6.6 AEROSOL INDIRECT EFFECTS AND RADIATIVE-DYNAMICAL FEEDBACKS IN THIN STRATOCUMULUS

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1. Introduction

Boundary-layer stratiform clouds are persistent and prevalent Klein and Hartmann (1993), imparting a strong negative forcing to the Earth’s radiative budget (Chen et al. 2000). The representation of these clouds in current climate models is relatively poor, leading to large uncertainty in climate projections [Randall et al. (2007); Intergovernmental Panel on Climate Change (IPCC)]. This problem is exacerbated by the sensitivity of stratiform clouds to perturbations in aerosol concentrations which impact cloud evolution and subsequently alter the Earth’s radiative budget. Twomey (1977) showed that an increase in cloud aerosol concentrations leads to higher cloud reflectivity and thus reducing the solar heating of the surface (the first aerosol indirect effect, or first AIE). Evidence for this effect has been found in satellite observations of stratocumulus impacted by aerosol and moisture from ship tracks (e.g., Coakley et al. 1987).

Twomey (1977) considered static clouds, however. When \( N_d \) increases the cloud can thin or thicken through various dynamic feedbacks, altering the cloud radiative response (second aerosol indirect effects). Albrecht (1989), for instance, hypothesized that an increase in \( N_d \) can reduce the efficiency of the collection process, reducing precipitation and increasing both vertically integrated cloud liquid water (liquid water path, or LWP) and cloud lifetime. Subsequent studies have shown that the opposite of Albrecht’s hypothesis can also occur. For example, increasing \( N_d \) can lead to reductions in LWP when low relative humidity air resides above the boundary layer (Ackerman et al. 2004).

While AIEs in stratiform clouds have been extensively studied with models (e.g., Ackerman et al. 2004; Lu and Seinfeld 2005; Wood 2007; Sandu et al. 2008; Hill et al. 2008) and observations (e.g., Durkee et al. 2000; Lu et al. 2007), AIEs for low LWP stratiform clouds have been essentially ignored. Unlike thicker stratocumulus, cloud top longwave cooling is sensitive to both LWP and \( N_d \) in low LWP (< 50 g m\(^{-2}\)) clouds (e.g. Garrett and Zhao 2006). Longwave cooling drives circulations that help maintain stratiform clouds (e.g., Lilly 1968; Nicholls 1984); therefore changes in longwave cooling caused by AIEs can alter the dynamic driving force and cloud lifetime. While some modeling studies of low LWP clouds have been done (e.g. Lee et al. 2009), none have examined radiative-dynamic feedbacks to AIEs.

AIE feedbacks could be climatically important because low LWP clouds are prevalent in both satellite and ground-based remote sensing data (Turner et al. 2007). Stratiform cloud layers with LWP \(< 50 \text{ g m}^{-2}\) have been observed in both mid (Lu et al. 2007; Brunke et al. 2010) and high latitudes (De Boer et al. 2009). The Arctic is frequently covered with low LWP clouds. Perturbations in Arctic cloud amounts have been linked to the recent changes in sea-ice extent (Francis and Hunter 2006; Kay et al. 2008), an important regulator of the Earth’s climate. An environment with low aerosol concentrations like the Arctic (Curry et al. 1996) could be susceptible to AIEs.

Here we explore the nighttime and daytime evolution of low LWP (< 50 g m\(^{-2}\)) stratiform clouds, and how this evolution varies with \( N_d \), using large-eddy simulation (LES). We find that an increase in \( N_d \) leads to cloud thinning at night and cloud dissipation during the day. These findings suggest a new second aerosol indirect effect, particular to low LWP stratiform clouds, that may play a role in Earth’s climate.

2. Cloud Integrated Radiative Heating

Evolution of stratocumulus is dependent, in part, on radiative heating which drives the cloud dynamics. Generally, greater longwave cooling drives stronger circulations and may produce thicker, longer lasting clouds whereas greater shortwave warming has the opposite effect (e.g., Nicholls 1984). Radiative heating feeds back on other cloud processes (e.g. drizzle) and alters cloud evolution (Bretherton et al. 2004). Radiative heating depends primarily on the cloud LWP and \( N_d \). Moreover, to first order cloud dynamics respond to cloud vertically integrated radiative heating (what we will call the total radiative warming or cooling) of the layer (e.g., Lilly 1968; Wood 2007). Computations of total radiative heating for static clouds, there-

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AIE effects related to this warming for the entire range of low LWP.

3.. Dynamical Atmospheric Model

We examine radiative-dynamical feedbacks induced by AIEs in low LWP clouds using the Regional Atmospheric Modeling System (RAMS) version 4.3.0 (Cotton et al. 2003) in LES mode (see Stevens et al. 1998; Jiang et al. 2002). We configure the model for 70 grid-points with 50 m horizontal spacing and 95 grid-points with 30 m vertical spacing, making a modeling domain of 3.40 km on a side and 2.79 km in height. Entrainment of cloud-free air into the stratocumulus topped boundary layer (STBL) plays a significant role in stratiform cloud dynamics (e.g., Lilly 1968). While relatively accurate representations of entrainment with LES require higher vertical resolution than used here, relative comparisons between our simulations are valid (Bretherton et al. 1999).

Our model time-step is 2 s and, following Xu and Randall (1995), we set our radiative time-step at 6 s to avoid spatial correlation errors between radiative heating and evolving cloud properties. We use the bulk cloud microphysical scheme of Meyers et al. (1997), predicting both mass and number concentration of drizzle. However, cloud drop number concentration is fixed during a simulation, hence feedbacks between changes in drop concentration and dynamics do not occur. As a consequence, all AIEs occur through an initial change in $N_d$. This is a limitation, but should provide a qualitative first-order estimate of the radiative-dynamical response.

We initialize the model with random thermal perturbations to an input sounding that produces thick stratiform clouds (Hartman and Harrington 2005). A low LWP stratiform cloud was created by lowering the vapor mixing ratio by 2.75 kg m$^{-2}$ above the boundary layer inversion. Entrainment warming and drying then quickly leads to a thin, broken cloud.

To illustrate longwave and shortwave AIEs, we conduct simulations of both a nocturnal and a diurnal (daytime) case. We emulate aerosol perturbations with three different constant cloud droplet concentrations of 50 cm$^{-3}$, 200 cm$^{-3}$ and 1000 cm$^{-3}$. The nocturnal case is six hours long whereas the daytime case is sixteen hours in length to capture the diurnal cycle in solar heating. We start the simulations with $N_d = 50$ cm$^{-3}$ and 1000 cm$^{-3}$ at the end of the first hour of the $N_d = 200$ cm$^{-3}$ simulation to avoid differences in model spin-up.

4.. Results - Nocturnal Simulations

Substantial differences occur in our nocturnal simulations due to varying $N_d$. Averaged over the last two hours of six hour simulation time the LWP for our simulations with 50 cm$^{-3}$, 200 cm$^{-3}$ and 1000 cm$^{-3}$ are 31.7 g m$^{-2}$, 20.4 g m$^{-2}$ and 14.9 g m$^{-2}$ respectively (Fig. 2a). Cloud fractions, defined as the ratio of model columns with LWP $> 10$ g m$^{-2}$ to the total number of model columns, also
decrease substantially as $N_d$ increases (Fig. 2a). As Fig. 1 indicates, the decrease in LWP leads to reductions in total longwave cooling (Fig. 2b). An additional reduction in total longwave cooling is due to averaging-in grid cells that lack clouds. Larger $N_d$ leads to lower LWPs and hence weaker cloud circulations, as the boundary-layer averaged Turbulent Kinetic Energy (TKE) shows (Table 1; tables are at the end of this document).

How do increases in $N_d$ produce decreases in LWP? Because the simulations diverge immediately, the first hour is critical to explaining the differences. A strong drizzle process can change TKE (Stevens et al. 1998); however the vertically integrated drizzle flux divergence (converted to an energy flux, Table 1) is small in comparison to the total longwave cooling (Fig. 2b). The drizzle fluxes are consequently weak and so do not play a role in altering the energetics of the cloud layer. Direct changes in total longwave cooling also cannot be the reason, because if they were the dominant process LWP would increase since more cooling leads to greater drop growth. The only process left that can explain the initial differences is entrainment.

Increasing $N_d$ in stratocumulus leads to smaller drops that can evaporate more readily. Wang et al. (2003) and Hill et al. (2008) have demonstrated that an increase in $N_d$ leads to increased evaporational cooling through entrainment and hence reductions in LWP. Hill et al. (2009) termed this process the “evaporation-entrainment feedback” and it is the reason for the initial reduction in LWP with increases in $N_d$. In our case, cloud fractions are also reduced through this feedback because LWP is low.

The model dynamics in our three simulations are identical when they begin and therefore the modeled entrainment is also identical. This means that initial mixing of air across the cloud-top interface is similar amongst the three simulations, and leads to larger reductions in LWP and cloud fraction as $N_d$ is increased. The LWPs and cloud fractions are reduced to different amounts concomitant with reductions in $N_d$ and are then associated with differing amounts of total longwave cooling. For $N_d = 1000 \text{ cm}^{-3}$ the total longwave cooling is sufficiently reduced to result in a thin broken cloud layer in which LWP decreases with time, whereas for $N_d = 50 \text{ cm}^{-3}$ the total longwave cooling is large enough to maintain cloud growth.

5. Results - Simulations of the Diurnal Cycle

In general, shortwave heating decreases LWP and STBL-averaged TKE in stratocumulus as compared to nighttime cases (e.g., Turton and Nicholls 1987). Moreover, as heating strengthens during the morning cloud top is at least partially decoupled from the surface, and this can also reduce TKE and LWP (e.g., Turton and Nicholls 1987; Lu and Seinfeld 2005; Sandu et al. 2008). Although these prior results are for larger LWP stratocumulus we find that shortwave heating has similar effects in our simulations (Fig. 3). For all $N_d$, LWP and cloud fraction decrease during the morning hours as the cloud layer warms and decouples from the sub-cloud layer (Fig. 3a). Production of drizzle and STBL-averaged TKE are concomitantly reduced (Table 1).

Reductions in LWP and cloud fraction are also associated with decreases in total longwave cooling (Fig. 3b). Because longwave cooling is sensitive to these reductions when LWP is low. The combination of increased shortwave warming and reduced longwave cooling consequently leads to strong reductions in LWP and in STBL-averaged TKE as $N_d$ rises (Table 1). Shortwave warming also causes some decoupling of the cloud layer from the atmosphere below which reduces fluxes of vapor into the cloud. The reduction in vapor flux from below and the reduction in overall cloud radiative cooling allow for greater net evaporation of the cloud through entrainment. Liquid water path and longwave cooling are further reduced causing the cloud layer to thin and break. These processes define a negative feedback loop for low LWP stratiform clouds causing either substantial reductions in LWP ($N_d = 50 \text{ cm}^{-3}$) or the dissipation of the cloud layer when $N_d$ is increased ($N_d = 200 \text{ cm}^{-3}$ and $1000 \text{ cm}^{-3}$).

Similar to the nocturnal simulations, differences in entrainment play a vital role in the evolution of the daytime simu-
6. Summary and Concluding Remarks

To test the robustness of our results we conducted ensembles of nocturnal and daytime simulations. We ran six additional realizations for each pair of droplet concentration ($N_d = 50 \text{ cm}^{-3}$ and $1000 \text{ cm}^{-3}$) and nighttime/daytime conditions, resulting in 24 additional simulations. We created these ensembles by perturbing the initial temperature field differently in each realization. All members of these ensembles displayed the same important qualitative features as described in the two previous sections, suggesting that our results are robust.

Our simulations suggest that increasing droplet concentration in low LWP stratiform clouds leads to decreases in LWP and possibly cloud dissipation, in contrast to the hypothesis of Albrecht (1989). The mechanism for this decrease in LWP is unlike that of Ackerman et al. (2004). Their study determined that increasing $N_d$ leads to slower drying of cloud droplets as they become smaller. Entrainment of dry air from above the cloud top interface can then lead to more efficient drying of the cloud, reducing the LWP. Since we do not account for sedimentation of cloud droplets, and our cloud drop concentrations are constant, our mechanism is different.

In our study we find, for low LWP cloud layers, that changes in both entrainment drying and radiative heating with droplet concentration drive a second AIE. The evaporative-entrainment feedback leads to rapid initial entrainment and evaporation of the cloud layer when $N_d$ is high. This efficient entrainment drying leads to more cloud breaks and lower LWP values for higher droplet concentrations. Because total longwave cooling is sensitive to LWP when LWP is low, there are consequent reductions in total longwave cooling and circulation strength. Therefore, for low values of LWP these cloud layers are more tenuous when droplet concentration is high and are more prone to dissipating during the day, and especially so when LWP < 20 g m$^{-2}$.

Perturbations in the aerosol concentrations of low-level, low LWP cloud layers can therefore substantially alter the radiative forcing of both the surface and atmosphere, and may have possible impacts on regional, and perhaps global, climate (Turner et al. 2007). An increase in aerosol can lead to increases in $N_d$, resulting in lower cloud LWP and cloud fraction, or even cloud dissipation. This change in cloud evolution with $N_d$ is associated with reduced downwelling longwave flux at the surface during both nighttime and daytime (Table 2). Reductions in the longwave flux are more substantial during the daytime because LWP varies more with $N_d$. Increasing $N_d$ from $50 \text{ cm}^{-3}$ to $1000 \text{ cm}^{-3}$ results in a 3% decrease in this flux during night and a 15% decrease during the day. The albedo of the cloud scene is also decreased as $N_d$ increases and LWP and cloud fraction decrease.

Geographic regions where low LWP clouds are frequent and aerosol concentrations are relatively low, such as the Arctic (Curry et al. 1996; De Boer et al. 2009), may be particularly susceptible to changes in aerosol concentrations. Francis and Hunter (2006) suggest that downwelling longwave flux at the surface plays a prominent role in the retreat or advance of the Arctic sea-ice edge. Moreover, Kay et al. (2008) showed that the 2007 minimum in sea-ice extent was related to cloud cover. It is quite possible that aerosol-induced changes in
Arctic cloud cover could have important ramifications for the energy budget of the Arctic surface. We model warm clouds in this study, and clouds in the Arctic are liquid phase during much of the warm season (Curry et al. 1996). However, the importance of this AIE with respect to ice and mixed-phase cloud processes in the Arctic should be investigated.

In general the global frequency of low-level, low LWP clouds will determine the importance of any attendant aerosol feedbacks to the radiative budget. Further examination of ground-based and satellite remote sensing data to determine the geographic and temporal frequency of low LWP clouds seems required.

Acknowledgement This research was supported by the Department of Energy's (DOE) Atmospheric Radiation Measurement Program, the National Aeronautics and Space Administration and the Pennsylvania Space Grant Consortium Research Fellowship Program. Computer resources were provided by the SHARCNET, a consortium of Canadian academic institutions.

References


Table 1. Horizontally and vertically averaged TKE and domain-averaged vertically integrated drizzle flux divergence, both temporally averaged over the fifth and sixth hours of each set of simulations.

<table>
<thead>
<tr>
<th>$N_d$ (cm$^{-3}$)</th>
<th>TKE (m$^2$ s$^{-2}$)</th>
<th>Drizzle Flux (W m$^{-2}$)</th>
<th>TKE (m$^2$ s$^{-2}$)</th>
<th>Drizzle Flux (W m$^{-2}$)</th>
</tr>
</thead>
<tbody>
<tr>
<td>50</td>
<td>0.053</td>
<td>6.3</td>
<td>0.028</td>
<td>3.3</td>
</tr>
<tr>
<td>200</td>
<td>0.035</td>
<td>1.1</td>
<td>0.021</td>
<td>0.7</td>
</tr>
<tr>
<td>1000</td>
<td>0.028</td>
<td>0.8</td>
<td>0.014</td>
<td>0.4</td>
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Table 2. Domain-averaged downwelling surface longwave flux and shortwave reflectivity (albedo) at top of modeling domain. For the nocturnal simulations fluxes are temporally averaged over the last two hours of simulation. For the simulations of the diurnal cycle data are temporally averaged over the seventh and eighth hours of simulation, when the sun is highest in the sky.

<table>
<thead>
<tr>
<th>$N_d$ (cm$^{-3}$)</th>
<th>Longwave Flux (W m$^{-2}$)</th>
<th>Longwave Flux (W m$^{-2}$)</th>
<th>Albedo</th>
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<tbody>
<tr>
<td>50</td>
<td>368.5</td>
<td>352.8</td>
<td>0.13</td>
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<td>200</td>
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<td>319.4</td>
<td>0.08</td>
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<td>1000</td>
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