Examination of effects of aerosols on a pyroCb and their dependence on fire intensity and aerosol perturbation using a cloud-system resolving model

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Abstract

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 via a modeling framework. Results from a case study show that when observed fire intensity is high, there is an insignificant impact of fire-produced aerosols on the convective development of the pyroCb and associated changes in water vapor and the amount of cirrus cloud near the tropopause. However, as fire intensity weakens, the effects of aerosols on microphysical variables and processes such as droplet size and autoconversion increase. Modeling results shown herein indicate that aerosol-induced invigoration of convection is significant for pyroCb with weak-intensity fires and associated weak surface heat fluxes. Thus, there is a greater aerosol effect on the transportation of water vapor to the upper troposphere and the production of cirrus cloud with weak-intensity fires, whereas these effects are muted with strong-intensity fires.
1. Introduction

A recent study by Kablick et al. (2018) has shown that pyrocumulonimbus (pyroCbs) can transport significant amounts of water vapor to the upper troposphere and the lower stratosphere (UTLS) and thus may have a role in seasonal UTLS water vapor budgets. Any change in water vapor in the UTLS has an exceptionally strong influence on the global radiation budget and thus Earth’s climate (Solomon et al., 2010). PyroCbs involve and control cirrus clouds around their tops that reach the UTLS. Changes in cirrus clouds in the UTLS are known to have a strong influence on the global radiation budget (Solomon et al., 2010). The examination of mechanisms through which pyroCbs affect water vapor and cirrus clouds in the UTLS can thus be a way of better understanding climate changes.

By definition, pyroCbs initiate over a fire, and the large surface energy release affects their dynamic, thermodynamic and microphysical development. The dynamics of these events has been shown to be mostly controlled by fire-induced latent and sensible heat fluxes at and near the surface. However, questions remain about what role the large concentration of cloud condensation nuclei (CCN) contained in smoke has on the vertical development and microphysical properties. Studies (e.g., Rosenfeld et al., 2008; Storer et al., 2010; Tao et al., 2012) have shown that aerosols affect cumulonimbus clouds, and this raises a possibility that fire-generated aerosols affect pyroCb development. As an example of aerosol impacts on cumulonimbus clouds, these studies have demonstrated that increases in aerosol loading can make the size of droplets (i.e., cloud-liquid particles) smaller. Individual aerosol particles act as seeds for the formation of droplets and thus increasing aerosol loading or increasing aerosol concentrations lead to more droplets formed. More droplets mean more competition among them for available water vapor needed for their condensational growth, and this more competition makes individual droplets smaller. Aerosol-induced smaller sizes of droplets reduce the efficiency of the growth of cloud-liquid particles to raindrops via autoconversion that is a collection process among cloud-liquid particles for them to grow to be raindrops, given that the efficiency is proportional to the sizes. This reduced efficiency leads to less cloud liquid converted to rain. More cloud liquid is thus available for transport to places above the freezing level by updrafts.
This eventually induces more freezing of cloud liquid, which enhances parcel buoyancy, and this enhancement invigorates updrafts and associated convection.

Compared to the research done on the role played by fire-generated heat fluxes in the development of pyroCbs and their effects on water vapor and cirrus clouds in the UTLS, the research on that role by fire-generated aerosols has been scarce. Motivated by this lack of understanding, this paper focuses on the role by those aerosols in the development of a pyroCb and its effects on water vapor and cirrus clouds in the UTLS. To examine that role, this study extends the previous modeling work that was described in Kablick et al. (2018). That modeling work compared effects of fire-generated heat fluxes on the development of a pyroCb and its impacts on the UTLS water vapor and cirrus clouds to those of fire-generated aerosols. In that comparison, those effects of fire-generated aerosols were shown to be negligible as compared to those effects of heat fluxes. However, aerosol effects on cloud development vary with cloud basic properties such as basic updrafts that are determined by environmental conditions (e.g., Khain et al., 2008; Lee et al., 2008; Tao et al., 2012). Basic updrafts are determined by environmental instability as represented by convective available potential energy (CAPE). Lee et al. (2008) have shown that different clouds with different basic updrafts, which are due to different CAPE, show different sensitivity of cloud microphysical and thermodynamic development to aerosol concentration. Hence, it is hypothesized that aerosol effects on the pyroCb development and its impacts on the UTLS water vapor and cirrus clouds can vary depending on the intensity of the pyroCb basic updrafts.

Based on this hypothesis, to examine the potential variation of aerosol effects on the pyroCb development and its impacts on the UTLS water vapor and cirrus clouds with the varying basic updrafts of pyroCbs, numerical simulations are performed. These simulations are for a case of a pyroCb which is identical to that in Kablick et al. (2018), and performed by using a cloud-system resolving model (CSRM) which is able to resolve cloud-scale dynamic, thermodynamic and microphysical processes. By resolving these processes that play a critical role in the development of clouds and their interactions with aerosols, we are able to obtain confident information on aerosol effects on the pyroCb development and its impacts the UTLS water vapor and cirrus clouds, and on associated dynamic, thermodynamic and microphysical mechanisms. The basic modeling 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 study, simulating microphysical conditions, detrained water vapor mixing ratios, and cirrus cloud properties only considering a basic updraft framework. The present study expands upon that work by performing sensitivity simulations in which basic updrafts in the pyroCb are allowed to vary, enabling us to ascertain the dependence of those aerosol effects on basic updrafts. Note that CAPE, which determines basic updrafts in convective clouds, are strongly dependent on surface latent and sensible heat fluxes (e.g., Houze, 1993), and in the case of pyroCb these fluxes are controlled by fire intensity. Hence, these sensitivity simulations in turn enable us to study the dependence of those aerosol effects on fire intensity. Here, we see that the pyroCb basic updrafts are controlled by fire intensity and thus the pyroCb basic updrafts are referred to as fire-driven updrafts, henceforth.

Aerosol effects on clouds are initiated by an increase in aerosol concentration, which can be caused by an increase in aerosol emission at and near the surface, and dependent on how much aerosol concentration increases, or on the magnitude of an increase in aerosol concentration, i.e., aerosol perturbation (e.g., Rosenfeld et al., 2008; Koren et al., 2012). This dependence has not been examined in Kablick et al. (2018) and this study examines this dependence by performing additional sensitivity simulations where the magnitude of aerosol perturbation varies.

2. CSRM

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 parameterizations have been included in all simulations by adopting the Rapid Radiation Transfer Model (RRTM; Mlawer et al., 1997; Fouquart and Bonnel, 1980).
To represent the 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 $k$th bin is determined as $m_k = 2m_{k-1}$.

The cloud-droplet nucleation parameterization, which is based on Köhler theory, is used to represent cloud-droplet nucleation. Arbitrary aerosol mixing states and arbitrary aerosol size distributions can be fed to this parameterization. To represent heterogeneous ice-crystal nucleation, the parameterizations by Lohmann and Diehl (2006) and Möhler et al. (2006) are used. In these parameterizations, contact, immersion, condensation-freezing, and deposition nucleation paths are all considered by taking into account the size distribution of IN. Homogeneous aerosol (or haze particle) and droplet freezing, based on the size distribution of droplets, is also considered following the theory developed by Koop et al. (2000).

3. Case description and simulations

3.1 Control run

The control run for an observed pyroCb case is performed over a forested site in the Canadian Northwest Territories (60.03° N, 115.45° W). Kablick et al. (2018) give details about the site and the pyroCb case. The control run is identical to the Full Simulation in Kablick et al. (2018) except for the different microphysical schemes between them; remember that this study uses a bin scheme, while Kablick et al. (2018) used a bulk scheme. The control run is performed for one day from 12:00 GMT on August 5th to 12:00 GMT on August 6th in 2014 and captures the initial, mature, and decaying stages of the pyroCb. This simulation is performed in a three dimensional domain with horizontal and vertical lengths of 300 km and 20 km, respectively. For the simulation, the horizontal resolution is
500 m and the vertical resolution is 200 m to resolve cloud dynamic, thermodynamic and microphysical processes.

Figure 1 shows a satellite image of the observed pyroCb when it is about to advance into its mature stage. In Figure 1, the red circle marks the fire spot whose spatial length is \( \sim 40 \) km. To emulate this in the simulation, at the center of the simulation domain, a fire spot with a diameter of 40 km is placed as shown in Figure 2. In the fire spot, the surface latent and sensible heat fluxes are set at 1800 and 15000 W m\(^{-2}\), respectively. In areas outside of the fire spot in the domain, the surface latent and sensible heat fluxes are set at 310 and 150 W m\(^{-2}\), respectively. These surface heat-flux values follow the previous studies which are Trentmann et al. (2006) and Luderer et al. (2006) and adopt boreal forest emissions. Following Kablick et al. (2018), the surface heat-flux values are prescribed with no temporal variation and no consideration of interactions between heat fluxes and the atmosphere in the control run. Hence, the setup for the surface heat fluxes is idealized and this enables a better isolation of aerosol effects themselves on the pyroCb development and its impacts on the UTLS water vapor and cirrus clouds for the given surface heat fluxes by excluding effects of interactions between the surface heat fluxes and atmosphere on those development and impacts.

For the selected pyroCb case, aerosol properties that can be represented by aerosol chemical composition, size distribution and concentration are unknown. Hence, in the fire spot for the first time step, the concentration of aerosols acting as CCN is prescribed to be 15000 cm\(^{-3}\) in the planetary boundary layer (PBL), and decreases exponentially with height above the PBL top. Outside of the fire spot for the first time step, the concentration of aerosols acting as CCN is prescribed to be 150 cm\(^{-3}\) in the PBL and also decreases exponentially with height above this layer. For the control run, the other aerosol properties are assumed to follow typical values determined in previous studies. For example, Reid et al. (2005) have shown that aerosol mass produced by forest fires is generally composed of \(~50-70\%\) of organic-carbon (OC) compounds, \(~5-10\%\) of black-carbon (BC) material, and \(~20-45\%\) of inorganic species. Based on those results, the approximate median value of each chemical component percentage range is used in the control run. Aerosol particles are assumed to be composed of 60\% OC, 8\% BC, and 32\% inorganic species. In the control run, OC is assumed to be water soluble and composed of (by mass) 18 \% levoglucosan
8 (C6H10O5, density = 1600 kg m\(^{-3}\), van’t Hoff factor = 1), 41% succinic acid (C4H6O4, density = 1572 kg m\(^{-3}\), van’t Hoff factor = 3), and 41% fulvic acid (C33H32O19, density = 1500 kg m\(^{-3}\), van’t Hoff factor = 5) based on typically observed chemical composition of OC compounds over fire sites (Reid et al., 2005). In the control run, the inorganic species is assumed to be ammonium sulfate, a representative inorganic species associated with fires (Reid et al., 2005). This chemical composition taken for aerosol particles is assumed to be spatiotemporally unvarying in the control run. According to Reid et al. (2005), Knobelspiessel et al. (2011), and Lee et al. (2014), it is reasonable to assume that the initial aerosol size distribution follows the unimodal lognormal distribution in fire sites. Hence, the control run adopts the unimodal lognormal distribution as an initial aerosol size distribution. Those studies have indicated that in general, median aerosol diameter and standard deviation of the distribution range from \(~0.01\) to \(~0.03\) μm and from \(~2.0\) to \(~2.2\), respectively. By taking the approximate median value of each of these ranges, median aerosol diameter and standard deviation of the adopted unimodal distribution are assumed to be 0.02 μm and 2.1, respectively, for the control run. The unimodal distribution, which is adopted by the simulation, in the PBL over the fire spot is shown in Figure 3 with log-scales for x- and y-axes. For the control run, aerosol properties of IN and CCN are assumed to be identical except that at the first time step, the IN concentration is 100 times lower than the CCN concentration. This is based on a general difference in concentration between CCN and IN (Pruppacher and Klett, 1978).

In Figure 1, the observed cirrus cloud at the top of the pyroCb is located to the northeast of the fire spot due to the northeastward winds at the altitude of the cirrus cloud. The cloud first formed around the fire spot. However, winds advected it northeastward. The extent of the observed cirrus cloud is \(~100\) km. Figure 2 shows the field of cloud-ice mass density at the top of the simulated pyroCb and at a time that corresponds to the satellite image in Figure 1. This field in Figure 2 represents the simulated cirrus cloud in the control run. As observed, the simulated cirrus is located to the northeast of the fire spot and the extent of the simulated cirrus cloud is \(~100\) km. Hence, we see that there is good agreement in the morphology of the cirrus cloud between the observation and the simulation.
Figure 4 shows the vertical distribution of the cloud reflectivity field in dBZ, which is observed by the Cloudsat and averaged over the Cloudsat path, and its simulated counterpart in the control run. The details of the reflectivity field are given in Kablick et al. (2018). There is good agreement between observed and simulated cloud reflectivity fields. The agreement in the observed and simulated cirrus cloud and reflectivity fields demonstrates that the pyroCb-case simulation is reasonable.

3.2 Low-aerosol run

To see the role played by fire-generated aerosols in the development of the pyroCb and its effects on water vapor and cirrus clouds in the UTLS, we repeat the control run by reducing aerosol concentration in the fire spot from 15000 cm$^{-3}$ to the background aerosol concentration (i.e., 150 cm$^{-3}$). This reduction removes fire-generated aerosols in the fire spot. The only difference is in aerosol concentration in the fire spot and there are no other differences in the simulation setup which is described in Section 3.1 between the control run and this repeated run. Hence, comparisons between the control run and this repeated run, which is referred to as the low-aerosol run, will identify the role played by fire-generated aerosols in the pyroCb development and its impacts on the UTLS water vapor and cirrus clouds. Here, the low-aerosol run is identical to the Low Aerosol Simulation in Kablick et al. (2018) except for the different microphysical schemes between them.

3.3 Additional runs

We examine the above-mentioned potential variation of effects of fire-generated aerosols on the pyroCb development and its impacts on the UTLS water vapor and cirrus clouds with varying fire intensity and associated fire-driven updrafts. For the examination, we repeat the control run by varying fire intensity. Remember that surface latent and sensible heat fluxes on which fire-driven updrafts in convective clouds are strongly dependent are controlled by fire intensity. Hence, fire intensity can be represented by fire-induced surface latent and sensible heat fluxes and the variation of fire intensity in the repeated control run is accomplished by the variation of these fire-induced surface heat fluxes. As a first step
for the examination, the control run is repeated by reducing fire-induced surface latent and sensible heat fluxes by a factor of 2. Then, the control run is repeated again by reducing these fluxes by a factor of 4. The first repeated run represents a case with medium fire intensity, while the second repeated run represents a case with weak fire intensity. Relative to these repeated runs, the control run represents a case with strong fire intensity. Henceforth, the first repeated run is referred to as “the medium run” and the second repeated run is referred to as “the weak run”. Then, to see effects of fire-generated aerosols on the pyroCb development and its impacts on the UTLS water vapor and cirrus clouds for each of those cases with different fire intensity, the medium run and the weak run are repeated with the identical initial aerosol concentration to that in the low-aerosol run. The repeated medium run and weak run are referred to as “the medium-low run” and “the weak-low run”, respectively. The control run, the medium run, and the weak run are the polluted-scenario runs, while the low-aerosol run, the medium-low run, and the weak-low run are the clean-scenario runs. Comparisons between the medium run and the medium-low run and those between the weak run and the weak-low run isolate those effects of fire-generated aerosols for the case of medium fire intensity and the case of weak fire intensity, respectively. Comparisons between the control run and the low-aerosol run identify those aerosol effects for the case of strong fire intensity.

Effects of fire-generated aerosols on the pyroCb development and its impacts on the UTLS water vapor and cirrus clouds can also be dependent on how large fire-induced increases in aerosol concentrations are or the magnitude of fire-induced aerosol perturbation in a fire spot. Motivated by this, the previously described simulations are repeated by varying the magnitude of aerosol perturbation in the fire spot. To test the sensitivity of results to the magnitude of fire-induced aerosol perturbation, for each fire intensity, we repeat the polluted-scenario run by increasing and reducing the magnitude by a factor of 2 in the fire spot but not outside of the fire spot. These simulations with the increased magnitude have an aerosol concentration of $30000 \text{ cm}^{-3}$ at the first time step over the fire spot in the PBL and are referred to as the control-30000 run, the medium-30000 run, and the weak-30000 run for strong, medium, and weak fire intensity, respectively. These simulations with the reduced magnitude have an aerosol concentration of $7500 \text{ cm}^{-3}$ at the first time step over the fire spot in the PBL and are referred to as the control-
7500 run, the medium-7500 run, and the weak-7500 run for strong, medium, and weak fire intensity, respectively. Motivated by the analysis described in Section 4.2, we additionally repeat the medium run and the weak run with aerosol concentrations of 2000 and 1000 cm$^{-3}$ at the first time step over the fire spot in the PBL, respectively. The repeated medium (weak) run is referred to as the medium-2000 (the weak-1000) run. Table 1 summarizes the simulations.

### 4. Results

Results from the control run and the low-aerosol run, which are equivalent to the Full Simulation and the Low Aerosol Simulation in Kablick et al. (2018), respectively, are described here. Kablick et al. (2018) mainly focused on comparisons themselves between aerosol effects and heat-flux effects on pyroCb development and its impacts on the UTLS water vapor and cirrus clouds. In this study, we expand upon the results of Kablick et al. (2018) by providing additional details of the simulation results by focusing on the impacts of pyroCb development on the UTLS water vapor and cirrus clouds, and aerosol effects on pyroCb development.

Figure 5 shows the vertical distributions of the averaged updraft mass fluxes over cloudy areas, i.e., areas where the sum of liquid-water content (LWC) and ice-water content (IWC) is non-zero, and over the simulation period between 17:00 GMT on August 5th and 12:00 GMT on August 6th in the control run and the low-aerosol run for the case of strong fire intensity. 17:00 GMT on August 5th is a time around which the pyroCb starts to form and 12:00 GMT on August 6th is the end of the simulation period. In this study, drops with radii smaller (greater) than 20 µm are classified as droplets (raindrops). For the calculation of LWC (IWC), we only considered droplets (ice crystals). Stated differently, droplet mass but not rain mass is used to obtain LWC and the mass of ice crystals but not the mass of snow aggregates, graupel and hail is used to obtain IWC.

The updraft mass flux is one of the most representative variables that are indicative of the cloud dynamic intensity and the magnitude of convective invigoration. As seen in Figure 5, the control run and the low-aerosol run for strong fire intensity have similar updraft mass fluxes. Table 2 gives the averaged values of updraft mass fluxes over cloudy
areas at all altitudes and over the simulation period between 17:00 GMT on August 5th and 12:00 GMT on August 6th for simulations. In Figure 5 and Table 2, updraft mass fluxes in the control run are only ~3% greater than those in the low-aerosol run. Given the hundredfold difference in aerosol loading over the fire spot between the runs, this 3% difference in updraft fluxes is negligibly small.

Water vapor around and above the tropopause or the UTLS plays an important role in the global radiation budget, thus garnering much attention from the climate-change community. Motivated by this, we examine the role played by the pyroCb in the UTLS water vapor. The vertical distributions of averaged water-vapor mass density around and above the tropopause (~ 13 km) are shown in Figure 6. The water-vapor mass density shown by colored lines in Figure 6 is averaged over cloudy grid columns that have the non-zero sum of liquid-water path (LWP) and ice-water path (IWP) and over the simulation period between 17:00 GMT on August 5th and 12:00 GMT on August 6th. For the calculation of LWP (IWP), we only considered droplets (ice crystals) as for the calculation of LWC (IWC). The black line in Figure 6 shows the average of water-vapor mass density over non-cloudy grid columns that have the zero sum of LWP and IWP and over the simulation period between 17:00 GMT on August 5th and 12:00 GMT on August 6th in the control run. This represents the background water-vapor mass density. Table 2 shows the averaged values of water-vapor mass density over altitudes between 13 and 16 km for simulations.

As seen in Figure 6, 16 km is an altitude to which the non-zero water-vapor mass density over cloudy columns extends. The comparison between water-vapor mass density over the cloudy columns and that over non-cloudy columns in the control run as shown in Figure 6 and Table 2 demonstrates that there is a substantial increase in the amount of water vapor in the UTLS due to the pyroCb. There is about five times greater water-vapor mass over the cloudy columns that represent the pyroCb area than in the background outside the pyroCb area in the control run.

Updrafts in the pyroCb transport water vapor to the UTLS, which leads to the substantial increase in the amount of water vapor in the UTLS over the pyroCb area. For the simulation period between 17:00 GMT on August 5th and 12:00 GMT on August 6th, the averaged water-vapor mass fluxes at the tropopause over cloudy and non-cloudy grid
columns are $8.30 \times 10^{-6}$ and $0.57 \times 10^{-6}$ kg m$^{-2}$ s$^{-1}$, respectively. Due to the presence of the pyroCb and associated updrafts in cloudy grid columns, there are substantial increases in water vapor fluxes at the tropopause over those cloudy grid columns as compared to those fluxes in the background over non-cloudy grid columns. This leads to larger water vapor mass in the UTLS over the pyroCb than in the background outside the pyroCb or the pyroCb area in the control run. It is also shown that the vertical extent of water vapor is extended further up to ~ 16 km by the pyroCb as compared to the extent of ~14 km in the background (Figure 6).

Similar to the situation with updraft mass fluxes, there is only a small (~2%) increase in the averaged water vapor mass in the UTLS, as shown in Figure 6 and Table 2, in the control run as compared to that in the low-aerosol run for strong fire intensity. This small variation in updraft mass fluxes between the control run and the low-aerosol run also results in a small variation in the transportation of water vapor to the UTLS and the averaged water-vapor fluxes at the tropopause between these two simulations. These averaged fluxes are over cloudy columns for the simulation period between 17:00 GMT on August 5th and 12:00 GMT on August 6th. The averaged water-vapor fluxes vary from $8.30 \times 10^{-6}$ kg m$^{-2}$ s$^{-1}$ in the control run to $8.21 \times 10^{-6}$ kg m$^{-2}$ s$^{-1}$ in the low-aerosol run.

In addition to water vapor in the UTLS, cloud ice or ice crystals around the tropopause play an important role in the global radiation budget. These ice crystals comprise cirrus clouds. To identify the role played by the pyroCb in cirrus clouds, Figure 7 shows the averaged cloud-ice mass density over cloudy areas for the simulation period of 17:00 GMT on August 5th to 12:00 GMT on August 6th in the simulations. The altitude of homogeneous freezing is at ~ 9 km, so cirrus clouds which are composed of ice crystals only are between ~ 9 km and ~13 km. Between ~ 9 km and ~13 km, the averaged cloud-ice mass density in the control run is shown in Figure 7, and ranges from ~ 0.028 to ~ 0.037 g m$^{-3}$. In non-cloudy areas and in the clear sky background outside the pyroCb, cloud-ice mass density equals zero. Hence in the control run, the pyroCb increases cloud-ice mass density between ~0.028 to ~0.037 g m$^{-3}$ over the background.

Updrafts in the pyroCb produce supersaturation, which leads to the generation of cloud-ice mass and associated cirrus clouds via deposition, the primary source of cloud-ice mass. Similar to the situation with updraft mass fluxes, comparisons between the control run and
the low-aerosol run for strong fire intensity show that there is only a small increase (~4%) in cloud-ice mass in the control run particularly between 9 km and 13 km, as shown in Figure 7 and Table 2, as compared to that in the low-aerosol run. Due to the negligible variation of updraft mass fluxes, there are negligible variations of supersaturation and deposition between the simulations as seen in Figure 8, and thus a negligible variation of cloud-ice mass between the control run and the low-aerosol run. Figure 8 shows the vertical distributions of the averaged deposition rate over cloudy areas for the simulation period of 17:00 GMT on August 5th to 12:00 GMT on August 6th.

In summary, the pyroCb and associated updrafts cause a substantial enhancement of the transportation of water vapor to the UTLS. They also produce cirrus clouds. The role, which is played by fire-generated aerosols and their effects on the pyroCb and its updrafts, in the enhancement of the transportation of water vapor to the UTLS and in the production of cirrus cloud is not significant for strong fire intensity.

4.1 Dependence of aerosol effects on fire intensity

The weak sensitivity of updrafts, water vapor, and cirrus clouds to aerosol loading in the pyroCb may be related to fire intensity. When fire-generated surface heat fluxes and fire intensity are increased, it is likely that in-cloud latent heat is also increased because a major source of in-cloud latent heating is surface heat flux. Therefore, the aerosol-induced perturbations of latent heating may be relatively small compared with large in-cloud latent heat contributed by surface fluxes with very intense burning. In other words, aerosol-induced increases in parcel buoyancy and updrafts are relatively small compared with the large buoyancy and strong fire-driven updrafts produced by strong fire intensity and the associated large in-cloud latent heat.

Considering that a major source of in-cloud latent heat is surface heat fluxes, when the fire-generated surface heat fluxes and the fire intensity are reduced, in-cloud latent heat is also likely to be smaller. Here, we are interested in how the magnitude of an aerosol-induced perturbation of latent heating for a pyroCb with weak fire intensity is compared to that with strong fire intensity. This is just based on a possibility that with background in-
cloud latent heat varying with fire intensity, the relative magnitude of aerosol-induced perturbation of latent heat to surface flux-dominated latent heat may vary.

### 4.1.1 Updrafts and the UTLS water vapor and cirrus cloud

Figure 9 shows the averaged updraft mass fluxes over cloudy areas at each altitude in the runs with different fire intensity for the simulation period of 17:00 GMT on August 5th to 12:00 GMT on August 6th. Here, the averaged updraft mass fluxes in the low-aerosol run, the medium-low run and the weak-low run represent fire-driven updrafts for strong, medium and weak fire intensity, respectively. Due to different fire intensity and associated CAPE, fire-driven updrafts vary between these runs. The variation of these fluxes between the low-aerosol run and the control run for strong fire intensity, between the medium-low run and the medium run for medium fire intensity, and between the weak-low run and the weak run for weak fire intensity, respectively, is induced by fire-generated aerosols. As seen in Figure 9 and Table 2, all of the cases of weak, medium and strong fire intensity show aerosol-induced increases in updraft mass fluxes. Of interest is that the percentage increase in updrafts in the case of weak fire from those in the weak-low run to those in the weak run is the greatest, while the percentage increase in the case of strong fire from updrafts in the low-aerosol run to those in the control run is the smallest. The percentage increase for the case of medium fire from updrafts in the medium-low run to those in the medium run is intermediate (Figure 9 and Table 2). Here, the percentage difference, including both the percentage increase and decrease, is the relative difference in the value of variables between the control run and the low-aerosol run for strong fire intensity, between the medium run and the medium-low run for medium fire intensity or between the weak run and the weak-low run for weak fire intensity. This percentage difference for strong fire intensity is obtained as follows in this study:

\[
\frac{\text{The control run minus the low-aerosol run}}{\text{The low-aerosol run}} \times 100 \text{ (\%)} \quad (1)
\]

The percentage difference for medium fire intensity is obtained by replacing the control run with the medium run and replacing the low-aerosol run with the medium-low run in
Equation (1). The percentage difference for weak fire intensity is obtained by replacing the control run with the weak run and replacing the low-aerosol run with the weak-low run in Equation (1). Associated with those greatest increases in updraft mass fluxes, the percentage increases in water vapor and cloud-ice mass in the UTLS, which are calculated by Equation (1), are the greatest in the case of weak fire as seen in Figures 10 and 11 and Table 2. Figures 10 and 11 show the averaged water-vapor and cloud-ice mass density at each altitude, respectively, over cloudy areas for the simulation period of 17:00 GMT on August 5th to 12:00 GMT on August 6th in the runs with different fire intensity. Associated with those smallest increases in updraft mass fluxes, the percentage increases in water vapor and cloud-ice mass in the UTLS are the smallest in the case of strong fire as seen in Figures 10 and 11 and Table 2. Associated with the medium increase in updraft mass fluxes, the percentage increases in water vapor and cloud-ice mass in the UTLS are intermediary in the case of medium fire as compared to the cases of strong and weak fire (Figures 10 and 11 and Table 2).

4.1.2 Volume mean radius of droplets ($R_v$)

**a. Cloud droplet number concentration (CDNC) and LWC**

The simulation period is divided into four sub-periods for this next analysis: period 1 between 17:00 and 19:00 GMT on August 5th, period 2 between 19:00 and 21:00 GMT on August 5th, period 3 between 21:00 GMT and 23:00 GMT on August 5th, and period 4 between 23:00 GMT on August 5th and 12:00 GMT on August 6th. The initial formation of the pyroCb corresponds with the beginning of period 1, and along with periods 2 and 3 correspond to the initial stages of cloud development, while period 4 corresponds to the mature and the decaying stages. As seen in Figure 12, CDNC, which is averaged over cloudy areas at all altitudes and over period 1, decreases as the fire intensity and updrafts decrease. However, the control run, the medium run, and the weak run have the much higher averaged CDNC than the low-aerosol run, the medium-low run, and the weak-low run, respectively. Remember that the control run, the medium run and the weak run have higher aerosol concentrations than the low-aerosol run, the medium-low run and the weak-
low run, respectively, over the fire spot (Table 1). Increasing CDNC enhances competition among droplets for a given amount of water, which is available for the condensational growth of droplets, in a cloud. Enhanced competition eventually curbs the condensational growth and reduces droplet size, which is represented by \( R_v \) in this study. This explains why \( R_v \), which is averaged over cloudy areas at all altitudes and over period 1, is smaller in the control run than in the low-aerosol run for strong fire intensity, in the medium run than in the medium-low run for medium fire intensity, and in the weak run than in the weak-low run for weak fire intensity, respectively, as seen in Figure 12. Of interest is that as fire intensity weakens, although the averaged CDNC reduces, which tends to lower the competition among droplets, the averaged \( R_v \) decreases not only among the control run, the medium run and the weak run with higher aerosol concentrations over the fire spot but also among the low-aerosol run, the medium-low run and the weak-low run with lower aerosol concentrations over the fire spot as shown in Figure 12. This is because \( R_v \) is proportional to \( \left( \frac{LWC}{CDNC} \right)^{1/3} \). Here, LWC represents the given amount of water which is available for the condensational growth of droplets. This proportionality means that for a given CDNC, a decrease in LWC also causes \( R_v \) to decrease, i.e., a decrease in the available amount of water for the condensational growth with no changes in CDNC induces a decrease in \( R_v \). As shown in Figure 12, LWC, which is averaged over cloudy areas at all altitudes and over period 1, also decreases with weakening fire intensity and updrafts not only among the control run, the medium run and the weak run but also among the low-aerosol run, the medium-low run and the weak-low run. Effects of LWC on \( R_v \), which weakens as fire intensity weakens, outweigh those of CDNC and this leads to the decrease in the averaged \( R_v \) with weakening fire intensity as seen in Figure 12.

Note that the averaged LWC in the control run is similar to that in the low-aerosol run for strong fire intensity, while the averaged LWC in the medium run is similar to that in the medium-low run for medium fire intensity. The averaged LWC in the weak run is also similar to that in the weak-low run for weak fire intensity. The averaged LWC reduces with weakening fire intensity from that in the control run to that in the weak run through that in the medium run during period 1 (Figure 12). This reduction is similar to the reduction in the averaged LWC from that in the low-aerosol run to that in the weak-low run through that in the medium-low run during period 1 (Figure 12). Considering this, the fact that the...
averaged CDNC is much higher in the control run than in the low-aerosol run leads to a situation where \( \frac{LWC}{CDNC} \), which is the base of \( \left( \frac{LWC}{CDNC} \right)^{\frac{1}{3}} \), and thus \( R_v \) are much smaller in the control run than in the low-aerosol run for strong fire intensity; the fact that the averaged CDNC is much higher in the medium run than in the medium-low run leads to a situation where \( \frac{LWC}{CDNC} \) and \( R_v \) are much smaller in the medium run than in the medium-low run for medium fire intensity; the fact that the averaged CDNC is much higher in the weak run than in the weak-low run leads to a situation where \( \frac{LWC}{CDNC} \) and \( R_v \) are much smaller in the weak run than in the weak-low run for weak fire intensity.

Using the averaged LWC and the averaged CDNC that are shown in Figure 12, \( \left( \frac{LWC}{CDNC} \right)^{\frac{1}{3}} \) is calculated. \( \frac{LWC}{CDNC} \) reduces from \( 4.27 \times 10^{-14} \) kg in the control run for strong fire intensity to \( 8.08 \times 10^{-15} \) kg in the weak run for weak fire intensity through \( 3.09 \times 10^{-14} \) kg in the medium run for medium fire intensity. \( \frac{LWC}{CDNC} \) reduces from \( 1.1 \times 10^{-12} \) kg in the low-aerosol run for strong fire intensity to \( 8.1 \times 10^{-13} \) kg in the weak-low run for weak fire intensity through \( 1.00 \times 10^{-12} \) kg in the medium-low run for medium fire intensity. Here, the absolute variation of \( \frac{LWC}{CDNC} \) with varying fire intensity is about one order of magnitude smaller among the control run, the medium run and the weak run than among the low-aerosol run, the medium-low run and the weak-low run during period 1. Despite this, the absolute and percentage variations of \( \left( \frac{LWC}{CDNC} \right)^{\frac{1}{3}} \) and thus the averaged \( R_v \) is greater among the control run, the medium run and the weak run than among the low-aerosol run, the medium-low run and the weak-low run during period 1. This is due to the characteristics of a function of the form “\( x^{\frac{1}{3}} \)” whose reduction with reducing x is greater when x is positive and smaller for an identical decrement in x. \( \left( \frac{LWC}{CDNC} \right)^{\frac{1}{3}} \) varies by \( 1.50 \times 10^{-5} \) kg from \( 3.50 \times 10^{-5} \) kg in the control run for strong fire intensity to \( 2.00 \times 10^{-5} \) kg in the weak run for weak fire intensity, while it varies by \( 9.80 \times 10^{-6} \) kg from \( 1.03 \times 10^{-4} \) kg in the low-aerosol run for strong fire intensity to \( 9.32 \times 10^{-5} \) kg in the weak-low run for weak fire intensity. Associated with this, the averaged \( R_v \), as seen in Figure 12, shows a 47 % reduction from 3.20 \( \mu \)m in the control run for strong fire intensity to 1.70 \( \mu \)m in the weak run for weak...
intensity, and the averaged $R_v$ shows a 10% reduction from 7.75 μm in the low-aerosol run for strong intensity to 6.98 μm in the weak-low run for weak intensity during period 1.

Figure 13 plots the function $y = x^{\frac{1}{3}}$ to demonstrate the behavior of $R_v(y)$ with respect to $\frac{LWC}{CDNC}(x)$. In Figure 13 the arbitrary values $x_{L1}$ and $x_{L2}$ are large, and $x_{S1}$ and $x_{S2}$ are small.

In Figure 13, a situation is assumed where the variation of $x$-value from $x_{L1}$ to $x_{L2}$ is greater than that from $x_{S1}$ and $x_{S2}$. The variation between $x_{S1}$ and $x_{S2}$ emulates the variation of $\frac{LWC}{CDNC}$ among the control run, the medium run and the weak run with higher aerosol concentrations over the fire spot, while the variation between $x_{L1}$ and $x_{L2}$ emulates the variation of $\frac{LWC}{CDNC}$ among the low-aerosol run, the medium-low run and the weak-low run with lower aerosol concentrations over the fire spot. Here, $x$ and $y$ correspond to $\frac{LWC}{CDNC}$ and $\frac{LWC}{CDNC}^{\frac{1}{3}}$, respectively. Despite the smaller variation between $x_{S1}$ and $x_{S2}$ than that between $x_{L1}$ and $x_{L2}$, the corresponding variation of $y$ is greater between $x_{S1}$ and $x_{S2}$ than between $x_{L1}$ and $x_{L2}$. This graphically explains the larger variation of $\frac{LWC}{CDNC}^{\frac{1}{3}}$ and thus the averaged $R_v$ in Figure 12 despite the smaller variation of $\frac{LWC}{CDNC}$ among the control run, the medium run and the weak run than among the low-aerosol run, the medium-low run and the weak-low run.

**b. Equilibrium supersaturation**

During period 1 between 17:00 and 19:00 GMT on August 5th, the lower value of the equilibrium supersaturation in rising air parcels in the control run than in the low-aerosol run for strong fire intensity, in the medium run than in the medium-low run for medium fire intensity, and in the weak run than in the weak-low run for weak fire intensity, respectively, aids the greater reduction in $R_v$ among the control run, the medium run and the weak run than among the low-aerosol run, the medium-low run and the weak-low run with weakening fire intensity as seen in Figure 12. Remember that the control run, the medium run and the weak run have higher aerosol concentrations than the low-aerosol run, the medium-low run and the weak-low run, respectively, over the fire spot (Table 1).
et al. (2009) and Lee and Penner (2010) have also shown that the lower equilibrium supersaturation in simulations with higher aerosol concentrations than simulations with lower aerosol concentrations. The rate of parcel supersaturation change is expressed by the following equation as shown in Khain et al. (2000):

\[
\frac{dS}{dt} = A_1 W - A_2 \frac{dq_L}{dt}
\]  

(2)

In Equation (2), \(S\) is parcel supersaturation, while \(W\) and \(q_L\) are the updraft speed and the mixing ratio of cloud-liquid particles (or droplets) in a rising air parcel, respectively. The coefficients \(A_1\) and \(A_2\) are functions of thermodynamic parameters such as temperature, latent heat of vaporization, air viscosity and pressure. The first term in Equation (2) represents the generation of supersaturation by adiabatic air cooling, while the second term represents the depletion of supersaturation by the consumption of water vapor via the condensational growth of unactivated aerosol particles. At first, supersaturation in a rising air parcel increases with time as the parcel rises from around the surface. However, eventually, as the parcel goes up further, the increasing trend of supersaturation stops and its decreasing trend starts in the rising air parcel, which is when \(\frac{dS}{dt}\) becomes zero and supersaturation reaches its maximum value or becomes equilibrium supersaturation (Rogers and Yau, 1991). Rogers and Yau (1991) have shown that a higher aerosol concentration and associated higher consumption of water vapor by aerosol particles before supersaturation reaches its equilibrium value induce lower equilibrium supersaturation for a given set of updraft speed, aerosol composition, a form of aerosol size distribution and thermodynamic condition in a rising air parcel by using an idealized conceptual model.

This is consistent with Lee et al. (2009) and Lee and Penner (2010) who used a large-eddy simulation model for a real case. Rogers and Yau (1991) have also shown that a lower updraft speed and associated less adiabatic cooling induce lower equilibrium supersaturation for a given set of aerosol concentration, aerosol composition and a form of aerosol size distribution. According to well-known Köhler theory, smaller aerosol particles have greater critical supersaturation for a given aerosol chemical composition and aerosol particles with critical supersaturation, which is lower than the equilibrium supersaturation,
are activated (Rogers and Yau, 1991). Hence, as equilibrium supersaturation lowers in a rising air parcel, the minimum size of activated aerosol particles increases for a given aerosol composition as shown in Rogers and Yau (1991). Among aerosol particles that are activated in the rising air parcel, aerosol particles with the minimum size have the largest critical supersaturation, hence, all the particles with larger sizes than the minimum size are activated.

During period 1, as fire intensity weakens and updraft speed decreases for the identical initial aerosol concentration, the identical aerosol composition, and the identical assumed form of initial aerosol size distribution, parcel equilibrium supersaturation lowers not only among the low-aerosol run for strong fire intensity, the medium-low run for medium fire intensity and the weak-low run for weak fire intensity but also among the control run for strong fire intensity, the medium run for medium fire intensity and the weak run for weak fire intensity. Also, during period 1, percentage differences in updraft speed, which is averaged over areas with positive updraft speed and period 1, between the control run and the low-aerosol run for strong fire intensity, between the medium run and the medium-low run for medium fire intensity, and between the weak run and the weak-low run for weak fire intensity, respectively, are less than 2% and thus negligibly small. These percentage differences are calculated by Equation (1). During period 1, differences, which is also calculated by Equation (1), in thermodynamic parameters (i.e., temperature, latent heat of vaporization, air viscosity and pressure), which are also averaged over areas with positive updraft speed and period 1, between the control run and the low-aerosol run, between the medium run and the medium-low run, and between the weak run and the weak-low run, respectively, are less than 5% and thus considered negligibly small. However, there are two orders of magnitude higher aerosol concentration over the fire spot in the control run than in the low-aerosol run for strong fire intensity, in the medium run than in the medium-low run for medium fire intensity, and in the weak run than the weak-low run for weak fire intensity, respectively, at the first time step. This is with the identical aerosol composition and the identical assumed form of initial aerosol size distribution between the control run and the low-aerosol run, between the medium run and the medium-low run, and between the weak run and the weak-low run, respectively. Hence, mostly due to differences in aerosol concentration between the control run and the low-aerosol run, between the
medium run and the medium-low run, and between the weak run and the weak-low run, respectively, the averaged equilibrium supersaturation and the averaged associated minimum size of activated aerosol particles over areas with positive updraft speed and period 1, are lower and higher, respectively, in the control run than in the low-aerosol run for strong fire intensity, in the medium run than in the medium-low run for medium fire intensity, and in the weak run than the weak-low run for weak fire intensity. Associated with this, as diagrammatically depicted in Figure 14, the increase in the averaged minimum size with weakening fire intensity and associated decreasing updraft speed and equilibrium supersaturation occurs in the size range that is closer to the right tail of the assumed unimodal aerosol size distribution among the control run, the medium run and the weak run than among the low-aerosol run, the medium-low run and the weak-low run. Figure 14 is the same as Figure 3 but with linear scales for x- and y-axes. In Figure 14, $D_{cs}$ and $D_{cw}$ represent the averaged minimum size in the low-aerosol run for strong fire intensity and in the weak-low run for weak fire intensity, respectively, while $D_{ps}$ and $D_{pw}$ represent the averaged minimum size in the control run for strong fire intensity and in the weak run for weak fire intensity, respectively. The averaged equilibrium supersaturation reduces from 0.21% in the control run for strong fire intensity to 0.10% in the weak run for weak fire intensity. Associated with this, the averaged minimum size in diameter increases from 0.09 $\mu$m in the control run for strong fire intensity to 0.12 $\mu$m in the weak run for weak fire intensity over period 1. The averaged equilibrium supersaturation reduces from 0.55% in the low-aerosol run for strong fire intensity to 0.31% in the weak-low run for weak fire intensity. Associated with this, the averaged minimum size increases from 0.04 $\mu$m in the low-aerosol run for strong fire intensity to 0.07 $\mu$m in the weak-low run for weak fire intensity over period 1.

The aerosol distribution as depicted in Figure 14 represents the averaged form of the distribution over the simulation domain and period. This indicates that on average, the overall initial form of aerosol size distribution is well maintained over the domain and period, although aerosol concentration in each size bin of the distribution evolves with time and space. The above-described size range, which is associated with the increase in the averaged minimum size with weakening fire intensity and decreasing updraft speed, is between the averaged minimum size with strong fire intensity and that with weak fire intensity over period 1.
intensity as seen in Figure 14. The concentration of aerosol particles with a size, which is closer to the right tail, is lower than that with another size, which is less close to the right tail, as long as these sizes are on the right-hand side of the distribution peak as seen in Figures 3 and 14; since most of aerosol activation occurs for aerosol sizes on the right-hand side of the peak, here we are only concerned with the size ranges on the right-hand side. Stated differently, a larger portion of total aerosol concentration is over a size range that is farther from the right tail than over the other range which is closer to the right tail, in case the size increment over the two ranges is similar as can be seen in Figure 14. Note that associated with similar updraft speeds between the control run and the low-aerosol run for strong fire intensity, between the medium run and the medium-low run for medium fire intensity, and between the weak run and the weak-low run for weak fire intensity, respectively, during period 1, the reduction in updraft speed with weakening fire intensity among the low-aerosol run, the medium-low run and the weak-low run is also similar to that among the control run, the medium run and the weak run during period 1. This contributes to a situation where the increment in the averaged minimum size, which is 0.03 micron, among the low-aerosol run, the medium-low run and the weak-low run is similar to that among the control run, the medium run and the weak run during period 1 as diagrammatically depicted in Figure 14. The increment is the averaged minimum size with weak fire intensity minus that with strong fire intensity.

All aerosol particles with size greater than the minimum-activation size contribute to the overall CDNC. Accordingly, as seen in Figure 14, the increase in the averaged minimum size as fire intensity weakens reduces the number of aerosol particles that can be activated and droplets. This reduction in the number of activated aerosol particles is equal to the number of aerosol particles in the size range between the averaged minimum size with strong and weak fire intensity. Figure 14 demonstrates that this increase in the minimum-activation size with weakening fire intensity occurs closer to the tail among the control run, the medium run and the weak run than among the low-aerosol run, the medium-low run and the weak-low run. Recall that a large portion of the total aerosol concentration is in the size range which is closer to the right tail of the assumed unimodal aerosol size distribution than that which is less close to the right tail as long as changes in the minimum size in these two size ranges are similar and these ranges are on the right-hand side of the
aerosol distribution. So, a similar increase in the averaged minimum-activation size for a weakened fire results in a smaller percentage reduction in the total activated aerosol concentration among the control run, the medium run and the weak run than among the low-aerosol run, the medium-low run and the weak-low run during period 1. As seen in Figure 12, CDNC, which is averaged over cloudy areas and period 1, varies by 8% from 850 cm\(^{-3}\) in the control run with strong fire intensity to 780 cm\(^{-3}\) in the weak run with weak fire intensity. The averaged CDNC varies by 76% from 33 cm\(^{-3}\) in the low-aerosol run with strong fire intensity to 8 cm\(^{-3}\) in the weak-low run with weak fire intensity. This contributes to greater reduction in \(\frac{\text{LWC}}{\text{CDNC}}\)\(^{\frac{1}{3}}\) and thus \(R_v\) as fire intensity weakens among the control run, the medium run and the weak run than among the low-aerosol run, the medium-low run and the weak-low run during period 1. This is for a similar LWC between the control run and the low-aerosol run for strong fire intensity, between the medium run and the medium-low run for medium fire intensity, and between the weak run and the weak-low run for weak fire intensity, respectively.

The contribution is easily understood if we assume a situation where CDNC and LWC in the control run, the low-aerosol run, the medium run, the medium-low run, the weak run and the weak-low run are identical to those in Figure 12 except for the fact that CDNC reduces by 82% from 850 cm\(^{-3}\) in the control run with strong fire intensity to 150 cm\(^{-3}\) in the weak run with weak fire intensity. In this situation, the CDNC percentage reduction is greater between the control run and the weak run than between the low-aerosol run and the weak-low run with weakening fire intensity. This causes a greater reduction in \(\frac{\text{LWC}}{\text{CDNC}}\)\(^{\frac{1}{3}}\) (and thus \(R_v\)) between the low-aerosol run and the weak-low run than between the control run and the weak run with weakening fire intensity. \(\frac{\text{LWC}}{\text{CDNC}}\)\(^{\frac{1}{3}}\) varies by 9.80×10^{-6} kg from 1.03×10^{-4} kg in the low-aerosol run for strong fire intensity to 9.32×10^{-5} kg in the weak-low run for weak fire intensity, and it varies by 3.00×10^{-7} kg from 3.50×10^{-5} kg in the control run for strong fire intensity to 3.47×10^{-5} kg in the weak run for weak fire intensity. Hence, the smaller percentage variation of CDNC plays a role in the greater reduction in \(\frac{\text{LWC}}{\text{CDNC}}\)\(^{\frac{1}{3}}\) and thus \(R_v\) with weakening fire intensity among the control run, the medium run and the weak run than among the low-aerosol run, the medium-low run and
the weak-low run during period 1. Remember that the control run, the medium run and the weak run constitute the polluted-scenario runs with higher aerosol concentrations over the fire spot, and the low-aerosol run, the medium-low run and the weak-low run constitute the clean-scenario runs with lower aerosol concentrations over the fire spot. This larger reduction in $R_v$ for the polluted-scenario case compared with the clean-scenario case—both with similar LWC and LWC reduction—is also reported in Reid et al. (1999). However, Reid et al. (1999) did not explain the cause of the greater $R_v$ reduction for the polluted case. The polluted case has 100 times higher aerosol concentration than the clean case in Reid et al. (1999), which is similar to the polluted-scenario runs as compared to the clean-scenario runs in this study.

4.1.3 Autoconversion, freezing, deposition and condensation

According to previous studies (e.g., Khairoutdinov and Kogan, 2000; Liu and Daum, 2004; Lee and Baik, 2017), autoconversion is strongly dependent on the size of cloud droplets and is proportional to the size of cloud droplets. This is explained by the fact that the efficiency of collection among droplets is proportional to droplet size (Pruppacher and Klett, 1978; Rogers and Yau, 1991). Larger droplets collect or coalesce with other droplets more efficiently. Autoconversion is a collection process among droplets to form raindrops, which means autoconversion rate and cloud droplet size is proportional. Due to the larger $R_v$ during period 1, the subsequent autoconversion rates, which are averaged over cloudy areas and over period 2, are higher in the low-aerosol run than in the control run for strong fire intensity, in the medium-low run than in the medium run for medium fire intensity, and in the weak-low run than in the weak run for weak fire intensity, respectively (Figure 15a). Due to the larger absolute and percentage reduction in $R_v$, as described in Section 4.1.2, there is a larger absolute and percentage reduction in autoconversion rate among the control run, the medium run and the weak run than among the low-aerosol run, the medium-low run and the weak-low run with weakening fire intensity during period 2 (Figure 15a). The averaged autoconversion rates over period 2 reduce from $3.61 \times 10^{-6}$ g m$^{-3}$ s$^{-1}$ in the control run with strong fire intensity to $0.93 \times 10^{-6}$ g m$^{-3}$ s$^{-1}$ in the weak run with weak fire intensity through $2.01 \times 10^{-6}$ g m$^{-3}$ s$^{-1}$ in the medium run with medium fire intensity by 74%. Those
averaged autoconversion rates reduce from $4.52 \times 10^{-6}$ g m$^{-3}$ s$^{-1}$ in the low-aerosol run with strong fire intensity to $3.94 \times 10^{-6}$ g m$^{-3}$ s$^{-1}$ in the weak-low run with weak fire intensity through $4.43 \times 10^{-6}$ g m$^{-3}$ s$^{-1}$ in the medium-low run with medium fire intensity by 14%.

Associated with this, differences in the averaged autoconversion rates between the weak run and the weak-low run for weak fire intensity are greater than those between the control run and the low-aerosol run for strong fire intensity over period 2; differences in the averaged autoconversion rates between the medium run and the medium-low run for medium fire intensity are greater than those between the control run and the low-aerosol run for strong fire intensity, and smaller than those between the weak run and the weak-low run for weak fire intensity during period 2 (Figure 15a).

Due to smaller autoconversion rates, there is more cloud liquid available for freezing in the control run than in the low-aerosol run for strong fire intensity, in the medium run than in the medium-low run for medium fire intensity, and in the weak run than in the weak-low run for weak fire intensity, respectively, particularly during period 2. Hence, the rate of cloud-liquid freezing, which is averaged over cloudy areas and period 2, is greater in the control run than in the low-aerosol run, in the medium run than in the medium-low run, and in the weak run than in the weak-low run, respectively (Figure 15a). Remember that the control run for strong fire intensity, the medium run for medium fire intensity and the weak run for weak fire intensity constitute the polluted-scenario runs with higher aerosol concentrations over the fire spot, and the low-aerosol run for strong fire intensity, the medium-low run for medium fire intensity and the weak-low run for weak fire intensity constitute the clean-scenario runs with lower aerosol concentrations over the fire spot. Differences in autoconversion rates between the polluted-scenario run and the clean-scenario run, which increase with weakening fire intensity, induce those differences in the amount of cloud liquid available for freezing to get greater with weakening fire intensity (Figure 15a). Thus, differences in the averaged rate of cloud-liquid freezing between the polluted-scenario run and the clean-scenario run over period 2 gets greater with weakening fire intensity (Figure 15a). Due to this, differences in freezing-related latent heat between the runs increase with weakening fire intensity. When fire intensity is strong, the difference in freezing-related latent heat, which is averaged over cloudy areas and period 2, between the polluted-scenario run, which is the control run, and the clean-scenario run, which is the
low-aerosol run, is $1.60 \times 10^{-4}$ J m$^{-3}$ s$^{-1}$. However, with medium fire intensity, that difference between the polluted-scenario run, which is the medium run, and the clean-scenario run, which is the medium-low run, is $6.98 \times 10^{-4}$ J m$^{-3}$ s$^{-1}$, while with weak fire intensity, that difference between the polluted-scenario run, which is the weak run, and the clean-scenario run, which is the weak-low run, is $7.94 \times 10^{-4}$ J m$^{-3}$ s$^{-1}$. This corresponds to the variation of the percentage differences, which are calculated by Equation (1), in the averaged freezing-related latent heat between the polluted-scenario run and the clean-scenario run from 9% with strong fire intensity to 83% with weak fire intensity through 51% with medium fire intensity over the period 2.

As shown in Lee et al. (2017), enhanced freezing-related latent heat strengthens updrafts in places where freezing occurs and this, in turn, enhances deposition and deposition-related latent heat. Hence, although deposition, which is averaged over cloudy areas and period 2, is slightly lower, due to those strengthened updrafts, the averaged deposition and deposition-related latent heat are greater in the control run than in the low-aerosol run for strong fire intensity, in the medium run than in the medium-low run for medium fire intensity, and in the weak run than in the weak-low run for weak fire intensity, respectively, during period 3 (Figures 15a and 15b). As seen in Figure 16, differences in the averaged freezing rate (and thus the averaged freezing-related latent heating) over cloudy areas between the control run and the low-aerosol run for strong fire intensity, between the medium run and the medium-low run for medium fire intensity, and between the weak run and the weak-low run for weak fire intensity, respectively, do not change much up to ~20:30 GMT after they start to appear around 18:30 GMT. However, after ~20:30 GMT, these differences start to increase as time goes by for each fire intensity. This is because as convection intensifies, the transportation of cloud liquid to places above the freezing level starts to be effective around 20:30 GMT. The greater freezing and thus freezing-related latent heat in the control run than in the low-aerosol run, in the medium run than in the medium-low run, and in the weak run than in the weak-low run, respectively, which start to be significant around 20:30 GMT as compared to those before 20:30 GMT, invigorates updrafts, which are represented by the averaged updraft mass fluxes over cloud areas. This subsequently causes updrafts to be stronger in the control run than in the low-aerosol run for strong fire intensity, in the medium run than in the medium-low run for...
medium fire intensity, and in the weak run than in the weak-low run for weak fire intensity, respectively, from ~21:00 GMT on (Figure 16). Then, the stronger updrafts induce deposition, which is averaged over cloudy areas, to be greater in the control run than in the low-aerosol run for strong fire intensity, in the medium run than in the medium-low run for medium fire intensity, and in the weak run than in the weak-low run for weak fire intensity, respectively. This is around 10-20 minutes after the stronger updrafts in the control run than in the low-aerosol run, in the medium run than in the medium-low run, and in the weak run than in the weak-low run, respectively, start to occur (Figure 16). Note that deposition-related latent heat is about one order of magnitude greater than freezing-related latent heat for a unit of mass of hydrometeors involved in phase-transition processes. This contributes to much greater differences in deposition-related latent heat during period 3 than those in freezing-related latent heat between the control run and the low-aerosol run for strong fire intensity, between the medium run and the medium-low run for medium fire intensity, and between the weak run and the weak-low run for weak fire intensity, respectively, during periods 2 or 3 (Figures 15a and 15b). To satisfy mass conservation, the enhanced updrafts above the freezing level, due to enhanced freezing and deposition, induce more updraft mass fluxes below the freezing level in the control run than in the low-aerosol run for strong fire intensity, in the medium run than in the medium-low run for medium fire intensity, and in the weak run than in the weak-low run for weak fire intensity, respectively. This leads to more convergence around and below cloud base, which is air flow from environment to cloud, in the control run than in the low-aerosol run for strong fire intensity, in the medium run than in the medium-low run for medium fire intensity, and in the weak run than in the weak-low run for weak fire intensity, respectively. The more mass fluxes and the more convergence below the freezing level, in turn, enhance condensation. Hence, condensation, which is averaged over cloud areas, starts to be greater when time reaches ~22:30 GMT in the control run than in the low-aerosol run for strong fire intensity, in the medium run than in the medium-low run for medium fire intensity, and in the weak run than in the weak-low run for weak fire intensity, respectively (Figure 16). This induces the averaged condensation and condensation-related latent heat to be greater in the control run than in the low-aerosol run for strong fire intensity, in the medium run than in the medium-low run for medium fire intensity, and in the weak run than in the
weak-low run for weak fire intensity, respectively, during period 4 (Figure 15c). Enhanced condensation in turn enhances updrafts, establishing a feedback between freezing, deposition, condensation, and updrafts and thus, enhancing freezing, deposition, condensation, and updrafts further. This enhancement due to feedback eventually determines the overall differences in the pyroCb properties and their impacts on the UTLS water vapor and cloud ice between the control run and the low-aerosol run for strong fire intensity, between the medium run and the medium-low run for medium fire intensity, and between the weak run and the weak-low run for weak fire intensity, respectively.

Differences in freezing-related latent heat between the polluted-scenario run and the clean-scenario run increase with weakening fire intensity, particularly during period 2. Recall that the control run for strong fire intensity, the medium run for medium fire intensity and the weak run for weak fire intensity are the polluted-scenario runs, while the low-aerosol run for strong fire intensity, the medium-low run for medium fire intensity and the weak-low run for weak fire intensity are the clean-scenario runs. Thus, percentage differences in freezing-affected updrafts and subsequently in deposition-related latent heat, which is averaged over cloudy areas and period 3, between the polluted-scenario run and the clean-scenario run also increase with weakening fire intensity (Figures 15a, 15b and 16). Those differences, as calculated by Equation (1), in deposition-related latent heat are 16%, 181%, and 417% for strong, medium, and weak fire intensity, respectively, as seen in Figures 15b and 16. Since percentage increases in deposition-related latent heat in the polluted-scenario run get greater with weakening fire intensity, the subsequent percentage increases in updrafts in the polluted-scenario run as compared to updrafts in the clean-scenario run get greater with weakening fire intensity, particularly during period 3 (Figure 16). During period 4, due to these greater increases in updrafts in the polluted-scenario run with weaker fire intensity, the percentage increases in condensation in the polluted-scenario run as compared to condensation in the clean-scenario run get greater with weakening fire intensity (Figures 15c and 16). Then, the increases in condensation, in turn, further enhance the increases in updrafts in the polluted-scenario run for each fire intensity. This enhancement is greater with weaker fire intensity due to the greater increases in condensation with weaker fire intensity. This leads to the greater overall effects of the pyroCb on the UTLS water vapor and ice with weaker fire intensity.
4.2 Dependence of aerosol effects on the magnitude of aerosol perturbation

Table 3 shows that for each of the strong-, medium-, and weak-fire cases, there are increases in the UTLS water-vapor mass and in the UTLS amount of cirrus clouds in the run with the fire-induced aerosol perturbations of 30000 or 7500 cm$^3$. These increases are relative to the mass and the amount in the low-aerosol run for the strong-fire case, in the medium-low run for the medium-fire case, and in the weak-low run for the weak-fire case, respectively, with no fire-induced aerosol perturbation. Note that for each of the three types of fire-induced aerosol perturbations of 30000, 15000 and 7500 cm$^3$, aerosol-perturbation-induced percentage increases in the UTLS water-vapor mass and the amount of UTLS cirrus clouds get greater as fire intensity weakens (Tables 2 and 3). The qualitative nature of results regarding the dependence of the percentage increases in the UTLS water-vapor mass and the amount of UTLS cirrus clouds on fire intensity thus does not depend on the magnitude of the fire-induced aerosol perturbation.

Until now, we considered the situation where the fire-induced aerosol perturbation does not vary with fire intensity. Note that so far, we have taken interest in the sensitivity to fire intensity of an aerosol perturbation on pyroCb development, UTLS water vapor, and cirrus clouds. Hence, to examine and isolate the sensitivity, we have shown comparisons among sensitivity simulations by varying only the fire intensity while maintaining a constant aerosol perturbation. While working well for the isolation aspect, this strategy does not reflect reality well. It may be that weaker fire intensity produces a smaller aerosol concentration. This possibility is not that unrealistic, since stronger fire likely involves more material burnt and more aerosols from it.

With this situation in mind, we make comparisons among three pairs of simulations: the low-aerosol run and the control-30000 run for strong fire vs. the medium-low run and the medium run for medium fire vs. the weak-low run and the weak-7500 run for weak fire. Hence, among these three pairs, the magnitude of fire-induced aerosol perturbation reduces with weakening fire, emulating the possibility that weaker fire intensity involves a less amount of aerosols. For strong fire, the perturbation-related aerosol concentration is 30000 cm$^3$, for medium fire, it is 15000 cm$^3$, and for weak fire, it is 7500 cm$^3$. As shown in
Tables 2 and 3, comparisons among these three pairs show that relative importance of aerosol effects on the pyroCb development and its impacts on UTLS water vapor and cirrus clouds increases for weaker fires, and it does not matter if the aerosol perturbation reduces or stays constant with weakening fire intensity. In these comparisons, it is also possible that when fire-induced aerosol perturbation is very low for medium or weak fire intensity, the latent heat perturbation by aerosol perturbation can be very low. This very low latent heat is not large enough to increase the relative importance of those aerosol effects with weakening fire intensity. Based on this, the medium run and the weak run are repeated again. The medium run is repeated with lower fire-induced aerosol perturbations than the perturbation of 15000 cm$^{-3}$, while the weak run is repeated with lower fire-induced aerosol perturbations than the perturbation of 7500 cm$^{-3}$. Recall that when the repeated medium run has the aerosol perturbation of 2000 cm$^{-3}$, the repeated medium run is referred to as the medium-2000 run; when the repeated weak run has the aerosol perturbation of 1000 cm$^{-3}$, the repeated weak run is referred to as the weak-1000 run. The percentage increases in the UTLS water vapor and the cirrus-cloud amount from the medium-low run to the medium-2000 run or from the weak-low run to the weak-1000 run are smaller than those increases, for the case of strong fire, from the low-aerosol run to the control-30000 run. This indicates that when fire-induced aerosol perturbation reduces too much with weakening fire intensity, the relative importance of aerosol effects on pyroCb development and its impacts on the UTLS water vapor and cirrus clouds no longer increases with the weakening fire intensity.

5. Summary and conclusion

This study investigates an observed case of a pyroCb using a modeling framework. In particular, this study focuses on effects of fire-produced aerosols on pyroCb development and its impacts on the UTLS water vapor and cirrus clouds. Results show that pyroCb updrafts transport water vapor to the tropopause and above efficiently. This leads to a much greater amount of water vapor around and above the tropopause (i.e., the UTLS) over the pyroCb as compared to that in the background outside the pyroCb. The pyroCb also generates a deck of cirrus cloud around the tropopause. It is found that the role played by fire-produced aerosols or the fire-induced aerosol perturbation in the water-vapor
transportation to UTLS and the production of cirrus cloud in the pyroCb gets more significant as fire intensity weakens.

As fire intensity weakens, due to the reduction in LWC, $R_v$ decreases despite the reduction in CDNC that tends to increase $R_v$. During the initial stage, there is a similar LWC between the polluted-scenario run (i.e., the control run for strong fire intensity, the medium run for medium fire intensity and the weak run for weak fire intensity with the fire-induced aerosol perturbation) and the clean-scenario run (i.e., the low-aerosol run for strong fire intensity, the medium-low run for medium fire intensity and the weak-low run for weak fire intensity with no fire-induced aerosol perturbation) for each fire intensity. The reduction in LWC with weakening fire intensity among the polluted-scenario runs (i.e., the control run, the medium run and the weak run) is also similar to that among the clean-scenario runs (i.e., the low-aerosol run, the medium-low run and the weak-low run). During the initial stage, there are much greater CDNC in the polluted-scenario run than in the clean-scenario run for each fire intensity, and the smaller CDNC reduction among the polluted-scenario runs than among the clean-scenario runs with weakening fire intensity. This situation during the initial stage induces $R_v$ to reduce much more among the polluted-scenario runs than among the clean-scenario runs with weakening fire intensity. This reduces autoconversion more among the polluted-scenario runs than among the clean-scenario runs with weakening fire intensity. This makes differences in autoconversion between the polluted-scenario run and the clean-scenario run enhance as fire intensity weakens. The enhancing difference in autoconversion between the polluted-scenario run and the clean-scenario run causes greater differences in freezing-related latent heat as fire intensity weakens. Through feedback between freezing, deposition, updrafts, and condensation, differences in freezing-related latent heat induce differences in updrafts between the polluted-scenario run and the clean-scenario run. Those greater differences in freezing-related latent heat also lead to greater differences in updrafts, producing the greater differences in the UTLS water vapor and cirrus clouds between the runs with weaker fire intensity. This means that the role of fire-produced aerosols in water-vapor transport to the UTLS and the production of cirrus cloud in the pyroCb becomes more significant as fire intensity weakens. This role which is more significant with weaker fire intensity is robust to the magnitude of the given fire-induced aerosol perturbation which
was assumed not to vary with varying fire intensity. This more significant role with weaker fire intensity is also robust to the variation of the fire-induced aerosol perturbation with the varying fire intensity unless the variation is very high.

It is true that the level of the understanding of a mechanism that controls the role played by fire-produced aerosols in the development of pyroCbs and their impacts on water vapor and cirrus clouds in the UTLS has been low. This study shows that fire-produced aerosols can invigorate convection and updrafts and thus cause enhanced transportation of water vapor to the UTLS and enhanced formation of cirrus clouds. This study finds that the mechanism that controls the invigoration of convection by aerosols in the pyroCb is consistent with the traditional invigoration mechanism which was proposed and detailed in Rosenfeld et al. (2008). However, this study shows that for pyroCbs produced by strong fires, the aerosol-induced invigoration and its effects on water vapor and cirrus clouds in the UTLS are insignificant. Note that traditional understanding generally focuses on effects of fire-produced heat and water vapor and their associated fluxes around the surface on the pyroCb and does not consider effects of fire-produced aerosols on the pyroCb, and this understanding adequately explains the mechanics for pyroCbs in association with strong fires. However, this study suggests that the role of fire-produced aerosols in pyroCb development and its effects on the UTLS water vapor and cirrus clouds should be considered for cases where pyroCbs form over weak-intensity fires, should one be observed in nature.

It is of interest to note that when fire-induced aerosol perturbations are strongly reduced for cases of weaker-intensity fires compared with strong-intensity fires, the significance of the role played by fire-produced aerosol perturbation does not increase any longer and starts to reduce with weakening fire. This suggests that there is a critical level of aerosol perturbation below which the increase in the significance with weakening fire intensity ceases.
Author contributions
SSL came up with the research goals and aims, performed the simulations, and wrote the manuscript. GK and ZL selected the case, analyzed observations, and provided data to set up the simulations while reviewing and providing comments on the manuscript.

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References


Lee, S. S., Feingold, G., Koren, I., Yu, H., Yamaguchi, T., and McComiskey, A.: Effect of


Reid, J. S., Koppmann, R., Eck T. F., and Eleuterio, D.: A review of biomass burning


FIGURE CAPTIONS

Figure 1. VIIRS visible image of fire, smoke and cirrus cloud which are associated with the selected pyroCb. The bright white represents cirrus cloud or anvil cloud at the top of the pyroCb, while the red circle marks the fire spot. The dark white represents smoke produced by fire. Adapted from Kablick et al. (2018).

Figure 2. The simulated fire spot and the field of cloud-ice mass density at the top of the simulated pyroCb when the pyroCb is about to advance to its mature stage. The red circle marks the simulated fire spot, while the field represents the simulated cirrus cloud.

Figure 3. Initial aerosol size distribution in the PBL over the fire spot. N represents aerosol number concentration per unit volume of air and D aerosol diameter.

Figure 4. The vertical distribution of the radar reflectivity which is averaged over the Cloudsat path.

Figure 5. Vertical distributions of the averaged updraft mass fluxes over cloudy areas, where the sum of liquid-water content (LWC) and ice-water content (IWC) is non-zero, and over the simulation period between 17:00 GMT on August 5th and 12:00 GMT on August 6th in the control run and the low-aerosol run.

Figure 6. Vertical distributions of the averaged water-vapor mass density in the control run and the low-aerosol run over altitudes above 13 km and over the simulation period between 17:00 GMT on August 5th and 12:00 GMT on August 6th. Colored lines represent the averaged values over cloudy grid columns where there is the non-zero sum of liquid-water path (LWP) and ice-water path (IWP) in the control run and the low-aerosol run, while the black line represents those values over non-cloudy columns where there is the zero sum of LWP and IWP in the control run.

Figure 7. Vertical distributions of the averaged cloud-ice mass density over cloudy areas, where the sum of liquid-water content (LWC) and ice-water content (IWC) is non-zero,
and over the simulation period between 17:00 GMT on August 5th and 12:00 GMT on August 6th in the control run and the low-aerosol run.

Figure 8. Vertical distributions of the averaged deposition rate over cloudy areas and over the simulation period between 17:00 GMT on August 5th and 12:00 GMT on August 6th in the control run and the low-aerosol run.

Figure 9. Same as Figure 5 but for all three types of fire intensity.

Figure 10. Same as Figure 6 but for all three types of fire intensity.

Figure 11. Same as Figure 7 but for all three types of fire intensity.

Figure 12. The averaged CDNC, Rv, and LWC over cloudy areas at all altitudes and over period between 17:00 and 19:00 GMT on August 5th.

Figure 13. Graphic depiction of y values as a function of “$x^{\frac{1}{3}}$”. $x_{L1}$ and $x_{L2}$ represent large x values, while $x_{S1}$ and $x_{S2}$ represent small x values. y values corresponding to $x_{L1}$, $x_{L2}$, $x_{S1}$ and $x_{S2}$ are $y_{L1}$, $y_{L2}$, $y_{S1}$ and $y_{S2}$, respectively. The variation of x value from $x_{L1}$ to $x_{L2}$ is greater than that from $x_{S1}$ to $x_{S2}$.

Figure 14. Diagrammatic depiction of the varying minimum size of aerosol activation with varying fire intensity in the unimodal aerosol size distribution which is assumed in this study. The details of the varying minimum size are described in Section 4.1.2. $D_{cs}$ and $D_{cw}$ represent the minimum size in the low-aerosol run for strong fire intensity and in the weak-low run for weak fire intensity, respectively, while $D_{ps}$ and $D_{pw}$ represent the minimum size in the control run for strong fire intensity and in the weak run for weak fire intensity, respectively. Here, the variation of the minimum size from $D_{cs}$ to $D_{cw}$ is identical to that from $D_{ps}$ to $D_{pw}$.
Figure 15. The averaged rates of condensation, deposition and cloud-liquid freezing over cloudy areas at all altitudes and over (a) periods 2, (b) period 3 and (c) period 4. In panel (a), the averaged autoconversion rates over cloudy areas at all altitudes and over periods 2 is additionally shown.

Figure 16. Time series of differences in the averaged values of variables, which are related to aerosol-induced invigoration of convection, over cloudy areas at all altitudes (a) between the control run and the low-aerosol run for strong fire intensity, (b) between the medium run and the medium-low run for medium fire intensity and (c) between the weak run and the weak-low run for weak fire intensity.
<table>
<thead>
<tr>
<th>Simulations</th>
<th>Surface sensible heat fluxes in the fire spot (W m⁻²)</th>
<th>Surface latent heat fluxes in the fire spot (W m⁻²)</th>
<th>Aerosol concentration in the PBL over the fire spot (cm⁻³)</th>
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<td>Weak run</td>
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<td>Weak-1000</td>
<td>3750</td>
<td>450</td>
<td>1000</td>
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Table 1. Summary of simulations
Table 2. The averaged updraft mass fluxes over cloudy areas at all altitudes, the averaged water-vapor mass density over altitudes between 13 and 16 km and over cloudy columns except for the averaged background water-vapor mass density which is also over altitudes between 13 and 16 km but over non-cloudy columns, and the averaged cirrus-cloud mass density over cloudy areas between 9 and 13 km. These averaged values are obtained over the simulation period between 17:00 GMT on August 5th and 12:00 GMT on August 6th. “Difference” is the percentage difference between the polluted-scenario run and the clean-scenario run for each fire intensity. Note that the control run for strong fire intensity, the medium run for medium fire intensity and the weak run for weak fire intensity constitute the polluted-scenario runs with higher aerosol concentrations over the fire spot, and the low-aerosol run for strong fire intensity, the medium-low run for medium fire intensity and the weak-low run for weak fire intensity constitute the clean-scenario runs with lower aerosol concentrations over the fire spot. The percentage difference is 

\[
\frac{\text{The polluted-scenario run minus the clean-scenario run}}{\text{The clean-scenario run}} \times 100 \%.
\]
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<th>Control-7500</th>
<th>Medium-30000</th>
<th>Medium-7500</th>
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<th>Weak-30000</th>
<th>Weak-7500</th>
<th>Weak-1000</th>
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<tbody>
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<td><strong>Water vapor mass</strong></td>
<td>2.38 (5%)</td>
<td>2.28 (0.9%)</td>
<td>1.87 (42%)</td>
<td>1.50 (14%)</td>
<td>1.36 (3%)</td>
<td>1.31 (125%)</td>
<td>0.75 (29%)</td>
<td>0.60 (3%)</td>
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<td><strong>13 and 16 km</strong></td>
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<td></td>
<td></td>
<td></td>
<td></td>
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<tr>
<td><strong>Cirrus cloud mass</strong></td>
<td>0.025 (9%)</td>
<td>0.023 (0.2%)</td>
<td>0.023 (92%)</td>
<td>0.014 (17%)</td>
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<tr>
<td><strong>9 and 13 km</strong></td>
<td>10^{-3} g m^{-3}</td>
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Table 3. The averaged water-vapor mass density over cloudy columns between 13 and 16 km and, the averaged cirrus-cloud mass density over cloudy areas between 9 and 13 km. These averaged values are obtained over the simulation period between 17:00 GMT on August 5\textsuperscript{th} and 12:00 GMT on August 6\textsuperscript{th}. The number in parenthesis is the percentage difference which is $\frac{\text{The control-30000 (or the control-7500) run minus the low-aerosol run}}{\text{The low-aerosol run}} \times 100$ (%) for strong fire intensity, $\frac{\text{The medium-30000 (or the medium-7500 or the medium-2000) run minus the medium-low run}}{\text{The medium-low run}} \times 100$ (%) for medium fire intensity, and $\frac{\text{The weak-30000 (or the weak-7500 or the weak-1000) run minus the weak-low run}}{\text{The weak-low run}} \times 100$ (%) for weak fire intensity.
Suomi VIIRS
21:15 UTC (15:15 MDT)

Figure 1
Figure 2
Figure 3
Figure 4
Figure 5

[Graph showing updraft mass fluxes vs. height for Control and Low-aerosol scenarios]
Figure 6
Figure 7
Figure 8
Figure 9
Figure 10
Figure 11
Figure 12

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

- **CDNC** (x 10^2 cm^-3)
- **Rv** (µm)
- **LWC** (x 10^2 g m^-3)
Figure 13
Figure 14
Figure 15
Figure 15

Period 4 (23 GMT on August 5th - 12 GMT on August 6th; mature and decaying stages)
Differences in the averaged values

a) Control run minus Low-aerosol run

- Freezing rate ($2.1 \times 10^6 \text{ g m}^{-3} \text{s}^{-1}$)
- Deposition rate ($10^5 \text{ g m}^{-2} \text{s}^{-1}$)
- Updraft mass fluxes ($25.0 \text{ kg m}^{-2} \text{s}^{-1}$)
- Condensation rate ($3 \times 10^{-3} \text{ g m}^{-2} \text{s}^{-1}$)

Time (h)

b) Medium run minus Medium-low run

- Freezing rate ($1.6 \times 10^5 \text{ g m}^{-3} \text{s}^{-1}$)
- Deposition rate ($4.0 \times 10^5 \text{ g m}^{-2} \text{s}^{-1}$)
- Updraft mass fluxes ($8.0 \text{ kg m}^{-2} \text{s}^{-1}$)
- Condensation rate ($8.1 \times 10^{-3} \text{ g m}^{-2} \text{s}^{-1}$)

Time (h)

Figure 16
Figure 16