

618   increases from the low-aerosol to control-30000 runs for the case of strong fire. This

619   indicates that when the fire-induced aerosol perturbation decreases too much with

620   weakening fire intensity, the relative importance of aerosol effects on pyroCb development

621   and its impacts on the UTLS water vapor and cirrus clouds no longer increases with

622   weakening fire intensity.

623   Results in this section show that the increasing impacts of the fire-induced aerosol

624   perturbations on the UTLS water vapor and cirrus clouds with weakening fire intensity are

625   robust whether or not aerosol perturbations vary with fire intensity unless their variation is

626   extremely high.

627

628   **5. Conclusions**

629

630   This study investigates an observed case of a pyroCb using a modeling framework. In

631   particular, this study focuses on effects of fire-produced aerosols on pyroCb development

632   and its impacts on the UTLS water vapor and cirrus clouds. Results show that the pyroCb

633   efficiently transports water vapor to the tropopause and above. This leads to a much greater

634   amount of water vapor around and above the tropopause (i.e., the UTLS) over the pyroCb

635   compared to that outside the pyroCb. The pyroCb also generates a deck of cirrus clouds

636   around the tropopause. The role of fire-produced aerosols or the fire-induced aerosol

637   perturbation in the water-vapor transport to the UTLS and the production of cirrus clouds

638   becomes significant as fire intensity weakens.

639   During the initial stage, there is a similar LWC between the polluted-scenario and

640   clean-scenario runs for each fire intensity. The reduction in LWC with weakening fire

641   intensity among the polluted-scenario runs is similar to that among the clean-scenario runs.

642   Much greater N_d is in the polluted-scenario run than in the clean-scenario run for each fire

643   intensity. N_d decreases are smaller among the polluted-scenario runs than among the clean-

644   scenario runs with weakening fire intensity. This situation during the initial stage induces

645   R_v to decrease much more among the polluted-scenario runs with weakening fire intensity.

646   This reduces autoconversion more among the polluted-scenario runs and increases

647   differences in autoconversion between the polluted-scenario and clean-scenario runs as fire

648   intensity weakens. The increasing differences in autoconversion between the runs cause

649   greater differences in freezing-related latent heat as fire intensity weakens. Through

650   feedback between freezing, deposition, updrafts, and condensation, differences in freezing-

651   related latent heat induce those in updrafts between the runs. Those greater differences in

652   freezing-related latent heat lead to greater differences in updrafts, producing the greater

653   differences in the UTLS water vapor and cirrus clouds between the runs with weaker fire

654   intensity. This means that the role of fire-produced aerosols in the water-vapor transport to

655   the UTLS and the production of cirrus clouds becomes more significant as fire intensity

656   weakens. This more significant role of fire-produced aerosols with weaker fire intensity is

657   robust to the magnitude of a given fire-induced aerosol perturbation, and the variation of

658   the fire-induced aerosol perturbation with fire intensity unless the variation is very high.

659   The level of understanding of the role played by fire-produced aerosols in the

660   development of pyroCbs and their impacts on the UTLS water vapor and cirrus clouds has

661   been low. This study shows that fire-produced aerosols can invigorate updrafts and

662   convection and thus enhance the transport of water vapor to the UTLS and the formation

663   of cirrus clouds. We find that the mechanism that controls the invigoration of convection

664   by aerosols in the pyroCb is consistent with the traditional invigoration mechanism

665   proposed and described by Koren et al. (2008) and Rosenfeld et al. (2008). However, this

666   study shows that for pyroCbs produced by strong fires, the aerosol-induced invigoration

667 and its effects on the UTLS water vapor and cirrus clouds are insignificant. Note that

668 traditional understanding generally focuses on effects of fire-produced heat and water

669 vapor and their associated surface fluxes on the pyroCb and does not consider effects of

670 fire-produced aerosols on the pyroCb. This understanding adequately explains the

671 mechanics for pyroCbs in association with strong fires. This study suggests that when

672 pyroCbs form over weak-intensity fires, those effects of fire-produced aerosols require

673 consideration.

674 Note that when fire-induced aerosol perturbations are strongly reduced for cases of

675 weaker-intensity fires compared with strong-intensity fires, the significance of the role of

676 the fire-produced aerosol perturbation no longer increases and starts to decrease with

677 weakening fire. This suggests that there may be a critical level of aerosol perturbation

678 below which the increase in this significance with weakening fire intensity ceases.

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Author contributions

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700 SSL generated the research ideas and goals, performed the simulations, and wrote the
701 manuscript. GK and ZL selected the case, analyzed observations, and provided data to set
702 up the simulations while reviewing and providing comments on the manuscript. CHJ, YSC,
703 JU and WJC revised the manuscript based on the reviewers' comments and performed
704 associated analyses of simulation and observation data.
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924 **FIGURE CAPTIONS**
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927 Figure 1. Visible image of the fire, smoke and cirrus cloud in association with the selected

928 pyroCb. This image was taken by the visible infrared imaging radiometer suite onboard the

929 Suomi spacecraft. Bright white represents cirrus (anvil) at the top of the pyroCb, and the

930 red circle marks the fire spot. Dark white represents smoke produced by the fire. Adapted

931 from Kablick et al. (2018).

932

933 Figure 2. The simulated fire spot (red circle) and the field of cloud-ice mass density (cirrus

934 cloud) at the top of the simulated pyroCb when it is about to enter its mature stage.

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936 Figure 3. The vertical distribution of the radar reflectivity averaged along the Cloudsat path.

937

938 Figure 4. Vertical distributions of the average updraft mass fluxes at all altitudes in cloudy

939 areas (i.e., where the sum of LWC and IWC is non-zero) over the simulation period

940 between 17:00 GMT on August 5th and 12:00 GMT on August 6th.

941

942 Figure 5. Vertical distributions of the average water-vapor mass density at altitudes above

943 13 km and over the simulation period between 17:00 GMT on August 5th and 12:00 GMT

944 on August 6th. Colored lines represent the average values over cloudy grid columns (non-

945 zero sum of LWP and IWP). The black line represents those values over non-cloudy

946 columns (zero sum of LWP and IWP) in the control run.

947

948 Figure 6. Vertical distributions of the average cloud-ice mass density at all altitudes in

949 cloudy areas (non-zero sum of LWC and IWC) over the simulation period between 17:00

950 GMT on August 5th and 12:00 GMT on August 6th.

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952 Figure 7. Same as Figure 6 but for deposition rate.

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954 Figure 8. Average N_d, R_v, and LWC at all altitudes in cloudy areas over the period between

955 17:00 and 19:00 GMT on August 5th.

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957 Figure 9. Average rates of condensation, deposition and cloud-liquid freezing at all
958 altitudes in cloudy areas and over periods (a) 2, (b) 3 and (c) 4. In panel (a), average
959 autoconversion rates are also shown.

960

961 Figure 10. Time series of differences in average values of variables related to aerosol-
962 induced invigoration of convection at all altitudes in cloudy areas between the (a) control
963 and low-aerosol runs for strong fire intensity, (b) medium and medium-low runs for
964 medium fire intensity and (c) weak and weak-low runs for weak fire intensity.

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Simulations	Surface sensible heat fluxes in the fire spot (W m^{-2})	Surface latent heat fluxes in the fire spot (W m^{-2})	Aerosol concentration in the PBL over the fire spot (cm^{-3})
Control run	15000	1800	15000
Low-aerosol run	15000	1800	150
Control-30000	15000	1800	30000
Control-7500	15000	1800	7500
Medium run	7500	900	15000
Medium-low run	7500	900	150
Medium-30000	7500	900	30000
Medium-7500	7500	900	7500
Medium-2000	7500	900	2000
Weak run	3750	450	15000
Weak-low run	3750	450	150
Weak-30000	3750	450	30000
Weak-7500	3750	450	7500
Weak-1000	3750	450	1000

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988   Table 1. Summary of simulations

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	Backg-round	Control	Low-aerosol	Difference (%)	Medium	Medium-low	Difference (%)	Weak	Weak-low	Difference (%)
Updraft mass fluxes (kg m^{-2} s^{-1})		1.23	1.19	3	0.89	0.70	27	0.42	0.21	100
Water-vapor mass density between 13 and 16 km (10^{-3} g m^{-3})	0.46	2.31	2.26	2	1.61	1.32	22	0.93	0.58	60
Cirrus-cloud mass density between 9 and 13 km (g m^{-3})		0.024	0.023	4	0.017	0.012	42	0.008	0.004	100

1008

1009 Table 2. Average updraft mass fluxes at all altitudes in cloudy areas, the average water-

1010 vapor mass density over altitudes between 13 and 16 km and over cloudy columns except

1011 for the average background water-vapor mass density which is also over altitudes between

1012 13 and 16 km but over non-cloudy columns, and the average cirrus-cloud mass density

1013 between 9 and 13 km in cloudy areas. 16 km is the altitude to which the non-zero water-

1014 vapor mass density over cloudy columns extends (Figure 5). These average values are

1015 obtained over the simulation period between 17:00 GMT on August 5th and 12:00 GMT on

1016 August 6th. "Difference" is the percentage difference between the polluted-scenario and

1017 clean-scenario runs for each fire intensity ($\frac{Polluted-scenario\ run\ minus\ clean-scenario\ run}{Clean-scenario\ run} \times$

1018 100 (%)).

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	Control-30000	Control-7500	Medium-30000	Medium-7500	Medium-2000	Weak-30000	Weak-7500	Weak-1000
Water vapor mass density between 13 and 16 km (10^{-3} g m^{-3})	2.38 (5%)	2.28 (0.9%)	1.87 (42%)	1.50 (14%)	1.36 (3%)	1.31 (125%)	0.75 (29%)	0.60 (3%)
Cirrus cloud mass density between 9 and 13 km (g m^{-3})	0.025 (9%)	0.023 (0.2%)	0.023 (92%)	0.014 (17%)	0.012 (3%)	0.013 (225%)	0.006 (50%)	0.004 (8%)

1026

1027 Table 3. Average water-vapor mass density between 13 and 16 km over cloudy columns
1028 and the average cirrus-cloud mass density between 9 and 13 km in cloudy areas over the
1029 simulation period between 17:00 GMT on August 5th and 12:00 GMT on August 6th. The
1030 numbers in parentheses are the percentage differences:
1031 $\frac{Control-30000\ (or\ control-7500)\ run\ minus\ low-aerosol\ run}{Low-aerosol\ run} \times 100$ (%) for strong fire intensity,
1032 $\frac{Medium-30000\ (or\ medium-7500\ or\ medium-2000)\ run\ minus\ medium-low\ run}{Medium-low\ run} \times 100$ (%) for
1033 medium fire intensity, and
1034 $\frac{Weak-30000\ (or\ weak-7500\ or\ weak-1000)\ run\ minus\ weak-low\ run}{Weak-low\ run} \times 100$ (%) for weak fire
1035 intensity.
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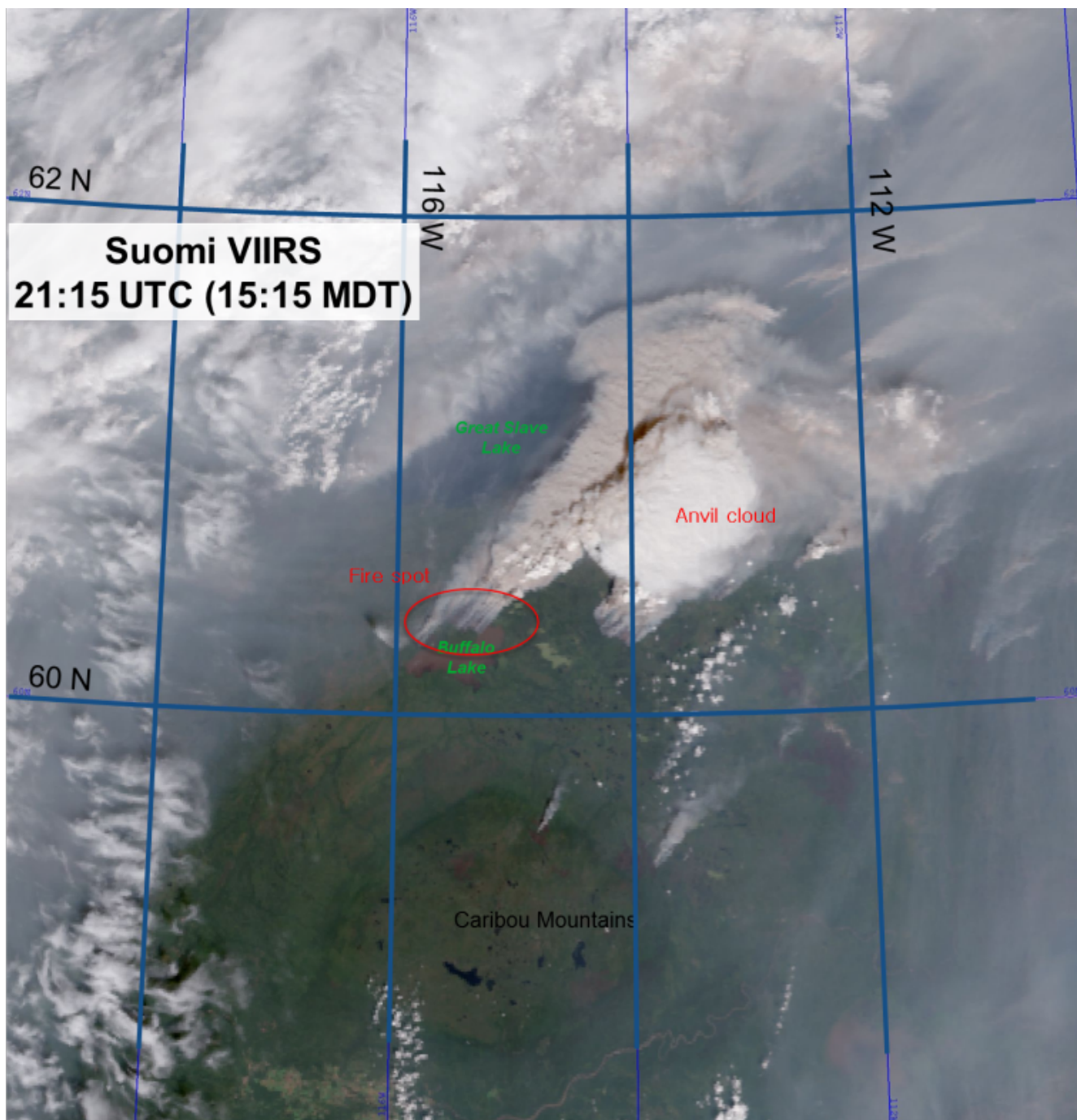
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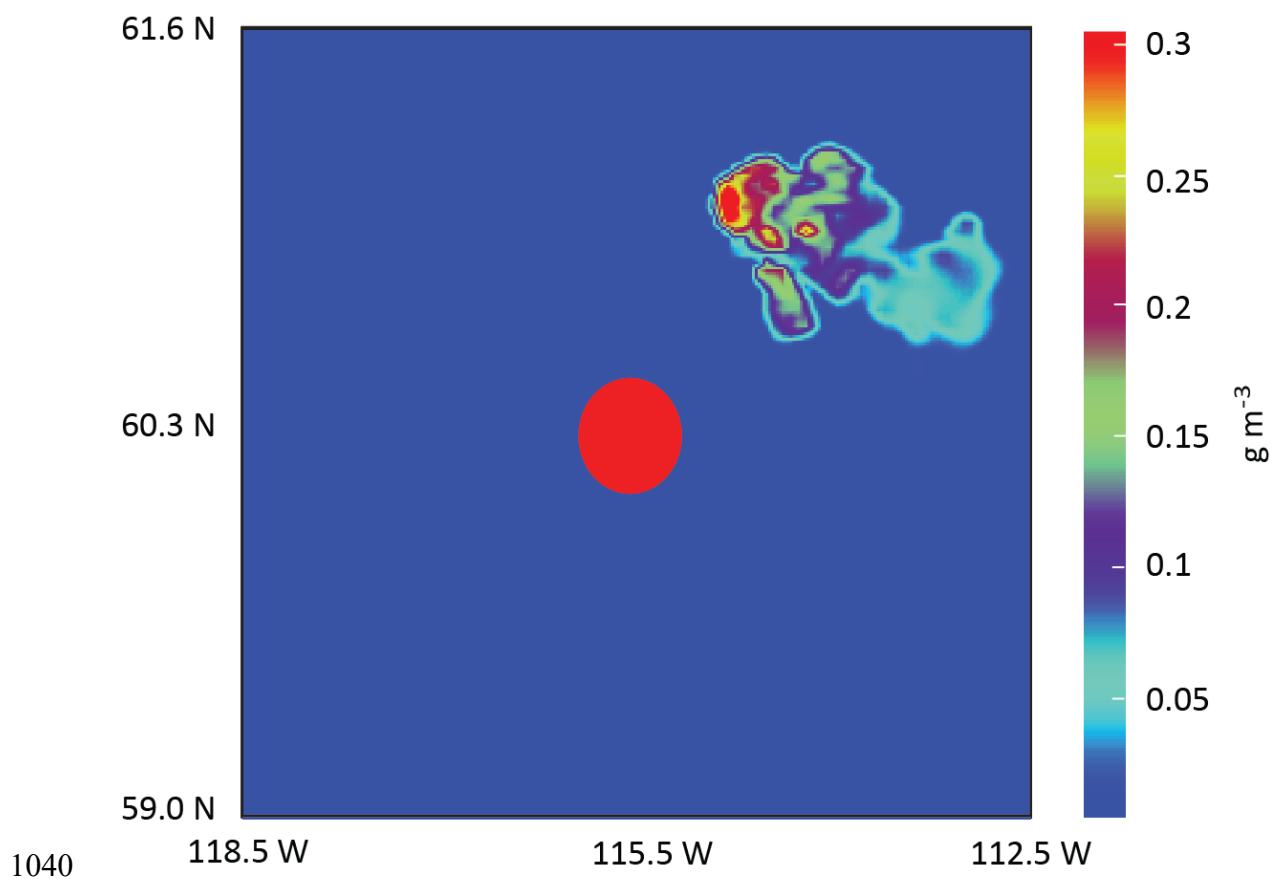
Figure 1

Figure 2

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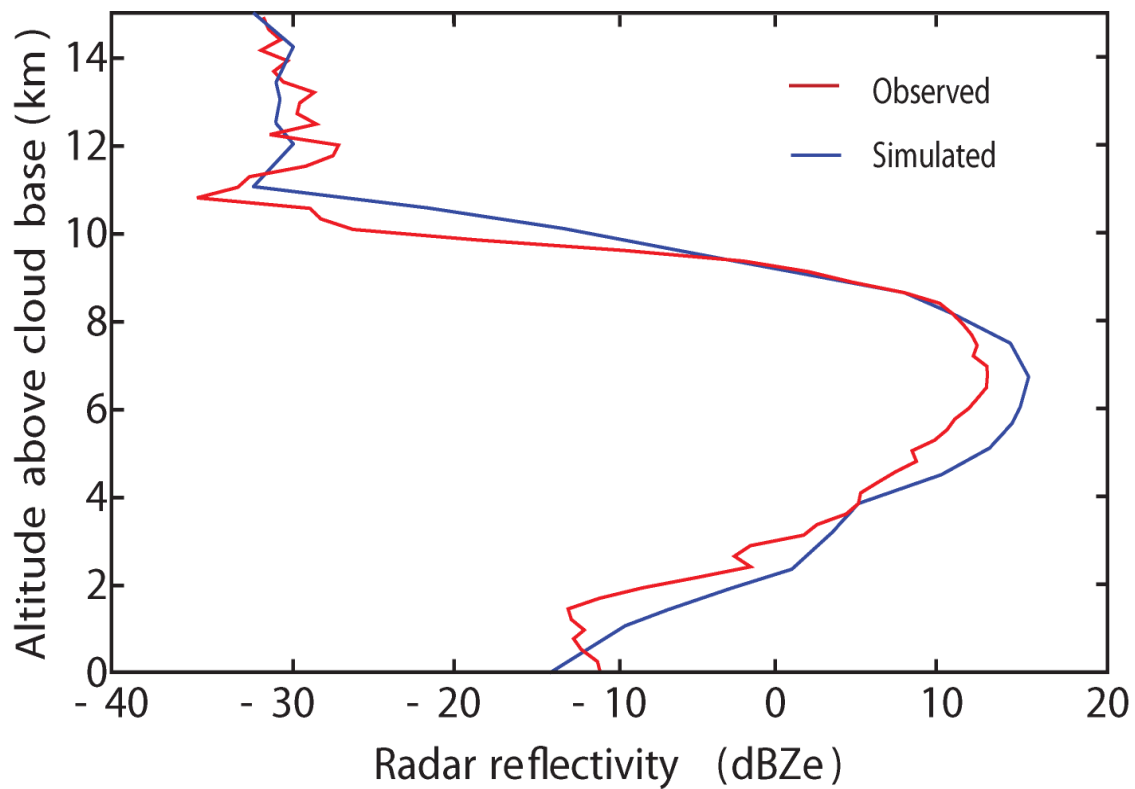
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Figure 3

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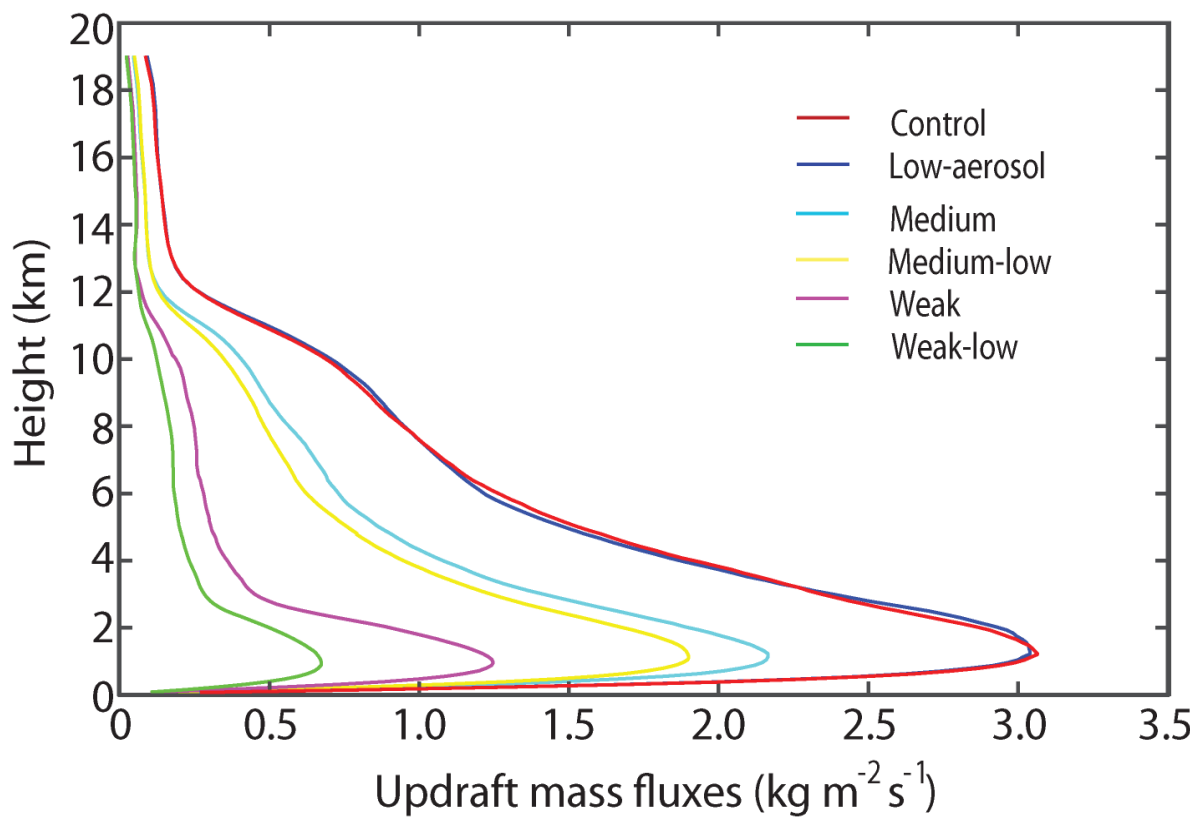
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Figure 4

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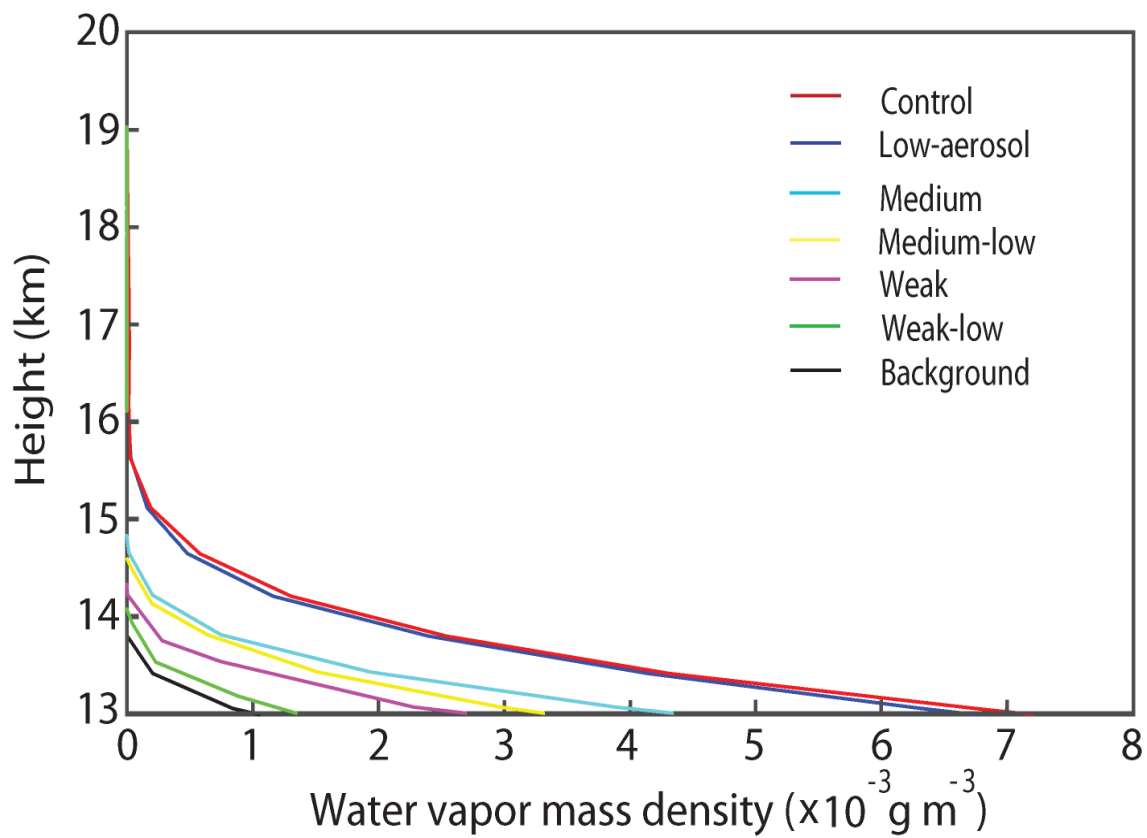
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Figure 5

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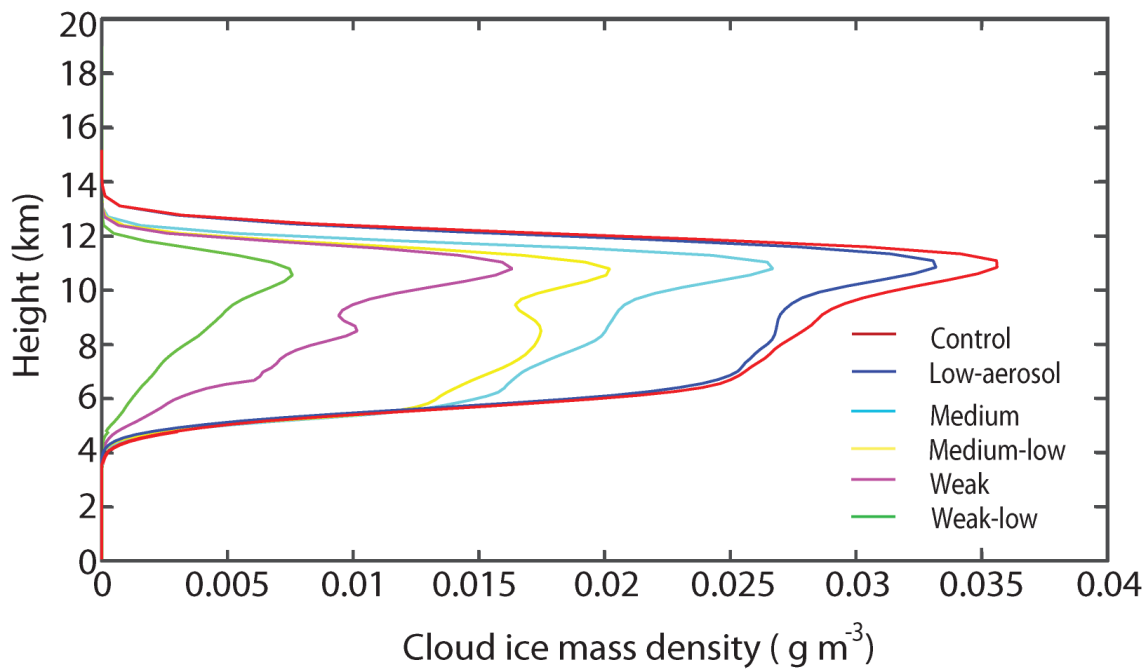
Figure 6

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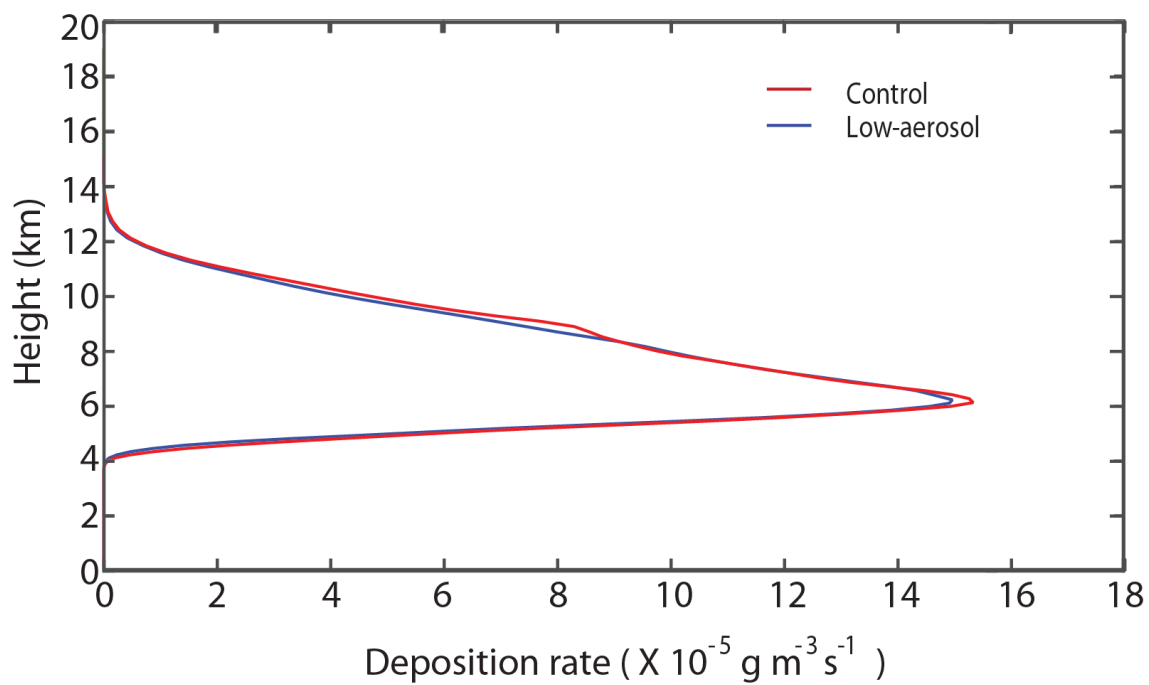
Figure 7

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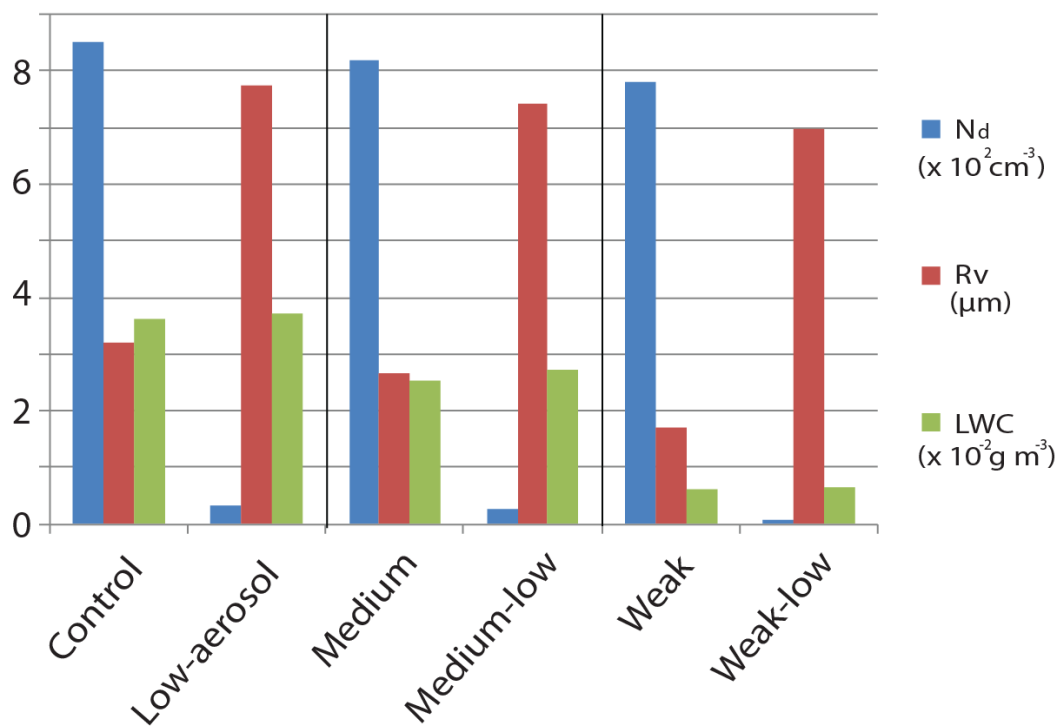
Period 1 (17 GMT - 19 GMT on August 5th; initial stage)

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Figure 8

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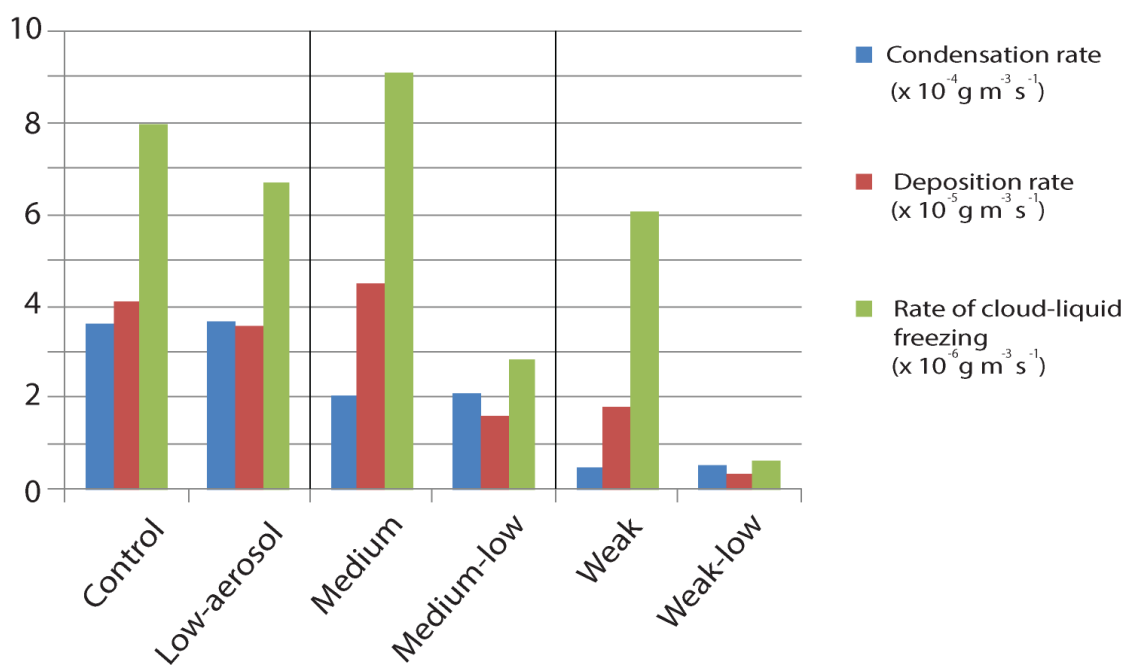
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1109 **Figure 9**

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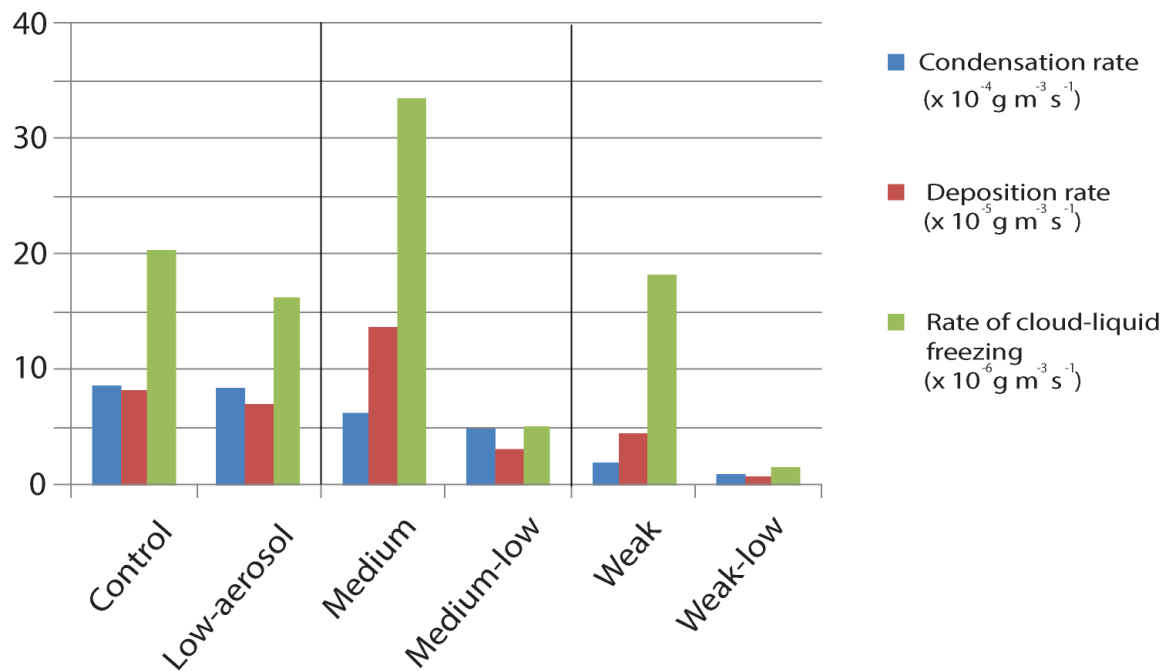
Figure 9

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Differences in the averaged values

1118 **Figure 10**

c Weak run minus Weak-low run
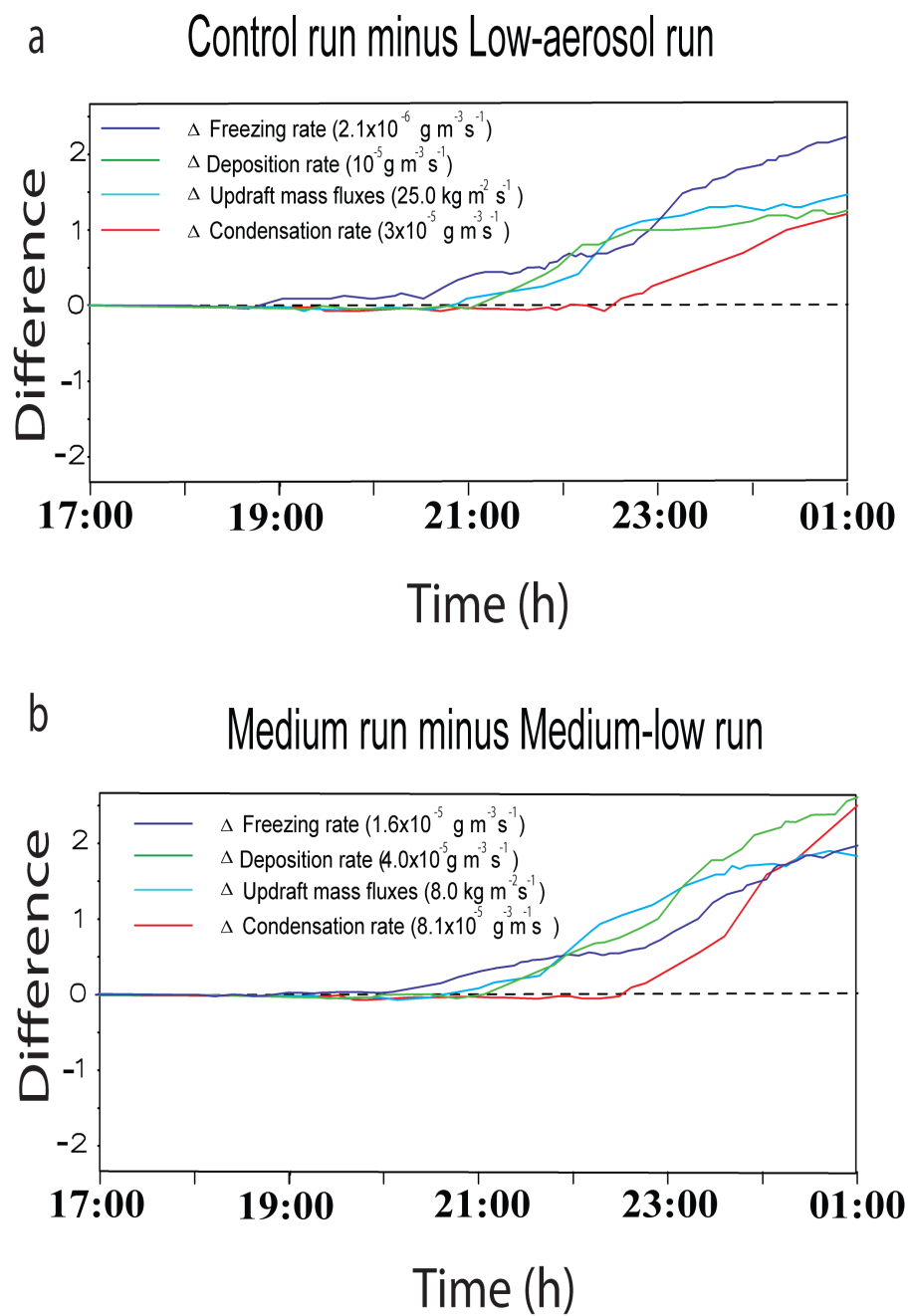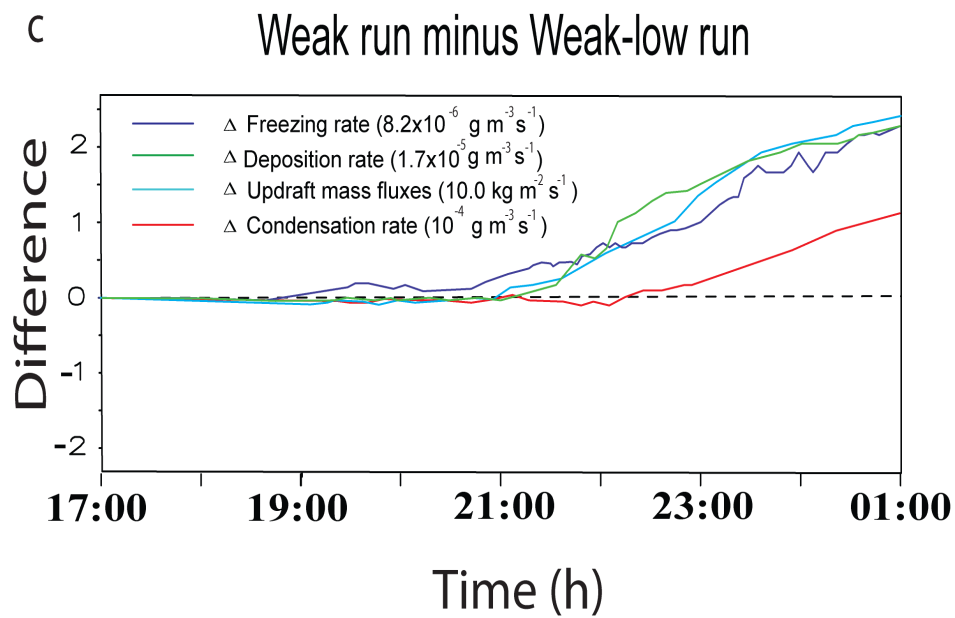
Figure 10

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