THE IMPACT OF RADIATIVE HEATING AND COOLING ON MARINE STRATOCUMULUS DYNAMICS

A Dissertation in
Meteorology
by
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Abstract

We investigate the impact of radiative heating on the dynamics of the stratocumulus-topped boundary layer (STBL). Radiative heating computations through one-dimensional static cloudy model atmospheres show us that both longwave and shortwave radiative heating are sensitive to droplet concentration ($N_d$) and liquid water path (LWP) when LWP is low (LWP < 20 g m$^{-2}$). For higher LWPs, longwave radiative heating is not sensitive to $N_d$ or LWP while shortwave radiative heating continues to be sensitive to both quantities.

We used large-eddy simulation to study the STBL dynamical response to radiative heating. Nocturnal LESs of the STBL are sensitive to $N_d$ when LWP is low and the free tropospheric air is dry. Entrainment and longwave radiative cooling lead to lower cloud fractions when $N_d$ is high as compared to when it is low. These low cloud fractions are associated with less longwave radiative cooling and weaker STBL circulations, and entrainment is able to suppress cloud growth. In contrast, when $N_d$ is low and cloud fractions are not as low, longwave radiative cooling is large enough to support stronger STBL circulations and the cloud layer grows against entrainment. We suggest that accounting for changes in longwave radiative heating with droplet concentration is important in simulating low level liquid water clouds. When LWP is not low, changes in drizzle strength with $N_d$ mitigate differences in nocturnal STBL dynamics owing to changes in longwave radiative heating with $N_d$. The dependence of longwave radiative heating on $N_d$ is not as significant for these LWPs.
Daytime simulations of the STBL revealed that shortwave radiative heating affects the STBL primarily through increasing thermodynamic stability and this effect increases as solar zenith angle (Θ) decreases. This increase in stability is associated with decreased LWP, slower entrainment, weakened circulations and strengthen decoupling of the cloud layer from the sub-cloud layer. When LWP is not low, this decoupling is stronger for high $N_d$ because shortwave absorption is stronger and drizzle is weak. For low $N_d$ a stronger drizzle process may aid in partially re-coupling the cloud and sub-cloud layers together through generation of conditional instability in the sub-cloud layer. For low LWP, increased shortwave warming also leads to reduced longwave cooling. This reduction in longwave cooling leads to even weaker circulations and stronger decoupling of the cloud layer from the sub-cloud than for STBLs with higher LWP. Because entrainment is more vigorous and circulations are weaker for high $N_d$, low LWP clouds in the STBL are more likely to dissipate over the diurnal cycle when $N_d$ is high as compared to when $N_d$ is low.

Radiative in the simulations described above was computed using the Independent Column Approximation (ICA). We tested the impact on STBL dynamics of using the ICA in shortwave radiative heating rate computation by coupling a three-dimensional Monte Carlo shortwave radiative transfer solver to our LES. Preliminary results show that the use of the ICA for shortwave radiative heating computation has a minimal impact on STBL dynamics.
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List of Symbols

LWP  Liquid water path
STBL Stratocumulus topped boundary layer
$\theta_l$ Liquid water potential temperature
TOA  Top of the atmosphere
$w'w'$ vertical velocity variance
$\Theta$ Solar zenith angle
ICA  Independent Column Approximation
ERM  Eddy-resolving model
LES  Large-eddy simulation
$N_d$ Droplet concentration
LWC  Liquid water content
$\mu_0$ Cosine of the solar zenith angle
MLS  Mid-latitude summer
$r_v$ Water vapor mixing ratio
\( r_l \)  Cloud water mixing ratio  

CKD  Correlated-k distribution  

\( T_{\text{rad}} \)  Radiative timestep  

\( \Delta X \)  Horizontal grid size  

\( \Delta \)  Change in a quantity  

\( U \)  Wind speed  

TKE  Turbulent kinetic energy  

\( z_i \)  Height of the boundary layer  

\( \overline{w'\theta'}_i \)  Vertical buoyancy flux  

\( w_e \)  Entrainment rate  

SHDOM  Spherical Harmonics Discrete Ordinate Method
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This perpetual motion machine she made today is a joke! It just keeps going faster and faster...Lisa! Get in here...In this house, we obey the laws of THERMODYNAMICS!

-Homer Simpson
Radiation interacts with the atmosphere over a wide range of spatial and temporal scales. On timescales of less than a day we find intricate interactions between radiation and the Earth-atmosphere system, particularly through the evolution of clouds and cloud systems. The ability to accurately simulate or at least parameterize these cloud processes in climate models is paramount to achieving better understanding of the earth’s climate. According to the Intergovernmental Panel on Climate Change (IPCC, 2007), cloud-climate feedbacks remain the largest source of uncertainty in climate sensitivity estimates (Randall et al., 2007). Representations of cloud and radiation interactions are recognized as a source of great uncertainty in estimates of climate sensitivity made through the use of large-scale atmospheric models (Cess et al., 1996). If we are to lessen this uncertainty it is crucial that we come to understand the interactions of clouds and radiation in much greater detail.

Broadband radiative transfer through cloudy air leads to diabatic warming or cooling of those regions, and this warming or cooling leads to both direct and indirect changes in buoyancy, altering cloud-scale motions. Here we will focus
on one particular interaction of clouds and radiation: the impact of surface and atmospheric radiative heating\textsuperscript{1} on cloud dynamics. The contribution of radiative heating to cloud dynamics has been studied for a variety of cloud systems, including cirrus (Starr and Cox, 1985; Harrington et al., 2009), altocumulus (Liu and Kreuger, 1998), cumulus (Guan et al., 1995; Guan et al., 1997; Schumann et al., 2002), cumulonimbus (Frame, 2008), and stratocumulus, where the cloud is composed entirely of liquid (e.g., Lilly, 1968; Nicholls, 1984; Harrington et al., 2000; Hartman and Harrington, 2005a, 2005b), and also where the cloud is mixed-phase (e.g., Harrington and Olsson, 2001).

In the present study we will focus on the interactions of atmospheric radiation and dynamics within marine stratocumulus clouds, which persistently cover wide swaths of the subtropical ocean (Klein and Hartmann, 1993). Stratocumulus layers at low latitudes impart a significant negative forcing on the Earth’s radiative budget (Chen et al., 2000). The temperature of these clouds is not dramatically different from the underlying ocean surface so they do not significantly alter the amount of longwave radiation emitted to space by the Earth. However, stratocumulus have a much higher albedo than the ocean’s surface so they reflect more shortwave radiation back to space than the ocean surface. Stratocumulus cloud albedo strongly varies with the cloud layer’s liquid water path (LWP) and cloud fraction, which in turn are dependent on a myriad of dynamical processes, including cloud-radiation interactions (Fig. 1.1). Better understanding of stratiform cloud layer dynamics is necessary for a better understanding of the magnitude of their impact on the Earth’s radiative budgets in a changing climate.

The stratocumulus topped boundary layer (STBL) has been extensively studied via a plethora of field experiments (e.g., Albrecht et al., 1995; Stevens et al., 2003; Bretherton et al., 2004). Because observations of radiative flux divergence through cloud layers are notoriously difficult to obtain (Curry, 1986), numerical modeling is typically used to investigate the interactions of clouds and radiation. The STBL has been simulated using a variety of models including mixed-layer theory (e.g., Schubert et al., 1979; Lock and MacVean, 1999; Wood, 2007), 2-D eddy resolving simulations (e.g., Jiang et al., 2002; Mechem et al., 2008), and

\textsuperscript{1}In this study, we use “heating” to refer to either a loss or gain of energy, so “radiative heating” refers to warming or cooling of air due to radiative flux divergence. “Warming” and “cooling” specifically refer to gains or losses in energy, respectively.
large-eddy simulation (e.g., Moeng et al., 1996; Stevens et al., 2005; Ackerman et al., 2009). This multitude of observations and modeling simulations helps us form a picture of the processes significant in STBL evolution.

Figure 1.1. The various physical processes associated with stratocumulus cloud layers. We show schematics for both a) well-mixed and b) decoupled stratocumulus topped boundary layers. Profiles of liquid water potential temperature ($\theta_l$) are shown to the right of the schematics. Note that well-mixed boundary layers are not always observed at night and decoupled boundary layers are not always observed during the day.

Simulations and observations suggest that the evolution of the STBL is modulated by several processes that interact and feedback on one another (Fig. 1.1). Longwave radiative cooling at cloudtop drives circulations in the STBL. At night
these circulations can result in a well-mixed boundary layer where total moisture content and thermal energy are relatively constant with height (Fig. 1.1a). These circulations bring moisture and energy from the surface layer up into the cloud layer and also work to entrain warm, dry air from above the cloud layer into the STBL. This entrainment both lifts and dries out the cloud layer, and can also directly weaken downdraft strength. Figure 1.2 shows physically how air is entrained into the STBL. As positively buoyant plumes in the boundary layer rise into the stably stratified air above, strong downward and horizontal motions along the interface are generated. These motions along the interface result in mixing of air across the interface and warm, dry air is pulled into the STBL (Sullivan et al., 1999).

Processes in the STBL can weaken these radiatively driven circulations and result in a cloud layer that is not directly coupled to the surface (Fig. 1.1b). Shortwave warming of the cloud layer and longwave warming at cloudbase pro-
vide cloudy parcels with positive buoyancy that can both strengthen updrafts and weaken downdrafts. Together, radiative warming and entrainment of warm air can stabilize the cloud layer with respect to the sub-cloud layer and cause possible decoupling of the cloud layer from the sub-cloud layer and surface. Drizzle can also stabilize the cloud layer with respect to the sub-cloud layer. Net formation of drizzle warms the cloud layer and evaporation of drizzle cools the sub-cloud layer. Drizzle also removes liquid water from the cloud layer and modifies entrainment across the cloudtop interface (Stevens et al., 1998a; Ackerman et al., 2009).

If the cloud layer is decoupled, processes in the STBL are altered in response. Cloud thinning can occur, changing radiative heating profiles and also affecting drizzle rates. All of the above processes are dependent on the cloud droplet concentration and droplet size distribution of the cloud layer. The feedbacks and interplay between these processes internal to the STBL, as well as those external to it, such as surface properties and the large-scale subsidence rate, make for a complex cloud system in which one process is difficult to untangle from another.

Longwave radiative cooling from the top of stratocumulus cloud decks has long been postulated to play a leading role in the dynamics of the cloud system (Lilly, 1968). The sharp discontinuities in temperature and amount of emitting matter at the cloudtop produce radiative cooling rates on the order of $10 \text{ K hr}^{-1}$. This cooling causes negative buoyancy for cloud elements near cloudtop, and as these cloud elements sink, mass continuity leads to circulations within the boundary layer. Nicholls (1989) found through in situ observations that not all cloud elements near cloudtop were negatively buoyant. He attributed this finding of occasional positive buoyancy to entrainment of warm air overlying the cloud and mixing of this air with radiatively cooling cloudy air. He determined that horizontal convergence and divergence directly drive individual vertical circulations. Moeng and Schumann (1991) echo this sentiment, noting that “STBL downdrafts are driven, at their root, dynamically by the convergence of the horizontal flow, rather than buoyantly.” However, the importance of cloudtop radiative cooling cannot be neglected, as it indirectly drives these vertical circulations and is needed to explain the vertical structure of the STBL. For instance Nicholls (1984), using a mixed-layer model, showed that accounting for radiative cooling at cloudtop was required to obtain realistic vertical fluxes of water and potential temperature.
Many studies have investigated the interaction of longwave radiative transfer with stratocumulus cloud layers via large-eddy simulation (e.g., Bretherton et al., 1999; Harrington and Olsson, 2001; Stevens et al., 2005; Verzijlbergh et al., 2009). Bretherton et al. (1999) noted that accurate representation of entrainment requires accurate representation of the distribution of longwave radiative cooling at cloudtop but did not try to untangle the two processes. Longwave radiative flux divergence through a cloud layer has long been assumed to be strongly dependent on LWP (Stephens, 1978). Following this assumption, longwave cloudtop radiative cooling and cloudbase warming within modeled stratiform cloud layers have routinely been parameterized as having an exponential dependence on LWP only and no dependence on cloud droplet concentration or droplet size distribution. This assumption has been shown to give reasonable accuracy as compared to detailed radiative transfer calculations (Stevens et al., 2005; Larson et al., 2007). Garrett and Zhao (2006) note that cloud longwave emissivity varies significantly when LWP is lower than 50 g m$^{-2}$, and that it varies with droplet concentration for low values of LWP.

Thin liquid water clouds are of particular importance to the Earth’s climate. Turner et al. (2007) highlighted two reasons for this importance$^{2}$. One is the high frequency of occurrence of thin liquid water clouds; satellite data analysis has revealed that the mean LWP of low level liquid water clouds is 51 g m$^{-2}$ and that they cover 27.5% of the globe. Two, longwave and shortwave radiative fluxes at the top of the atmosphere (TOA) and at the surface (and hence cloud radiative forcing) are very sensitive to changes in LWP when LWP is $< 50$ g m$^{-2}$.

Stevens et al. (2005), in their simulations of the STBL, accounted for the reduction in radiative cooling when modeled LWP decreased below 50 g m$^{-2}$. They found that this reduction was correlated with reduction in boundary layer turbulence and increased decoupling of the cloud layer from the surface, but entrainment rates remained largely unchanged. Stevens et al. (2005) did not comment on droplet concentration impacts on longwave radiative heating. Given Garrett and Zhao’s (2006) findings, the dependence of longwave radiative heating on droplet concentration and size spectrum may have implications for modeled stratiform

$^{2}$In their study, Turner et al. (2007) defined thin liquid water clouds are those with LWP less than 100 g m$^{-2}$. 
cloud dynamics, but whether it does has yet to be tested.

To date numerical modeling has not been used to understand shortwave radiation-cloud dynamical interactions in detail. In many studies shortwave radiation is neglected altogether to simplify interpretation of modeling results. However, this is not the only reason cloud systems are commonly modeled as if they exist only at night. Modeling shortwave radiative heating adds a great deal of complexity and computational expense as compared to modeling longwave radiative heating through cloudy layers. Shortwave radiative flux divergence is not only dependent on LWP but also on the geometric depth of the cloud layer, modeled cloud droplet number concentration and possibly the droplet size distribution (Boers and Mitchell, 1994). Simple exponential models of radiative transfer (e.g., Stevens et al., 2005) are not likely to correctly capture profiles of shortwave radiative flux divergence through stratiform cloud layers.

When present, shortwave radiative heating plays as large a role in STBL dynamics as longwave radiative heating. Diurnal variation in stratocumulus cloud layer thickness has been observed, with the layer deepening during night and thinning during the day (Albrecht et al., 1990). Absorption of shortwave radiation by the stratocumulus layer plays an important role in this cycle as it works to offset the longwave cooling at cloudtop and warm the cloud interior during the day. Nicholls (1984) observed daytime STBLs that were not well-mixed from cloudtop to the surface and posited that shortwave warming of the cloud layer could lead to stabilization of the cloud layer with respect to the sub-cloud layer, “decoupling” the two layers. This phenomenon was further investigated by Turton and Nicholls (1987), who looked at the diurnal variation of the STBL using mixed-layer modeling. If decoupling occurred during modeling simulations, two mixed-layer models were used to simulate the boundary layer: one for each layer now separated from the other. Echoing the findings of Nicholls (1984), realistic values of integrated buoyancy flux for the STBL were not obtained by Turton and Nicholls (1987) when decoupling due to shortwave insolation was not accounted for. These studies are limited by the assumptions implicit in a one-dimensional mixed-layer modeling framework, most notably the assumptions of horizontal homogeneity and of the layers having “well-mixed” vertical structures (Turton and Nicholls, 1987).

Shortwave warming weakens buoyantly driven circulations and reduces mois-
ture transport from the sub-cloud layer into the cloud layer, allowing stratiform cloud layers to dry out due to entrainment from above (Bretherton et al., 2004). This explanation has been buoyed through large-eddy simulation (Ackerman et al., 2004; Duynkerke et al., 2004; Hartman and Harrington, 2005a, 2005b; Lu and Seinfeld, 2005). Lu and Seinfeld (2005) found that both turbulent kinetic energy and vertical velocity variance $\overline{w'w'}$ decrease in daytime simulations of stratocumulus as compared to nocturnal simulations. They also found thinner model cloud layers, owing to weaker turbulent motions and entrainment drying. Through their large-eddy simulations, Sandu et al. (2007) investigated the impact of perturbations in modeled cloud droplet number concentration on the diurnal cycle of stratocumulus. Their findings corroborated those of Lu and Seinfeld (2005). These studies’ focus was not the impact of shortwave radiative warming on stratocumulus cloud dynamics per se, but rather the impact on the aerosol indirect effects on these clouds. Hartman and Harrington (2005a, 2005b) studied how droplet growth and the collection process in the STBL were modified when both shortwave and longwave radiative impacts on cloud droplet growth and cloud dynamics were accounted for. Using large-eddy simulations they found that the in-cloud residence times of air parcels increased with decreasing solar zenith angle $\Theta$, as strong shortwave warming led to weaker circulations more confined to the cloud layer. However, like most previous studies the impact of shortwave radiative heating on the dynamics of the STBL was not the focus of this study.

Most recently, Caldwell and Bretherton (2009) used large-eddy simulation to simulate the evolution of the STBL over several diurnal cycles, but did not discuss shortwave radiative-cloud interactions in much detail. They instead focused on diurnal variability in liquid water path, entrainment rates, and drizzle rates. To our knowledge no studies exist detailing the response of stratocumulus topped boundary layer dynamics to shortwave warming. What are the principle dynamical and thermodynamical effects associated with shortwave warming in the STBL? How does shortwave warming interact with and modify other important processes in the STBL, such as entrainment and precipitation? We advance tentative answers to these questions in our study.

All previous modeling studies that included shortwave warming neglected horizontal transport of shortwave radiation. For these studies, radiative transfer was
computed using the Independent Column Approximation (ICA). The ICA assumes that each vertical column in an atmospheric model is plane parallel and independent of the other model columns so that radiation is not exchanged between them. The use of the ICA introduces errors in computed radiative heating rates that can be significant when horizontal grid resolutions are as small as those typically used in large-eddy simulation (Cahalan et al., 2005). The horizontal transport of long-wave radiation has been accounted for in a few small-scale atmospheric modeling studies (Guan et al., 1997; Mechem et al., 2008), but to our knowledge no one has investigated whether biases in computed shortwave radiative warming due to the ICA have an impact in modeled daytime stratocumulus cloud dynamics. What role does the horizontal transport of shortwave radiation and its associated warming have on the STBL? Does the use of the Independent Column Approximation (ICA) allow us to capture the aspects of shortwave radiative transfer important for STBL dynamics?

Before answering our questions regarding shortwave radiative-cloud interactions we must first understand longwave radiative-cloud interactions within our large-eddy simulations. Although longwave radiative-cloud interactions have been studied more frequently and in greater detail than their shortwave counterpart, a sound understanding of our own nocturnal STBL simulations will give us a firm platform with which to gain knowledge about shortwave impacts on the STBL. Additionally, we wished to explore the dependence of longwave radiative heating on droplet concentration and the droplet size distribution and whether or not it is necessary to account for them in simulations of the STBL.

In Chapter 2 we examine how cloud radiative heating varies with LWP, droplet concentration and cloud depth independent of a dynamical atmospheric model. We use the results in this chapter to guide and inform our dynamical simulations of the STBL. Chapter 3 details our atmospheric model and its configuration for our study of radiative-cloud interactions. Chapter 4 details our eddy-resolving model (ERM) simulations. We use these simulations to understand the first-order STBL dynamic response to a wide range of microphysical and radiative variations, and to limit the number of large-eddy simulations (LESs) we run. In Chapter 5 we discuss the influences of longwave radiative heating on nocturnal marine stratocumulus evolution as revealed by our LESs. In Chapter 6 we use more LES
results in discussion of shortwave radiative-cloud interactions and daytime marine stratocumulus. Chapter 7 discusses the use of the ICA in computation of shortwave radiative heating and its applicability to modeling STBL dynamics. In Chapter 8 we discuss our conclusions and potential avenues for future work.
Offline Radiative Heating Calculations

Solar radiative flux divergence within cloud layers is a function of several variables, the most important of which are liquid water path (LWP), cloud droplet number concentration \( N_d \) and geometric cloud depth (Boers and Mitchell, 1994). The cloud droplet size distribution also plays a role in broadband shortwave radiative transfer. In principle, longwave radiative transfer within cloud layers depends on the same factors as shortwave radiative transfer but is thought to be primarily dependent on LWP and only weakly dependent on the other factors (Stephens, 1978; Harrington and Olsson, 2001). Garrett and Zhao (2006) challenge this assumption. To better understand radiative heating in our dynamical modeling framework we found it useful to first compute shortwave, longwave and broadband radiative transfer through cloudy layers outside of our dynamical modeling framework. In this way we can study how cloud radiative heating varies with LWP, \( N_d \), particle size and cloud depth without including radiative-dynamical feedbacks.

We are interested in the vertical distribution of radiative heating within a cloud
layer; hence we must also know the vertical distribution of cloud liquid water content (LWC). For the stratocumulus topped boundary layer (STBL) average cloud LWC can increase adiabatically with height when entrainment and precipitation are negligible (Brenguier et al., 2000). In general entrainment and precipitation are not negligible, leading to some reduction in LWC from the adiabatic amount at various heights (Brenguier et al., 2000). Average cloud LWC in the STBL generally increases adiabatically as the altitude increases from cloudbase and then increases subadiabatically as cloudtop is neared, but there are many possible profiles of cloud LWC (Brenguier et al., 2000). For simplification we limit our offline radiative transfer computations to clouds with an adiabatically-varying profile of LWC throughout and investigate variations in cloud integrated radiative heating only. Cloud integrated radiative heating is used as a first-order measure of how much the cloud layer (or the STBL if it is well-mixed) warms or cools due to radiative processes (e.g. Lilly, 1968; Wood, 2007).

We computed shortwave and longwave cloud integrated radiative heating for one-dimensional plane-parallel STBL layers at 51 solar zenith angles from $\Theta = 0^\circ$ (overhead sun) to $90^\circ$ (nighttime) in increments of 0.02 $\mu_0$ (cosine of the solar zenith angle). Shortwave and longwave cloud integrated radiative heating are summed together to give total broadband cloud integrated radiative heating. For these calculations the surface shortwave albedo was set to zero. We created these STBL layers by modifying the mid-latitude summer (MLS) McClatchey et al. profile (1971) to incorporate a cloud layer with base at 1 km altitude and tops varying from 1.02 km altitude to 1.50 km altitude in 0.02 km increments. Liquid water path (LWP) varies commensurately and monotonically increases with cloud geometric depth to a maximum of 253 g m$^{-2}$ for a 0.50 km thick cloud layer.

We used the following procedure to modify the MLS profile to include an adiabatic cloud. The temperature profile is dry adiabatic from the surface to cloudbase and moist adiabatic through the cloud, maintaining the MLS profile surface temperature. The water vapor mixing ratio $r_v$ from cloudbase to the surface is the equilibrium mixing ratio value at cloudbase. In cloud $r_v$ is at the equilibrium value at each height and the cloud water mixing ratio $r_l$ is the difference between this value and $r_v$ at cloudbase. To avoid spurious radiative divergence at cloudtop we smoothed the temperature and mixing ratio values from cloudtop to the
MLS profile values at 2 km. We provide an example of a profile modified with this procedure here, plotting temperature, cloud LWC and vapor mixing ratio for a profile where we added an adiabatic cloud of 0.20 km thickness (Fig. 2.1). We compute radiative fluxes and heating rates from the surface to an assumed top of atmosphere of 41 km, but only display the portion of the profile modified.

![Thermodynamic profile from modified MLS sounding, surface to 3km. We added an adiabatic cloud of 0.20 km thickness to the standard MLS profile (McClatchey, 1971) with the procedure we described in the text.](image)

**Figure 2.1.** Thermodynamic profile from modified MLS sounding, surface to 3km. We added an adiabatic cloud of 0.20 km thickness to the standard MLS profile (McClatchey, 1971) with the procedure we described in the text.

### 2.1 Radiative Transfer Model

To distinguish the model atmospheres of the STBL in our dynamical modeling framework from those atmospheres used here for radiative heating computations,
we define those here as “static model atmospheres.” Because our static model atmospheres are plane-parallel we employed a two-stream solver. Our two-stream solver is described in Harrington (1997) and is based on the two-stream solvers of Ritter and Geleyn (1992) and Zdunkowski et al. (1980). To obtain better accuracy we updated the spectral band model from that described in Harrington (1997) to the correlated-k distribution (CKD) method as described in Cole (2005). The 27 shortwave CKD wavelength intervals match Kato et al. (1998) below 0.498 µm and Fu and Liou (1992) above 0.498 µm. The longwave portion is modeled with the CKD specified in Fu and Liou (1992) and includes the CKDv2.4 model for continuum absorption (Mlawer et al., 1999).

Cloud optical properties for the shortwave and longwave CKD bands are modeled using the binned approach described in Harrington and Olsson (2001). We assume a gamma droplet size distribution with a shape parameter $\nu$ at 10. A shape parameter of 1 gives an exponential distribution of droplet sizes. As $\nu$ increases the droplet size mode increases while the number of droplets at that size decreases, resulting in a broader distribution. Calculations with $\nu$ set to 15 and 6 showed minimal change in our results. The graybody approximation is used within each wavelength interval, meaning that cloud optical properties are constant over each interval.

### 2.2 Results

Figure 2.2 displays the variation of cloud integrated longwave radiative cooling with LWP for three droplet concentrations (50 cm\(^{-3}\), 200 cm\(^{-3}\), and 1000 cm\(^{-3}\)) fixed throughout the cloud layer. For all droplet concentrations, cloud integrated longwave radiative cooling increases quickly (i.e. radiative heating becomes more negative) up to a LWP of about 50 g m\(^{-2}\) and increases slowly with LWP thereafter. For large LWPs (greater than 50 g m\(^{-2}\)), droplet concentration has a small impact on integrated longwave cooling. Differences of 2.2 W m\(^{-2}\) and 2.5 W m\(^{-2}\) (3%) exist for LWP values of 50 g m\(^{-2}\) and 250 g m\(^{-2}\), respectively, between the integrated cooling values for 50 cm\(^{-3}\) and 1000 cm\(^{-3}\). The amount of cooling decreases slightly with increasing droplet concentration.

We focus on the variation of cloud integrated longwave radiative cooling for
Figure 2.2. Integrated cloud longwave radiative heating (negative values mean cooling) as a function of liquid water path for low-altitude stratiform clouds with an adiabatic liquid water content. Liquid water path varies from 0 to 250 g m$^{-2}$.

For LWP $< 50$ g m$^{-2}$ we observe larger differences in integrated cooling as compared to those seen for larger LWP values. The difference maximizes at 7.8 W m$^{-2}$ (15%) at a LWP of 6.8 g m$^{-2}$.

Figure 2.3. Integrated cloud longwave radiative heating as a function of liquid water path for low-altitude stratiform clouds with an adiabatic liquid water content. Liquid water path varies from 0 to 50 g m$^{-2}$.

Differences in cloud integrated longwave radiative cooling with droplet concentration are explained by differences in the total absorption cross section of each collection of droplets. For a homogeneous cloud layer, increasing droplet concentra-
tion while keeping LWC constant results in a larger total absorption cross section and hence a higher emissivity and absorptivity in the longwave spectrum. For low LWP clouds LWP < 20 g m$^{-2}$, where broadband longwave emissivities are less than unity throughout the cloud layer, the integrated radiative cooling increases with droplet concentration. For thicker clouds the emissivities at cloudtop are very close to unity and hence longwave flux divergence is not significantly dependent on droplet concentration. At cloudbase, however, the emissivities are not close to unity and so the amount of longwave radiation absorbed from the surface depends on droplet concentration regardless of cloud LWP. These two combined effects cause integrated longwave radiative cooling to decrease with increased droplet concentration for thicker clouds. As we show in Chapter 5, these changes in integrated longwave radiative cooling with increased droplet concentration produce dynamical responses in simulated stratiform cloud systems. These dynamical responses are particularly pronounced when LWP is low$^3$.

We examine the variation of cloud integrated shortwave radiative heating with Θ and LWP for $N_d = 50$ and 1000 cm$^{-3}$ in Fig. 2.4. For both drop concentrations

Figure 2.4. Integrated cloud shortwave radiative heating as a function of LWP and Θ for low-altitude stratiform clouds with an adiabatic liquid water content. a) $N_d = 50$ cm$^{-3}$ b) $N_d = 1000$ cm$^{-3}$. Contours in W m$^{-2}$.

$^3$In this study we use “low LWP” to mean LWP < 20 g m$^{-2}$. Below this value of LWP integrated longwave cooling is especially sensitive to LWP and $N_d$, and above this value of LWP integrated longwave cooling is not as sensitive to both LWP and $N_d$. 
Figure 2.5. Difference in integrated cloud shortwave radiative heating as a function of LWP and Θ for low-altitude stratiform clouds with an adiabatic liquid water content. Here we subtract the results for $N_d = 50 \text{ cm}^{-3}$ from those for $N_d = 1000 \text{ cm}^{-3}$. Contours in $\text{W m}^{-2}$.

Integrated shortwave radiative heating increases as Θ decreases and LWP increases, that is, as the sun rises in the sky and as the cloud thickens. There are notable differences, however.

Figure 2.5 shows the difference in cloud integrated shortwave radiative heating between $N_d = 1000 \text{ cm}^{-3}$ and $N_d = 50 \text{ cm}^{-3}$. Note that for all solar zenith angles and $\text{LWP} > 110 \text{ g m}^{-2}$, there is less integrated radiative heating for $N_d = 1000 \text{ cm}^{-3}$. This difference decreases to $-16 \text{ W m}^{-2}$ for $\Theta = 0^\circ$ (overhead sun) and the thickest cloud layer. For clouds of lower LWP, however, the sign of the difference in integrated radiative heating is dependent on Θ. For large solar zenith angles ($\Theta > 66^\circ$), there is less integrated radiative heating for the higher droplet concentration. As Θ decreases, there is more integrated radiative heating for a droplet concentration of $N_d = 1000 \text{ cm}^{-3}$ and this difference reaches a maximum of $11 \text{ W m}^{-2}$ for a LWP of $34 \text{ g m}^{-2}$.

We can explain the pattern of differences in integrated shortwave radiative heating observed in Fig. 2.5 through two competing physical effects. An increase in cloud droplet number concentration for fixed LWP leads to an associated increase in cloud optical depth with respect to incoming solar radiation. This increase in cloud optical depth leads to an increase in reflectivity of the cloud, known as the Twomey effect (Twomey, 1977). The increase in cloud droplet concentration
also increases the chance for solar radiation that enters the cloud to be absorbed. The probability of absorption increases because multiple-scattering increases with a decrease in droplet size (for realistic values of $N_d$). As a consequence radiation is more likely to be absorbed before it exits the cloud.

When droplet concentration increases for large LWP values and short radiation paths through the cloud ($\Theta$ large), the Twomey effect wins out. Consequently cloud integrated shortwave radiative heating decreases with increases in $N_d$. For shorter radiation paths ($\Theta$ small) and lower LWP values, the opposite occurs and cloud integrated shortwave radiative heating increases with increasing $N_d$.

Boers and Mitchell (1994) computed cloud absorptivity for adiabatic stratiform cloud layers where $N_d$ was fixed at two different values. They computed differences in cloud absorptivities for two different values of $N_d$ as a function of cloud geometric depth and solar zenith angle and found a similar pattern in absorptivity differences to what we see in Fig. 2.5, giving credence to our findings.

![Figure 2.6](image)

**Figure 2.6.** Integrated cloud shortwave heating and longwave cooling as a function of LWP for low-altitude stratiform clouds with an adiabatic liquid water content, $N_d = 50$ and $1000$ cm$^{-3}$. Integrated shortwave radiative heating computed with $\Theta = 0^\circ$. If we compare the cloud integrated shortwave and longwave radiative heating when LWP is low ($< 20$ g m$^{-2}$) we find that integrated longwave cooling varies more with LWP than integrated shortwave heating for both $N_d = 50$ and $1000$ cm$^{-3}$ (Fig. 2.6. We set $\Theta$ at $0^\circ$ to exhibit the largest possible integrated shortwave heating values for a fixed LWP (Fig. 2.4). Both integrated shortwave heating and longwave cooling are zero when LWP is zero. When LWP increases to $15.3$ g m$^{-2}$
for $N_d = 50 \text{ cm}^{-3}$, integrated shortwave heating rises to 16.4 W m$^{-2}$ and integrated longwave cooling rises $-58.9$ W m$^{-2}$. We find similar change for $N_d = 1000 \text{ cm}^{-3}$.

The differences in change of integrated radiative heating with LWP between the longwave and shortwave can be attributed to differences in absorption efficiencies for small cloud droplets. For these values of LWP, cloud droplet effective radii are low ($< 2 \mu$m). Marquis and Harrington (2005) noted significant differences in absorption efficiency over the shortwave and longwave spectrums for cloud droplets of these sizes. Cloud droplets of these sizes efficiently absorb and emit radiation in the longwave but scatter shortwave radiation more efficiently than they absorb it. Their absorption efficiency also increases faster in the longwave than the shortwave.

Figure 2.6 indicates that, for low LWP clouds (LWP $< 20 \text{ g m}^{-2}$), changes in integrated longwave cooling with LWP might be more significant than changes in integrated shortwave heating in impacting daytime STBL dynamics. Additionally, because integrated shortwave heating is a fraction of the integrated longwave cooling for a constant LWP, longwave radiative heating may prove more important to STBL dynamics for thin clouds than integrated shortwave heating, regardless of $\Theta$.

![Figure 2.6](image_a.png)

![Figure 2.6](image_b.png)

**Figure 2.7.** Integrated cloud broadband radiative heating as a function of LWP and $\Theta$ for low-altitude stratiform clouds with an adiabatic liquid water content. a) $N_d = 50 \text{ cm}^{-3}$ b) $N_d = 1000 \text{ cm}^{-3}$. Contours in W m$^{-2}$.

We show the variation of cloud integrated broadband radiative heating with solar zenith angle and LWP for $N_d = 50$ and 1000 cm$^{-3}$ in Figure 2.7. For most
Figure 2.8. Difference in integrated cloud broadband radiative heating as a function of LWP and Θ for low-altitude stratiform clouds with an adiabatic liquid water content. Here we subtract the results for $N_d = 50\,\text{cm}^{-3}$ from those for $N_d = 1000\,\text{cm}^{-3}$. Contours in $\text{W m}^{-2}$.

LWPs and solar zenith angles we observe net cooling of the stratiform cloud layer. Only for large LWP and low Θ do we observe net heating. This figure suggests that for many stratiform cloud systems shortwave heating can be viewed as reducing the integrated longwave cooling and weakening the associated circulations but this is not always the case.

Figure 2.8 shows the difference in cloud integrated broadband radiative heating between $N_d = 1000\,\text{cm}^{-3}$ and $N_d = 50\,\text{cm}^{-3}$. Differences in radiative heating have three local maxima. For LWP = 6.8 g m$^{-2}$ and Θ = 75°, the difference is largely due to differences in integrated longwave radiative cooling. For the other two maxima, LWP = 41.9 g m$^{-2}$, Θ = 0° and LWP = 253.0 g m$^{-2}$, Θ = 0°, differences in integrated shortwave radiative heating play the more primary role. This figure suggests that portions of (LWP,Θ) parameter space around these three local maxima should be investigated for potential aerosol indirect effects related to radiative heating.
For our study of cloud-radiative interactions in a dynamical framework we use the Regional Atmospheric Modeling System (RAMS) version 4.3.0 (Cotton et al., 2003). RAMS is a multipurpose numerical model that can be configured for simulations of atmospheric phenomenon from the general circulation (order of 10000 km) to large-eddies in clouds (order of 100 m). RAMS has been used as a framework for numerous studies of the Arctic cloudy boundary layer (e.g. Harrington and Olsson 2001; Prenni et al. 2007) and the marine STBL (e.g. Stevens et al. 1998; Jiang et al. 2002).

In the present study we employ RAMS to model the atmosphere at the smaller end of the spatial scales (order of 100 m). We conduct modeling simulations in both two dimensions (one vertical and one horizontal) as an eddy-resolving model (ERM) and three dimensions for large-eddy simulation (LES). Using RAMS as an eddy-resolving model has the advantage of requiring far less computational time as compared to large-eddy simulation, but as real atmospheric motion occurs in three dimensions these simulations have limitations. Some of these limitations will be detailed in the next chapter. Regardless, two-dimensional simulations of the atmo-
spheric boundary layer have been found to exhibit behavior observed during field experiments (Krueger and Bergeron 1994; Bretherton et al. 1999). Here we use two-dimensional simulations to gain a wide view of shortwave radiation’s impact on stratocumulus dynamics so as to guide our choice of specific three dimensional cases.

3.1 Model Configuration

We configure RAMS to run with 68 gridboxes (70 gridpoints) of 50 m spacing in the horizontal and 93 gridboxes (95 gridpoints) of 30 m spacing in the vertical, making for a modeling domain 3.40 km on a side and 2.79 km in height. Bretherton et al. (1999) found that higher vertical resolutions of 5 m to 10 m are required to accurately resolve both air motions at the stratocumulus top interface and entrainment rates. Models with vertical grid spacing such as ours over-entrain as compared to those with the higher vertical resolutions. However, Bretherton et al. (1999) also state that profiles of turbulence statistics can be well represented even when entrainment is too efficient. This statement gives us confidence that relative comparisons of statistics between simulations within our study will be valid.

We do not model large scale divergence and this is not anticipated to significantly influence our simulations owing to the strong initial capping inversion on the boundary layer. Because our horizontal and vertical resolutions are of comparable size, we use the Deardorff (1974) isotropic diffusion scheme to model subgrid scale turbulent kinetic energy. Surface sensible and latent heat fluxes are prescribed and set to zero throughout all simulations to simplify our analysis. Lateral boundary conditions are cyclic and the model top is rigid with a Rayleigh friction absorbing layer to eliminate spurious reflection of gravity waves off the model top. The model bottom corresponds to an ocean surface with a specified non-varying temperature of 288 K.

Both eddy-resolving and large-eddy simulations are run for 6 hours and model timesteps of 2 s. For eddy-resolving simulations the radiative timestep are also set at 2 s as the simulations are not computationally expensive. For large-eddy simulations it is prudent to increase the radiative timestep so as to reduce processing time. Xu and Randall (1995) suggest that eliminating errors in spatial correlation
between radiative heating rates and cloudy elements requires that the radiative timestep $\delta T_{\text{rad}}$ meet the following criterion:

$$\delta T_{\text{rad}} \leq \frac{1}{2} \frac{\Delta X}{U},$$

(3.1)

where $\Delta X$ is the horizontal grid size and $U$ is the maximum wind speed. For our case, the maximum initial windspeed in any direction is 4 m/s so a radiative timestep of 6 seconds suffices and is used.

We use the bulk cloud microphysical scheme of Meyers et al. (1997), and only warm microphysical processes are considered. We prescribe fixed cloud droplet number concentrations while allowing cloud LWC to vary as in a single-moment scheme. When drizzle is allowed, rain drop number concentrations and rain water content are both prognosed as in a double-moment scheme. We assume gamma size distributions for both cloud droplets and rain, setting the shape parameter $\nu$ at 6 for cloud droplets (narrow spectrum) and at 2 for rain drops (broad spectrum). We account for sedimentation of rain drops but not cloud droplets. Sedimentation of cloud droplets has been found to reduce entrainment drying at the cloudtop interface (Bretherton et al., 2007) and consequently been shown to impact both precipitation rates (Ackerman et al., 2009) and entrainment rates (Caldwell and Bretherton, 2009). Our neglect of this process is a noteworthy limitation of our study.

### 3.2 Radiative Transfer Model

Because the focus of our study is cloud-radiation interactions, we desired accuracy in our radiative heating rate computations and were willing to sacrifice computational time to do so. However, accurately accounting for the high spectral variation and three-dimensionality of radiative transfer within an atmospheric model is an enormous computational burden that is not realizable at present. We use the two-stream solver and correlated-k distribution (CKD) spectral band model described in Chapter 2 for computation of both shortwave and longwave radiative heating rates in most of our RAMS simulations. An implicit assumption of this radiative transfer solver is the Independent Column Approximation (ICA), which implic-
itly assumes no net horizontal transport between model columns. To answer the question of whether or not horizontal transport of shortwave radiation can be significant to modeled stratocumulus dynamics, we coupled a shortwave Monte Carlo radiative transfer solver with a CKD spectral band model to RAMS for a small set of simulations. This solver and associated results will be detailed in Chapter 6. For all simulations the surface shortwave albedo is 0.05, a value typical of the ocean.

Numerical singularities can exist in two-stream solutions to solar radiative transfer, and most commonly occur in the more strongly absorbing water vapor bands in the near-infrared (King and Harshvardhan, 1986). These singularities can lead to spurious solar heating computations which we serendipitously avoided in our offline calculations but were unable to avoid in our RAMS simulations. To eliminate this problem, we note when negative solar radiative fluxes or heating rates are computed for a specific band during a modeling simulation. We then recalculate solar radiative fluxes for that band with Beer’s Law using the total extinction optical thickness, assuming that absorption dominates radiative transfer and scattering is negligible. We computed cloud solar radiative heating for these bands using the absorption optical thickness only, removing the scattering portion. We then summed the solar radiative flux divergence over all spectral intervals to compute the solar radiative heating rates for a model column, as usual.

3.3 Case Study

Our input soundings are modified from those used in Hartman and Harrington (2005a, 2005b), which are loosely based on Moeng et al. (1996). Because solar heating and its dynamical influence on the STBL may depend on cloud structure, we used two soundings. Fig. 3.1 displays the potential temperature and water vapor mixing ratio for those two soundings. The OVERCAST sounding produces a well-mixed STBL capped by a strong inversion, with a cloud fraction of unity and initial cloudbase and top at 450 m and 710 m, respectively. To produce a broken stratocumulus cloud field we modified the OVERCAST sounding by lowering the vapor mixing ratio by 2.75 g kg$^{-1}$ for all levels above the boundary layer inversion. When the model is initialized with this sounding, called the BROKEN sounding,
entrainment drying and warming at the inversion lead to a cloud fraction less than unity within one hour of simulation time.

![Graphs showing potential temperature and water vapor mixing ratio profiles.](image)

**Figure 3.1.** Soundings used to initialize RAMS. a) Potential temperature profile. b) Water vapor mixing ratio profile. Bold line in b) shows the moisture profile for the OVERCAST sounding and the dashed line shows the moisture profile for the BROKEN sounding.

Brost et al. (1982) demonstrated the importance of vertical wind shear in stratocumulus dynamics through measured turbulence budgets in the STBL off the California coast. In the present study we are interested in cloud-radiation interactions which affect cloud dynamics through buoyancy alterations. Therefore, to isolate these effects as best we can we focus our efforts on stratocumulus topped boundary layers with minimal wind shear. The initial u-wind and v-wind components are 2 and -4 m s\(^{-1}\), respectively, and do not vary with height.
4.1 Purpose

We are interested in how variations in cloud radiative heating owing to changes in droplet concentration \( N_d \) and solar zenith angle \( \Theta \) elicit different dynamic responses in the STBL. Large-eddy simulations (LESs) are computationally expensive. Six-hour simulations of nocturnal stratocumulus take 40 hours while simulations of daytime stratocumulus take 54 hours of runtime. Simulating the variation in dynamical responses of the STBL to shortwave and longwave radiative heating through numerous LESs would take an inordinate amount of time. In lieu of this we turn to two-dimensional eddy-resolving model (ERM) simulations. ERM simulations take far less computational time to complete than LESs, and so we can run a large number of ERM simulations within a small fraction of the time necessary to run a commensurate number of LESs.

The two-dimensionality of ERM simulations limits the validity of results derived from these simulations. Although ERMs can represent “turbulent-like” eddies, a two-dimensional representation of turbulent flow is fundamentally different from a
three-dimensional representation as the flow in one horizontal direction is assumed uniform. In three-dimensional turbulent flows a transfer of energy exists, from the largest scales of the flow down to the smallest scales at which viscous dissipation occurs. This transfer of energy is known as the eddy cascade and is driven by vortex stretching (Tennekes and Lumley, 1972). Vortex stretching is absent from two-dimensional simulations of turbulent flow and as such the energy cascade is represented incorrectly.

Stevens et al. (1998b) discussed differences between modeling turbulent flow in two- and three-dimensions. For LES of homogeneous turbulence, so long as the resolved flow is in the inertial subrange, the rate of turbulent kinetic energy (TKE) generation and energy dissipation is independent of the model grid-scale chosen. That this important property of modeled turbulent flow is not dependent on grid resolution is comforting. This independence of grid-scale is a property not found in ERM simulations. An increase in grid resolution results in an increase in modeled TKE.

Because the energy cascade is represented incorrectly in ERM simulations, however, energy is dissipated via other physical mechanisms, such as interaction with rigid boundaries. This property of two-dimensional representations of turbulent flows is known as the “up-scale energy cascade” and is originally discussed in Fjørtoft (1953). The up-scale energy cascade is directly related to the lack of vortex stretching in two-dimensional simulations of turbulent flow. As grid resolution is increased more of the energy-containing eddies are resolved, as for LES, but this energy is not driven to smaller scales via the energy cascade in ERM simulations and TKE increases.

Moeng et al. (1996) conducted an intercomparison of a set of LES and ERM codes to observe how these codes modeled the nocturnal STBL. The time evolution of cloud-top height, cloud amount, LWP and layer averaged TKE as represented by the ERM simulations was similar to that of the LESs. However, the vertical profiles of momentum flux and TKE were very different between the two- and three-dimensional simulations. These similarities and differences indicate that some first-order quantities can be simulated with reasonable accuracy with ERM simulations but we must be careful in deeper analysis.

Because of these issues with ERM simulations and two-dimensional representa-
tions of turbulent flow in general, we confine ourselves to examining only integrated quantities, such as LWP and cloud-layer radiative heating, and eschew studying or interpreting vertical profiles. Our purpose is to obtain a general picture of how the STBL responds to microphysical and radiative variations and how sensitive this response is. We also advance some general understanding of the physics observed in these simulations but will discuss those physics in more detail using LES of selected cases.

4.2 Results

We conducted six-hour ERM simulations using both the OVERCAST and BROKEN stratocumulus soundings. We modeled fixed values of $N_d$ at 50, 100, 200, 500 and 1000 cm$^{-3}$ covering a realistic range of cloud droplet concentrations. After one hour of simulation spin-up time, when shortwave radiation was not modeled, we applied fixed shortwave forcings in each simulation by fixing $\Theta$ from 0° (overhead sun, maximum shortwave forcing) to 90° (no shortwave forcing) in 15° increments. Solar zenith angles were fixed at one value so that modeled STBL dynamics would not be altered by variations in shortwave forcing, as was done in Hartman and Harrington (2005b). Presented here are domain-averaged vertically integrated quantities, temporally averaged over the last two hours of the simulation. We investigate the simplest set of cases first, i.e. the OVERCAST simulations with precipitation not allowed (Fig. 4.1).

4.2.1 OVERCAST case, non-drizzling

For all ERM simulations of the OVERCAST case the cloud fraction was unity and LWP is distributed fairly evenly throughout the modeling domain. We find that the cloud fraction is not unity for ERM simulations of the BROKEN case, and the existence of breaks in the cloud layer has a significant impact on STBL dynamics (see below). Figure 4.1a shows the variation of LWP with $\Theta$ and $N_d$ for the OVERCAST case. For any $N_d$, LWP decreases monotonically with decreasing $\Theta$ and increased shortwave forcing. This relationship is expected from the findings of previous modeling studies (e.g. Turton and Nicholls, 1987); the cloud layer thins
Figure 4.1. Eddy resolving model output for the OVERCAST sounding with drizzle not allowed. Quantities are domain averaged, vertically integrated and temporally averaged over the last two hours of the simulations.
as it is warmed and becomes more decoupled from the sub-cloud layer. Describing
the relationship of LWP with $N_d$ for a fixed $\Theta$ is more complicated.

When we first add shortwave forcing one hour into these simulations, LWP is
44 g m$^{-2}$, a value at which shortwave absorption increases with $N_d$ (Fig. 2.5). In
response to this shortwave absorption, LWP decreases, and this decrease is greater
for larger $N_d$. Boers and Mitchell (1994) suggested that geometrical thickness of a
cloud layer (and hence its LWP) would decrease with time in response to shortwave
heating if cloud shortwave absorption was directly proportional to $N_d$. We see the
result of this dynamic response in Fig. 4.1a, as LWP at the end of the simulation
monotonically decreases as $N_d$ increases for a fixed value of $\Theta$. We also note
that the variation of LWP at the end of the simulation with droplet concentration
increases as $\Theta$ decreases. This greater variation in LWP is commensurate with a
greater variation in amount of cloud solar absorption at this time (Fig. 4.1d).

Figure 4.1b shows the variation of STBL-average TKE with $\Theta$ and $N_d$. Like
LWP, average TKE decreases and boundary layer circulations weaken with de-
creasing $\Theta$, as found in previous studies (Lu and Seinfeld, 2005). For constant
$\Theta$ average TKE decreases with increasing $N_d$ and is associated with changes in
entrainment and radiative heating that are described below.

Our simulated cloud layers all have LWP $> 20$ g m$^{-2}$ and are not low LWP
clouds so integrated longwave cooling, depicted in Fig. 4.1c, varies little with either
$\Theta$ or $N_d$. The slight decrease in integrated longwave cooling with increasing droplet
concentration similar to the offline computations displayed in Fig. 2.2; emissivities
at cloudbase are not close to unity and longwave radiative transfer there depends on
$N_d$. Longwave cooling decreases slightly with an decrease in $\Theta$. This relationship
can be attributed to the cloud layer thinning slightly with decreasing $\Theta$. As the
cloud layer ascends less with increased shortwave forcing (see Fig. 4.1e), clear-sky
emission from the atmosphere above the cloud to the cloud increases.

Figure 4.1d shows integrated shortwave warming and its variation with $\Theta$ and
$N_d$ averaged over the last two hours of the simulations. Integrated shortwave heat-
ing increases with decreasing $\Theta$, as expected, but it also decreases with increasing
$N_d$. This variation is in contrast to what we found in our offline radiative com-
putations, where integrated shortwave warming increases with increasing $N_d$ at a
fixed $\Theta$ (Fig. 2.5) for similar LWPs. Clearly there are important feedbacks between
radiative heating and STBL dynamics that are leading to lower LWP at the end of the simulations with larger $N_d$. These feedbacks, which are related to entrainment (see below), in turn lead to decreasing shortwave warming with increased droplet concentration. We also note that this relationship is not monotonic; for low solar zenith angles comparing the $N_d = 500 \text{ cm}^{-3}$ and $N_d = 1000 \text{ cm}^{-3}$ simulations shows that shortwave warming increases slightly with an increased number of cloud droplets.

We can infer relative rates of entrainment through the average height of the boundary layer top ($z_i$) for the last two hours of simulation. Because all simulations started with the same initial conditions, a higher value of $z_i$ at the end of the simulation indicates a larger average rate of entrainment for that simulation. Here we define the height of the boundary layer top for a model timestep as the model layer with the maximum $\theta_l$ gradient. In Fig. 4.1e we see that $z_i$ decreases as $\Theta$ decreases, indicating that entrainment rates are decreased as radiative heating increases. This decrease in entrainment rate is likely related to the decrease in TKE. Entrainment rates have long been associated with the strength of STBL circulations in mixed-layer theory (e.g. Lock and MacVean, 1999) and in large-eddy simulation of the STBL (e.g. Caldwell and Bretherton, 2009); stronger circulations generally lead to stronger entrainment and vice versa. As $N_d$ increases for a fixed $\Theta$ $z_i$ increases, showing that entrainment rates increase when there are more smaller cloud droplets. For $N_d = 500 \text{ cm}^{-3}$ and $N_d = 1000 \text{ cm}^{-3}$ the entrainment rates are very similar for all values of $\Theta$.

To explain our modeled relationship of LWP with solar warming and how it changes with droplet concentration we must have some understanding of the interplay between the significant drivers in these ERM simulations. For this non-drizzling overcast case STBL evolution is driven primarily by radiative heating and entrainment. Integrated longwave cooling is similar for all simulations so the variation of LWP and TKE can be explained by shortwave warming and entrainment.

Entrainment is more vigorous for the simulations with higher values of $N_d$, meaning that more warming and drying is taking place via this process and would be expected to reduce LWP. To understand the impact of shortwave warming we must consider how LWP changes during the simulations. When solar warming is initially applied one hour into these ERM simulations, the LWP is $44 \text{ g m}^{-2}$. At this
value of LWP shortwave warming increases with increasing number concentration (Fig. 2.5). Taken together the increased shortwave absorption and entrainment for larger values of \( N_d \) during the simulations result in thinner cloud layers at the end for those droplet concentrations, as suggested by Boers and Mitchell (1994).

### 4.2.2 OVERCAST case, drizzling

![Figure 4.2](image)

**Figure 4.2.** Eddy resolving model output for the OVERCAST sounding with drizzle allowed. Quantities are domain averaged, vertically integrated and temporally averaged over the last two hours of the simulations. Here we display average rainfall rate and integrated drizzle flux divergence.

Next we turn to ERM results for the OVERCAST simulations with precipitation allowed. As previously stated, drizzle removes liquid water from the cloud layer, modifies entrainment across the cloudtop interface and changes the thermodynamic stability of the STBL (Stevens et al., 1998a; Ackerman et al., 2009). Numerous studies have found that, depending on the vertical profile of drizzle production and its evaporation, entrainment into and convection within the STBL can be either enhanced or reduced (Ackerman et al., 2009). For example, Stevens et al. (1998a) found that drizzle stabilizes the cloud layer with respect to the sub-cloud layer through net condensational heating of the cloud layer and net evaporative cooling of the sub-cloud layer. Drizzle also dries out downdrafts, making them “potentially buoyant,” further weakening circulations within the STBL (Stevens et al., 1998a). Following on Stevens et al. (1998a), Jiang et al. (2002) found
that evaporative cooling of drizzle in the sub-cloud layer can destabilize that region, leading to cumuliform convection that helps to couple the sub-cloud and cloud layers back together. When there was little evaporative cooling in Jiang et al.’s (2002) simulations, however, cumuliform convection in the sub-cloud layer did not develop. As we mentioned before, the strength of STBL circulations and entrainment rates are related. Consequently entrainment and drizzle are coupled processes as well.

How does the addition of the drizzle process affect our simulations? We expect differences between results for the non-drizzling (those depicted in Fig. 4.1) and drizzling cases only where drizzle is significant. The strength of drizzle production is displayed in Figure 4.2, which shows rainfall rate and cloud integrated drizzle flux divergence for these cases. Figure 4.2a shows that drizzle rates decrease with increasing $N_d$, as expected. This reduction in drizzle rate is the aerosol indirect effect as proposed by Albrecht et al. (1989); an increase in cloud droplet concentration leads to smaller droplets, precipitation suppression and an increase in cloud lifetime. The one exception to this trend is found for $N_d = 50 \text{ cm}^{-3}$ and $N_d = 100 \text{ cm}^{-3}$ when there is no shortwave forcing. In this case the drizzle rate is higher for the larger value of $N_d$. Though droplet size is an important factor in determining precipitation rate, the time these droplets spend both in-cloud and near cloudtop also play a significant role (Hartmann and Harrington, 2005b) and could explain this reversal in the dominant trend.

Figure 4.2a shows that drizzle rates decrease with decreasing $\Theta$ (increasing shortwave forcing). This reduction in drizzle rate is expected as the cloud layer warms due to more absorption of shortwave radiation causing the cloud and sub-cloud layers to decouple (Turton and Nicholls, 1987; Lu and Seinfeld, 2005). The change in drizzle rate with shortwave forcing is strongest for values of $N_d$ which produce the greatest nocturnal drizzle rates. The largest drizzle rates of 0.32 and 0.33 mm day$^{-1}$ occur for nocturnal cloud layers with $N_d$ fixed at 50 cm$^{-3}$ and 100 cm$^{-3}$, respectively and drop to 0.15 and 0.05 mm day$^{-1}$, respectively when the sun is fixed overhead. For nocturnal simulations where $N_d \geq 200 \text{ cm}^{-3}$ the drizzle rate is less than 0.10 mm day$^{-1}$. Drizzle rates in our eddy-resolving simulations are similar to those in the LES intercomparison in Ackerman et al. (2009), where the interquartile range of drizzle rates was between 0.0 and 0.4 mm day$^{-1}$. 
Figure 4.3. Eddy resolving model output for the OVERCAST sounding with drizzle allowed. Quantities are domain averaged, vertically integrated and temporally averaged over the last two hours of the simulations.
The flux divergence of drizzle can be converted into an energy flux for comparison to radiative fluxes. As cloud droplets are converted to drizzle and drizzle falls out of the cloud layer, net latent warming of the cloud occurs. Drizzle that evaporates and cools the sub-cloud layer increases the temperature gradient between the cloud layer and subcloud layer and is included in the drizzle flux divergence. Cloud-integrated drizzle flux divergence for this set of simulations is shown in Fig. 4.2b. Unsurprisingly, drizzle flux divergence, like drizzle rate, decreases with increasing shortwave forcing and decreases with increasing $N_d$. The two quantities do not mirror each other, however, because drizzle that evaporates before reaching the surface is not included in the drizzle rate. Comparing Fig. 4.2b to Fig. 4.3d we see that drizzle’s contribution to the energy budget of the STBL is of the same magnitude as solar radiative forcing, indicating that drizzle has a significant impact on STBL dynamics.

This impact can be discerned by comparing Fig. 4.1 to Fig. 4.3. We first look at the impact of drizzle on the simulations when $N_d > 200 \text{ cm}^{-3}$. For these simulations drizzle rate and drizzle flux divergence vary little with $N_d$ and drizzle flux divergence is 1 to 5 W m$^{-2}$ for all simulations. We see that LWP, STBL-averaged TKE and integrated cloud shortwave warming decrease slightly if at all. The drizzle process modifies these simulations little for all values of $\Theta$, indicating that drizzle does not play a significant role.

In Fig. 4.2b we see that drizzle flux divergence values for $N_d = 50 \text{ cm}^{-3}$ and $N_d = 100 \text{ cm}^{-3}$ vary from 25.3 and 15.8 W m$^{-2}$ respectively for nocturnal clouds to 7.3 and 2.5 W m$^{-2}$ respectively when the sun is fixed overhead. These fluxes are significant fractions of the shortwave and longwave radiative fluxes, especially when $\Theta$ is large, indicating that we might expect drizzle to significantly modify these simulations.

Comparisons of integrated quantities in Fig. 4.1 and Fig. 4.3 for $N_d = 50 \text{ cm}^{-3}$ and $N_d = 100 \text{ cm}^{-3}$ demonstrate that this is indeed the case. LWP is reduced by more than 15 g m$^{-2}$ for $N_d = 50 \text{ cm}^{-3}$ at a $\Theta = 90^\circ$ and TKE is reduced by almost half. When drizzle is allowed for these droplet concentrations integrated shortwave heating (Fig. 4.3d) decreases, associated with the reduction in LWP. Integrated longwave cooling (Fig. 4.3c) is slightly reduced for two reasons; one, LWP is reduced, and two, if the lower boundary layer top (Fig. 4.3e) is lower there
is more atmospheric mass above the cloudtop as compared to the non-drizzling case. More atmosphere above the cloud means that there is more emission of longwave radiation down to cloudtop, reducing longwave cooling there.

Through further examination of Fig. 4.1 and Fig. 4.3 we can see that drizzle reduces LWP, TKE, and radiative heating at all solar zenith angles modeled but its impact decreases with decreasing solar zenith angle. At \( \Theta = 90^\circ \) drizzle reduces LWP by 10 g m\(^{-2}\) and TKE by a little more than 30% for \( N_d = 50 \text{ cm}^{-3} \). The drizzle rate decreases with decreasing \( \Theta \) owing to increased solar absorption by the cloud layer. This increased solar warming leads to more prominent decoupling of the cloud layer from the sub-cloud, reducing moisture fluxes into the cloud from below and consequently LWP and drizzle strength. Values of \( z_i \) shown in Fig. 4.3e are essentially the same between the drizzling and non-drizzling sets of ERM simulations for all fixed solar zenith angles and droplet concentrations, indicating that for the OVERCAST case the entrainment rate is not sensitive to drizzle.

For both drizzling and non-drizzling sets of OVERCAST ERM simulations, the integrated quantities have a relatively monotonic dependence on \( \Theta \) and droplet concentration. This indicates that we can limit the number of fixed values of \( \Theta \) and \( N_d \) for which we conduct LESs and we expect that our findings will apply to intermediate values of \( \Theta \) and \( N_d \) that we do not simulate.

### 4.2.3 BROKEN case, non-drizzling

Figure 4.4 shows ERM output from our BROKEN non-drizzling simulations. Entrainment of dry air from above the inversion leads to considerably lower LWPs (compare Fig. 4.4a to Fig. 4.1a). The largest value of LWP is 32.3 g m\(^{-2}\) for \( N_d = 50 \text{ cm}^{-3} \) when there is no shortwave forcing, which is a little more than a third of the LWP simulated in the commensurate OVERCAST case. As \( \Theta \) decreases the increased shortwave warming leads to a thin broken cloud layer. For \( N_d = 500 \text{ cm}^{-3} \) and a \( \Theta = 0^\circ \) the cloud layer is almost non-existent. Liquid water path generally decreases with increased shortwave absorption and increasing droplet concentration but this is not always the case as it was for the OVERCAST set of simulations.
Figure 4.4. Eddy resolving model output for the BROKEN sounding with drizzle not allowed. Quantities are domain averaged, vertically integrated and temporally averaged over the last two hours of the simulations.
Figure 4.4f displays cloud fractions. Here we define cloudy gridcells as those with a column-integrated LWP greater than $10 \text{ g m}^{-2}$. For the OVERCAST simulations the cloud fraction was unity and hence not displayed. For the BROKEN cases we see that for all $N_d$ cloud fractions are above 0.9 for the nocturnal cases and decrease as $\Theta$ decreases. The decrease in cloud fraction with $\Theta$ varies greatly with $N_d$ and suggests that these cloud layers are sensitive to concurrent changes in radiative heating and entrainment with $N_d$. For $\Theta = 0^\circ$, the cloud fraction for $N_d = 50 \text{ cm}^{-3}$ is 0.76 and for $N_d = 500 \text{ cm}^{-3}$ the cloud fraction is 0.02. This sensitivity will be explored in later chapters.

Average STBL TKE (Fig. 4.4b) is generally larger for this set of simulations, and especially so for the simulations where $\Theta$ is large. Why there is more TKE for this set of simulations is not immediately clear. Studies have shown that energy can be produced in the STBL through cloud-top entrainment instability (Lock and MacVean, 1999). This theory was developed by Randall (1980) and Deardorff (1980) and states that if the jump in equivalent potential temperature across the boundary layer interface is negative and of a large enough magnitude, entrainment and subsequent mixing can lead to the production of negatively buoyant downdrafts and enhanced convection. However, neither our OVERCAST nor our BROKEN simulations fit this instability criterion for the last two hours of runtime.

In the following two chapters we describe large-eddy simulations (LESs) for both the OVERCAST and BROKEN cases. For fixed solar zenith angles and droplet concentrations we modeled, STBL-averaged TKE for LESs was similar between the two cases. It is possible that the modeled increase in STBL-averaged TKE we see by comparing Fig. 4.4b to Fig. 4.1b is an artifact of our ERM simulations.

Integrated longwave cooling is depicted in Fig. 4.4c. For the nocturnal simulations the emissivity of these cloud layers is close to one as LWP is not low (LWP $> 20 \text{ g m}^{-2}$). Because of the reduction of vapor content above and consequent reduction in clear-sky emission to the cloud layer as compared to the OVERCAST simulations, integrated longwave cooling is larger for the BROKEN nocturnal simulations. As $\Theta$ decreases and shortwave forcing increases, the LWPs drop to below $20 \text{ g m}^{-2}$ and the cloud fractions drop below unity. As the cloud layer thins and breaks the cloud layer average emissivities drop below one and
integrated longwave cooling is reduced, sometimes considerably.

Figure 4.4d shows that integrated shortwave heating for the BROKEN simulations is also reduced as compared to the OVERCAST simulations (Fig. 4.1d) and this reduction increases as Θ decreases. Reduction in integrated shortwave heating can also be explained by the thinning of the simulated cloud layers due to entrainment of overlying dry air and shortwave heating. As LWP decreases shortwave heating will also decrease if other factors controlling shortwave heating are fixed. Here we see a negative feedback for STBL evolution involving shortwave heating of the cloud layer; as shortwave forcing increases the cloud layer thins and breaks up, reducing total shortwave heating of the cloud layer.

We find that the height of the boundary layer $z_i$ is generally higher for the BROKEN simulations as compared to the OVERCAST simulations (Fig. 4.4e). Because of these difference in $z_i$ we can infer that entrainment is a more vigorous process for the BROKEN simulations. This increased vigor can be associated with the increased average TKE and circulation strength in the STBL.

Figures 4.4b - f clearly show that these thin cloud layers exhibit greater variability in TKE, radiative heating, entrainment rate and cloud fraction as compared to the OVERCAST simulations (Fig. 4.1) when Θ and drop concentration are varied. Although there are general trends, the relationships between Θ, $N_d$ and these quantities are not as monotonic as the OVERCAST simulations when drizzle is not allowed.

### 4.2.4 BROKEN case, drizzling

We reran the same set of simulations with the BROKEN sounding for drizzling clouds. Figure 4.5 displays rainfall rate and cloud-integrated drizzle flux divergence for these cases. The maximum drizzle rate of 0.11 mm day$^{-1}$ occurs for $N_d = 50$ cm$^{-3}$ when shortwave forcing is absent and is a third of the maximum drizzle rate seen in the OVERCAST cases. The associated drizzle flux divergence for this case is 6.5 W m$^{-2}$. The change of drizzle strength with Θ is highly nonlinear, and especially so for $N_d = 50$ cm$^{-3}$ and $N_d = 100$ cm$^{-3}$. For other droplet concentrations the drizzle flux divergence does not exceed 1.5 W m$^{-2}$. In general drizzle is weak because the cloud layers are thin and broken with low LWPs. While
Figure 4.5. Eddy resolving model output for the BROKEN sounding with drizzle allowed. Quantities are domain averaged, vertically integrated and temporally averaged over the last two hours of the simulations. Here we display average rainfall rate and integrated drizzle flux divergence.

The peculiar variation of drizzle rate with $\Theta$ shown in Fig. 4.5 gives us reason for caution in interpretation of output for these cases. Like the modeled differences in STBL-averaged TKE between the BROKEN and OVERCAST cases we discussed earlier, this peculiar variation of drizzle rate reminds us of potential shortcomings in the use of ERM simulations. As we stated in the beginning of this chapter, we wish only to obtain a general picture of how the STBL responds to microphysical and radiative variations and how sensitive this response is. Therefore, in lieu of detailing the impact of drizzle on the BROKEN case, we simply note that the drizzle process, though not vigorous, alters the results somewhat as compared to that seen in Fig. 4.4. However, drizzle does not change the general overall trends in the integrated quantities we investigate here.

Although not as monotonic as we found for the OVERCAST case, integrated quantities investigated for our BROKEN case also have a relatively monotonic dependence on $\Theta$ and droplet concentration. This gives us confidence that we can limit the number of fixed values of $\Theta$ and $N_d$ for which we conduct LESs, as we do for the OVERCAST case. Similarly, we expect that our findings will apply to
intermediate values of $\Theta$ and values of $N_d$ that we do not simulate.
5.1 Purpose

Through our eddy-resolving model (ERM) simulations we established a general understanding of the impact of radiative heating on stratocumulus-topped boundary layer (STBL) dynamics. Now we investigate the impacts of radiative heating in detail using a limited set of large-eddy simulations (LES). First we look at LESs of the nocturnal STBL to model the impact longwave radiative transfer has on STBL dynamics. We do this first step for two reasons. First, as we said in our introduction, a sound understanding of our nocturnal STBL simulations will give us a firm platform with which to explore solar impacts on the STBL. Second, we wish to investigate possible interactions between longwave radiative heating, droplet concentration, droplet size distribution and cloud dynamics and whether or not it is necessary to account for these in simulations of the STBL. Recall that in Chapter 2 we determined that variations in the shape parameter, $\nu$, of our assumed
gamma droplet size distribution had a minimal impact on computed radiative flux divergence. Because of this minimal impact we focus our efforts on the dependence of longwave radiative heating on droplet concentration.

As Garrett and Zhao (2006) noted and our offline calculations in Chapter 2 corroborate, cloud longwave emissivity varies significantly with droplet concentration when LWP is low. Our ERM results described in Chapter 4 suggest that the dependence of longwave radiative heating on droplet concentration could have implications for STBL dynamics (see Fig. 4.4 and attendant discussion). This is especially plausible for the simulations of the BROKEN case, where domain-averaged LWP was 32.3 g m\(^{-2}\) or lower. Lee et al. (2009) investigated the effects of varying aerosol concentration on thin stratocumulus clouds using ERM simulations. They found that increased aerosol concentration and concurrent increase in cloud droplet concentration leads to increased condensation, latent heating and subsequently stronger circulations. Together, increased condensation and stronger circulations led to an increase in LWP when precipitation reached the surface and a decrease in LWP when precipitation did not reach the surface. Changes of radiative fluxes and heating with aerosol concentration and the subsequent impact of these changes on STBL dynamics were not a focus of the study by Lee et al. (2009).

By using LES throughout the rest of our study we free ourselves from some of the limitations of ERM simulations discussed in the previous chapter. From our LESs of the STBL we will explore and interpret STBL dynamics through vertical profiles of atmospheric properties in addition to time series of integrated quantities. We acknowledge that LES, though it is the best model we have at present, gives far from a perfect representation of the atmosphere.

Stevens et al. (2005) studied the ability of a suite of LES codes to simulate nocturnal stratocumulus observed during the second Dynamics and Chemistry of Marine Stratocumulus (DYCOMS-II) field study. They found that using the most basic configuration in most LES codes tested led to an overestimate of mixing at the cloudtop interface. This overentrainment led in turn to underestimation of turbulent intensities and LWP. This outcome is sobering as LWP is of first-order importance in describing the STBL. We emphasize the use of LES only as a tool to investigate the potential impact of radiative heating in the STBL. We will suggest
a potential field study in which these impacts could be observed in the atmosphere at the end of this work.

We conducted six hour LESs of the nocturnal STBL using both the OVERCAST and BROKEN stratocumulus soundings. We ran four LESs with each sounding, setting $N_d$ to fixed values of 50 and 1000 cm$^{-3}$ and varying whether or not drizzle is allowed. The ERM simulations described in the previous chapter give us confidence that simulations with these low and high fixed values of $N_d$ will bracket the dynamic responses we can expect to see for intermediate values of $N_d$.

5.2 Results: OVERCAST case, non-drizzling

As for our ERM simulations, we investigate the simplest set of cases first, i.e. the OVERCAST simulations without drizzle. We examine time series output first and vertical profiles next.

5.2.1 Time Series

For all OVERCAST LESs the cloud fractions were unity throughout (not shown). Fig. 5.1 shows time series output for the two LESs using the OVERCAST sounding and no drizzle. The first hour of “spin-up” simulation time is omitted from these plots. In Fig. 5.1a we see that domain-averaged LWP depends on $N_d$ and is higher throughout the simulations for $N_d = 50$ cm$^{-3}$. The domain-averaged LWP averaged over the last two hours of simulation time is 93.7 g m$^{-2}$ for $N_d = 50$ cm$^{-3}$ and is 83.6 g m$^{-2}$ for $N_d = 1000$ cm$^{-3}$. As in the ERM simulations of the nocturnal STBL, longwave radiative heating and entrainment are the primary drivers of these LESs and we anticipate that these processes play a role in the LWP dependence on droplet concentration.

The time evolution of the average height of the boundary layer top $z_i$ can be used to infer entrainment rates as we did in the previous chapter, and is displayed in Fig. 5.1b. In this plot we see that for both LESs $z_i$ is constant and then jumps rapidly by 30m where it again remains constant. This change of boundary layer height with time is an artifact of the discretization of the atmosphere in LES. This phenomenon is commonly termed “grid-hopping” or “staircasing” and has
Figure 5.1. Large-eddy simulation output for the OVERCAST sounding without drizzle. Time series of quantities are domain averaged and vertically integrated.

been noted in other LES studies (e.g. Bretherton et al., 1999; Lock and MacVean, 1999). The vertical grid-spacing of our simulations is 30 m and so the cloud layer top and associated maximum in liquid water potential temperature, \( \theta_l \), gradient can ascend or descend in discrete values of 30 m for a model column. When this ascension occurs for all model columns \( z_i \) increases by 30 m.

Because this grid-hopping of \( z_i \) occurs earlier and more frequently when \( N_d = 1000 \text{ cm}^{-3} \), the modeled entrainment is more vigorous in that simulation compared to when \( N_d = 50 \text{ cm}^{-3} \). To investigate why modeled entrainment increases with droplet concentration we will examine vertical profiles of relevant quantities in Fig. 5.2 below.
Figure 5.1c shows the time evolution of integrated cloud longwave cooling for these two simulations. When $N_d = 1000 \, \text{cm}^{-3}$ there is less integrated longwave cooling throughout the simulation. The integrated cooling averaged over the domain and the last two hours of simulation time is $-60.6 \, \text{W m}^{-2}$ for $N_d = 50 \, \text{cm}^{-3}$ and $-58.9 \, \text{W m}^{-2}$ for $N_d = 1000 \, \text{cm}^{-3}$. This difference can be explained with the same reasoning used for Fig. 2.2; at cloudbase emissivities are not close to unity and increase with increased droplet concentration. This leads to more cloudbase absorption of longwave radiation emitted by the surface when the droplet concentration is high, decreasing the cloud integrated longwave cooling. Liquid water path is also larger for $N_d = 50 \, \text{cm}^{-3}$ and integrated longwave cooling increases slightly with LWP.

Figure 5.1d shows the time evolution of domain-averaged STBL turbulent kinetic energy (TKE) from which we can infer the comparative strength of boundary layer circulations. Apart from a short period during the fifth hour of simulation time, TKE is higher throughout the simulation for $N_d = 50 \, \text{cm}^{-3}$ indicating that circulations are generally stronger. We had said previously that in both mixed-layer models and LESs strong boundary layer circulations are generally associated with strong entrainment and vice versa. However, the two are not always related. In these LESs entrainment is more vigorous for $N_d = 1000 \, \text{cm}^{-3}$ but there are weaker circulations as compared to when $N_d = 50 \, \text{cm}^{-3}$ (compare Fig. 5.1b to Fig. 5.1d). That the simulation with less TKE has more vigorous entrainment suggests that changes in entrainment with droplet concentration may be associated with processes at the STBL interface instead of the strength of the largest circulations.

### 5.2.2 Vertical Profiles

To further our understanding of these simulations, we show the vertical profiles of several quantities in Fig. 5.2. Following Ackerman et al. (2009), we first normalized five minute horizontally-averaged vertical profiles by the height of the boundary layer top, $z_i$, defined as the model layer with the maximum $\theta_l$ gradient. We next interpolated these normalized data to a uniform vertical grid. These uniformly gridded data were then temporally averaged over the last two hours of simulation.
Figure 5.2. Large-eddy simulation output for the OVERCAST sounding without drizzle. Horizontally averaged vertical profiles. Data were temporally averaged over five minutes, and then were normalized by the height of the boundary layer top, $z_i$. These data were then interpolated to a uniform vertical grid. These uniformly gridded data were then temporally averaged over the last two hours of simulation time.
Cloud liquid water content (LWC) is shown in Fig. 5.2a. Although the simulated cloud layer for \( N_d = 50 \text{cm}^{-3} \) has a lower cloudbase and a higher LWC throughout the cloud layer than the simulated cloud layer for \( N_d = 1000 \text{cm}^{-3} \), the profiles of horizontally averaged LWC have similar structure. Liquid water content increases almost linearly from cloudbase to near cloudtop then drops off rapidly towards zero at the top of the STBL.

Figure 5.2b depicts both shear and buoyancy production of TKE for these two simulations. Because we initialized our simulations with soundings that contained no vertical wind shear, TKE in our STBL is generated primarily through buoyancy production. Except in the surface layer shear production of TKE is near zero.

We can discern a few properties of the STBL through examination of the buoyancy production profiles for these two LESs. First we see that for both droplet concentrations positive buoyancy production of TKE is maximized near cloudtop. The positive maxima correspond to longwave radiative cooling from cloudtop (Fig. 5.2e) that drives the STBL circulations, on average generating negative buoyancy in air parcels that subsequently sink to lower altitudes (Nicholls, 1989). Above these positive maxima there is a peak of negative buoyancy production, or buoyancy destruction, for both droplet concentrations. For LES, this negative peak at the STBL top has been associated with the entrainment process (e.g. Stevens et al., 1999). Warm air mixed into the STBL from above the cloudtop interface increases positive buoyancy for air parcels at the STBL top. This increase in positive buoyancy detracts somewhat from the strength of circulations generated through longwave cooling.

In the cloud layer the buoyancy production curves are similar for the two droplet concentrations modeled. Below cloudbase, however, the buoyancy production curves are dissimilar. For both droplet concentrations there is a notable decrease in buoyancy production at and below cloudbase near 500 m. For \( N_d = 50 \text{cm}^{-3} \), however, buoyancy production first increases before it decreases as we descend to the surface and there is positive buoyancy production of TKE throughout the STBL. Such a curve indicates that downdrafts generated by longwave cooling at cloudtop are likely to reach the surface and that the STBL is well mixed. For \( N_d = 1000 \text{cm}^{-3} \), there is negative buoyancy production, or destruction, of TKE.
below the cloud layer from about 250 m down to the surface. Excepting the entrainment zone, negative buoyancy production anywhere within the STBL suggests some amount of decoupling between the cloud layer and sub-cloud layer (Stevens, 2000). This negative buoyancy production also shows that downdrafts originating at cloudtop are hindered in their journey to the surface as mechanically-forced mixing is required, reducing TKE (Schubert et al., 1979).

We can find further evidence for the extent of decoupling within the STBL by examining vertical profiles of vertical velocity variance $w'w'$ (Fig. 5.2c). The location of this maximum indicates where the strongest variation in updrafts and downdrafts, and hence circulations, lies for this value of $N_d$. For $N_d = 50 \text{ cm}^{-3}$, the maximum in $w'w'$ is in the middle of the STBL. Because the strongest vertical component of TKE is below the cloud layer and $w'w'$ is not small except at the boundary layer interfaces, we can infer that this STBL is not decoupled. For $N_d = 1000 \text{ cm}^{-3}$ the maximum in $w'w'$ is near cloudbase and is shifted to higher altitudes as compared to $N_d = 50 \text{ cm}^{-3}$. Because the vertical velocity variance is weaker in the sub-cloud layer, we can infer that vertical motion there is weak and hence the STBL is partially decoupled. The larger values of $w'w'$ for $N_d = 50 \text{ cm}^{-3}$ as compared to $N_d = 1000 \text{ cm}^{-3}$ also show us that circulations for the former are stronger than circulations for the latter.

Fig. 5.2b we see a local minimum in buoyancy production at cloudbase for $N_d = 50 \text{ cm}^{-3}$. This local minimum indicates that the model STBL may be close to decoupling at that droplet concentration. Consequently, these vertical profiles of buoyancy production may suggest that this case could be particularly sensitive to potential changes in thermal structure related to radiative transfer or entrainment. Therefore, the sensitivities we find in our study are possibly dependent on the soundings we use to initialize our simulations, and are not necessarily representative of those that would be found with different initializations.

### 5.3 Discussion: OVERCAST case, non-drizzling

In addition to a LWP dependence on droplet concentration these two LESs exhibit markedly different dynamical structures related to $N_d$. How does changing droplet concentration in our model STBL alter its LWP and dynamical structure? First we
examine modeled differences in entrainment. An increase in warming and drying of air within the entrainment zone could explain decreases in LWP. We expect that, for similar amounts of dry and moist air, evaporative cooling should be more rapid if \( N_d \) is larger. Since the droplets are more numerous, they are smaller and there is a larger surface area over which net evaporation or condensation can take place. Additionally, increased mixing of warm air into the STBL from above with increasing \( N_d \) could be associated with decreases in circulation strength and decoupling of the cloud layer from the sub-cloud layers (e.g. Schubert et al., 1979).

Recall that entrainment is more rapid for \( N_d = 1000 \text{ cm}^{-3} \) (Fig. 5.1b). Although there is evidence of some decoupling for this value of \( N_d \), both simulations have profiles of \( \overline{w'w'} \) that indicate they are mixed down to the surface and mixed layer theory may apply to some extent. For mixed layers, entrainment rate can be related to the strength of the inversion in terms of the liquid water potential temperature. Liquid water potential temperature, \( \theta_l \), accounts for the contribution of enthalpy of phase change of water to changes in potential temperature \( \theta \). Because of this accounting, vertical profiles of \( \theta_l \) are useful in determining how easily air can move vertically when sensible and latent heating occur.

In mixed-layer theory, buoyancy flux at the STBL interface is related to the entrainment rate and inversion strength by

\[
\overline{w'\theta_l'} = w_e \Delta \theta_l,
\]

where \( w_e \) is the entrainment velocity and \( \Delta \theta_l \) is the jump in liquid water potential temperature across the inversion (Lilly 1968). The buoyancy fluxes at the STBL interface as depicted in Fig. 5.2b are very similar between the two simulations. If these buoyancy fluxes are taken as the same, then the entrainment rate should decrease as inversion strength increases.

Fig. 5.2d shows vertical profiles of \( \theta_l \). We can see that \( \theta_l \) is virtually independent of height within the STBL up to the inversion and that the STBL is well-mixed for both droplet concentrations. The liquid water potential temperature is slightly higher for \( N_d = 1000 \text{ cm}^{-3} \) and this could be related to entrainment and/or longwave radiative transfer. For this droplet concentration, either less integrated longwave cooling, or more mixing in of warm, dry air from above, could explain the higher value of \( \theta_l \) for the STBL as compared to \( N_d = 50 \text{ cm}^{-3} \). We
can also see that the STBL inversion is stronger for \( N_d = 50 \text{ cm}^{-3} \); \( \theta_l \) increases more with height for the lower droplet concentration. This difference in inversion strength is likely part of the reason for the difference in entrainment rates.

Because radiative heating and entrainment are the primary drivers within these simulations, we next turn to vertical profiles of longwave radiative heating for the two LESs (Fig. 5.2e). We see clear differences in both longwave cooling at cloudtop and longwave warming at cloudbase between the two simulations. The maximum in longwave cooling for \( N_d = 1000 \text{ cm}^{-3} \) is \(-4.1 \text{ K hr}^{-1}\), compared to \(-3.5 \text{ K hr}^{-1}\) for \( N_d = 50 \text{ cm}^{-3} \). The maximum cooling rate near cloudtop is larger for \( N_d = 1000 \text{ cm}^{-3} \) but radiative cooling for \( N_d = 50 \text{ cm}^{-3} \) extends further below cloudtop. Maxima in longwave warming near cloudbase are \(0.3 \text{ K hr}^{-1}\) for \( N_d = 1000 \text{ cm}^{-3} \) and \(0.2 \text{ K hr}^{-1}\) for \( N_d = 50 \text{ cm}^{-3} \), and the warming for the \( N_d = 1000 \text{ cm}^{-3} \) simulation extends further above cloudbase.

These differences in profiles of modeled longwave radiative heating could be explained by: a) differences in the model cloud structure and LWC owing to dynamical responses, as we see in Fig. 5.2a, b) by changes in longwave radiative transfer with droplet concentration, as we discussed in Chapter 2, or c) some combination of a) and b). To separate the two effects as much as possible we compute longwave radiative heating profiles for both droplet concentrations offline for a static model atmosphere as we did in Chapter 2. For this comparison we choose a cloud of 320 m geometric depth and LWP of 105.9 g m\(^{-2}\), for two reasons. One, this static cloud layer’s emissivity at cloudtop is unity like those simulated here, and two, this LWP is comparable to those at the end of our LESs. Figure 5.3 shows the longwave radiative heating profiles for this static atmosphere.

Cloud integrated longwave cooling is not sensitive to \( N_d \) for clouds of this LWP (Fig.2.2). The computed cloud integrated longwave cooling for this static atmosphere is \(-69.6 \text{ W m}^{-2}\) for \( N_d = 1000 \text{ cm}^{-3} \) and \(-72.1 \text{ W m}^{-2}\) for \( N_d = 50 \text{ cm}^{-3} \), a 3% difference. We see similar differences in integrated cooling in Fig. 5.1c. However, Fig. 5.3 shows us that the in-cloud longwave radiative heating profile depends on the modeled droplet concentration in significant ways. Between these two droplet concentrations, the peak longwave cooling rate at cloudtop is different by 31% (\(-14.4 \text{ K hr}^{-1}\) for \( N_d = 1000 \text{ cm}^{-3} \) and \(-10.5 \text{ K hr}^{-1}\) for \( N_d = 50 \text{ cm}^{-3} \)). As in our LESs, longwave cooling extends further into the cloud for the lower droplet
Figure 5.3. In-cloud longwave radiative heating rates for $N_d = 50$ and 1000 cm$^{-3}$ for the static model atmosphere described in Chapter 2. Cloud added to mid-latitude summer (MLS) profile extends from 1.0 km at base to 1.32 km at top. Liquid water path is 105.9 g m$^{-2}$.

concentration. The peak longwave warming rate near cloudbase is 0.8 K hr$^{-1}$ for $N_d = 1000$ cm$^{-3}$ and 0.6 K hr$^{-1}$ for $N_d = 50$ cm$^{-3}$, a 28% difference.

Figure 5.3 shows us that, independent of dynamical feedbacks, profiles of cloud longwave heating change when the modeled droplet concentration changes. Could these differences in longwave heating related to changing $N_d$ help to explain the differences in entrainment, and consequently LWP and STBL dynamical structure, that we see in our two LESs? To examine this possibility, we ran a LES where $N_d = 50$ cm$^{-3}$ but use 1000 cm$^{-3}$ as the value of $N_d$ for computation of radiative heating within the cloud. We will refer to this LES as $N_d = 50$ cm$^{-3}$ with 1000 cm$^{-3}$ for radiation. If the LWP and STBL dynamical structure modeled with this LES configuration are better aligned with those from our LES where $N_d = 1000$ cm$^{-3}$, we can infer that the radiative heating difference for the two droplet concentrations plays a significant role in the divergence of the two runs.

Figure 5.4 compares results from the LES where $N_d = 50$ cm$^{-3}$ with 1000 cm$^{-3}$ for radiation to those with $N_d = 50$ cm$^{-3}$ and $N_d = 1000$ cm$^{-3}$. Figure 5.4a and Fig. 5.4b display time series of LWP and STBL-averaged TKE, respectively. The use of 1000 cm$^{-3}$ for radiative computations while using $N_d = 50$ cm$^{-3}$ for all other processes brings the time evolution of both integrated quantities into closer agreement with the LES where $N_d = 1000$ cm$^{-3}$. Figure 5.4c shows the
Figure 5.4. Large-eddy simulation output for the OVERCAST sounding without drizzle. Three LESs: $N_d = 50 \text{ cm}^{-3}$, $N_d = 1000 \text{ cm}^{-3}$ and $N_d = 50 \text{ cm}^{-3}$ with $1000 \text{ cm}^{-3}$ for radiation. a.) to c.) Time series of quantities that are domain averaged and vertically integrated. d.) to f.) Horizontally averaged vertical profiles. Data were temporally averaged over five minutes, and then were normalized by the height of the boundary layer top, $z_i$. These data were then interpolated to a uniform vertical grid. These uniformly gridded data were then temporally averaged over the last two hours of simulation time.
time evolution of $\Sigma_i$ for these two LESs and shows that entrainment rates are also similar for the $N_d = 50 \text{ cm}^{-3}$ with 1000 cm$^{-3}$ for radiation and $N_d = 1000 \text{ cm}^{-3}$ LESs.

Figure 5.4d and Fig. 5.4e show vertical profiles of $\overline{w'w'}$ and buoyancy production of TKE, respectively. As with the time series output, the profiles for $N_d = 50 \text{ cm}^{-3}$ with 1000 cm$^{-3}$ for radiation and $N_d = 1000 \text{ cm}^{-3}$ are more similar than the profiles from $N_d = 50 \text{ cm}^{-3}$. Though the maxima in $\overline{w'w'}$ are located at different normalized heights, the strength of circulations exhibited in Fig. 5.4d for $N_d = 50 \text{ cm}^{-3}$ with 1000 cm$^{-3}$ for radiation is almost the same as for $N_d = 1000 \text{ cm}^{-3}$ This similarity in circulations is also indicated by buoyancy destruction below cloudbase for both simulations.

The increase in droplet concentration for longwave radiative heating computation while using the lower droplet concentration for all other processes results in lower LWPs, more vigorous entrainment, less vigorous circulations and buoyancy destruction in the sub-cloud layer, all of which are characteristic of the $N_d = 1000 \text{ cm}^{-3}$ LES. Note that we have not strictly isolated feedbacks between radiative heating and other processes in our model STBL. Differences in droplet concentration for modeled evaporation, condensation and entrainment in the two LESs ($N_d = 50 \text{ cm}^{-3}$ and $N_d = 1000 \text{ cm}^{-3}$) lead to differences in radiative heating through these processes. Nevertheless, the results shown in Fig. 5.4 give us confidence that changes in longwave radiative heating with changes in $N_d$ are an important reason why the two LESs ($N_d = 50 \text{ cm}^{-3}$ and $N_d = 1000 \text{ cm}^{-3}$) diverge substantially.

Physically, how can $N_d$-induced changes in longwave radiative heating profiles lead to these modeled changes in STBL evolution? Recall that for the two droplet concentrations modeled, we saw that the inversion at STBL top was weaker for $N_d = 1000 \text{ cm}^{-3}$ (Fig. 5.2e). In Fig. 5.5 we see that the inversion strength for the two simulations $N_d = 1000 \text{ cm}^{-3}$ and $N_d = 50 \text{ cm}^{-3}$ with 1000 cm$^{-3}$ for radiation are similar, suggesting that their similarities in longwave radiative heating are related to similarities in inversion strength and entrainment. Here we discuss two possible reasons to relate longwave radiative heating and inversion strength together.

First, changes in inversion strength with droplet concentration could be related
Figure 5.5. Large-eddy simulation output for the OVERCAST sounding without drizzle. 3 LES: $N_d = 50$ cm$^{-3}$, $N_d = 1000$ cm$^{-3}$ and $N_d = 50$ cm$^{-3}$ with 1000 cm$^{-3}$ for radiation). Vertical profile of $\theta_l$, horizontally averaged. Data were temporally averaged over five minutes, and then were normalized by the height of the boundary layer top, $z_i$. These data were then interpolated to a uniform vertical grid. These uniformly gridded data were then temporally averaged over the last two hours of simulation time.

to changes in integrated longwave cooling. Recall that there is less overall cloud layer cooling for the higher droplet concentration (Fig. 5.1c). Over time this would lead to a higher value of $\theta_l$ in the STBL and, everything else equal, a weaker inversion. Keep in mind that more rapid entrainment in itself can lead to a weaker inversion, as mixing of warmer air from above would also raise the value of $\theta_l$ in the STBL.

Second, changes in radiative cooling rate with $N_d$ at the STBL interface could explain the consequent decrease in $\theta_l$ inversion strength with $N_d$. Similar to how warming the air in the STBL inversion weakens the inversion, radiative cooling within the inversion layer can also weakens the inversion. It has been theorized that cloudy thermals penetrating into the inversion layer lead to radiative cooling in areas where those penetrations occur, promoting entrainment (e.g. Lock and MacVean, 1999). The penetrating thermals that radiatively cool the inversion would likely be optically thin and their emissivities would be sensitive to droplet concentration. Increases in droplet concentration can subsequently lead to increased radiative cooling in these penetration areas. This increased radiative cooling could further weaken the inversion strength in these areas and subsequently entrainment could be more efficient.
These arguments are difficult to provide evidence for as both radiative cooling and entrainment are inextricably linked together and feedback on each other. Nevertheless, these three LESs suggest that the LWP and dynamical structure of the STBL is dependent on the droplet concentration and associated changes in longwave radiative heating.

We ran an additional LES of the nocturnal STBL with an intermediate value of $N_d = 200\,\text{cm}^{-3}$. The purpose of this additional simulation was to check if the differences in nocturnal STBL evolution and associated explanations for these differences apply to intermediate values of $N_d$. Figure 5.6 shows us that this is indeed the case. For $N_d = 200\,\text{cm}^{-3}$ vertical profiles of cloud LWC, buoyancy production of TKE, $w'w'$ and longwave radiative heating rate all exhibit behaviors that lies between the higher and lower droplet concentrations.

5.4 Results: OVERCAST case, drizzling

Drizzle influences nocturnal STBL evolution and the impact that longwave radiative heating has on cloud dynamics. To examine these influences we ran two LESs with the same values of $N_d$ as before, i.e. $N_d = 50$ and $1000\,\text{cm}^{-3}$, with drizzle. From our OVERCAST drizzling ERM simulations detailed in Chapter 4 we expect drizzle to play a significant role in STBL evolution and dynamics for $N_d = 50\,\text{cm}^{-3}$ and drizzle to be less important in STBL evolution and dynamics for $N_d = 1000\,\text{cm}^{-3}$.

5.4.1 Time Series

We first turn to time series output in Figure 5.7. Figure 5.7a shows the time evolution of LWP for the LESs in which drizzle is and is not allowed. We can see that the drizzle process reduces LWP for both droplet concentrations. For $N_d = 50\,\text{cm}^{-3}$ LWP at the end of the simulation time is reduced from 110 to 80 g m$^{-2}$, a change of 27.2% when drizzle is included. For $N_d = 1000\,\text{cm}^{-3}$ LWP at the end of the simulation time is reduced from 95 to 87 g m$^{-2}$ for a change of 13.6%.

Average rainfall rate and integrated drizzle flux divergence as shown in Fig. 5.7b and c clearly show drizzle to be a more significant process for the lower droplet concentrations.
Figure 5.6. Large-eddy simulation output for OVERCAST sounding without drizzle. 3 LES: \( N_d = 50 \text{cm}^{-3}, N_d = 200 \text{cm}^{-3} \) and \( N_d = 1000 \text{cm}^{-3} \). Horizontally averaged vertical profiles. Data were temporally averaged over five minutes, and then were normalized by the height of the boundary layer top, \( z_i \). These data were then interpolated to a uniform vertical grid. These uniformly gridded data were then temporally averaged over the last two hours of simulation time.
Figure 5.7. Large-eddy simulation output for the OVERCAST sounding with drizzle. Time series of quantities are domain averaged and vertically integrated.
concentration, as expected (Albrecht, 1989; Sandu et al., 2007). Integrated drizzle flux divergence averaged over the last two hours of the simulation is 18.5 W m\(^{-2}\) for \(N_d = 50 \text{ cm}^{-3}\) and 4.0 W m\(^{-2}\) for \(N_d = 1000 \text{ cm}^{-3}\). The associated average drizzles rates are 0.29 mm day\(^{-1}\) for \(N_d = 50 \text{ cm}^{-3}\) and 0.11 mm day\(^{-1}\) for \(N_d = 1000 \text{ cm}^{-3}\). The net effect of the drizzle process in both cases is to warm the STBL, and we might expect this to weaken the strength of the \(\theta_l\) inversion. We will return to this possibility when we examine vertical profiles of \(\theta_l\).

To look at how entrainment and STBL-averaged TKE are affected by the drizzle process we look at time series output from the four LESs, \(N_d = 50\) and 1000 cm\(^{-3}\) in which drizzle is and is not allowed. Fig. 5.7d shows that \(\overline{z_t}\) ascends slower for both droplet concentrations when drizzle is allowed and indicates that, for this case, the drizzle process reduces entrainment rates. Fig. 5.7e shows that STBL-average TKE is reduced through the addition of drizzle for both droplet concentrations. As we noted in the previous chapter, drizzle has been found to both enhance and reduce STBL entrainment rates and circulation strength (Ackerman et al., 2009). We will endeavor to understand why drizzle reduces the vigorousness of entrainment and convection for our particular case through the analysis of vertical profiles.

### 5.4.2 Vertical Profiles

Figure 5.8a shows the domain-averaged vertical distribution of LWC. Echoing what we found in the time evolution of LWP displayed in Fig. 5.7a, LWC is reduced for the drizzling cases and this reduction is larger for \(N_d = 50 \text{ cm}^{-3}\). While LWP changes when drizzle is allowed, the basic average vertical structure of LWC does not change.

Figure 5.8b and c show the resolved scale buoyancy production of TKE and \(\overline{w'w'}\), respectively. Recall that the vertical profiles of buoyancy production of TKE and \(\overline{w'w'}\) in Fig. 5.2 indicated that the STBL is partially decoupled when \(N_d = 1000 \text{ cm}^{-3}\) for the non-drizzling case but is coupled when \(N_d = 50 \text{ cm}^{-3}\). For \(N_d = 1000 \text{ cm}^{-3}\) we see that the addition of drizzle slightly increases buoyancy production of TKE within the cloud layer and slightly reduces it in the sub-cloud layer. Vertical velocity variance is reduced for this droplet concentration through-
Figure 5.8. Large-eddy simulation output for OVERCAST sounding with drizzle. Horizontally averaged vertical profiles. Data were temporally averaged over five minutes, and then were normalized by the height of the boundary layer top, \( z_i \). These data were then interpolated to a uniform vertical grid. These uniformly gridded data were then temporally averaged over the last two hours of simulation time.
out the STBL and the maxima are shifted to a slightly higher normalized altitude. These reductions in $\overline{\omega'w'}$ can be substantial, maximizing at 23% at 0.5 $z_i$.

Considering the reductions in LWC, these reductions suggest that though the drizzle process may be weak for $N_d = 1000$ cm$^{-3}$, it is not necessarily negligible.

For $N_d = 50$ cm$^{-3}$ the addition of drizzle results in a substantial reduction of buoyancy production of TKE in the sub-cloud layer, leading to buoyancy destruction (Fig. 5.8b). We did not see this buoyancy destruction when drizzle was not allowed and as we previously noted, buoyancy destruction observed anywhere in the STBL can indicate some amount of decoupling (Stevens, 2000). Changes in buoyancy production of TKE within the cloud layer owing to drizzle are negligible. These reductions in buoyancy production could be relate to changes in vertical thermodynamical structure and “potential buoyancy” (Stevens et al., 1998a), and are discussed below.

Figure 5.8c shows that $\overline{\omega'w'}$ is significantly reduced for $N_d = 50$ cm$^{-3}$ when drizzle is allowed, as has been demonstrated in previous LES studies (e.g. Stevens et al., 1998a; Ackerman et al., 2009). The maximum in $\overline{\omega'w'}$ is reduced by 35% and is shifted up into cloudbase, again indicating that the STBL is not as well coupled as when drizzle is not allowed. How is drizzle changing the dynamical structure of our model STBL? Changes in longwave radiative heating and entrainment owing to drizzle are expected to be important in changing STBL dynamics, as is warming and cooling associated with net formation and evaporation of drizzle.

There are small reductions in integrated longwave cooling between the drizzling and non-drizzling cases for these droplet concentrations; averaged over the last two hours of simulation these reductions are less than 0.5 W m$^{-2}$ (not shown). In Fig. 5.8d we examine the longwave radiative heating profiles from the four LESs. The distribution of longwave radiative heating is altered slightly because of changes in the LWC distribution. For both droplet concentrations the maxima in longwave cooling are reduced in the drizzling case, and this reduction is more significant for $N_d = 50$ cm$^{-3}$. This change in longwave heating profile could be related to the modeled differences in entrainment rate. Reductions in longwave cooling in the entrainment zone may reduce local weakening of the inversion and hence reduce entrainment rate, as discussed for the non-drizzling case.

Figure 5.8e shows the net latent heating rates associated with drizzle formation.
and evaporation. The flux divergence of drizzle can be converted into an energy
flux and can then be used to compute flux divergences and heating rates, as is
commonly done with radiative fluxes. For \( N_d = 50 \text{ cm}^{-3} \) we see that there is
warming due to net formation of drizzle inside the cloud layer and net cooling due
to net evaporation of drizzle in the sub-cloud layer. The warming maximizes at
0.25 K hr\(^{-1}\) near the top of the STBL. Net evaporation of drizzle occurs throughout
the sub-cloud layer, maximizing at under 0.05 K hr\(^{-1}\) near the surface. For \( N_d =
1000 \text{ cm}^{-3} \) warming owing to net formation of drizzle maximizes at 0.05 K hr\(^{-1}\)
near the top of the STBL. There is negligible net evaporation of drizzle modeled
for this droplet concentration below cloudbase.

Warming and cooling values associated with drizzle are an order of magnitude
lower than longwave radiative heating rates. Nevertheless these drizzle heating
rates can help to explain the reduction in buoyancy production and TKE when
drizzle is allowed. By warming the cloud layer drizzle reduces buoyancy production
of TKE at cloudtop owing to radiative cooling. It is also possible that warming in
the entrainment zone owing to drizzle could act in concert with the reductions in
radiative cooling there, strengthening the local inversion and hence reducing the
entrainment rate.

Furthermore, by warming the cloud layer and cooling the sub-cloud layer drizzle
also increases the temperature difference between cloudbase and the sub-cloud
layer. The increased temperature difference produces stability between the cloud
and sub-cloud layers which leads to reduced mixing between the two layers, an
effect of drizzle seen in other LES studies (Stevens et al., 1998a). We detailed this
effect of drizzle in an earlier schematic (Fig. 1.1b).

Drizzle can also lead to circulations within the cloud layer coupled together
with circulations in the sub-cloud layer. If cooling and moistening of the sub-
cloud layer by net evaporation occurs near cloudbase, this can lead to conditional
instability and possibly cumulus cloud formation (Stevens et al., 1998a; Jiang et
al., 2002). This conditional instability is not found at all horizontal for some areas
below cloudbase; drizzle stabilizes the cloud with respect to the cloud in some areas
and results in instability in other areas. When this instability is resolved, vertical
motion is generated below cloud and the stratiform layer and the sub-cloud layer
can be recoupled. Figure 5.9 shows how drizzle can lead to partial recoupling of
the boundary layer from some locations at cloud base.

Figure 5.9. Schematic diagrams showing how the drizzle process can lead to partial, intermittent coupling of the boundary layer. a) In the decoupled STBL, drizzle evaporates below cloud base and generates instability. b) The instability is resolved and circulations in the sub-cloud and cloud layers reconnect.

In our nocturnal simulations this generation of conditional instability may be occurring, but buoyancy generation of TKE below could decreases in the drizzling simulations. This decrease suggests the increased temperature difference between the cloud and sub-cloud layer owing to drizzle is more important for these nocturnal simulations. We find that drizzle may lead to partial coupling of the STBL during the daytime, and this is why Fig. 5.9 indicates daytime conditions. We discuss this partial coupling in the following chapter.

Figure 5.8f shows the vertical profiles of $\theta_l$. We can see that the addition of drizzle increases the median value of $\theta_l$ in the STBL. The net effect of drizzle is to warm the STBL through a slight reduction in integrated longwave cooling and net formation of drizzle that does not evaporate before it reaches the surface (Fig. 5.7b). This increase is more significant for $N_d = 50 \text{ cm}^{-3}$ where drizzle is stronger. This increase in $\theta_l$ and associated weaker inversion do not correspond with faster entrainment, however. Entrainment instead slows when drizzle is added. This suggests processes might be warming the inversion layer and increasing the strength of the inversion. This warming in the inversion layer might be associated with penetrating thermals, in which net formation of drizzle and less radiative cooling (because of reduced LWC) occurs.
5.4.3 “Potential Buoyancy”

Stevens et al. (1998a) theorized that the inclusion of drizzle into LES could result in reduction in buoyancy fluxes at cloudbase and below through a “potential buoyancy” mechanism. This effect is a consequence of the uneven production of drizzle between updrafts and downdrafts. Drizzle formed in updrafts results in a reduction of liquid water in downdrafts, meaning that the lifting condensation level for updrafts is lower than it is for a paired downdraft. Because the lifting condensation level is higher in downdrafts when drizzle is present, circulations completing a circuit require more mechanical energy. This reduction in buoyant energy, though originating when drizzle forms near the top of a cloud, is not realized until cloudbase, and this is why it is known as a “potential buoyancy” mechanism.

Stevens et al. (1998a) provided evidence for this effect through the partitioning of buoyancy fluxes into updrafts and downdrafts. They found that most of the reduction in buoyancy flux was found in downdrafts, consistent with the theory they postulated. This reduction in buoyancy flux within downdrafts has been found in other LES studies (Ackerman et al., 2009). We examine buoyancy fluxes conditionally sampled over updrafts and downdrafts in Fig. 5.10. Updrafts and downdrafts are defined as gridcells where the vertical velocity is greater than 0 m s\(^{-1}\) or less than \(-0.1\) m s\(^{-1}\) respectively.

Figures 5.10a and b show conditionally sampled buoyancy flux for \(N_d = 50\) cm\(^{-3}\) for both drizzling and non-drizzling simulations. For these simulations we find that the drizzle process reduces buoyancy flux in the sub-cloud layer for both updrafts and downdrafts but the reduction in buoyancy flux for downdrafts is more substantial. We also note reductions in buoyancy flux owing to drizzle for \(N_d = 1000\) cm\(^{-3}\) in Fig. 5.10c and d. The drizzle process is not as strong for this droplet concentration so the reductions are not as large as for \(N_d = 50\) cm\(^{-3}\), but again we find larger reductions in downdrafts as compared to updrafts. For both droplet concentrations this potential buoyancy effect may also play a role in the reduction in buoyancy flux and consequent reduction in circulation strength.

From our plots of LWP (Fig. 5.7a), buoyancy production of TKE (Fig. 5.8b) and \(\overline{w'w'}\) (Fig. 5.8d), we can see that the addition of drizzle to our LES mitigates the impact that interactions of longwave radiative heating and droplet concentration have on STBL structure. For this OVERCAST case, accounting for drizzle leads
Figure 5.10. Large-eddy simulation output for the OVERCAST sounding with and without drizzle. Buoyancy fluxes are conditionally sampled over updrafts and downdrafts, defined as gridcells where the vertical velocity is greater than 0.1 m s\(^{-1}\) or less than \(-0.1\) m s\(^{-1}\) respectively. Horizontally averaged vertical profiles. Data were temporally averaged over five minutes, and then were normalized by the height of the boundary layer top, \(z_i\). These data were then interpolated to a uniform vertical grid. These uniformly gridded data were then temporally averaged over the last two hours of simulation time.

to smaller differences in LWP and STBL dynamical structure between the two droplet concentrations. The drizzle process effectively mitigates the differences in the two simulations associated with differences in longwave radiative heating.
5.5 Modeling STBL Longwave Radiative Transfer

Whether or not interactions of longwave radiative heating and droplet concentration need to be accounted for in LES is dependent on the purpose of the simulations. If estimates of first-order statistics, such as LWP and integrated longwave cooling are desired, our simulations suggest that, for buoyancy-driven nocturnal STBLs where cloud fraction is unity and \( \text{LWP} > 20\, \text{g m}^{-2} \), accounting for changes in longwave radiative heating with droplet concentration does not appear necessary. For these STBLs, a reasonably accurate description of STBL first-order dynamical structure can also be obtained without accounting for these interactions. If more detail or accuracy is desired, however, accounting for interactions of longwave radiative heating and droplet concentration may be necessary to account for as they alter processes in the STBL entrainment zone.

5.6 Results: BROKEN case, non-drizzling

As with the OVERCAST case, we ran four LESs with the BROKEN sounding, with and without drizzle and setting \( N_d \) to fixed values of 50 and 1000 cm\(^{-3}\). Preliminary LESs, not shown here, suggested to us that this BROKEN case was rather sensitive to changes in droplet concentration. We were concerned that this sensitivity could be related to model spin-up; that is, the first hour of simulation time in which cloud scale motions develop. To avoid this possibility, we ran the first hour of time with \( N_d = 200\, \text{cm}^{-3} \) and then began the two simulations with different droplet concentrations from the end of that simulation’s first hour. Using this procedure ensures that the one hour of spin-up time in our LESs is not unduly influencing our model STBL evolution. We begin by describing the non-drizzling simulations.

5.6.1 Time Series

Figure 5.1 shows time series output for the two non-drizzling LESs using the BROKEN sounding. Time series of domain-averaged LWP (Fig. 5.11a) show that for
Figure 5.11. Large-eddy simulation output for the BROKEN sounding without drizzle. Time series of quantities are domain averaged and vertically integrated.
these simulations the cloud layer is substantially thinner than those modeled in the OVERCAST case. Domain-averaged LWP continues to increase with time for $N_d = 50 \text{ cm}^{-3}$ but decreases slowly for $N_d = 1000 \text{ cm}^{-3}$. Liquid water path averaged over the last two hours of simulation time is 37.4 g m$^{-2}$ for $N_d = 50 \text{ cm}^{-3}$ and 14.3 g m$^{-2}$ for $N_d = 1000 \text{ cm}^{-3}$. Substantial reductions in LWP as compared to the OVERCAST case are expected as the air entrained from above is drier in this case (see Fig. 3.1) and leads to more evaporation. For these two droplet concentrations there is also a larger disparity between LWP values averaged over the last two hours of the simulations as compared to the OVERCAST non-drizzling case. The ratio of LWP($N_d = 50 \text{ cm}^{-3}$)/LWP($N_d = 1000 \text{ cm}^{-3}$) over this time period is 2.61 for the BROKEN case, compared to 1.07 for the OVERCAST case.

Another result of dry air entrainment is that the cloud fraction is not unity throughout the BROKEN simulations and the use of different droplet concentrations results in a significant difference in cloud fraction values (Fig. 5.11b). Recall that we define cloudy gridcells as those with a column-integrated LWP greater than 10 g m$^{-2}$. Averaged over the last two hours of simulation time, cloud fraction is 0.96 for $N_d = 50 \text{ cm}^{-3}$ and 0.59 for $N_d = 1000 \text{ cm}^{-3}$. We also see a periodic increase and decrease in cloud fraction with time. Patterns with similar periodicity exist in the other time series as well (Fig. 5.11c, d and e) and we will explain the relationship between these patterns below.

Figure 5.11c shows the time evolution of the height of the boundary layer $z_i$ which we again use to infer relative rates of entrainment. This figure shows us that entrainment rates are more rapid for this case as compared to the OVERCAST case. The entrainment rate for $N_d = 1000 \text{ cm}^{-3}$ is faster than for $N_d = 50 \text{ cm}^{-3}$ at first but becomes slower as the simulations progress.

Figure 5.11d displays cloud integrated longwave cooling for these simulations and we see significant differences for the two droplet concentrations. The integrated coolings averaged over the last two hours of simulation time are $-69.3 \text{ W m}^{-2}$ and $-53.3 \text{ W m}^{-2}$ for $N_d = 50$ and 1000 cm$^{-3}$ respectively. Average LWP for these two droplet concentrations lie on either side of 20 g m$^{-2}$ and integrated longwave cooling decreases rapidly when LWP is low (Fig. 2.3). This disparity in average LWP partially explains the large difference in integrated longwave cooling but the difference in cloud fractions also plays an important role. The amount of longwave
radiation emitted from a clear grid-cell where LWP < 10 g m\(^{-2}\) is considerably less than that from a cloudy grid-cell. If cloud fractions decrease and there are more breaks in the clouds we expect the average integrated longwave cooling from the cloud layer to be considerably smaller.

Because STBL TKE in our non-drizzling nocturnal simulations is generated primarily by radiative cooling, we would expect that the differences in integrated longwave cooling we see in Figure 5.11d to be associated with differences in STBL-averaged TKE. Figure 5.11e shows that there is indeed a wide disparity in the amount of TKE for the two droplet concentrations. Averaged over the last two hours of simulation time, STBL-averaged TKE is more than twice as large for \(N_d = 1000 \text{ cm}^{-3}\) as compared to \(N_d = 50 \text{ cm}^{-3}\).

The oscillations of cloud fraction, integrated longwave cooling and STBL-averaged TKE with time are related to the time evolution of the boundary layer top \(z_i\). Earlier in this chapter we described grid-hopping and that the modeled change of boundary layer height with time is an artifact of the discretization of the atmosphere in LES. Properties of our model STBL are dependent on when this grid-hopping is occurring. Through examination of our time series we can see that during grid-hopping periods, cloud fraction increases, integrated longwave cooling decreases and STBL-averaged TKE decreases. When the cloud layer is ascending, moist air mixed up into the dry layer leads to net condensation in the layer above. At that time model gridcells in that layer become cloudy, breaks in the cloud fill in and cloud fraction increases. As cloud fraction increases and there are fewer model columns with low LWP, the average cloud layer emissivity increase and leads to increased longwave radiative cooling. When the cloud layer is not ascending, warm dry air mixed into the cloud layer can lead to thinned cloud areas, decreasing the cloud fraction. As the cloud fraction drops, average cloud layer emissivity decreases and integrated longwave cooling lessens.

We may expect that increased radiative cooling during grid-hopping periods should lead to increased STBL-averaged TKE but this not the case for our simulations. A larger portion of STBL TKE drives entrainment at resolved scales during these periods so STBL TKE actually decreases during grid-hopping. After the cloud layer stabilizes at a higher model level STBL TKE slowly increases until the next grid-hopping event.
5.6.2 Vertical Profiles

Figure 5.12 displays vertical profiles for LESs of the BROKEN non-drizzling case, normalized and averaged as for the OVERCAST case. Figure 5.12a shows the domain-averaged vertical distribution of LWC. As with the time evolution of LWP displayed in Fig. 5.11a, LWC is substantially smaller than for the OVERCAST case, and there are substantial differences in LWC for the two modeled droplet concentrations. For $N_d = 50 \text{ cm}^{-3}$ the maximum LWC is $0.25 \text{ g kg}^{-1}$ and for $N_d = 1000 \text{ cm}^{-3}$ it is $0.11 \text{ g kg}^{-1}$. Liquid water content maximizes further below cloudtop than for the OVERCAST case and indicates that entrainment of dry air is reducing LWC near the cloudtop. Because cloud fractions are not unity for the BROKEN case, the horizontal variation of LWC is expected to be important. This horizontal variation of LWC is related to vertical air motions and will be discussed further below.

In Fig. 5.12b we show vertical profiles of buoyancy and shear production of TKE. As with our OVERCAST case, shear production of TKE is negligible except in the surface layer, hence circulations are buoyantly driven. For both droplet concentrations there is buoyancy destruction in the sub-cloud layer, indicating some amount of decoupling (Stevens, 2000). For $N_d = 1000 \text{ cm}^{-3}$ there is evidence of stronger decoupling because there is less buoyancy generation in the cloud layer and more buoyancy destruction below cloud for that value of $N_d$. We can also see that buoyancy destruction in the entrainment zone is noticeably different for the two droplet concentrations; for $N_d = 50 \text{ cm}^{-3}$ it is larger. Buoyancy generation of TKE in the cloud layer is also significantly higher for $N_d = 50 \text{ cm}^{-3}$.

With these vertical profiles of buoyancy generation of TKE we would expect circulations for $N_d = 1000 \text{ cm}^{-3}$ to be weaker and more confined to the cloud layer, and Fig. 5.12c showing $\bar{w}'w'$ indicates this is indeed correct. The maximum value of $\bar{w}'w'$ for $N_d = 1000 \text{ cm}^{-3}$ is less than half that of $N_d = 50 \text{ cm}^{-3}$ and is shifted more towards the STBL top. Vertical velocity variance also is close to zero below $0.25 \bar{z}_i$ for $N_d = 1000 \text{ cm}^{-3}$ but is larger above the surface for $N_d = 50 \text{ cm}^{-3}$. These profiles show that circulations are weaker and that there is a larger amount of decoupling for $N_d = 1000 \text{ cm}^{-3}$ as compared to $N_d = 50 \text{ cm}^{-3}$.

There are two notable differences in vertical profiles of $\theta_l$ (Fig. 5.12d) between the two droplet concentrations. First, $\theta_l$ changes with height in the STBL for
Figure 5.12. Large-eddy simulation output for the BROKEN sounding without drizzle. Horizontally averaged vertical profiles. Data were temporally averaged over five minutes, and then were normalized by the height of the boundary layer top, $z_i$. These data were then interpolated to a uniform vertical grid. These uniformly gridded data were then temporally averaged over the last two hours of simulation time.
\(N_d = 1000 \text{ cm}^{-3}\) indicating that the STBL is not as well-mixed as in the corresponding OVERCAST simulation. Second, the inversion at STBL top is stronger for \(N_d = 50 \text{ cm}^{-3}\). For the OVERCAST case we correlated relative differences in entrainment rate with relative differences in inversion strength but this correlation may not be appropriate for the BROKEN case. First, based on the profiles of \(\overline{w'w'}\) and \(\theta_i\), the \(N_d = 1000 \text{ cm}^{-3}\) simulation does not appear well-mixed and mixed-layer theory will be less appropriate in this case. Second, unlike the OVERCAST case, there are differences in buoyancy fluxes in the entrainment zones for these two droplet concentrations and these relative differences could also be correlated with relative differences in entrainment.

Horizontally averaged profiles of longwave radiative heating (Fig. 5.12e) show that there is more cooling at STBL top and more warming at cloudbase for \(N_d = 50 \text{ cm}^{-3}\). As before, differences in profiles of modeled longwave radiative heating might be explained by a) differences in the model cloud structure and LWC owing to dynamical responses or b) changes in longwave radiative transfer with droplet concentration, or some combination of the two. In Chapter 2 we detailed how cloud longwave emissivity increases with droplet concentration, and does so rapidly for thinner clouds. Figure 5.13 shows this effect.

More average LWC for the lower droplet concentration results in a larger emissivity for the cloud layer as compared to the higher droplet concentration. The disparity in LWC offsets the increases in emissivity with droplet concentration we expect when LWC is held constant (Fig. 5.13). This figure shows computed longwave radiative heating profiles for a static model atmosphere using the two droplet concentrations in our LESs, similar to Fig. 5.3. Here the cloud is 100 m in geometric depth, giving a LWP of 10.6 g m\(^{-2}\). This choice gives us thin clouds for which there are vivid differences in longwave radiative heating related to droplet concentration. Between these two droplet concentrations, the peak longwave cooling rate at cloudtop is \(2.4 \text{ K hr}^{-1}\) higher for \(N_d = 1000 \text{ cm}^{-3}\). For this cloud there is more longwave warming at cloudbase (\(0.2 \text{ K hr}^{-1}\)) for \(N_d = 1000 \text{ cm}^{-3}\) and virtually none for \(N_d = 50 \text{ cm}^{-3}\).

Both cloudtop cooling and cloudbase warming are higher for \(N_d = 50 \text{ cm}^{-3}\) in Fig. 5.12e, and this indicates that differences in LWC are driving these differences in longwave radiative heating. The increase in emissivity with droplet concentration
Figure 5.13. In-cloud longwave radiative heating rates for $N_d = 50$ and $1000\text{ cm}^{-3}$ for the static model atmosphere described in Chapter 2. Cloud added to MLS profile extends from 1.0 km at base to 1.1 km at top. Liquid water path is $10.6\text{ g m}^{-2}$.

we find when LWC is unchanged (Fig. 5.13) is more than offset by the larger LWC for $N_d = 50\text{ cm}^{-3}$ (Fig. 5.12a).

5.7 Discussion: BROKEN case, non-drizzling

Even more so than for the OVERCAST case, the use of different model droplet concentrations results in different STBL structure and dynamics for the BROKEN case. To what physical mechanisms do we attribute the divergence in the two LESs? As with the OVERCAST case, longwave radiative heating and entrainment are important processes. A significant difference between these simulations and those for the OVERCAST case is that cloud fractions are not unity and hence horizontal inhomogeneity may be important.

Entrainment of warm, dry air from above the STBL typically takes place where updrafts and downdrafts work together to distort the STBL interface (Sullivan et al., 1999), and we might expect this process to lead to lower values of LWC in downdrafts. From his observations, Nicholls (1989) found that STBL downdrafts are on average drier than updrafts, and attributed this difference to downdrafts containing entrained air. Gerber et al. (2005) corroborated this correlation of lower values of LWC and downdrafts. It is possible to have a higher than average
LWC in any individual downdraft, however.

Figure 5.14. Large-eddy simulation output for the BROKEN sounding without drizzle. Data are conditionally sampled over updrafts and downdrafts, defined as gridcells where the vertical velocity is greater than 0.1 m s\(^{-1}\) or less than −0.1 m s\(^{-1}\) respectively. Horizontally averaged vertical profiles. Data were temporally averaged over five minutes, and then were normalized by the height of the boundary layer top, \(z_i\). These data were then interpolated to a uniform vertical grid. These uniformly gridded data were then temporally averaged over the last two hours of simulation time.

Because we believe entrainment to play an important role in thinning cloud layers in the BROKEN case as compared to the OVERCAST case, and because cloud fractions are not unity, we conditionally sample LWC and longwave radiative heating over updrafts and downdrafts (as in Fig. 5.10). If there are notable differences in these properties between updrafts and downdrafts we can associate these differences with entrainment drying.
Figure 5.14 shows the conditionally sampled vertical profiles of LWC and long-wave radiative heating. In Fig. 5.14a and b we see that for both droplet concentrations LWC is significantly higher in updrafts as compared to downdrafts. For $N_d = 50 \text{ cm}^{-3}$ the maximum LWC is $0.26 \text{ g kg}^{-1}$ in updrafts and $0.16 \text{ g kg}^{-1}$ in downdrafts, compared to $0.16 \text{ g kg}^{-1}$ for updrafts and $0.06 \text{ g kg}^{-1}$ for downdrafts for $N_d = 1000 \text{ cm}^{-3}$. Note that LWC is clearly subadiabatic near cloudtop; LWC does not increase linearly with height up to the top of the STBL. On average entrainment appears to be significant in drying out the cloud in downdrafts. Total liquid water path in downdrafts is reduced by 34% from its value in updrafts for $N_d = 50 \text{ cm}^{-3}$ and is reduced by 67% for $N_d = 1000 \text{ cm}^{-3}$.

In Fig. 5.14c and d we see that these differences in LWC between updrafts and downdrafts lead to differences in longwave radiative heating. Within updrafts the longwave radiative heating profiles are similar between the droplet concentrations. Liquid water content is higher for $N_d = 50 \text{ cm}^{-3}$ but cloud emissivity is sensitive to droplet concentration for these thin cloud layers. For $N_d = 1000 \text{ cm}^{-3}$ the cloud layer is thinner but the increase in emissivity with droplet concentration offsets the decrease in emissivity associated with a lower LWC. The maximum longwave cooling rate is lower in downdrafts as compared to updrafts for both droplet concentrations and is reduced more for $N_d = 1000 \text{ cm}^{-3}$. In downdrafts the increase in emissivity with droplet concentration does not offset the decrease in emissivity with LWC and the longwave cooling rate for $N_d = 50 \text{ cm}^{-3}$ is higher than for $N_d = 1000 \text{ cm}^{-3}$.

From these profiles we see that entrainment is associated with a reduction in LWC and longwave radiative cooling for both droplet concentrations in downdrafts as compared to updrafts. This reduction in radiative cooling is especially large for $N_d = 1000 \text{ cm}^{-3}$ where cloud water is almost nonexistent in downdrafts. Longwave radiative cooling is the primary driver of circulations in these simulations and the substantial decrease in this cooling with increased droplet concentration may help to explain changes in STBL structure and dynamics.

Over the last two hours of simulation time the domain-averaged LWP decreases slightly with time (Fig. 5.11a) for $N_d = 1000 \text{ cm}^{-3}$. Entrainment continuously works to evaporate cloud for both droplet concentrations but is more rapid for $N_d = 1000 \text{ cm}^{-3}$. For this simulation, entrainment drying leads to very thin
cloud in downdrafts for this droplet concentration, substantially reducing longwave radiative cooling there. This reduction in longwave cooling is associated with reductions in circulation strength such that the STBL is not well-mixed down to the surface (Fig. 5.12c). Consequently there is less water vapor flux from the sub-cloud layer to resupply the cloud layer with water. Moreover, less radiative cooling in the STBL also leads to less net cloud droplet formation, and subsequently the cloud layer is not maintained against entrainment drying from above for $N_d = 1000 \text{ cm}^{-3}$.

In contrast, for $N_d = 50 \text{ cm}^{-3}$ the domain-averaged LWP increases with time over the last two hours of simulation (Fig. 5.11a). Entrainment drying for $N_d = 50 \text{ cm}^{-3}$ also leads to lower LWC in downdrafts and reductions in longwave radiative cooling. For the lower droplet concentration the drying process is not as rapid, however, and the consequent reduction in cooling does not lead to as much decoupling as it does for $N_d = 1000 \text{ cm}^{-3}$. The magnitude of radiative cooling is large enough to both mix the STBL to the surface and cool the STBL, aiding in the formation of cloud. The cloud layer is subsequently able to maintain itself against entrainment drying.

The results above show that the BROKEN case is sensitive to the interplay between radiative heating and entrainment, and changes in model droplet concentration cause the simulated STBL to evolve along different pathways. The decrease in cloud fraction and longwave cooling for an increased droplet concentration appears to be important in explaining the bifurcation in the two LESs. From the time series data (Fig. 5.11), this bifurcation in LWP, cloud fraction and integrated longwave radiative cooling is readily apparent by the second hour of simulation. Therefore the interplay of radiative heating and entrainment near the beginning of the simulation may be important in driving the simulations apart and we investigate this possibility next.

Holding other factors constant, mixing of uniform portions of cool, moist air and warm, dry air leads to more rapid net evaporation when droplet concentration increases. Entrainment drying of the cloud layer near the beginning of the simulation could be more rapid for $N_d = 1000 \text{ cm}^{-3}$ and, in large part, lead to differences in the modeled STBL. For $N_d = 1000 \text{ cm}^{-3}$ the entrainment rate for the second hour of simulation time is faster than for $N_d = 50 \text{ cm}^{-3}$ (Fig. 5.11c). We also
know that longwave radiative heating is dependent on droplet concentration, and especially so for low LWP clouds (LWP < 20 g m\(^{-2}\)) such as those modeled in the BROKEN case (Fig. 5.13). Differences in longwave radiative cooling near the beginning of the simulation could be leading to differences in circulation strength and entrainment that are important to driving the observed divergence in the simulations.

As we said in the previous section regarding the OVERCAST case, radiative cooling and entrainment are linked together and isolating one from the other is exceedingly difficult. Following what we did for the the OVERCAST case, we can compute longwave radiative heating with the same droplet concentration in both LESs. We used \(N_d = 200 \text{ cm}^{-3}\) for spin-up for both simulations. We now run two additional LESs for this case where \(N_d = 50 \text{ cm}^{-3}\) and \(N_d = 1000 \text{ cm}^{-3}\) but use 200 cm\(^{-3}\) as the value of \(N_d\) for computation of radiative heating within the cloud. These additional LESs do not strictly isolate feedbacks between radiative heating and entrainment in our model STBL. However, they give us some measure of how important differences in radiative heating owing to different droplet concentrations are in producing the bifurcation in our LESs. We will refer to these LESs as \(N_d = 50 \text{ cm}^{-3}\) with 200 cm\(^{-3}\) for radiation and \(N_d = 1000 \text{ cm}^{-3}\) with 200 cm\(^{-3}\) for radiation.

Figure 5.15 compares results from the LES where \(N_d = 50 \text{ cm}^{-3}\) with 200 cm\(^{-3}\) for radiation to that with \(N_d = 1000 \text{ cm}^{-3}\) with 200 cm\(^{-3}\) for radiation. Figures 5.15a through d shows time series of domain-averaged LWP, cloud fraction, integrated longwave radiative cooling and STBL-averaged TKE. For all four integrated quantities the time evolution of the two simulations are in better agreement as compared to those in Fig. 5.11. Vertical profiles of \(w'w'\) and buoyancy production of TKE averaged over the last two hours of simulation time are displayed in Fig. 5.15d and e respectively. As with the time series data, these profiles are more alike than those in Fig. 5.12.

These simulations show that changes in radiative heating with \(N_d\) are important in explaining the divergence in the BROKEN simulations. However, there are notable differences in the two simulations depicted in Fig. 5.15, indicating that differences in radiative heating are not the only reason for this observed divergence. Differences in entrainment with droplet concentration must also be important.
Figure 5.15. Large-eddy simulation output for the BROKEN sounding without drizzle. \( N_d = 50 \text{ cm}^{-3} \) with 200 cm\(^{-3} \) for radiation and \( N_d = 1000 \text{ cm}^{-3} \) with 200 cm\(^{-3} \) for radiation and. a.) to c.) Time series of quantities, domain averaged and vertically integrated. d.) to f.) Horizontally averaged vertical profiles. Data were temporally averaged over five minutes, and then were normalized by the height of the boundary layer top, \( z_i \). These data were then interpolated to a uniform vertical grid. These uniformly gridded data were then temporally averaged over the last two hours of simulation time.
Because the time series data shown in Fig. 5.11 diverge early in the two simulations, entrainment and longwave radiative heating appear to initially work together in different ways depending on modeled droplet concentration. Together the two processes lead to differing values of cloud fraction and LWP in the second hour of the simulation. From Fig. 5.11c we know the entrainment rate was more rapid for $N_d = 1000 \text{ cm}^{-3}$ during this time period. Also, at the beginning of the second hour, integrated longwave radiative cooling is also greater for $N_d = 1000 \text{ cm}^{-3}$.

To determine if differences in longwave radiative heating between the two droplet concentrations might be important, we examine longwave radiative heating and $\theta_l$ profiles averaged over the second hour of simulation time in Fig. 5.16.

![Longwave heating rate](a)

![Liquid water potential temperature](b)

**Figure 5.16.** Large-eddy simulation output for the BROKEN sounding without drizzle. Horizontally averaged vertical profiles. Data were temporally averaged over five minutes, and then were normalized by the height of the boundary layer top, $z_i$. These data were then interpolated to a uniform vertical grid. These uniformly gridded data were then temporally averaged over the last two hours of simulation time.

For the OVERCAST case we suggested that an increase in radiative cooling in penetrating thermals within the inversion layer could be associated with an increased entrainment rate by cooling the inversion layer and promoting mixing across the STBL interface. Figure 5.16a shows that there is more radiative cooling at the STBL interface for $N_d = 1000 \text{ cm}^{-3}$ throughout the first hour of simulation. Additionally we see that the $\theta_l$ inversion is just a bit weaker for $N_d = 1000 \text{ cm}^{-3}$ for the same time period.

Together Fig. 5.16 and Fig. 5.11 suggest the following hypothesis for the diver-
gence of these two simulations for the BROKEN case. When the simulations begin with different droplet concentrations, radiative cooling and entrainment are larger for \( N_d = 1000 \text{ cm}^{-3} \) as compared to \( N_d = 50 \text{ cm}^{-3} \), and radiative cooling may be aiding in this larger entrainment rate. This larger entrainment rate allows for more efficient drying, causing a larger initial decrease in cloud fraction and a concurrent initial decrease in integrated radiative cooling for \( N_d = 1000 \text{ cm}^{-3} \). Subsequently circulations weaken for \( N_d = 1000 \text{ cm}^{-3} \) faster than for \( N_d = 50 \text{ cm}^{-3} \) and further entrainment drying suppresses cloud growth for the rest of the simulation. In contrast the initial decreases in cloud fraction and integrated radiative cooling when \( N_d = 50 \text{ cm}^{-3} \) are not enough to suppress cloud growth.

Horizontal inhomogeneities in cloud liquid water for this case suggest that horizontal transport of longwave radiation could be important, and our radiative heating computations do not account for this. Mechem et al. (2008) investigated the importance of horizontal transport of longwave radiation in boundary layer cloud systems. Using an eddy-resolving model coupled together with the Spherical Harmonics Discrete Ordinate Method (SHDOM) radiative transfer model (Evans, 1998), they modeled two cases with and without horizontal transport of longwave radiation. The two cases they modeled were based on observations of a solid stratocumulus cloud deck and a broken trade cumulus field. For these two cases they found that using the Independent Column Approximation was sufficient to accurate model cloudy boundary layer evolution, but Mechem et al. (2008) note that their results do not necessarily apply to all boundary layer cloud fields. Our BROKEN case is dissimilar from those Mechem et al. (2008) modeled because our LWP values are less than half of those Mechem et al. (2008) modeled for their solid stratocumulus deck. Our cloud fractions are significantly higher than those for the broken trade cumulus case they modeled. Investigation of the importance of horizontal transport of longwave radiation for our BROKEN case may prove worthwhile but is beyond the scope of this work.

5.8 Results: BROKEN case, drizzling

For the OVERCAST case we found notable differences in nocturnal STBL evolution and dynamics that are related to changes in longwave radiative heating with
droplet concentration, but also found that the drizzle process mitigates these differences. To examine the effect of drizzle on our simulations of the BROKEN case we ran two LESs with the same values of \( N_d \) as above, \( N_d = 50 \) and \( 1000 \text{ cm}^{-3} \), with drizzle. From our BROKEN drizzling ERM simulations detailed in Chapter 4 we expect drizzle not to play as significant a role in STBL evolution and dynamics as compared to the OVERCAST case.

Since drizzle depends strongly on \( N_d \), we ran an additional LES with an intermediate value of \( N_d = 200 \text{ cm}^{-3} \). As for our BROKEN non-drizzling simulations, we first ran the LES with \( N_d = 200 \text{ cm}^{-3} \). The LESs with \( N_d = 50 \) and \( 1000 \text{ cm}^{-3} \) were started from one hour into this simulation so that model spin-up did not unduly influence STBL evolution. With these three simulations, we simultaneously examine the importance of drizzle and the applicability of our results to intermediate values of \( N_d \).

### 5.8.1 Time Series

Time series data for these three LESs clearly demonstrate that the dependence of the model STBL on droplet concentration we found for our BROKEN non-drizzling simulations is still readily evident when drizzle is present (Fig. 5.17). Additionally, the behavior for \( N_d = 200 \text{ cm}^{-3} \) lies between the simulation results for the other two droplet concentrations, and suggests that our results are applicable to intermediate droplet concentrations.

To better understand the effect of drizzle on the BROKEN case we first examine time series of integrated drizzle flux divergence in Fig. 5.17c. Drizzle flux divergence is small for \( N_d = 200 \text{ cm}^{-3} \) and \( N_d = 1000 \text{ cm}^{-3} \); averaged over the last two hours of simulation, the drizzle flux divergence is 1.1 and 0.8 W m\(^{-2}\) respectively. These small values indicate that the drizzle process is not significant for these droplet concentrations and echoes what we found in our ERM simulations (Fig. 4.5a). For \( N_d = 50 \text{ cm}^{-3} \) the integrated drizzle flux divergence averaged over the last two hours is 6.3 W m\(^{-2}\). These values suggest that drizzle is more important for this droplet concentration but the effects are still small compared to the amount of integrated radiative cooling.

Through comparison of output from our non-drizzling simulations of the BRO-
Figure 5.17. Large-eddy simulation output for BROKEN sounding with drizzle. 3 LES: $N_d = 50 \text{ cm}^{-3}$, $N_d = 200 \text{ cm}^{-3}$ and $N_d = 1000 \text{ cm}^{-3}$. Time series of quantities are domain averaged and vertically integrated.
KEN case (Fig. 5.11) to drizzling simulations (Fig. 5.17) we find that drizzle has the same general effect on the STBL as it does in the OVERCAST case for $N_d = 50 \text{ cm}^{-3}$. Liquid water path is reduced (Fig. 5.17a) and entrainment rate is slowed (Fig. 5.17e). Because the cloud layer is thinner in this case this reduction in LWP is associated with a decrease in cloud fraction (Fig. 5.17b and consequent reduction in integrated longwave radiative cooling (Fig. 5.17e). We did find this reduction in integrated longwave cooling in the OVERCAST case when comparing non-drizzling simulations to drizzling simulations. STBL-averaged TKE (Fig. 5.17f) is also reduced and is likely related to the reduction in integrated longwave radiative cooling. For $N_d = 1000 \text{ cm}^{-3}$ drizzle has a minimal impact on time series of integrated quantities.

5.9 Modeling STBL Longwave Radiative Transfer - Low LWP

Although the drizzle process alters STBL evolution and dynamics in notable ways for the BROKEN case, the addition of drizzle does not diminish the divergence in simulations owing to droplet concentration as it did in the OVERCAST case. Turner et al. (2007) found that low level liquid water clouds such as those modeled here are frequently found in satellite and ground-based data. They also noted that cloud radiative forcing is sensitive to changes in LWP when LWP is $< 50 \text{ g m}^{-2}$, hence these types of clouds play an important role in the Earth’s radiative budget.

For low LWP clouds we find significant inhomogeneities in the horizontal distribution of cloud liquid water in our LESs and cloud breaks can exist. The cloud fraction is dependent on entrainment and radiative heating, and those processes are dependent on droplet concentration. In Chapter 2 we found that integrated longwave radiative cooling is sensitive to changes in droplet concentration when LWP low ($< 20 \text{ g m}^{-2}$). Integrated radiative cooling is also sensitive to changes in cloud fraction, and because it is the driver for STBL dynamics, changes in integrated radiative cooling can lead to changes in first-order STBL evolution and dynamics. Our LESs suggest that accounting for changes in longwave radiative heating with droplet concentration is important for simulating low level liquid water clouds.
CHAPTER 6

Large-eddy Simulations - Shortwave Radiative Impact

6.1 Purpose

In the last chapter we examined nocturnal LESs of our OVERCAST and BROKEN cases in detail. Now that we have an understanding of these nocturnal simulations we turn to LESs of the stratocumulus-topped boundary layer (STBL) when shortwave radiative heating is present. As we stated in our Introduction, there have been several modeling studies of the daytime STBL, but none to our knowledge have investigated the response and feedbacks of STBL dynamics to shortwave heating through LES as solar zenith angle (\(\Theta\)) varies. From these previous studies we expect that the addition of shortwave forcing to our LESs could further decouple the cloud layer from the sub-cloud layer (Nicholls, 1984; Turton and Nicholls, 1987) causing vertical velocity variance \(\overline{w'w'}\) and STBL-averaged turbulent kinetic energy to weaken (Lu and Seinfeld, 2005; Caldwell and Bretherton, 2009). Liquid water path could consequently decrease because of entrainment drying (Lu and
Seinfeld, 2005), but the entrainment process is also likely modified by the addition of shortwave warming. Decrease in LWP and decoupling of the cloud layer from the sub-cloud may also lead to a decrease in drizzle rate (e.g. Sandu et al., 2007) but droplet in-cloud residence time is also likely to play a role (e.g. Hartman and Harrington, 2005b).

Our ERM simulations in Chapter 4 bear out some of these responses to shortwave forcing: For our OVERCAST case we found in our ERM simulations that, as $\Theta$ decreases, LWP and STBL-averaged TKE decrease (Fig. 4.1 and Fig. 4.3). Entrainment and drizzle rates are also reduced as solar zenith angle decreases (Fig. 4.2).

Through ERM simulations of our BROKEN case we found the same relationships of LWP, STBL-averaged TKE and entrainment rate to shortwave heating as found for the OVERCAST case, but with greater variation (Fig. 4.4). Cloud fraction also decreases with decreasing $\Theta$. Because we found that variations in cloud fraction were associated with variations in LWP, STBL-averaged TKE and integrated longwave cooling in our nocturnal LES simulations (Fig. 5.11), we anticipate that changes in cloud fraction with shortwave heating may be important for our daytime LESs. Possibly due to the shortcomings of ERM simulations, we found a peculiar variation of drizzle rate with solar zenith angle for the BROKEN case (Fig. 4.5). Therefore, our ERM simulations do not appear to be suitable for guiding expectations in similarly configured LESs.

Following our ERM simulations and the modeling study of Hartman and Harrington (2005b), we will first examine daytime LESs where $\Theta$ is fixed. With this procedure modeled STBL dynamics are not altered by diurnal variations in shortwave forcing and we can determine the STBL response to varying amounts of this forcing. As evidence for the STBL response we examine time series and vertical profiles of atmospheric properties as in the previous chapter.

Because the sun is not fixed in the sky, we also simulate the diurnal cycle of the STBL for both cases to better understand how varying incoming shortwave radiation impacts STBL dynamics. There have been several studies of the daytime STBL and its diurnal cycle using LES (Lu and Seinfeld, 2005; Sandu et al., 2007; Caldwell and Bretherton, 2009), but in contrast to those studies we focus on how shortwave radiative heating alters STBL dynamics. Because we simulate both
cases with high and low droplet concentrations we can also comment on the aerosol indirect effect on the STBL diurnal cycle. Sandu et al. (2007) have studied the impact of modeled droplet concentration on the diurnal cycle of stratocumulus in detail, and we will also examine the extent to which our simulations corroborate their findings.

6.2 Results: OVERCAST case, drizzling, Fixed Solar Zenith Angles, low $N_d$

We conducted six hour LESs using both the OVERCAST and BROKEN stratocumulus soundings. Following our nocturnal LESs we fixed values of $N_d$ at 50 and 1000 cm$^{-3}$. After one hour of simulation spin-up time where shortwave radiation was not modeled, we applied shortwave forcings in each simulation by fixing $\Theta$ at 0°, 30°, 45° and 60°. Using these values of $\Theta$ covers a significant range of shortwave heating (see Chapter 2). We first investigate the OVERCAST case with drizzle allowed, eschewing discussion of non-drizzling simulations. The variation of time series output and domain-averaged vertical profiles with fixed $\Theta$ does change in a quantitative fashion when drizzle is disabled, but the qualitative relationships are the same. Solar radiative-drizzle feedbacks are found to be well behaved for this case. Though the BROKEN case exhibits more sensitivity than the OVERCAST case, drizzle is not a vigorous process and solar-radiative feedbacks are also found to be well behaved.

To demonstrate the effect of shortwave forcing on the STBL we examine results for five different values of $\Theta$ (the four above plus the nocturnal simulation from Chapter 4) and for one fixed droplet concentration at a time. In Fig. 6.1 and Fig. 6.2 we show time series output and vertical profiles, as in the previous chapter, for $N_d = 50$ cm$^{-3}$. We examine the time series output first.

6.2.1 Time Series

Visual inspection of both figures reveals that, as $\Theta$ decreases and incoming shortwave radiation increases, LESs of the OVERCAST drizzling case respond in a monotonic fashion. The change of LWP with time decreases with solar zenith
Figure 6.1. Large-eddy simulation output for the OVERCAST sounding with drizzle at five fixed solar zenith angles, $N_d = 50 \text{ cm}^{-3}$. Time series of quantities are domain averaged and vertically integrated.
angle (Fig. 6.1a), shifting from negative for $\Theta < 45^\circ$ to positive for $\Theta > 45^\circ$. LWP is nearly unchanged with time when $\Theta = 45^\circ$. Consequently, values of LWP averaged over the last two hours of simulation time vary over a wide range, from $71.5 \, \text{g m}^{-2}$ at $\Theta = 90^\circ$ to $32.0 \, \text{g m}^{-2}$ at $\Theta = 0^\circ$. We find that maximum shortwave forcing reduces the LWP by more than half that of the nocturnal simulation.

Entrainment rates, as inferred from the time evolution of $\overline{z_i}$ and STBL-averaged TKE, decrease with decreasing $\Theta$ (Fig. 6.1b and e). These results suggest that entrainment rates and circulation strength may be associated with each other for these simulations. While it is difficult to relate inversion strength and the interactions of radiative heating and drizzle to the entrainment rate, we further investigate this relationship below. Reductions in STBL LWP and entrainment rate during daytime have been observed (e.g. Caldwell and Bretherton, 2009) and modeled using both mixed-layer models (Turton and Nicholls, 1987) and LES (e.g. Sandu et al., 2007; Caldwell and Bretherton, 2009).

Integrated shortwave radiative heating increases monotonically with increasing LWP and decreasing $\Theta$, everything else being equal (Fig. 2.4), and we can see evidence of these relationships in our LESs. Figure 6.1c shows that integrated shortwave heating increases with decreasing $\Theta$, and this is especially evident at the beginning of hour one, when LWP is identical for all simulations. Integrated shortwave heating then varies as LWP varies with time at each solar zenith angle. For example, integrated shortwave heating increases with time for $\Theta = 60^\circ$ as LWP increases but decreases with time for $\Theta = 0^\circ$ as LWP decreases.

For all times and values of $\Theta$ the simulated cloud layers have a LWP $> 30 \, \text{g m}^{-2}$. For a fixed droplet concentration integrated longwave cooling varies only slightly with these values of LWP (Fig. 2.3). Nevertheless, as the simulations evolve with time we find that integrated longwave cooling diverges, ranging from $-60.5 \, \text{W m}^{-2}$ at $\Theta = 90^\circ$ to $-58.5 \, \text{W m}^{-2}$ at $\Theta = 0^\circ$ (Fig. 6.1d). This divergence is associated with the slight change in longwave cooling with LWP but is also associated with the entrainment rate and STBL ascension. As the STBL ascends, the cloud layer receives less longwave radiation from the atmosphere above because the atmospheric mass above is smaller.

Integrated drizzle flux divergence (Fig. 6.1f) decreases with decreasing $\Theta$ and changes in time commensurate with LWP, like integrated shortwave heating. Driz-
zle flux divergence drops considerably from $\Theta = 90^\circ$ to $\Theta = 60^\circ$ and from $\Theta = 60^\circ$ to $\Theta = 45^\circ$ but is fairly similar for the other solar zenith angles. The other plots in Fig. 6.1 exhibit the same pattern of changes with solar zenith angle: relatively similar for $\Theta \leq 45^\circ$, but relatively different for $\Theta \geq 45^\circ$. We will show that these similarities and differences can be associated with the extent of decoupling in the STBL, and how strongly it changes with shortwave heating (Fig. 6.2).

### 6.2.2 Vertical Profiles

Though the LWPs of these five simulations are dependent on $\Theta$, how that LWP is vertically distributed within the STBL is not. Peak LWC values near the top of the STBL decrease with decreasing $\Theta$ but the vertical profiles are similar (not shown) and do not complicate the interpretation of other vertical profiles. Examination of the buoyancy production of TKE (Fig. 6.2a) and its dependence on $\Theta$ reveals sharp decreases in buoyancy production near the STBL top as $\Theta$ decreases from $90^\circ$ to $45^\circ$ with relatively small decreases thereafter. Concurrently, buoyancy production below cloudbase is reduced sharply from $\Theta = 90^\circ$ to $\Theta = 60^\circ$ and remains fairly constant with subsequent drops in $\Theta$. Together these decreases in buoyancy production throughout the STBL indicate that, as $\Theta$ decreases, generation of downdrafts near cloudtop is weakened and more mechanical energy is required for these downdrafts to penetrate the sub-cloud layer (Schubert et al., 1979). Therefore we can expect stronger decoupling of the cloud layer from the sub-cloud and a less well-mixed STBL as $\Theta$ increases (e.g. Sandu et al., 2007).

Profiles of $\overline{w^\prime w^\prime}$ and their changes with $\Theta$ (Fig. 6.2b) exhibit the vertical circulation structure we anticipate from our interpretation of Fig. 6.2a. As $\Theta$ decreases, the maximum in $\overline{w^\prime w^\prime}$ decreases and shifts upward towards the STBL top. Values of $\overline{w^\prime w^\prime}$ in the sub-cloud layer also decrease as $\Theta$ decreases and are near zero above the surface when $\Theta \leq 45^\circ$. These profiles indicate, for $\Theta \leq 45^\circ$, circulations originating in the cloud layer are dissipating within the sub-cloud layer and the STBL is likely not well-mixed. Profiles of liquid water potential temperature ($\theta_l$, Fig. 6.2c) further corroborate this suggestion. For $\Theta < 45^\circ$, $\theta_l$ increases with height near the surface and this stable layer is not characteristic of a well-mixed
Figure 6.2. Large-eddy simulation output for the OVERCAST sounding with drizzle at five fixed solar zenith angles, $N_d = 50 \text{ cm}^{-3}$. Horizontally averaged vertical profiles. Data were temporally averaged over five minutes, and then were normalized by the height of the boundary layer top, $z_i$. These data were then interpolated to a uniform vertical grid. These uniformly gridded data were then temporally averaged over the last two hours of simulation time.
STBL.

The reduction in entrainment rate with decreasing $\Theta$ does not appear correlated with inversion strength at STBL top. As $\Theta$ decreases the inversion strength decreases (Fig. 6.2c) owing to the stronger shortwave heating of the STBL. The relationship from Lilly (1968) we cite in Chapter 5 suggests that for a well-mixed layer, entrainment rate should increase with decreasing inversion strength. That relationship is considered valid for a well-mixed layer but some of our simulations are not well-mixed and this relationship may not apply.

Hartman and Harrington (2005b) used LES to explore how vertical profiles of $w'w'$, buoyancy production of TKE and broadband radiative heating rates change with fixed values of $\Theta$. To initialize their simulations they used a sounding similar to the OVERCAST sounding used here. Although they used a different and less accurate radiative transfer model (Harrington, 1997), the patterns of changes with $\Theta$ they found in their vertical profiles are in agreement with our findings. Additionally, through coupling their LES output with a trajectory ensemble model, they found that for $\Theta \leq 45^\circ$, air parcels were confined to the cloud layer and for $\Theta > 45^\circ$ circulations could penetrate the depth of the boundary layer. In other words, for $\Theta \leq 45^\circ$ their STBL was not well-mixed and the cloud layer was significantly decoupled from the cloud layer. For $\Theta \geq 45^\circ$ the STBL was well-mixed.

These findings generally agree with those in our LESs, but we do not find as strong decoupling for $\Theta \leq 45^\circ$ as Hartman and Harrington (2005b) found; circulations penetrate the sub-cloud layer in our simulations. Hartman and Harrington (2005b) did not allow drizzle in their simulations, and in the next section we discuss the possibility that drizzle maintains partial coupling between the cloud and sub-cloud layers. Note that $\Theta = 45^\circ$, which appears to demarcate well-mixed and significantly decoupled STBLs, is dependent on thermodynamic structure and is not necessarily applicable to other STBLs.
6.3 Discussion: OVERCAST case, drizzling, Fixed Solar Zenith Angles, low $N_d$

Clearly the addition of shortwave radiative heating, and its changes with $\Theta$, is the reason for variation in STBL dynamical structure we find in Fig. 6.2a,b and c. To better understand how shortwave radiative heating modifies STBL dynamics, we must first examine how drizzle, longwave radiative heating and shortwave radiative heating change with $\Theta$. Longwave cooling at cloudtop is virtually identical for all simulations (Fig. 6.2e). As $\Theta$ decreases and the cloud thins, longwave warming at cloudbase becomes shallower and lessens in magnitude but these changes are negligible in comparison to cloudtop longwave cooling.

Figure 6.2f shows us shortwave radiative heating increases with decreasing $\Theta$. The distribution of this heating in-cloud also changes with $\Theta$. The peak in shortwave heating is found near STBL top regardless of $\Theta$ and the magnitude of heating there increases as $\Theta$ decreases, as expected. As $\Theta$ decreases, the magnitude of heating closer to cloudbase increases faster relative to the maxima in heating near STBL top. These differences could be related to different thermodynamic profiles for each simulation in addition to differences in $\Theta$.

To isolate changes in shortwave radiative transfer with $\Theta$ from those owing to dynamical responses, we compare shortwave radiative heating profiles for different values of $\Theta$ for a static model atmosphere, as we did in Chapter 2. We compute these profiles for the four $\Theta$s above and $N_d = 50 \text{ cm}^{-3}$. We choose a cloud of 220 m geometric depth and LWP of 50.7 g m$^{-2}$ because these values are similar to those simulated in our LESs.

Figure 6.3 shows the variation of shortwave radiative heating profiles with $\Theta$ for this static atmosphere. The shape of in-cloud shortwave heating profile does not depend strongly on $\Theta$ when the thermodynamical profiles are identical; Shortwave heating increases as $\Theta$ decreases for all vertical layers, above, in and below cloud. The shape of the shortwave radiative heating profile is sensitive to $\Theta$ within our LESs, and hence suggests there is a shortwave radiative-dynamical feedback in our LESs. These When our cloud layers in LES are thinned with decreasing $\Theta$, more shortwave radiative heating near cloudbase results, but, because the cloud thins, there is less integrated (total) absorption of shortwave radiation. This relationship
has been suggested before (Turton and Nicholls, 1987; Sandu et al., 2007), and our time series of integrated shortwave heating (Fig. 6.1c) also suggests this negative feedback in shortwave radiative heating.

Absorption of shortwave radiation in updrafts at cloudbase and in the cloud interior generates positive buoyancy in those portions of the cloud. At the same time cloudtop shortwave heating at cloudtop warms negatively buoyant downdrafts in those portions of the cloud, reducing downdraft strength. We find reductions in buoyancy generation of TKE in-cloud owing to decreasing $\Theta$ (Fig. 6.2b), suggesting that the latter effect is more significant than the former. Cloud shortwave radiative heating is not preferentially stronger in updrafts as compared to downdrafts (not shown), this suggesting that average warming of the cloud layer with respect to the sub-cloud across the model domain is more important for TKE reduction.

**Figure 6.3.** In-cloud shortwave radiative heating rates for $N_d = 50 \text{ cm}^{-3}$ for the static model atmosphere described in Chapter 2. Cloud added to mid-latitude summer (MLS) profile extends from 1.0 km at base to 1.22 km at top. Liquid water path is 50.7 g m$^{-2}$.

For daytime simulations we find that the drizzle process weakens with decreasing $\Theta$. Because cloud LWC and LWP decrease with decreasing $\Theta$, such a change in drizzle strength is not unexpected. Consequently, the warming associated with the formation of drizzle in cloud and the subsequent evaporation of this drizzle below cloud decreases in magnitude as $\Theta$ decreases (Fig. 6.1d). For our nocturnal LESs of the OVERCAST case, the drizzle process reduced entrainment and led to some decoupling of the cloud layer from the sub-cloud through warming of the cloud.
layer and cooling the sub-cloud layer. Circulation strength was also reduced, in
conjunction with the further decoupling and possibly through the “potential buoy-
ancy” mechanism (Stevens et al., 1998a). We expect that the magnitude of these
impacts is also reduced with decreasing $\Theta$. However, the increase in cloud warming
with decreasing $\Theta$ owing to absorption of shortwave radiation more than offsets
the reduction in magnitude of heating associated with drizzle and also strengthens
decoupling.

The net evaporation of drizzle in the sub-cloud layer also cools and moist-
ens that region and can lead to intermittent conditional instability and possibly
cumulus cloud formation, as we noted in the previous chapter (Stevens et al.,
1998a; Jiang et al., 2002). Although the drizzle process is reduced in the presence
of strong shortwave warming, the associated generation of conditional instability
may be important in keeping the cloud and sub-cloud layers partially coupled when
$\Theta$ is low.

As cloud solar radiative warming increases with decreasing $\Theta$, entrainment
rate slows and drizzle weakens while longwave radiative heating remains virtually
unchanged. The change in these processes suggests the following pathway for how
shortwave radiative heating modifies STBL dynamics for this case and droplet
concentration. As $\Theta$ decreases, the cloud layer warms with respect to the sub-cloud
layer, strengthening thermal stratification in the STBL. Subsequently this thermal
stratification leads to stronger decoupling of the cloud layer from the sub-cloud
as cloud shortwave absorption increases. This stronger decoupling is associated
with weaker STBL circulations, and this weakening could be associated with the
decrease in entrainment rate (e.g. Lock and MacVean, 1999). It is also possible
that solar warming near the cloudtop interface somewhat offsets the longwave
radiative cooling of penetrating thermals there, leading to reduced mixing along
the interface (Lock and MacVean, 1999). When the cloud layer is decoupled owing
to thermal stratification, entrainment of drier air from above the STBL interface
and reduced water vapor fluxes from the sub-cloud to the cloud layer both result
in LWP reduction and cloud thinning. Finally, drizzle may cause some re-coupling
through increases in conditional instability in the sub-cloud layer (Stevens et al.,
1998a).
6.4 OVERCAST case, drizzling, Fixed Solar Zenith Angles, high $N_d$: Differences

Increasing modeled droplet concentration for our nocturnal OVERCAST drizzling simulations in Chapter 5 resulted in a faster entrainment rate, a suppressed drizzle process and more partial decoupling of the cloud layer from the sub-cloud layer, but LWP was not significantly changed. How does our model STBL respond differently to shortwave radiative heating when $N_d$ is increased to 1000 cm$^{-3}$? We first look back to Fig. 2.5 to examine how cloud shortwave absorption changes when we change droplet concentration at different values of $\Theta$. We can use this information as a rough predictor of how our daytime STBL will respond to changes in droplet concentration.

OVERCAST simulations with $\Theta < 90^\circ$ and $N_d = 50$ cm$^{-3}$ exhibit LWP values between 25 and 55 g m$^{-2}$. For LWPs of this range, Fig. 2.5 shows us that increasing $N_d$ to 1000 cm$^{-3}$ results in less than a 1 W m$^{-2}$ change in cloud shortwave absorption when $\Theta = 60^\circ$, and this change is likely not significant. As $\Theta$ decreases differences in cloud shortwave absorption continue to increase towards a maximum of 11 W m$^{-2}$ at $\Theta = 0^\circ$ and LWP = 34 g m$^{-2}$. These increases in cloud shortwave absorption with increased droplet concentration suggest that we will find the largest changes in response when $\Theta$ is small.

Figure 6.4a shows us that integrated shortwave warming is indeed larger when $N_d = 1000$ cm$^{-3}$ as compared to $N_d = 50$ cm$^{-3}$ (Fig. 6.1c) when $\Theta \leq 45^\circ$. This increase is especially evident at the beginning of the simulations when LWP across all simulations for both droplet concentrations is identical. These increases in shortwave radiative heating are most likely the main reason for lowering LWP at the higher droplet concentration (Fig. 6.4b). Averaged over the last two hours of simulation, the LWP is 6.8% higher for $N_d = 1000$ cm$^{-3}$ as compared to $N_d = 50$ cm$^{-3}$ when there is no shortwave radiation modeled. In contrast, the percent difference in LWP between the two values of $N_d$ changes sign and becomes increasingly negative as $\Theta$ decreases. Liquid water path is 27.5% lower for $N_d = 1000$ cm$^{-3}$ as compared to $N_d = 50$ cm$^{-3}$ when $\Theta = 0^\circ$. As the cloud layers thin over time for this droplet concentration, integrated shortwave radiative heating decreases in response. Integrated longwave radiative cooling changes minimally with varying
Figure 6.4. Large-eddy simulation output for the OVERCAST sounding with drizzle at five fixed solar zenith angles, $N_d = 1000 \text{ cm}^{-3}$. a) - c) Time series of quantities are domain averaged and vertically integrated. d) - f) Horizontally averaged vertical profiles. Data were temporally averaged over five minutes, and then were normalized by the height of the boundary layer top, $z_i$. These data were then interpolated to a uniform vertical grid. These uniformly gridded data were then temporally averaged over the last two hours of simulation time.
Entrainment rate is also reduced as $\Theta$ decreases (Fig. 6.4c), as we found for $N_d = 50 \text{ cm}^{-3}$. The variation in entrainment is larger for $N_d = 1000 \text{ cm}^{-3}$ as compared to $N_d = 50 \text{ cm}^{-3}$ and could be associated with the wider variation in the cloud integrated shortwave heating over simulation time. Similar to what we found for $N_d = 50 \text{ cm}^{-3}$, STBL-averaged TKE decreases with decreasing $\Theta$ (not shown). This reduction in circulation strength could be caused by decoupling of the cloud and sub-cloud layers (Fig. 6.4d), and also might be associated with the reductions in entrainment rate. We might anticipate that reductions in entrainment rate would lead to less warming and drying of the STBL, increasing LWP. Increases in shortwave radiative heating most likely lead to greater overall heating and evaporation of cloud, overshadowing this entrainment response.

Vertical profiles of buoyancy production of TKE (Fig. 6.4d) and vertical velocity variance (Fig. 6.4e) for $N_d = 1000 \text{ cm}^{-3}$ reveal that, as $\Theta$ decreases, the cloud layer becomes more decoupled from the sub-cloud. Without shortwave radiative heating, there is some buoyancy destruction below cloudbase, a sign of some decoupling, but the profile $\tilde{w}'\tilde{w}'$ shows us that circulations generated near STBL top are reaching the surface. As $\Theta$ decreases, buoyancy generation in-cloud decreases and the maxima in buoyancy destruction rises higher in the STBL. The maxima in $\tilde{w}'\tilde{w}'$ also rises higher in the STBL with decreasing $\Theta$, and becomes smaller. For $\Theta = 60^\circ$, $\tilde{w}'\tilde{w}'$ is non-zero throughout the STBL but becomes very small near the surface, indicating that the STBL might still be well-mixed.

At lower values of $\Theta$, however, $\tilde{w}'\tilde{w}'$ is very small below $0.5 \bar{z}_i$, indicating that circulations generated at STBL top are not reaching the surface and the cloud layer is strongly decoupled from the sub-cloud. The decoupling is much stronger for $N_d = 1000 \text{ cm}^{-3}$ as compared to the daytime simulations for $N_d = 50 \text{ cm}^{-3}$. Vertical profiles of $\tilde{w}'\tilde{w}'$ for $N_d = 50 \text{ cm}^{-3}$ indicate that the STBL is well-mixed for all values of $\Theta$ (Fig. 6.2b). In contrast to $N_d = 50 \text{ cm}^{-3}$, drizzle is weak for this droplet concentration (not shown), hence drizzle does not play a significant role in either decoupling the cloud layer from the sub-cloud layer through warming the cloud layer or through cooling the sub-cloud layer. The drizzle process also does not work to recouple the two layers together through intermittent generation of conditional instability in the sub-cloud layer (Stevens et al., 1998a). These differ-
ences in drizzle strength might be important in explaining why stronger decoupling is observed for $N_d = 1000 \text{ cm}^{-3}$ as compared to $N_d = 50 \text{ cm}^{-3}$.

Because drizzle does not play an important role in heating the STBL, we examine changes in broadband radiative heating profile to help explain how STBL dynamics change with decreasing $\Theta$ (Figure 6.1f). As $\Theta$ decreases, we find that radiative cooling at cloudtop decreases and radiative warming in-cloud and at cloudbase increases. From $\Theta = 90^\circ$ to $\Theta = 0^\circ$ the maximum in cooling is reduced 47%, from 4.02 to 2.12 K hr$^{-1}$. The maximum in warming increases 71% from 0.35 to 0.60 K hr$^{-1}$ and rises to 0.85 $\overline{\zeta}$ from 0.70 $\overline{\zeta}$. These profiles show that shortwave radiative heating not only offsets the longwave radiative cooling at cloudtop which drives STBL circulations, but also promotes generation of positive buoyancy throughout the cloud layer. Because radiative heating profiles are not significantly different between updrafts and downdrafts (not shown), we infer that the direct effect of radiative heating on updrafts and downdrafts is similar.

For all values of $\Theta$, circulations are weaker for $N_d = 1000 \text{ cm}^{-3}$ as compared to $N_d = 50 \text{ cm}^{-3}$ (compare Fig. 6.2b to Fig. 6.4f). For our nocturnal simulations we attributed this difference in circulation strength to differences in entrainment rate and the strength of the drizzle process. We also suggested that longwave cooling at cloudtop for $N_d = 1000 \text{ cm}^{-3}$ promotes faster entrainment of warm dry air from above, resulting in shallower, weaker circulations. However, for $\Theta < 90^\circ$ broadband radiative cooling at cloudtop does not correlate with entrainment rate across the two modeled droplet concentrations. Regardless of the magnitude of broadband radiative cooling at cloudtop, entrainment rates are larger for all values of $\Theta$ for $N_d = 1000 \text{ cm}^{-3}$.

In their LESs of the STBL diurnal cycle, Caldwell and Bretherton (2009) found that $\overline{w'w'}$ from 0.6 $\overline{\zeta}$ to the top of the STBL correlates well with entrainment rate. They suggest that these quantities are well correlated because $\overline{w'w'}$ is a measure of the strength of motions responsible for entrainment. For our LESs, $\overline{w'w'}$ decreases as entrainment rate decreases with decreasing $\Theta$ for each fixed droplet concentration, but this relationship does not hold across simulations with different values of $N_d$.

Differences in the extent of decoupling in our simulations may help to explain why entrainment rate is not consistently related to either cloudtop radiative cooling
or values of $\overline{w'w'}$ near STBL top. When a cloud layer is partially decoupled from the sub-cloud layer, eddies of different strengths penetrate the sub-cloud layer to different depths, intermittently coupling the two layers to various extents (Stevens, 2000). Stronger eddies can promote stronger entrainment and vice versa for weaker eddies, and this variation in eddy strength and frequency is not visible in our two-hour averaged vertical profiles. Additionally, recent studies have shown cloud droplet sedimentation to play a significant role in the entrainment process (e.g. Bretherton et al., 2007; Ackerman et al., 2009). Because sedimentation of cloud droplets depends strongly on droplet concentration, our neglect of this process may complicate our attempts to correlate entrainment to radiative cooling or $w'w'$ at cloudtop.

As when $N_d = 50 \text{ cm}^{-3}$, increasing thermal stratification of the cloud layer and the sub-cloud appears to be the significant pathway by which shortwave radiation alters STBL dynamics as $\Theta$ decreases for $N_d = 1000 \text{ cm}^{-3}$. This increasing thermal stratification leads to stronger decoupling of the cloud layer from the sub-cloud and circulations weaken. However, there are notable differences between simulations with the two droplet concentrations. A stronger drizzle process for $N_d = 50 \text{ cm}^{-3}$ may be leading to more re-coupling of the cloud layer to the sub-cloud when $\Theta$ is low, as compared to $N_d = 1000 \text{ cm}^{-3}$ where the drizzle process is weak and decoupling is stronger.

The variations of LWP with $\Theta$ are also markedly different for the two droplet concentrations. For $N_d = 50 \text{ cm}^{-3}$, drizzle reduces LWP when $\Theta$ is large and shortwave heating is low, but shortwave heating reduces LWP when $\Theta$ is small and drizzle is weak. For the higher droplet concentration we find a larger variation in LWP because integrated shortwave heating varies more with $\Theta$ and the drizzle process is weak for all values of $\Theta$.

### 6.5 BROKEN case, drizzling, Fixed Solar Zenith Angles, low $N_d$

We found our nocturnal LESs of the BROKEN case to be more sensitive to droplet concentration than the OVERCAST case, and our ERM simulations from Chapter
suggest that these LESs will also prove more sensitive to shortwave radiation. When LWP is low, as it is for these simulations, integrated shortwave radiative heating does not increase with LWP as fast as integrated longwave radiative cooling (Fig. 2.6). Even with overhead sun and maximum shortwave forcing, integrated shortwave warming is only a fraction of the integrated longwave cooling for these low LWPs. The nocturnal LESs exhibited some decoupling for both values of \( N_d \) (Fig. 5.12b), and any additional shortwave warming of the cloud layer may strengthen this decoupling and lead to some amount of cloud evaporation. We first examine the STBL response to \( \Theta \) for \( N_d = 50 \text{ cm}^{-3} \).

### 6.5.1 Time Series

Similar to the OVERCAST case, our time series output exhibits gradual changes as \( \Theta \) decreases. For nocturnal simulations the cloud layer is considerably thinner for the BROKEN case as compared to the OVERCAST case, and increases in shortwave warming lead to even thinner cloud layers for this case. The variation of integrated shortwave radiative heating with \( \Theta \) is largest at the beginning of the first hour (Fig. 6.5c). This variation lessens with time as LWP decreases in response to strong shortwave warming when \( \Theta \) is small (Fig. 6.5a). Comparing LWP during the third and fourth hours of the simulation to LWP over the last two hours of the simulation, we find the cloud layer is slowly moistening for \( \Theta > 45^\circ \), unchanging at \( \Theta = 45^\circ \) and is slowly drying for \( \Theta < 45^\circ \). These trends in LWP with \( \Theta \) are associated with decoupling and we explore the extent of decoupling in more detail below (Fig. 6.6).

Because LWPs are low for the BROKEN case, decreases in LWP lead to decreases in cloud fraction (Fig. 6.5b). Cloud fraction was less than unity with no shortwave radiative heating, and as \( \Theta \) decreases cloud fraction is further reduced. The oscillations in cloud fraction, integrated longwave radiative cooling (Fig. 6.5b and d), and STBL-averaged TKE (not shown) are associated with grid-hopping and entrainment as we found in our nocturnal LESs of the BROKEN case. These oscillations are evident for all modeled values of \( \Theta \). The entrainment rate slows as \( \Theta \) decreases (Fig. 6.5e), and so the periodicity of these oscillations also decreases.

Like we found for the OVERCAST case, increases in integrated shortwave
Figure 6.5. Large-eddy simulation output for the BROKEN sounding with drizzle at five fixed solar zenith angles, $N_d = 50\, \text{cm}^{-3}$. Time series of quantities are domain averaged and vertically integrated.
radiative warming offset integrated longwave radiative cooling, reducing both circulation strength and the formation of cloud through cooling the STBL. However, unlike the OVERCAST case, increases in shortwave warming lead to decreases in longwave cooling itself because integrated longwave cooling is sensitive to LWP for low LWP clouds. This reduction in integrated longwave cooling is associated with further weakening of STBL circulation strength and reduced formation of cloud through STBL cooling. Liquid water path slowly decreases with time when $\Theta < 45^\circ$ and suggests that there is some minimum of integrated radiative cooling necessary to maintain the cloud layer against drying owing to entrainment and shortwave warming.

Integrated drizzle flux divergence is an order of magnitude less than integrated radiative cooling for $\Theta = 90^\circ$ and does not have a primary role in STBL dynamics. As $\Theta$ decreases the role of drizzle in modifying STBL dynamics is further reduced (Fig. 6.5f).

### 6.5.2 Vertical Profiles

For the OVERCAST case, the vertical distribution of LWC does not vary much with $\Theta$. Liquid water contents for the BROKEN case are reduced as $\Theta$ decreases but the vertical distribution of LWC varies as well, unlike for the OVERCAST case (Fig. 6.6a). In the nocturnal simulation of the OVERCAST case, LWC maximizes below cloudtop because entrainment of dry air reduces LWC near cloudtop. As $\Theta$ decreases, the entrainment rate is reduced (Fig. 6.5e) and mixing of dry air across the cloudtop interface is reduced commensurately. This reduction in entrainment with decreasing $\Theta$ leads to the LWC maxima becoming co-located cloudtop. There is more LWC in updrafts as compared to downdrafts for all values of $\Theta$, as we found for the nocturnal simulations. However, cloud in both updrafts and downdrafts is sufficiently thin to lead to significant reductions in radiative heating in both parts of the core circulation, hence we focus on domain averages.

As in the OVERCAST case we find that the buoyancy production of TKE is reduced throughout the cloud layer as $\Theta$ decreases (Fig. 6.6b). The largest reduction occurs as $\Theta$ decreases from $90^\circ$ to $60^\circ$, and the reductions lessen as $\Theta$ decreases further. These reductions in buoyancy production are caused by reductions in
Figure 6.6. Large-eddy simulation output for the BROKEN sounding with drizzle at five fixed solar zenith angles, $N_d = 50 \text{ cm}^{-3}$. Horizontally averaged vertical profiles. Data were temporally averaged over five minutes, and then were normalized by the height of the boundary layer top, $z_i$. These data were then interpolated to a uniform vertical grid. These uniformly gridded data were then temporally averaged over the last two hours of simulation time.
radiative cooling throughout the cloud layer (Fig. 6.6e). There is buoyancy destruction below cloudbase at all values of $\Theta$, indicating that there is some amount of decoupling (Stevens, 2000). Vertical profiles of the buoyancy production of TKE do not vary much with $\Theta$ below cloud.

As we found for the OVERCAST case, the maximum in $\overline{w'w'}$ decreases as $\Theta$ decreases from $90^\circ$ to $30^\circ$ (Fig. 6.6c). This relationship is anticipated as in-cloud buoyancy production decreases with $\Theta$. Curiously, as $\Theta$ decreases from $30^\circ$ to $0^\circ$, the trend is reversed; $\overline{w'w'}$ increases. We explain this reversal in trend through changes in entrainment. We suggest in Chapter 5 that entrainment of warm, dry air into the STBL is associated with reductions in circulation strength for nocturnal simulations of the BROKEN case. At $\Theta = 0^\circ$ entrainment is possibly reduced enough that mixing of warm dry air into the STBL does not weaken downdrafts generated by radiative cooling as much as it does for other values of $\Theta$.

For all values of $\Theta$ we find that $\overline{w'w'}$ does not become very small in the sub-cloud layer, indicating that the STBLs are at least partially coupled. Vertical velocity variance below cloud does decrease with decreasing $\Theta$, however, and could indicate that the cloud layer becomes more decoupled from the sub-cloud as $\Theta$ decreases. We do not find this decrease in $\overline{w'w'}$ below cloud as $\Theta$ decreases from $30^\circ$ to $0^\circ$. For these values of $\Theta$ vertical motions are similar and weak.

Vertical profiles of $\theta_l$ (Fig. 6.6d) corroborate this relationship of $\Theta$ to strength of decoupling. As $\Theta$ decreases, $\theta_l$ increases more with height in the STBL, and the STBL becomes warmer as integrated shortwave warming is increased and integrated radiative cooling is reduced. The inversion strength of $\theta_l$ decreases with decreasing $\Theta$ except when $\Theta$ decreases from $30^\circ$ to $0^\circ$. This trend in inversion strength is not well correlated with the entrainment rate as Lilly (1968) has suggested, and the discrepancy may be because our STBL is not always well-mixed. The strengthening of the inversion when $\Theta = 0^\circ$ might be related to the reversal in trend we found in vertical profiles of $\overline{w'w'}$ as $\Theta$ decreases.
6.6 Discussion: BROKEN case, drizzling, Fixed Solar Zenith Angles, low $N_d$

We find significant changes in both longwave and shortwave radiative heating as $\Theta$ decreases (Fig. 6.6e and f). Because the clouds are thin in this case, longwave radiative cooling occurs throughout the bulk of the cloud layer. As $\Theta$ decreases to 45° and the cloud thins we find that the small amount of longwave warming at cloudbase is eliminated. For low LWP clouds both longwave and shortwave radiative heating are sensitive to LWP, but longwave radiative heating is more sensitive (Fig. 2.6). In the cloud layer, the shortwave heating profile varies little from $\Theta = 45^\circ$ to $\Theta = 0^\circ$ while the maximum in longwave radiative cooling decreases from $-3.0 \text{K hr}^{-1}$ to $-2.3 \text{K hr}^{-1}$. These results suggest that, for low LWP stratus, changes in longwave radiative heating driven by shortwave warming play an important role in the STBL response to shortwave warming. The decrease in longwave radiative cooling and increase in shortwave radiative warming at cloudtop with $\Theta$ might help explain the reductions in entrainment rate (Fig. 6.5e). Both changes in radiative heating lead to less cooling of cloudy air at the interface and could lead to reduced mixing there (Lock and MacVeans, 1999).

For the BROKEN case, entrainment rate slows and drizzle weakens as cloud solar radiative warming increases with decreasing $\Theta$, as we found for the OVERCAST case. In contrast with the OVERCAST case, longwave radiative cooling is also reduced because the cloud layer is thin, and this is a key difference between the STBL responses to shortwave warming in the two cases. Shortwave radiative heating modifies STBL dynamics by warming the cloud layer with respect to the sub-cloud layer and strengthening the decoupling process, as in the OVERCAST case. Stronger decoupling leads to thinner clouds through net drying owing to shortwave warming and a reduction in water flux from the sub-cloud layer. Because LWP decreases, longwave radiative cooling at cloudtop also decreases. Because this cooling is the primary driver of STBL circulations, circulation strength also decreases and the cloud layer becomes further decoupled from the surface. Formation of cloud owing to longwave radiative cooling also decreases. Entrainment of warm, dry air from above, though reduced, works in concert with shortwave warming to further dry the cloud. When $\Theta$ becomes sufficiently low ($< 45^\circ$ in
this case), longwave radiative cooling can no longer maintain the cloud against entrainment and shortwave warming.

6.7 Differences: BROKEN case, drizzling, Fixed Solar Zenith Angles, high $N_d$

We now examine how increases in droplet concentration modify the STBL response to changes in $\Theta$. For nocturnal simulations of the BROKEN case, increasing $N_d$ from $50 \text{ cm}^{-3}$ to $1000 \text{ cm}^{-3}$ resulted in an increased entrainment rate and hence increased mixing of warm dry air into the STBL. This increased mixing in turn led to decreases in both cloud fraction and LWP (Fig. 5.11). There were consequent decreases in integrated longwave cooling, and together these changes resulted in a thin cloud layer in which formation of cloud through radiative cooling of the STBL was balanced against evaporation owing to entrainment. Circulation strength was considerably weaker as compared to $N_d = 50 \text{ cm}^{-3}$ and the cloud layer was partially decoupled from the sub-cloud (Fig. 5.12). For $N_d = 1000 \text{ cm}^{-3}$ and no shortwave forcing, LWP remains fairly constant for the last hours of the simulation. How does shortwave forcing modulate STBL dynamics for this case when $N_d = 1000 \text{ cm}^{-3}$?

For $\Theta < 90^\circ$ we find that LWP decreases with time and the rate becomes more negative with decreasing $\Theta$ (Fig. 6.7a). Cloud fraction follows the same pattern as LWP, decreasing with time for $\Theta < 90^\circ$. The rate of decrease in cloud fraction also increases as shortwave forcing increases (not shown). As for $N_d = 50 \text{ cm}^{-3}$, as $\Theta$ decreases integrated shortwave warming at the beginning of the simulation increases, but as the cloud thins and breaks up in response, integrated shortwave heating also decreases (Fig. 6.7b). Integrated longwave cooling also decreases with $\Theta$ (Fig. 6.7c) and because $\text{LWP} < 20 \text{ g m}^2$ these decreases become larger as LWP decreases (Fig. 2.6). Entrainment is also slowed as $\Theta$ decreases (Fig. 6.7d) but still plays a role in drying the cloud.

For $\Theta \leq 45^\circ$ longwave radiative cooling no longer supports the cloud layer in the face of strong solar heating and the cloud dissipates. Because surface fluxes are neglected a dry boundary layer with weak circulations results by the end of
Figure 6.7. Large-eddy simulation output for the BROKEN sounding with drizzle at five fixed solar zenith angles, $N_d = 1000\text{ cm}^{-3}$. a) - d) Time series of quantities are domain averaged and vertically integrated. e) - f) Horizontally averaged vertical profiles. Data were temporally averaged over five minutes, and then were normalized by the height of the boundary layer top, $z_i$. These data were then interpolated to a uniform vertical grid. These uniformly gridded data were then temporally averaged over the last two hours of simulation time.
the six hour simulation. Vertical profiles of buoyancy production of TKE and \( \bar{w'}v' \) averaged over the last two hours of simulation (Fig. 6.7e and f) also exhibit this decrease in circulation strength with decreasing \( \Theta \). For \( \Theta = 30^\circ \), longwave radiative cooling at STBL top generates circulations, but the cloud layer is decoupled from the surface. For \( \Theta \leq 45^\circ \) buoyancy production and \( \bar{w'}w' \) throughout the STBL become negligible as the cloud layer dries and dissipates.

Our simulations of the BROKEN case suggest that low LWP stratus layers under strong inversions can not be maintained in the presence of strong shortwave forcing, especially when drop concentrations are high. However, when \( N_d \) is low cloud layers are better maintained in the presence of strong shortwave forcing.

6.8 Results: OVERCAST, drizzling, Diurnal Cycle

To better understand how the variation of incoming shortwave radiation impacts our modeling cases, we conducted sixteen hour LESs where \( \Theta \) varies diurnally. We ran four LESs with the two initial soundings (OVERCAST and BROKEN) and the two values of \( N_d \) (50 cm\(^{-3}\) and 1000 cm\(^{-3}\)) as before. Our simulations of the STBL diurnal cycle were configured as if we were modeling an STBL off of the Baja California coast near the summer solstice. The initial time of simulation was 1200 Z on June 22, 2004, and the latitude and longitude of the center of the modeling domain was 30\(^\circ\)N, –125\(^\circ\)W. This configuration results in sunrise occurring at 1330 Z (one and a half hours into the simulation), the sun nearly overhead at solar noon (2018 Z, more than eight hours into the simulation), and sunset occurring at 300 Z the following day (one hour before the end of the simulation). By using this solar configuration we maximize the variation in incoming solar radiation throughout the day. We use our fixed solar zenith angle results as a guide for interpreting our diurnal results.

We compare these results to those from Sandu et al. (2007) and Caldwell and Bretherton (2009). In both of these studies the investigators used LES to study the diurnal cycle of stratocumulus cloud systems and the response of these cloud systems to changes in fixed droplet concentrations. Caldwell and Bretherton...
modeled droplet concentrations of 25 cm\(^{-3}\) and 100 cm\(^{-3}\). Because these values are dissimilar to our high droplet concentration (1000 cm\(^{-3}\)), we do not focus on their findings with respect to changes in droplet concentrations. Sandu et al. (2007) modeled droplet concentrations more in line with those in our study and so their findings should compare better to ours.

Sandu et al. (2007) report their modeled cloud fraction as close to unity throughout their simulations, hence our simulations of the OVERCAST case will bear more resemblance to their simulations. Caldwell and Bretherton (2009) simulated several diurnal cycles of stratocumulus in part to determine how well their LESs could represent cloud layers observed during the East Pacific Investigation of Climate (EPIC) field campaign (Bretherton et al., 2004). Over these diurnal cycles they simulated both an overcast and broken stratocumulus cloud layer. We can compare their results to ours for both cases, depending on their model cloud fraction at particular times in their simulations. We examine the OVERCAST case first using time series output (Fig. 6.8), starting from one hour into the simulation to neglect simulation spin up.

The variation of the integrated quantities in this figure exhibit the pattern we might expect based on our fixed-\(\Theta\) simulations. The first few hours of these diurnal simulations are not unlike our nocturnal simulations (Fig. 5.7), and their dependence on \(N_d\) is also similar. Liquid water paths are similar for the two droplet concentrations (Fig. 6.8a). Stronger entrainment and entrainment drying for \(N_d = 1000\) cm\(^{-3}\) (Fig. 6.8e) modulate the LWP for that simulation and for \(N_d = 50\) cm\(^{-3}\) the stronger drizzle process reduces LWP (Fig. 6.8f). The increased entrainment rate also leads to lower STBL-averaged TKE for \(N_d = 1000\) cm\(^{-3}\) as warm, dry air from above reduces circulation strength (Fig. 6.8b). For \(N_d = 1000\) cm\(^{-3}\) the cloud layer is also partially decoupled from the sub-cloud layer.

As \(\Theta\) decreases and shortwave radiative heating increases (Fig. 6.8c), there are reductions in both LWP and STBL-averaged TKE for both droplet concentrations. The reductions are larger for \(N_d = 1000\) cm\(^{-3}\), and we can use the reasoning from our simulations with fixed values of \(\Theta\) to help explain differences in the two diurnal simulations. Figure 6.8c shows us that there is more absorption of shortwave radiation for \(N_d = 1000\) cm\(^{-3}\), consistent with our previous daytime simulations. This increased shortwave warming reduces the entrainment rate for both droplet
Figure 6.8. Large-eddy simulation output for the OVERCAST sounding with drizzle. Sixteen hours of the diurnal cycle are modeled, from before one and a half hours before sunrise to one hour after sunset. Solar zenith angle reaches 0° at its maximum. Time series of quantities are domain averaged and vertically integrated.
concentrations but entrainment remains more rapid for $N_d = 1000 \text{ cm}^{-3}$ during the day. Together with a weak drizzle process, the increased shortwave warming weakens STBL circulations and results in strong decoupling of the cloud and sub-cloud layers. This decoupling allows for the entrainment process, though somewhat reduced in the presence of shortwave forcing, to efficiently dry the cloud.

The increased shortwave warming also results in decoupling for $N_d = 50 \text{ cm}^{-3}$, but not to the same extent as for $N_d = 1000 \text{ cm}^{-3}$. The $N_d = 50 \text{ cm}^{-3}$ simulation was better coupled before shortwave warming becomes strong, and as $\Theta$ decreases there is less shortwave warming for $N_d = 50 \text{ cm}^{-3}$ as compared to $N_d = 1000 \text{ cm}^{-3}$. Additionally, the drizzle process is weakened for $N_d = 50 \text{ cm}^{-3}$ during the day but may be causing intermittent conditional instability below cloud that keeps the cloud layer better coupled to the sub-cloud layer (Stevens et al., 1998a). Re-coupling of the cloud and sub-cloud layers may produce a larger vapor flux from the sub-cloud to the cloud layer as compared to $N_d = 1000 \text{ cm}^{-3}$. This vapor flux, together with less efficient entrainment drying, leads to higher LWP for $N_d = 50 \text{ cm}^{-3}$ in the presence of strong shortwave heating.

Unsurprisingly LWP, STBL-averaged TKE and drizzle increase as $\Theta$ increases and cloud shortwave warming decreases. Entrainment rates also increase. Throughout the simulation, variation in integrated longwave cooling and the dependence of this cooling on $N_d$ is small compared to changes in shortwave radiative heating and drizzle flux divergence. This result gives us further confidence that, when investigating STBL first-order characteristics such as LWP for clouds of sufficient thickness, the dependence of integrated longwave cooling on droplet concentration can be neglected.

How do results from our diurnal simulations compare to those from Sandu et al. (2007) and Caldwell and Bretherton (2009)? In their respective studies LWP, TKE and drizzle rates decrease with increasing shortwave radiation and vice versa, in agreement with our findings. These STBL responses to shortwave radiation have also been observed during field campaigns (e.g. Albrecht et al., 1990; Bretherton et al., 2004). Caldwell and Bretherton (2009) found their model overcast boundary layer to exhibit some decoupling at all times, during both day and night. This decoupling may be associated with the drizzle process (e.g. Stevens et al., 1998a). Additionally, the nighttime boundary layer modeled by Caldwell and Bretherton
(2009) is several hundred meters deeper than our nighttime boundary layer and may partially explain this difference in decoupling as well. As a boundary layer deepens and mass increases, stronger circulations are required to maintain vertical mixing throughout.

Sandu et al. (2007) found similar results to ours for the daytime STBL dynamical response to varying droplet concentrations. Using the fixed values of \( N_d = 50 \text{ cm}^{-3} \) and \( N_d = 600 \text{ cm}^{-3} \), they found that their simulation with the lower droplet concentration remained partially coupled to the surface during the day while the simulation with the higher droplet concentration did not. Subsequently, LWP decreases more for \( N_d = 600 \text{ cm}^{-3} \) during daytime. They ascribe this difference in the extent of decoupling, in part, to differences in drizzle rate, as we suggest.

Sandu et al. (2007) also found that entrainment rate decreases with increased droplet concentration in the presence of shortwave heating. They explain that, for \( N_d = 600 \text{ cm}^{-3} \), circulations are weakened as compared to \( N_d = 50 \text{ cm}^{-3} \) and do not support as rapid entrainment. In our study we find that entrainment rates are larger for the higher droplet concentration, though the difference in entrainment rate with droplet concentration grows smaller as \( \Theta \) decreases. There are several possible reasons for the differences in entrainment rates between ours and Sandu et al.’s (2007) simulations.

First, Sandu et al. (2007) account for sedimentation of cloud droplets and we do not. As droplet concentrations increase, the sedimentation rate of cloud droplets decreases, and this decrease near the STBL interface has been associated with changes in entrainment (Bretherton et al., 2007; Ackerman et al., 2009). Second, our model vertical grid spacings are different; they use 10 m and our simulations have 30 m. Stevens et al. (2005) found that entrainment rates in LES are sensitive to model vertical resolution, increasing as vertical grid spacing decreases. Third, changes in radiative cooling at cloudtop can interact with and influence entrainment, as we have emphasized in our study. Sandu et al. (2007) do not display radiative heating profiles, but they use a different radiative transfer model with a longer radiative timestep. Differences in radiative heating computation may also play a role in altering entrainment rates. Lastly, the initial soundings used in our simulations are not identical to those Sandu et al. (2007) used, and STBL
thermodynamic structure influences entrainment (e.g. Lilly, 1968; Caldwell and Bretherton, 2009).

6.9 Results: BROKEN, drizzling, Diurnal Cycle

As we find for the OVERCAST case, variation of integrated quantities with Θ and \(N_d\) for the BROKEN case exhibits patterns we might expect from our fixed Θ simulations. The fixed-Θ simulations of the BROKEN case suggested that low LWP stratus layers are better maintained in the presence of strong shortwave heating when the droplet concentration is low. For higher droplet concentrations low LWP stratus layers are not well maintained and can dissipate in the presence of strong shortwave heating. Figure 6.9 shows that the same results apply when a full diurnal cycle is simulated.

The variation of incoming shortwave radiation with time is the same as for the OVERCAST case, but because the LWP is smaller for the BROKEN case, the magnitude of integrated shortwave heating is less (Fig. 6.9c). For \(N_d = 50\text{cm}^{-3}\), LWP decreases during the day, reaching a minimum of 8 g m\(^{-2}\) 10.2 h into the simulation (Fig. 6.9a). The cloud layer is thin so cloud fraction drops considerably with increasing shortwave heating and reaches a minimum around the tenth hour of simulation (Fig. 6.9f). Concurrent with these decreases in LWP and cloud fraction, integrated longwave radiative cooling decreases (Fig. 6.9d).

Both the entrainment rate and STBL-averaged TKE decrease as Θ decreases (Fig. 6.9b and e). As Θ increases and shortwave heating decreases these trends reverse in the last four hours of the simulation. As we found in both our nocturnal and fixed-Θ simulations, oscillations in TKE are strongly correlated with the ascension of the STBL (see Fig. 5.11). The circulations are relatively weak for this case, and when the cloud layer ascends the increased mixing of entrained air leads to considerable decreases in STBL-averaged TKE.

For \(N_d = 1000\text{cm}^{-3}\), the cloud layer dissipates before shortwave forcing reaches its maximum. In the first few hours the combined effects of entrainment drying and shortwave radiative heating lead to more rapid decreases in LWP as compared to the lower droplet concentration. Concurrently, cloud fraction and longwave radiative cooling also decrease rapidly and circulations are rapidly weakened. Conse-
Figure 6.9. Large-eddy simulation output for the BROKEN sounding with drizzle. Sixteen hours of the diurnal cycle are modeled, from before one and a half hours before sunrise to one hour after sunset. Solar zenith angle reaches 0° at its maximum. Time series of quantities are domain averaged and vertically integrated.
sequently the cloud layer can not be maintained against further entrainment drying and shortwave warming; by the seventh hour of the simulation, LWP has decreased to zero. The cloud layer does not reform after sundown, although it might have had we ran the simulation longer.

The large differences in our BROKEN simulations as compared to Caldwell and Bretherton’s (2009) simulations make it difficult to compare the two. We focused on the role longwave and shortwave radiative heating play in regulating the evolution of the STBL. Caldwell and Bretherton (2009) do not comment on changes in longwave radiative cooling or shortwave radiative heating with changes in LWP, cloud fraction or droplet concentration, instead focusing on changes in entrainment and cloud microphysics. At times their modeled LWPs are $< 20 \text{ g m}^{-2}$ and changes in radiative heating could be important in helping to explain the evolution of their simulation.

Because their LES simulation was many times longer than ours, Caldwell and Bretherton (2009) forced their LES with observations from the EPIC field campaign to account for large-scale environmental changes over time. Modeled cloud fraction drops substantially three days into their simulation and they attribute this decrease to advection of dry air into their domain at that time. The large-scale environment for our LESs is constant, and hence this effect is not possible in our simulations.
In the previous chapter we gained some understanding of how the STBL evolves diurnally in our two cases. For those simulations, along with all others in this work, both longwave and shortwave radiative transfer were computed with the Independent Column Approximation (ICA). The ICA assumes that each vertical column in an atmospheric model is plane parallel and independent of the other model columns so that radiation is not exchanged between them. This approximation introduces errors into computed radiative heating rates, and these errors can be significant for the horizontal grid resolutions we used (Cahalan et al., 2005). The impact of longwave radiative heating errors owing to the use of the ICA has been tested in a few studies (e.g. Guan et al., 1997; Mechem et al., 2008). Mechem et al. (2008) investigated the STBL and we briefly commented on their study in Chapter 5. To our knowledge no one has tested the impact of shortwave radiative heating rate errors owing to the use of the ICA in the STBL.

The use of the ICA does not account for heating or cooling of cloud sides or
horizontal diffusion of radiation through clouds, and leads to incorrect shadowing of the surface. O’Hirok and Gautier (2005) calculated shortwave radiative fluxes with and without the ICA for two-dimensional static cloud fields of varying grid cell resolutions. For the smallest grid cell resolution they modeled, 200 m, use of the ICA generated maximum shortwave heating rate errors of more than 1 K hr$^{-1}$ and maximum shortwave surface flux differences of up to 500 W m$^{-2}$. Frame et al. (2009) computed shortwave radiative transfer with and without the ICA for a cumulonimbus cloud field and found similar radiative heating rate and surface flux errors. In this study we couple a three-dimensional radiative transfer (3DRT) solver to RAMS to test the impact of these sorts of errors on model STBL dynamics for the BROKEN case.

There are two widely used methods for computing 3DRT. They are the Monte Carlo method and the Spherical Harmonics Discrete Ordinate Method (SHDOM) (Cahalan et al., 2005). We use the former method. In general, Monte Carlo techniques are used to make probabilistic models of processes. When applied to radiative transfer, the Monte Carlo method is used to compute the trajectories of numerous “energy bundles” throughout a medium, solving the radiative transfer equation statistically. By converting probability distributions of the physical processes involved in radiative transfer (e.g. scattering and photon path length) to uniform probability distributions, uniform random numbers can be picked to determine each energy bundle’s trajectory probabilistically (Bohren and Clothiaux, 2006). Through tracking a large number of energy bundles as they interact with a medium, the three-dimensional radiative field is computed.

We are interested in computing shortwave 3DRT and so we simulate energy bundles as entering the top of Earth’s atmosphere from the sun. To obtain radiative quantities in a static cloud field through the Monte Carlo method, the fraction of energy bundles that meet a certain fate is calculated (Cahalan et al. 2005). For example, the radiative heating rate of a grid cell in a model domain is directly proportional to the ratio of energy bundles absorbed in that grid cell to the number of energy bundles tracked in the domain. The accuracy of Monte Carlo approaches applied to atmospheric radiative transfer is predictable through examination of the variances of solutions found with subsets of tracked energy bundles (Cahalan et al. 2005). In general, as the number of modeled energy bundles increases, the accuracy
of this method increases. An advantage of the Monte Carlo method is that it leads to unbiased errors, unlike errors owing to the use of the ICA (Bohren and Clothiaux, 2006). We replace the radiative transfer model within RAMS described in Chapter 2 with that developed by Cole (2005). This radiative transfer model employs a Monte Carlo radiative transfer solver for shortwave radiation and a two-stream solver for longwave radiation. Gaseous absorption is computed using the Li and Barker (2005) correlated-k spectral band model and cloud optical properties are computed using polynomial expressions dependent on cloud droplet effective radius. The coefficients of these polynomial fits are from Dobbie et al. (1999) for the shortwave spectrum and Lindner and Li (2000) for the longwave spectrum. The Monte Carlo solver can also be configured to use the ICA. When the ICA method is to be used, the horizontal grid resolution is made very coarse within the radiative transfer model ($10^{18}$ m). This changes the probability that energy bundles will be exchanged between model columns to near zero. This ICA configuration option makes this model suitable for testing the impact of shortwave radiative heating rate biases owing to the use of the ICA.

Computing shortwave 3DRT interactively within an atmospheric model is computationally expensive. Cole (2005) parallelized his radiative transfer model using the Message Passing Interface (MPI), reducing the time needed for radiative heating computations. For this study we run all other routine in RAMS on a single processor and use 64 processors in parallel to compute radiative heating. The use of 64 processors to track $10^7$ energy bundles in computation of shortwave radiative heating with this Monte Carlo solver requires roughly the same amount of processing time as the two-stream solver we described in Chapter 2 on a single processor. Because of the similarity in processing time we use $10^7$ energy bundles in our simulations.

Using this parallelized model requires that the number of gridpoints must be evenly divisible by the number of processors. Therefore we configured RAMS to run with 70 gridboxes (72 gridpoints) of 50 m spacing in the horizontal instead of 68 gridboxes (70 gridpoints), making for a slightly larger horizontal modeling domain (3.50 km on a side compared to 3.40 km previously). Otherwise we configure RAMS as described in Chapter 3.

Although biases in shortwave radiative heating owing to the use of the ICA can
be large, this heating is but one of the many processes that influence STBL dynamics (Fig. 1.1). We could test the impact of these heating rate errors in ensemble studies using a large number of LESs of the STBL, varying initial thermodynamic sounding and solar position, and find that these errors have no significant impact. With such a finding we could not conclusively state that the ICA is acceptable for shortwave radiative heating computation, as there might be a set of inputs, yet to be tested, for which it is not an acceptable approximation. In an attempt to sidestep this issue, we conceive of atmospheric model configurations in which these biases in shortwave radiative heating owing to the use of the ICA have a chance of impacting STBL dynamics. We then conduct a select number of LESs with these configurations, with and without the ICA. If these simulations show that the ICA does not significantly alter STBL dynamics, we have some confidence that, for grid resolutions typical of those used in LES, the ICA is a reasonable approximation to use in numerical models of the STBL.

Because the ICA assumes no net horizontal exchange or radiation between model columns, for perfectly horizontally homogeneous cloud fields its use results in no error. As horizontal inhomogeneity increases net horizontal transport of radiation increases and the use of the ICA results in larger errors in radiative computation. For shallow clouds like those we model here, studies have shown that errors in radiative fluxes owing to the ICA are larger for cloud fields of low cloud fraction as compared to high cloud fraction (e.g. Zuidema et al., 2008). Therefore we tested importance of shortwave horizontal transport with the BROKEN case, where cloud fractions are lower and there is significant horizontal variation in LWC, as evidenced by differences in LWC between updrafts and downdrafts (Fig. 5.14).

Surface fluxes are already neglected in all of our simulations, and so STBL dynamics for this test will be driven primarily by entrainment and radiative heating in both the shortwave and longwave. We did not find drizzle to qualitatively alter STBL dynamics for the BROKEN case, but we turn it off so that drizzle can not mask any impact owed to the use of the ICA in shortwave radiative heating computation. Cloud droplet concentration was set at 100 cm$^{-3}$. Our simulations were six hours long and modeled as occurring for the same day and geographical location as used in Chapter 6. After one hour of simulation spin-up time, when shortwave radiation was not modeled, we applied shortwave forcing according to
the time of day, allowing solar zenith and azimuth angle to vary.

As \( \Theta \) increases, the proportion of horizontal solar irradiance increases as the downward shortwave irradiance decreases. We wished to maximize both the magnitude of shortwave heating and its horizontal transport. In an attempt to do so we modeled two different five hour periods of the diurnal cycle, after the one hour of spin-up. First, we set the initial time after spin-up to 2030 Z. With this time, the sun is modeled as almost directly overhead when shortwave forcing is first applied, leading to a rapid decrease in cloud fraction. As time progresses, the solar zenith angle increases and a larger portion of shortwave radiation travels horizontally. From the beginning of the first hour to the end of these simulations \( \Theta \) ranges from 14.5° to 78.1°.

Second, we set the initial time after spin-up to 2230 Z. At the beginning of the first hour of these simulation, \( \Theta \) is 40.0° and decreases to 90°, with the sun on the horizon, four hours later. With this simulation we have more horizontal transport of shortwave radiation initially without as rapid a decrease in cloud fraction.

We modeled these two scenarios with and without the ICA. Time series from these two sets of simulations are shown in Fig. 7.1 and Fig. 7.2 respectively.

In both scenarios modeling horizontal transport of shortwave radiation resulted in minimal differences. In Fig. 7.1 we find no differences until hour four of the simulation, and those changes occur when the LWP < 5 g m\(^{-2}\). Figure 7.2 shows some small differences in cloud fraction from the beginning of hour two to the end of the simulation, but LWP is virtually unchanged. For both scenarios, integrated shortwave radiative heating changes minimally when horizontal transport of shortwave radiation is modeled. Domain averaged albedos for these cases are not appreciably altered either.

We ran another pair of simulations, with and without the use of the ICA, but using the OVERCAST sounding. These simulations were configured like the simulations above, excepting a modification to solar geometry. The initial time after spin-up was set to 1500 Z. From the beginning of the first hour to the end of these simulations, \( \Theta \) ranges from 57.7° to 11.2°. This solar geometry also results in a large proportion of horizontal solar irradiance and total shortwave heating. The results of this pair of simulations (not shown) revealed that the ICA and its associated error in radiative heating had a minimal impact. Because the OVER-
CAST case had cloud fractions of unity and clouds that were relatively horizontally homogeneous, we anticipated that the use of the ICA would be appropriate.

These simulations suggest that the ICA is acceptable for use in computation of shortwave radiative heating for the STBL, and we are left to speculate why this is the case. Clouds in the STBL are limited in vertical extent because of the strong inversion under which they exist. The largest shortwave heating rate errors due to the use of the ICA are typically found for clouds of large vertical extent and horizontal inhomogeneity, such as cumulus or cumulonimbus (O’Hirok and
Figure 7.2. Large-eddy simulation output for the BROKEN sounding without drizzle, $N_d = 100 \, \text{cm}^{-3}$. Longwave radiative heating is computed with the ICA for both simulations and shortwave radiative heating is computed with and without the ICA. $\Theta$ varies from $40.0^\circ$ at start of hour one to sundown at start of hour six. Time series of quantities are domain averaged and vertically integrated.

Gautier, 2005; Frame et al., 2009). Shortwave radiative heating rate errors in the STBL may simply not be large enough to appreciably alter the dynamics of the cloud system. Perhaps the changes in cloud radiative heating rates are dissipated by small-scale turbulence before they can influence the larger eddies driving the STBL.

Pincus and Stevens (2009) suggest that errors in computed radiative heating owing to the use of the ICA might be akin to random noise in radiative heating rates. Using a few LESs of the STBL, they found that their simulations were not
sensitive to uncorrelated random noise in radiative heating rates. Though it was not the primary focus of their study, they speculate that “[the ICA] may turn out to be a perfectly useful approximation for large-eddy simulations.”

We designed our tests in this chapter specifically to limit the number of simulations to execute. Three pairs of simulations are too small of a sample size with which to make a firm conclusion on the applicability of the ICA to the STBL. Further testing with other thermodynamic inputs and solar geometry is clearly needed. That said, we would not be surprised to find that other simulations show that the ICA is suitable for computation of shortwave radiative heating in LESs of the STBL.

Further, we recommend testing the applicability of ICA in shortwave radiative heating computation in altocumulus cloud systems (e.g. Liu and Kreuger, 1998) or cirrus cloud systems, for which surface fluxes are negligible. Simulations show cirrus uncinus to be partially driven by shortwave radiative heating (e.g. Starr and Cox, 1985; Harrington et al., 2009). Given that uncinus can have considerable vertical extent (e.g. Heymsfield, 1975), biases in computed shortwave radiative heating rates owing to the use of the ICA may prove significant to their modeled structure and dynamics.
Summary, Conclusions and Future Research

8.1 Summary and Conclusions

In this study we investigated the impact of radiative heating on stratocumulus-topped boundary layer (STBL) dynamics using large-eddy simulation (LES). The feedbacks and interactions between STBL processes, both internal and external, make for a complex cloud system in which one process is difficult to untangle from another. Interactions of clouds and radiation in the STBL are complex so we first took steps to enrich our understanding of these interactions before interpreting results from LESs.

First, we examined the variation of cloud longwave and shortwave radiative heating with liquid water path (LWP), droplet concentration ($N_d$) and cloud depth. We conducted this analysis outside of our large-eddy simulations, removing the potential for cloud-radiative feedbacks. This examination revealed that for low LWP clouds (LWP $< 20$ g m$^{-2}$) vertically integrated longwave radiative cooling
is sensitive to $N_d$, but this sensitivity is reduced for clouds of larger LWP. We also found that vertically integrated shortwave radiative heating is sensitive to $N_d$ and this sensitivity is dependent on $\Theta$ and LWP. Summing vertically integrated longwave and shortwave radiative heating together to obtain broadband radiative heating revealed portions of (LWP,$\Theta$) parameter space where aerosol indirect effects related to radiative heating may be important. Finally, we found that for low LWP clouds integrated longwave radiative cooling is more sensitive to LWP than integrated shortwave heating. This finding suggests that, for cloud layers of low LWP, changes in longwave cooling could prove more important to changes in STBL dynamics, even during the daytime.

Second, we ran the Regional Atmospheric Modeling System (RAMS) as a two-dimensional eddy-resolving model (ERM) to understand the first-order STBL dynamic response to a wide range of microphysical and radiative variations. We used two different thermodynamic profiles to initialize our ERM simulations: a) the OVERCAST sounding, where cloud fraction is unity, LWP is not low and integrated longwave cooling is not sensitive to LWP, and b) the BROKEN sounding, where cloud fraction is not unity and LWP is sufficiently low such that integrated longwave cooling is sensitive to LWP. These simulations revealed that first-order, integrated quantities depend on $\Theta$ and $N_d$ relatively monotonically. Consequently we were able to limit the number of fixed values of $\Theta$ and $N_d$ for which we conducted LESs, expecting that our findings with LES apply to intermediate values of $\Theta$ and $N_d$.

Through these preliminary steps we gained a general understanding of cloud-radiation interactions within the STBL, and this understanding guided us in configuration and interpretation of our LESs. Using RAMS configured for LES, we first studied longwave radiative-cloud interactions and their impact on dynamics using nocturnal simulations of the STBL, and we did so for two reasons. First, we wished to explore the dependence of longwave radiative heating on droplet concentration and its impact on STBL dynamics. Second, a sound understanding of these simulations provided a foundation upon which we could build knowledge regarding shortwave impacts on STBL dynamics.

We ran several nocturnal LESs, varying droplet concentration ($N_d = 50$ and $1000 \text{ cm}^{-3}$), the initial thermodynamic sounding (OVERCAST and BROKEN cases), and the
inclusion or neglect of drizzle. From simulations of the OVERCAST case, where LWP is not low and cloud fraction is unity, we found the following (Fig. 8.1).

![Diagram](image)

**Figure 8.1.** A conceptual diagram of the nocturnal STBL when the LWP is not low, independent of droplet concentration.

- Entrainment rate increases with $N_d$ and this increase may be associated with increases in cloudtop radiative cooling with $N_d$.

- For a non-drizzling STBL, this increased entrainment rate leads to partial decoupling of the cloud layer from the sub-cloud for high $N_d$, while for low $N_d$ the STBL is well-mixed.

- A stronger drizzle process leads to partial decoupling of the cloud-layer from the sub-cloud for low $N_d$. Drizzle is weak for high $N_d$ and consequently does not significantly alter nocturnal STBL dynamics.

- The drizzle process effectively mitigates the differences in simulations with different droplet concentrations.

- When the cloud layer is not thin, a reasonably accurate description of STBL first-order dynamical structure can be obtained without accounting for droplet concentration effects on longwave radiative heating. Accounting for these effects may be necessary to accurately capture detailed nocturnal STBL structure.
From our LESs of the BROKEN case, where LWP is low and cloud fraction is not unity, we found the following (Fig. 8.2).

Figure 8.2. A conceptual diagram of the nocturnal STBL when the LWP is low, a) low $N_d$, b) high $N_d$.

- Entrainment rate increases with $N_d$, and, as for the OVERCAST case, this increase in entrainment rate may be associated with increases in cloudtop radiative cooling with $N_d$.

- At the beginning of the simulations this difference in entrainment rate leads to different cloud fractions and LWPs.

- These differences in cloud fractions with $N_d$ lead to different reductions in domain-averaged integrated longwave cooling.

- For high $N_d$, integrated longwave cooling is reduced such that cloud growth is suppressed and the cloud layer is partially decoupled, as circulations are weak. For low $N_d$, integrated longwave cooling is reduced, but not enough to suppress cloud growth. Circulations are also stronger and the cloud layer is not as decoupled.

- The divergence of this set of LESs with droplet concentration indicates that accounting for changes in longwave radiative heating with droplet concentration is important for simulating low level liquid water clouds.
We added shortwave forcing to our nocturnal LESs of the STBL by fixing the sun at four solar zenith angles. These fixed solar zenith angle simulations, together with our simulations of the diurnal cycle, suggest the following for our OVERCAST case (Fig. 8.3):

- Shortwave radiative heating affects the STBL primarily through increasing thermodynamic stability. Shortwave heating causes the cloud layer to warm, circulations driven by longwave radiative cooling weaken, the cloud layer becomes more decoupled and entrainment slows.

- Liquid water path consequently decreases, but not enough to result in decreased longwave cooling.

- As $\Theta$ decreases, LWP decreases, the circulations weaken further and the decoupling becomes stronger. The extent of cloud thinning and decoupling is dependent on $N_d$.

- For high $N_d$, cloud shortwave absorption is large, leading to stronger decoupling and further cloud thinning as $\Theta$ decreases. The drizzle process is weak for high $N_d$ and can not play a role in coupling the cloud and sub-cloud layers together.

Figure 8.3. A conceptual diagram of the daytime STBL when the LWP is not low, a) low $N_d$, b) high $N_d$. 
• For low $N_d$, cloud shortwave absorption is not as large as for high $N_d$. As $\Theta$ decreases, decoupling is not as strong and cloud thinning is not as great. The drizzle process is stronger for low $N_d$ and could play a role in coupling the cloud and sub-cloud layers together through generation of conditional instability in the sub-cloud layer.

• Over the diurnal cycle, for high $N_d$ the cloud layer thins more and becomes more decoupled in the presence of strong shortwave heating. The cloud layer is maintained throughout the day regardless of $N_d$.

Our daytime LESs of the BROKEN case revealed an important difference in how shortwave warming influences the STBL, as compared to the OVERCAST case (Fig. 8.4):

![Figure 8.4. A conceptual diagram of the daytime STBL when the LWP is low, low $N_d$. For high $N_d$ the cloud layer dissipates in the presence of strong shortwave heating.](image)

• When nighttime LWP is low, subsequent cloud thinning related to shortwave warming leads to decreases in cloud fraction and concurrent reductions in longwave cooling. Reductions in longwave cooling with $\Theta$ can be larger than concurrent increases in shortwave warming.

• These changes in longwave and shortwave radiative heating lead to steep reductions in STBL circulation strength and entrainment rate, and lead to strong decoupling as $\Theta$ decreases.
• As $\Theta$ decreases for high $N_d$, partially decoupled circulations weaken further, formation of cloud through STBL cooling decreases and the cloud dissipates through entrainment drying.

• For low $N_d$, as $\Theta$ decreases radiatively driven circulations also weaken further and formation of cloud through STBL cooling decreases. Because circulations are stronger and entrainment is not as vigorous as when $N_d$ is high, the cloud layer is maintained against entrainment drying for a wider range of $\Theta$.

• Over the diurnal cycle low LWP clouds in the STBL are more likely to dissipate under strong shortwave warming when $N_d$ is high. For low $N_d$ the cloud layer can be maintained throughout the day.

Finally, we tested the impact on STBL dynamics of using the ICA in shortwave radiative heating rate computations. To accomplish this we coupled a radiative transfer solver to RAMS that accounts for horizontal transport of shortwave radiation (Cole, 2005). We simulated a portion of the diurnal cycle of the STBL with and without the ICA for both the OVERCAST and BROKEN cases. We ran three pairs of LES, with and without the ICA, and found that the ICA had no significant impact. We emphasize both the tentativeness of this conclusion and the need for further testing. Furthermore we suggest testing the impact of the ICA on the modeling of other cloud systems such as altocumulus or cirrus.

8.2 Future Research

Although we have learned a great deal about cloud-radiation interactions in the STBL, there are many questions yet to be answered. The importance of cloudtop longwave radiative cooling in the STBL is well known (e.g. Lilly, 1968), and in LES a large proportion of longwave radiative flux divergence occurs near cloudtop (e.g. Stevens et al., 2005; Ackerman et al., 2009). Is the vertical distribution of shortwave radiative heating as important as it is for longwave radiative heating? Are model simulations of the STBL dependent on the vertical profiles of solar radiative warming or simply the integrated solar radiative flux divergence (i.e., the total absorbed energy)?
The importance of low-altitude low LWP clouds has been realized only recently (e.g. Turner et al., 2007), and further study of processes surrounding these low LWP clouds is recommended (e.g. Lee et al., 2009). For low LWP clouds in the STBL we found differences in radiative heating profiles between updraft and downdraft regions. Are model simulations of the STBL dependent on this horizontal inhomogeneity in radiative heating? If we applied a domain-averaged vertical profile of radiative heating to all model columns we might save computational processing time, but would this procedure elicit a different model response? How about placing the total domain radiative flux divergence in just updraft or downdraft regions?

We have focused on the domain-averaged STBL dynamical response to radiative heating. How does the physical structure of STBL circulations change as $\Theta$ changes? For example, because circulation strength decreases as $\Theta$ decreases, corresponding slower downdraughts might broaden in response (Moeng and Schumann, 1991). How might this change correspondingly influence the updrafts? How are convergence regions below cloud and divergence regions at the STBL interface altered when $\Theta$ increases (Schubert et al., 1979)?

Using currently available numerical models, the above questions are answerable to some degree and should be investigated. Another path of study we recommend involves radiative heating and entrainment at the STBL interface. Our own difficulties in relating entrainment to other quantities in our LESs suggest that we need to better understand how radiative processes interact with and modify the entrainment process. To our knowledge no one has attempted to untangle radiative heating and entrainment using a modeling framework. In the year 2000, Jerry Harrington and Bjorn Stevens proposed a series of numerical simulations to gain understanding of how differential heating, such as from radiative transfer, modifies fluid flow with and without an entraining interface (Harrington 2007, personal communication). Such a set of simulations would further the community’s understanding of entrainment in the radiatively driven STBL, a well-studied but poorly understood process (Stevens, 2005).

Finally, numerical models can only be used as a tool to investigate the potential impact of radiative heating in the STBL. Observations of radiative heating throughout the STBL, along with observations of the STBL response to radiative
heating, would allow us to evaluate and revise the conclusions above.

Observations of cloud properties can be used to better constrain the radiative heating profiles computed by models (e.g. McFarlane et al., 2008). In our introduction we noted that observations of radiative flux divergence through stratus cloud layers are notoriously difficult to obtain (Curry, 1986), and this lack of measurements calls into question the ability of radiative transfer models to accurately represent radiative heating within cloud layers. A large portion of longwave radiative cooling occurs within the top 10 m of a stratus cloud and to our knowledge there are no observations of these rapid irradiance changes with height near cloud-top. Obtaining such detailed longwave irradiance measurements using aircraft would be difficult indeed.

In the proposed field campaign Arctic Lower Troposphere Observed Structure (ALTOS), a tethered balloon instrument platform will be used to obtain in-situ observations of the Arctic cloudy boundary layer. Longwave irradiance measurements will be taken as the balloon is raised and lowered through a low-level cloud layer (Harrington 2009, personal communication). Co-located with thermodynamic and cloud property observations, these measurements of longwave irradiance can be used to evaluate the performance of radiative transfer models in computing longwave radiative heating within the STBL.

How might we be able to observe the STBL response to radiative heating that our simulations suggest? The Millimeter Cloud Wavelength Radar (MMCR) may prove a useful instrument in observing elements of this response. For example, our simulations suggest that a drizzling STBL might remain better coupled to the surface during daytime than a non-drizzling STBL. Observations of drizzle occurrence and the variation of $w'w'$ with height, and how they vary with the diurnal cycle, could be used to evaluate this assertion.

The MMCR has been used to obtain velocity spectra within cloud layers, and can be used to separate cloud droplets from drizzle (Serpetzoglou et al., 2008). With MMCR observations of the STBL we could determine a) the occurrence of drizzle, and b) the vertical velocity variance ($w'w'$) inside the cloudy region of the STBL. Ghate (2009, personal communication), in a separate study, recently derived these particular quantities with MMCR observations of stratocumulus layers at the Atmospheric Radiation Measurement Program (ARM) Southern Great Plains
(SGP) site. Effective use of such observations could lead to helpful evaluation of the conclusions herein.
References


Vita

Jonathan L. Petters

Jonathan Petters was born in Raleigh, North Carolina in 1978 where he spent his formative years. He studied meteorology at North Carolina State University, and received his Bachelor of Science in August 2000 after a year studying abroad at the University of Hull in the United Kingdom. He then continued studies at North Carolina State and received his Master of Science in Atmospheric Sciences in August of 2002. Upon completion of his M.S. he accepted a scientific modeling support position as a contractor for the US Environment Protection Agency in the Research Triangle Park, North Carolina. After two years of employment he decided that six years of college education was really not enough, and he applied to doctoral programs in atmospheric science across the country. He accepted an assistantship under Dr. Eugene Clothiaux at the Pennsylvania State University in August of 2004, and worked closely with both Dr. Clothiaux and Dr. Jerry Harrington.

Jonathan has accepted a postdoctoral position, to begin in January 2010, with Dr. Patrick Chuang at the University of California at Santa Cruz.