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Atmospheric Chemistry and Physics Discussions

Interactive comment on "Cloud and aerosol effects on radiation in deep convective clouds: comparison with warm stratiform clouds" *by* S. S. Lee et al.

S. S. Lee et al.

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8. The impacts of surface fluxes and wind shear

The following is added in the summary and conclusion to discuss the impact of the surface fluxes and wind shear.

(LL1054-1079 in p35-36 in the new manuscript)

The identical surface fluxes from observation are prescribed for high- and low-aerosol runs. Therefore, surface fluxes do not contribute to different near-surface convergence and radiation. In this study, we focused on how aerosols affect clouds and radiation for an identical observed net heat and moisture supplied to or removed from the domain

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by large-scale flow and surface fluxes. Although feedbacks from differences in clouds onto the large-scale flow and surface fluxes cannot be captured by this design, this isolates interactions between aerosols, microphysics, and local dynamics (e.g., convergence fields within the model domain with zero domain-mean average) and enables the identification of microphysics-aerosol interactions on the scale of cloud systems. Grabowski (1999) reported the increase in the intensity of low-level convergence of deep convective clouds with increasing surface fluxes. Hence, the cloud mass is expected to be higher (lower) with higher (lower) surface fluxes through the more (less) intense near-surface convergence, leading to more (less) offset of SCF by LCF than in DEEP in each of the high- and low-aerosol runs. Lee et al. (2008b) reported that in general, increasing wind shear led to the increasing intensity of near-surface convergence and thus cloud mass; also with larger wind shear, the magnitude of cloud-mass increase with increasing aerosols was larger. However, they also showed that the intensity of the convergence could decrease with increasing wind shear at extremely high wind shear. When they increased wind shear to around 0.013 s-1 (a maximum value used in Weisman and Klemp (1982)) as a extremely high shear, the convergence weakened leading to smaller cloud mass at both high and low aerosol. This also led to smaller increase in cloud mass with increasing aerosols than that shown with a comparatively low shear. Hence, it is expected that increasing (decreasing) wind shear increases (decreases) the offset of SCF by LCF in each of high- and low-aerosol runs and that of increasing negative SCF by increasing LCF at high aerosol unless the shear is extremely high.

9. The sensitivity to CAPE

The following is added to indicate an uncertainty of the method to vary CAPE in idealized cases of convective clouds:

(LL974-979 in p32-33 in the new manuscript)

A given value of CAPE is not unique with respect to thermodynamic structure. For example, CAPE can be increased by increasing near-surface humidity or by increasing the middle-tropospheric lapse rate. The former approach has been used to generate the idealized CAPE variations in this paper. Increasing the CAPE in this way particularly favors increased condensate production with increasing aerosols and the subsequent interactions described here.

10 . Comparison with other studies

The following is added in the summary and conclusion to compare this study with other studies.

(LL843-851 in p28 in the new manuscript)

Lohmann (2008) examined the effects of changes in greenhouse gas since industrialization on precipitation using a GCM coupled with double-moment microphysics for both convective and stratiform clouds. She reported the invigoration of convective clouds in a warmer present-day climate, leading to increased precipitation in convective regions. Hence, her results appear to support the hypothesis about the changing relation between CAPE and the convection intensity (and thus cloud-top height) with global warming, suggested above. However, the implications for large-scale aspects of this study will require further study with larger-domain models which is coupled with advanced microphysics and able to resolve convective cells.

(LL910-919 in p30-31 in the new manuscript)

Lee et al. (2008a) showed that differences in the mass of ice particles (and thereby the offset of SCF by LCF) between the high- and low-aerosol runs were not significant before stronger updrafts were triggered by enhanced evaporative cooling of cloud liquid at high aerosol. The more intense feedback between updrafts and depositional heating after the development of stronger updrafts played a crucial role in the substantially increased ice mass at high aerosol. However, this does not preclude other interactions as controls on the responses of ice mass to aerosols. For example, Lohmann and

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Diehl (2006) indicated that different interactions between IN and nucleation (per se) in mixed-phase and ice clouds can lead to the significant variation of the offset of SFC by LCF with aerosol increases. The role of those interactions in responses of ice particles to aerosols deserves the further study.

(LL920-943 in p32 in the new manuscript)

Cui et al.'s (2006) study indicated that the immersion freezing is most dominant among ice-nucleation paths and less vigorous near the top of clouds at their mature stages at high aerosol due to more rapid evaporation of smaller drops in the CCOPE case. This process reduced the buoyancy at cloud top and produced stronger downdrafts flanking the updraft core of the high-aerosol clouds, cutting off the inflow within the boundary layer to lead to weaker near-surface convergence, updrafts and cloud mass at high aerosol. Cui et al.'s (2006) study simulated clouds existing predominantly below the homogeneous freezing level whereas convective clouds in this study grow above the homogeneous freezing level except for DEEP (LOW-CU) where clouds grow to just below the homogeneous freezing level as shown in Figures 3 and 9. This study found greater homogeneous freezing of aerosol (haze) particles and droplets contributed to larger mass of ice crystals around the top of clouds at their mature stages by boosting the buoyancy and, thereby, deposition more at high aerosol than at low aerosol. Also, it should be stressed that Cui et al.'s (2006) study considered the case of weak wind shear. As simulated by Cui et al. (2006) and found by Weisman and Klemp (1982), when wind shear was weak, downdrafts destroyed the updrafts. However, in this study with moderate wind shear according to Bluestein's (1993) definition, downdraft regions were separated from updrafts cores, as shown in Figure 8 in Lee et al. (2008a) (who simulated the same ARM case as in DEEP). This led to the updraft-increasing mechanisms via the developments of stronger downdrafts and thereby low-level convergence, initiating larger condensation in liquid clouds and deposition in ice clouds (mostly around and above the homogeneous freezing level) at high aerosol. Although homogeneous freezing is absent in DEEP (LOW-CU) as in the CCOPE case

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in Cui et al. (2006), these mechanisms enable the increased cloud mass in DEEP (LOW-CU) with increasing aerosols mostly by increasing condensation.

Response to additional comment 1

The sentence in the abstract pointed out here is removed, following the major comment #1.

Response to additional comment 2

The following sentence (LL3-4 in p15294 in the old manuscript) is removed:

‘‘stratiform clouds are considered resolved by GCM grids ‘‘

Response to additional comment 3

Following the comment here, the following sentences are removed:

The following sentence (LL3-7 in p15294 in the old manuscript) is removed:

‘‘However, stratiform clouds are considered to be resolved by GCM grids and thus represented more explicitly via microphysics parameterization than deep convective clouds. This enables the simulation of changes in the properties of stratiform clouds caused by green house gases and aerosols in a more realistic way as compared to that in sub-grid deep convective clouds.‘‘

Response to additional comment 4

The following is added in the summary and conclusion to discuss about the impacts of resolution on results:

(LL1036-1044 in p34-35 in the new manuscript)

As does the choice of two dimensions, the choice of resolution (2 km horizontal, 500 m vertical) affords substantial computational advantages. Donner et al. (1999) reported a series of test calculations with a similar cloud-system model with resolutions ranging from 500 m to 5 km. They found basic features of the integrations (e.g., patterns of

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vertical velocity) were similar for horizontal resolutions of 2 km or finer for convective clouds. Simulations in DEEP are repeated with the vertical resolution of 100 m to test the sensitivity of results to the vertical resolution. It is found that the principal aspects of results with the 100-m vertical resolution are similar to those with the 500-m vertical resolution.

The following is added in the summary and conclusion to discuss about the Hong and Pan PBL scheme at a horizontal resolution of 2 km:

(LL1045-1053 in p35 in the new manuscript)

When Hong and Pan's (1996) PBL scheme is used for the horizontal resolution of 4 km or finer, it is reported that the PBL heights are predicted deeper by around 500 to around 600 m as compared to those predicted by the turbulent kinetic energy scheme (Deng and Stauffer, 2006). This implies that the cloud-base height may have been overestimated in the cases of convective clouds. However, the uncertainty of the location of cloud bases nearly within the 500m-layer is not likely to affect the qualitative nature of the results here. This is because condensation and deposition from mid-level of liquid clouds (about 2-3 km above the bases) to the top of ice clouds play crucial roles in the determination of cloud mass and thus radiative properties of convective clouds.

Response to additional comment 5

Paragraphs at p.15298 are not about the model description but about the setup of domain size, resolutions and prescription of large-scale forcing. Hence, the statement suggested by the reviewer here is not included at 15298. Instead, it is included in the introduction where the model is briefly introduced.

Response to additional comment 6

Additional comparisons for the effective size and LWP are added:

(LL272-288 in p10 in the new manuscript)

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Differences in individual upward and downward fluxes between the high-aerosol run and observation in DEEP are within around 10 % of observed fluxes. The size and path of cloud particles (i.e., cloud liquid and cloud ice) play important roles in determining the impacts of clouds on radiative fluxes. Hence, one or both of simulated size and path of cloud particles are compared to the observation, depending on the availability of observed data. The domain-averaged liquid-water path (LWP) is 51 g m-2. This LWP is within around 10 % of the observed LWP (55 g m-2); the comparisons for the size of cloud liquid and the size and path of cloud ice are not viable here, since the ARM observation does not provide those data. Hence, clouds in DEEP can be considered to be simulated reasonably well for the calculation of radiation. Simulated LWP and effective diameter in SHALLOW are compared to observation by the Moderate Resolution Imaging Spectroradiometer (MODIS) to assess ability of the model to simulate stratiform clouds; comparisons for radiative fluxes are not viable here, since ECMWF data do not provide observed fluxes. The domain-averaged simulated LWP is 56.20 g m-2 and MODIS-observed LWP at the location of simulation is 59.35 g m-2. In-cloud average effective size of simulated cloud liquid is 18.56 μm and MODIS-observed size is 17.10 μm. Hence, differences are within around 10 %, demonstrating clouds in SHALLOW are reasonably well simulated.

Response to additional comment 7

Done.

Response to additional comment 8

The followings are added to describe the determination of cloud regions, calculation of in-cloud averages and the domain-averaged cumulative values:

(LL360-369 in p12-13 in the new manuscript)

For the calculation of in-cloud averaged values and cloud fraction, it is needed to determine the grid points in cloud. Grid points are assumed to be in cloud if the number

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concentration and volume-mean size of droplets is typical for clouds and fogs (1 cm-3 or more, 1μm or more; Pruppacher and Klett, 1997). To calculate the in-cloud average of a variable of interest, first, the conditional average over the grid points in cloud is obtained at each time step; the conditional average is the arithmetic mean of the variable over collected grid points in cloud (grid point in clear air is excluded from the collection). Then, those conditional averages are collected and averaged over time to obtain the in-cloud average in this study; only time steps with non-zero conditional averages are included in the collection over time.

(LL 414-420 in p14 in the new manuscript)

The domain-averaged cumulative value (denoted by $<>$) of any variable (denoted by A) in this study is calculated using the following formulation:

(unfortunately, the Interactive Discussion system is unable to read the formulation. Please, refer to the revised manuscript for the formulation)

where Lx is the domain horizontal lengths, which are 168 and 26 km for DEEP and SHALLOW, respectively, and is the air density.

Response to additional comment 9

Lee et al. (2008a) (who simulated the same case using the same model setup as in this study) showed that the role of rain did not play a role in the initiation of the updraft-increase mechanism at high aerosol in deep convective clouds. As shown in figure 7 and described in section 4.3 in Lee et al.(2008a), the mechanism is triggered by the increased cloud-liquid evaporation.

Response to additional comment 10

Removed:

(LL13-14 in p15310 in the old manuscript)

It is expected that environmental conditions do not contribute to these different re-

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sponses of radiation significantly in this study.

The paragraph (LL11-20 in p15310 in the old manuscript) is rewritten as follows:

(LL556-573 in p19 in the new manuscript)

Different cloud depth and cloud-top height primarily determine the different offset of SCF by LCF and of increased negative SCF by increased LCF at high aerosol between SHALLOW (IDEAL) and DEEP. Analysis here shows that differences in responses of radiation to clouds and aerosols between SHALLOW and DEEP are similar to those between SHALLOW (IDEAL) and DEEP. This is despite different environmental conditions between SHALLOW and SHALLOW (IDEAL). SHALLOW and SHALLOW (IDEAL) adopt the different initial and large-scale humidity and temperature conditions, surface fluxes, and surface albedo although they both have the inversion layer and the similar low wind shear (based on the wind variation from the surface to the cloud-top) and CAPE, favorable for the development of stratiform clouds; the maximum wind shear and CAPE during high- and low-aerosol simulations are around 0.0004 s-1 and around 300 J kg-1, respectively, for both SHALLOW and SHALLOW (IDEAL). This indicates different responses of radiation between deep clouds and low-level shallow clouds are fairly robust to surface conditions and overall atmospheric temperature and humidity conditions of stratiform clouds. The presence of the inversion layer in the cases of shallow clouds (leading to low CAPE and the formation of stratiform clouds with smaller cloud depth and lower cloud-top height than those in deep convective clouds) plays a key role in those different responses between deep convective and stratiform clouds.

Response to additional comment 11

The sentences (LL24 in p15311 - LL2 in p15312 in the old manuscript) is rewritten as follows:

(LL605-616 in p20-21 in the new manuscript)

The negative forcing at the lowest level in DEEP (CU) and DEEP (LOW-CU) lowers water vapor at the lowest level by offsetting the effect of strong surface positive moisture flux on water vapor at the lowest level prior to 16:40 UTC on 29 June; the negative forcing (positive surface moisture flux) acts to decrease (increase) the waver vapor at the lowest level. Around 16:40 UTC, the humidity forcing at the lowest level becomes zero losing its ability to offset the effect of strong positive surface moisture flux. Hence, the water-vapor-increase effect of the surface moisture flux begins to predominantly control the lowest-level water vapor. This causes vapor mixing ratio at the lowest level to begin to rise around 16:40 UTC. Note that identical surface fluxes are prescribed in DEEP, DEEP (CU), and DEEP (LOW-CU). Hence, after the humidity forcing becomes zero, the mixing ratio in DEEP (CU) and DEEP (LOW-CU) stabilizes to a value lower than that in DEEP around 16:30 UTC (Figures 8a and 8b).

Response to additional comment 12

In this study, IN as well as CCN increases at high aerosol.

Section 3.6 is added to explain the separate roles of CCN and IN as follows:

(LL764-787 in p26 in the new manuscript)

In this study, CCN and IN are varied simultaneously between the high- and low-aerosol runs in DEEP, which makes it difficult to separate the effects of CCN from those of IN and vice versa. However, as shown in Lee et al. (2008a), the mechanism producing stronger updrafts is triggered by the increased evaporation of cloud liquid at high aerosol. They showed the role of ice particles in triggering the mechanism was negligible as compared to that of liquid particles (See section 4.3 and 4.4 in Lee et al. (2008a) for more detail). Also, as can be seen in figure 3b, large portion of mass of cloud ice is concentrated around or above the level of homogeneous freezing (around 10 km) where it is found that homogeneous freezing of haze and droplet particles (formed on CCN particles) accounts for most of cloud-ice number. Hence, increased CCN induces the stronger updrafts by increasing cloud-liquid evaporation as well as contribute to most of increases in the number of ice particles around and above the homogeneous

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freezing level. The increased updrafts increase condensation and deposition and the large portion of the deposition occurs on the ice particles formed by homogeneous nucleation which are located around or above the level of homogeneous freezing. Hence, it is likely that the qualitative nature of results of this study does not depend on IN variation. To confirm this, simulations in DEEP are repeated with no variation of aerosols acting as CCN but with the 10-fold variation of aerosols acting as IN between the high- and low-aerosol runs. These simulations show that 2 % of the increased negative SCF is offset by increased LCF at high aerosol. This offset is around one order of magnitude smaller than those shown in simulations with both of the CCN and IN variations considered where the offset is 28 %. However, another set of repeated simulations only with CCN variation (with no variation of IN) shows the offset of around 25 %. Hence, these repeated simulations demonstrate that the results here are strongly sensitive to CCN variation and their dependence on IN variation is negligible.

The following is added in the summary and conclusion to discuss about the separate roles of CCN and IN and the points 2, 3 and 4 in the major comments 3.

(LL900-909 in p30 in the new manuscript)

Additional simulations were performed. Those simulations examined the sensitivity of results here to parameters in the parameterization of the ice-crystal fall speed, the threshold snow mixing ratio for the conversion of rimed snow to graupel, and the size distribution of precipitable hydrometeors. They showed that results in this study were robust to those parameters. Also, it was found that the role of CCN was much more important than that of IN in the presented results here. Increased cloud-liquid evaporation, near-surface convergence and thus updrafts and ice formation around and above the level of homogeneous freezing determined the cloud-mass increase at high aerosol. Those processes were mostly controlled by the CCN increase and the IN increase played a negligible role in changing cloud mass with changing aerosols.

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‘‘This generates stratiform clouds developing under nearly similar environment to that in DEEP.‘‘ (LL1-2 in p15309 in the new manuscript)

is rewritten as follows:

(LL515-518 in p17-18 in the new manuscript)

This generates stratiform clouds developing under more similar environment to that in DEEP than those in SHALLOW, although the inversion layer from the positive temperature forcing leads to lower maximum CAPE (around 300 J kg) than that (around 2500 J kg-1) in DEEP.

‘‘An additional idealized simulation of warm stratiform clouds (SHALLOW (IDEAL)) with the nearly same environmental conditions as in those in DEEP was carried out.‘‘ (LL 3-4 in p15315 in the old manuscript)

is rewritten as follows:

‘‘An additional set of idealized simulations of warm stratiform clouds (SHALLOW (IDEAL)) with the similar environmental conditions (except for the CAPE level due to the presence of the inversion layer imposed to generate the shallow clouds) to those in DEEP was carried out.‘‘ (LL814-817 in p27 in the old manuscript)

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