UNIVERSITY OF CALIFORNIA
Los Angeles
College of Letters and Science

Large eddy simulation study of the effect of drizzle on the dynamics of the stratocumulus topped boundary layers

A thesis topic proposal
as a part of the requirements for the degree of
Doctor of Philosophy in Atmospheric and Oceanic Sciences

by

Verica Savic-Jovcic

November 30, 2004

1 Introduction

Stratocumulus clouds (Sc) are a significant component of the Earth's climate system. By increasing the albedo and negligibly affecting the outgoing long-wave radiation, Sc cool the underlying surface. Additionally, because Sc cover large areas (about a third of the Earth's oceans) this cooling effect has global implications. Moreover, the impact on the climate is emphasized with an abundance of Sc over subtropical oceans, where the Earth annually receives large amounts of incoming solar radiation. Modeling studies suggest that the global cooling resulting from a modest increase in the Sc global coverage could offset the expected warming from doubling the concentration of carbon dioxide in the atmosphere (Slingo, 1990). The climatological importance of Sc encourages a fundamental understanding of the processes involved in the formation, sustainability and dissipation of these clouds. Steady progress has been made in this direction during the past half a century. However, many questions remain.

One such question is related to the appearance of areas of relatively cloud-free air embedded in an otherwise homogeneous cloud field. As satellite images in Figs. 1 and 2 depict, these 'clearings' (sharply bounded darker regions on the figures) actually represent clouds with a different organizational pattern. While Sc, the high albedo region, can be characterized as organizing in a closed cellular pattern, with brighter (thicker) clouds at the center of the cell and dimmer (thinner) or no cloud at the cell edges, the 'clearings' are more characteristic of an open cellular pattern where clouds define the cell boundaries and the cloud-free regions the cell center. Although these formations have attracted the interest of satellite meteorologists for a long time (see reviews by Agee *et al.*, 1973; Garay *et al.*, 2004), the analysis of *in situ* data collected in their vicinity has only recently been conducted (Stevens *et al.*, 2004). To emphasize both the compact structure of these features in reference to the surrounding Sc decks and their distinctive convective organization Stevens *et al.* (2004) named them 'pockets of open cells (POCs)', a terminology we adopt here. *In situ* data suggest a link between POCs and precipitation (Stevens *et al.*, 2004; vanZanten *et al.*, 2004), but the nature of the link and other possible processes which initiate and maintain POCs are still objects of speculation.

Precipitation (in the form of drizzle) is an interesting phenomenon in its own right. Drizzle can reduce both the cloud thickness and the horizontal continuity (Pincus and Baker, 1994; Albrecht, 1989), which are essential for the Sc albedo. Through this process, drizzle plays a key role in the second indirect effect of aerosols on climate. Furthermore, drizzle also

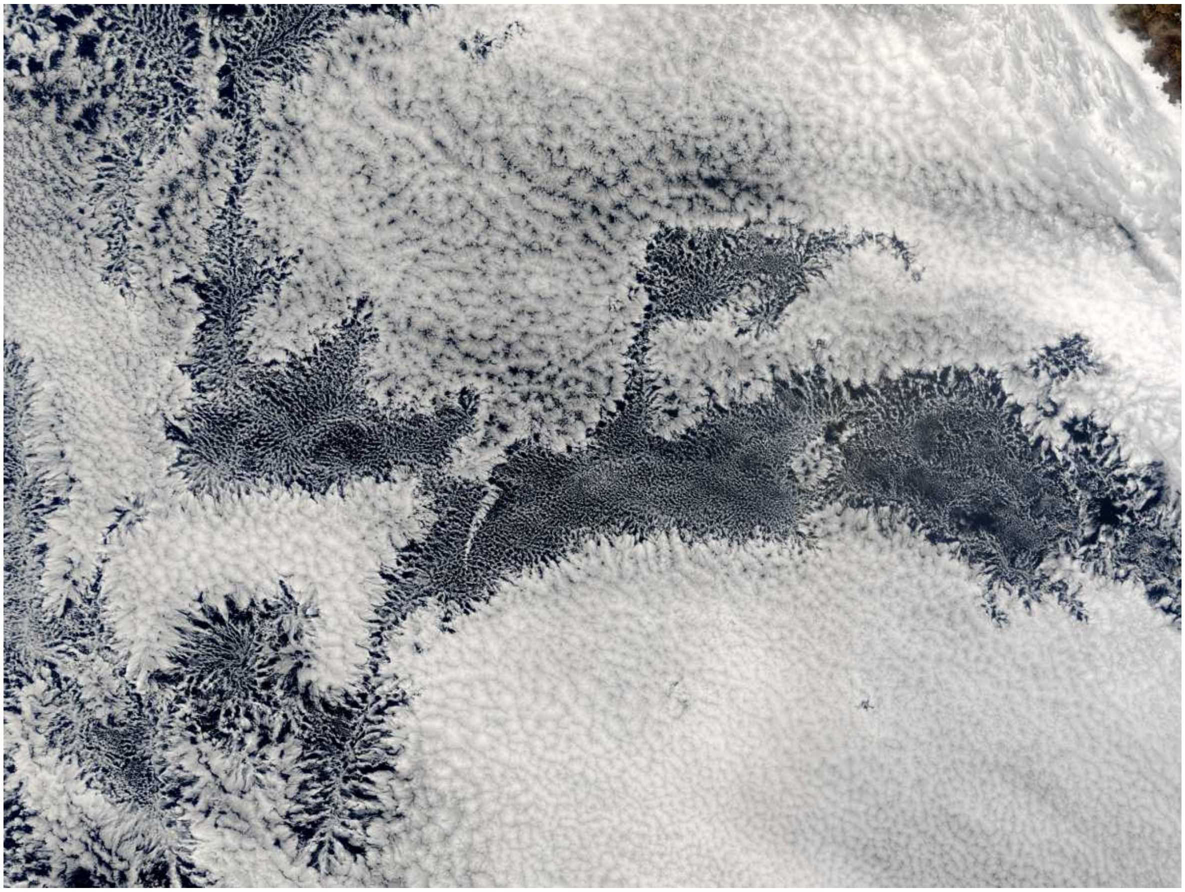

Figure 1: Example of POCs (darker regions) embedded in the broader regions of Sc cloud deck (lighter regions) off the coast of Peru.

affects the low-level flow characteristics by changing the thermodynamic properties of the stratocumulus topped boundary layer (STBL). Previous studies suggested that drizzle causes a transition from Sc- to Cumulus-type of dynamics (Paluch and Lenschow, 1991; Wang and Wang, 1994; Stevens *et al.*, 1998), but no consensus as to the mechanism of that change has been established.

The idea that drizzle can help initiate and maintain POCs is supported by the theoretical work of Wang and Wang (1994) and Stevens *et al.* (1998), who argued that drizzle causes the transition of Sc-type convection to the cumulus-type. Nevertheless, these studies have some limitations and warrant further study. In particular, Wang and Wang (1994) used a model that does not resolve the full 3D dynamics, whereas Stevens *et al.* (1998) used a 3D model,

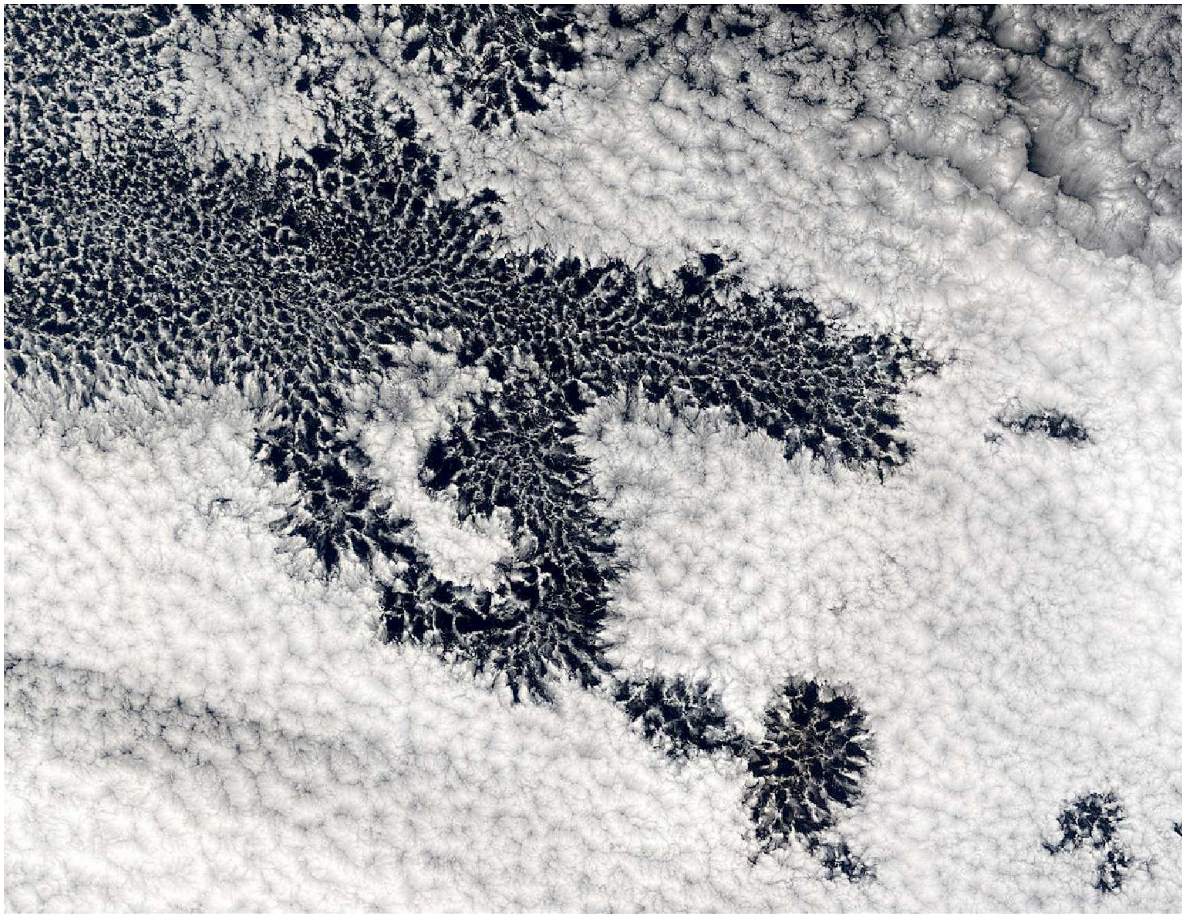

Figure 2: More detailed view of the cellular structure of POCs. A region of open cellular convection has dark cell interiors, with bright cell walls, whereas the closed cellular convection of Sc has bright cells with darker cell walls. (from Stevens, 2004)

but on a domain that is too small to allow open cells to emerge.

In the study proposed here, we would like to shed more light on the relationship between drizzle and the dynamics within the STBL, which is closely related to the morphology and organization of the clouds within the planetary boundary layer. As a start, we would like to explore if the drizzle induces circulation changes. The results of this initial work would be a basis for examining what particular aspect of drizzle affects the dynamical evolution of the layer, as well as exploring the mechanism of the formation and longevity of POCs. Our long-term goal is to achieve a better understanding of the differences in the dynamics between the precipitating and non-precipitating cloud topped boundary layer, as well as mechanisms for initiating POC-like features in the STBL.

Our method of addressing the above issues is to use large eddy simulations (LES) of the idealized nocturnal STBL. LES three-dimensionally resolves the most energetic eddies in the flow, and therefore requires the least assumptions regarding the flow (the parameterizations are for the sub-filter scale, microphysical and radiative processes). The particular LES that will be used in this study is the UCLA LES (the latest version is described in Stevens *et al.*, 2004). It has recently been configured to permit very large domain integrations capable of representing the scales of motion evident in the development of the open cellular convection that compose POCs. The previous version of this LES, which was restricted to a smaller domain, was the basis for the currently most detailed study of the dynamics of the drizzling STBL, *i.e.*, Stevens *et al.* (1998), which gives us confidence in the application of this LES for our study. To reduce the computational costs and focus on the effect of drizzle, rather than drizzle formation, we will parameterize drizzle using simple models and empirical relationships between drizzle rates and cloud depth in part motivated by the observations analyzed by vanZanten *et al.* (2004).

In the following, we review the ideas pertinent to this study. In section 2 the focus will be on the processes that determine the structure of the STBL, the impact of drizzle on the STBL, and description of POCs, where we will refine the goals that are outlined here. In section 3 we provide the description of the LES and in section 4 the experimental setup of the simulations. In section 5 we describe our plans for future work and in section 6 the proposed timeline for the study.

2 Background

Stratocumulus are low-level clouds with a generally stratiform appearance and underlying cellular structure. They develop at the top of thermodynamically distinct maritime atmospheric boundary layers. Such boundary layers form in conditions where the overlying free troposphere is much warmer than the underlying cold ocean. Tending to be more similar in characteristics to the latter they are often capped by a strong temperature inversion. These conditions are typically met in the eastern regions of subtropical oceans. Here the upwelling in the ocean brings cold water to the surface and in the atmosphere subsidence enhances the warmth of the overlying air contributing to the thermal contrast between the ocean and the overlying atmosphere. Klein and Hartmann (1993) used the difference between the potential temperature, θ, at 700 mb and its value at the surface, to quantify the ocean-atmosphere thermal contrast. They called this difference the lower tropospheric stability (LTS), and showed that the Sc prevalence correlates well with the LTS on the seasonal and interannual time scales. In addition, Klein *et al.* (1995) suggested that the local cloud amount is better correlated with the LTS 24 h upwind than with the local LTS, which indicates the importance of cold advection not only for the Sc formation (which was suggested by Paluch and Lenschow, 1991) but also for their sustainability. The correlation with the upwind conditions also indicates the existence of memory in the system, which could play a role in the POCs formation as well.

2.1 Typical Mean Structure of the STBL

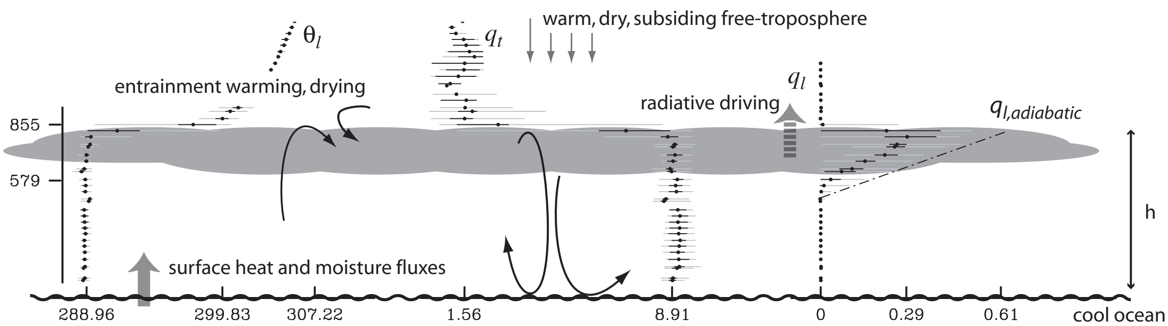

Figure 3: Cartoon of well mixed, non-precipitating, stratocumulus topped boundary layer, overlaid with profiles of θ_l, q_t and q_l. The profiles, as well as the heights of cloud base and top, are constructed from data from RF01 of DYCOMS-II. The dash-dot line represents the adiabatic liquid water content. (from Stevens, 2004)

Most of the conceptual and theoretical descriptions of the STBL involve a well mixed, radiatively driven and non-precipitating STBL, as depicted in a cartoon in Fig. 3. The cartoon illustrates the mean structure, environmental conditions and the most important processes that occur within the STBL. Data presented in the cartoon were collected during the first research flight (RF01) of DYCOMS-II (the second DYnamics and Chemistry Of the Marine Stratocumulus field study described in detail in Stevens *et al.*, 2003). Displayed are adiabatic invariants (liquid water potential temperature, $\theta_l \sim \theta \exp(\frac{-q_l L}{c_p T})$, which describes the thermal structure, and total water mixing ratio, q_t, which represents the moisture content within the layer; L, c_p and T are the latent heat of vaporization, isobaric specific heat and temperature, respectively), as well as the liquid water mixing ratio, q_l, which indicates the presence of the cloud.

As shown in Fig. 3, the STBL is relatively shallow, cool and moist, and capped by a warmer, drier and gently subsiding free atmosphere. The transition of θ_l and q_t between their boundary-layer and free-tropospheric values is sharp, with a strong increase in temperature (temperature inversion) and decrease in moisture and liquid water content. The inversion, which is very stable, acts against mixing of the STBL with free-tropospheric air and is the largest constituent of the lower tropospheric stability defined by Klein and Hartmann (1993) and discussed above.

Profiles of θ_l and q_t in Fig. 3 show that the STBL is vertically well mixed, which results from convective turbulence within the STBL. The main source of turbulence in the STBLs is the infrared radiative cooling at the top of the cloud, unlike in the dry convective boundary layers (DCBLs), where the dominant source of turbulence is the surface heat flux. This essential difference in the sources of the turbulent motion leads to the difference in the peak time for the maximum strength of turbulence between the two regimes. In the STBL turbulence is at its maximum during the night, when the cooling is the strongest due to the lack of offset from the solar radiation, while for the DCBL it peaks in the daytime, when the land surface is warmest. Many modeling and theoretical studies have taken advantage of the absence of the solar radiation in the nocturnal conditions preferred by the STBL, while most of the observations were performed during the day. This discrepancy between the observations and theory has recently been bridged during DYCOMS-II, whose data are presented in the cartoon.

Basic processes within the STBL are also illustrated in Fig. 3. In addition to mixing of STBL air, turbulent motions within the STBL entrain quiescent free tropospheric air into the STBL by engulfing and subsequently mixing it into the STBL air. From the perspective of the STBL mass budget, the diabatic growth of the STBL by entrainment counteracts the large-scale subsidence. From the heat budget point of view, entrainment warming competes with the cloud-top cooling and surface heat fluxes. As for the moisture budget, entrainment acts against the surface moisture fluxes and dries the STBL. Note that both the source (radiative cooling) and the sink (entrainment) of turbulence act at the same interface – cloud top – which makes the study of the STBL challenging in many respects.

Surface heat and moisture fluxes are additional sources of turbulence, but generally less important. As we will see later, when these sources become important, our view of the well mixed STBL becomes questionable.

2.2 Mixed-layer Framework

The well-mixed state of adiabatic invariants within the STBL is a foundation for our theoretical understanding of the STBL. Mixed layer theory, originally developed by Lilly (1968) and with only modest elaborations in the meantime, is still providing advantageous insights in the STBL properties. We use it too, as a starting point upon which we build further in describing the effect we expect for drizzle to have on the STBL.

In mathematical terms, the state of the STBL can be described simply in terms of the height of the layer (h), and the layer mean (bulk) values of θ_l and q_t, which we denote by a hat. The evolution of these quantities satisfies the following equations:

$$\frac{dh}{dt} = W + E \tag{1}$$

$$\frac{d\hat{\theta}_l}{dt} = \frac{1}{h}[V(\theta_{l,0} - \hat{\theta}_l) + E(\theta_{l,+} - \hat{\theta}_l) - \Delta F_{\theta_l}] \tag{2}$$

$$\frac{d\hat{q}_t}{dt} = \frac{1}{h}[V(q_{t,0} - \hat{q}_t) + E(q_{t,+} - \hat{q}_t) - \Delta F_{q_t}]. \tag{3}$$

Here: $W = -Dh$ is the large-scale subsidence at the top of the layer; D is the large-scale divergence, which is assumed to be independent of height within the boundary layer; E is the entrainment rate, which is used to parameterize the entrainment fluxes; V is the surface exchange velocity, which parameterizes the surface fluxes; ΔF_{θ_l} and ΔF_{q_t} are the total diabatic flux divergences across the layer depth and are related to the radiation and drizzle.

In the equations (1)-(3), which represent the mixed-layer model (MLM), all the parameters, except perhaps E, can with confidence be expressed in terms of the large-scale or bulk quantities. For instance, ΔF_{θ_l} and ΔF_{q_t} are related to the cloud depth through the liquid water path, which is straightforward to be determined in the MLM because the cloud top coincides with the top of the STBL and the cloud base is at the height where $q_t = q_s$ and $q_l = 0$, with q_s being the saturation water vapor mixing ratio that depends on the temperature and pressure.

The parameterization of E, necessary to close the system, has been a topic of ongoing research ever since the first formulation of the MLM. Most of the attempts to parameterize entrainment are related to the ability of the system to do work, the stability of the capping inversion layer and the effect of non-turbulent processes in deepening the layer (Stevens, 2002):

$$E = \mathcal{A}\frac{\mathcal{W}}{\Delta_{-}b} + \mathcal{D}. \tag{4}$$

Here, $\mathcal{A}$ is an efficiency factor that depends on the state of the STBL, $\mathcal{W}$ is a working rate that depends on the forcing, $\Delta_{-}b = \frac{g}{\theta_0}\Delta_{-}\theta_v$ is an isentropic buoyancy jump across the interface, that is used to measure the interface stability, and $\mathcal{D}$ represents non-turbulent processes (*e.g.*, radiative cooling of the air just above the inversion layer); also, $\theta_v = \theta(1 + 0.608q_v - q_l)$, q_v and θ_0 are the virtual potential temperature, water vapor mixing ratio and reference potential temperature, respectively. To the extent that buoyancy dominates the energetics of the STBL, the rate of working can be measured by the vertically integrated buoyancy flux, $\hat{\mathcal{B}} = \frac{g}{\theta_0}\int_0^h \overline{w'\theta_v'}dz$. The complexity of the problem, however, arises from the dependency of $\hat{\mathcal{B}}$ on the properties of the entrainment interface. The easiest way to explain the entrainment parameterization, then, would be to express $\mathcal{W}$ as the amount of work that the system can do if driven only by the surface fluxes and the radiative cooling at the top, without any entrainment. In this case we can expect $\mathcal{A} < 1$ because the system does work as it entrains. From this point of view, the question is how to parameterize $\mathcal{A}$. Another complexity arises from the presence of the condensate just below the entrainment interface, due to its evaporation in the process of mixing. If evaporation is strong enough (*i.e.*, evaporative cooling exceeds the entrainment warming), a mixed parcel becomes negatively buoyant, which is known as a buoyancy reversal. In this case evaporation can be seen as an additional source of turbulence kinetic energy, and entrainment can be a source rather than a sink of energy. Many authors have related the buoyancy reversal to the instability of the cloud top interface (known as a cloud top entrainment instability, or CTEI) and thus proposed it as a mechanism for breaking up the Sc deck (*e.g.*, Lilly, 1968). However, more recent studies (*e.g.*, Turton and Nicholls, 1987; Bretherton and Wyant, 1997; Wyant *et al.*, 1997) associate the Sc break up with decoupling, a process we discuss further below.

If the time scale of variation in the forcings is long compared to the turnover time scale within the STBL, the well mixed STBL tends to stay well mixed and therefore be characterized as *quasi-steady* state. A quasi-steady state is characterized by time invariability of the profiles of conserved variables (*e.g.*, in case of θ_l: $\partial_t\partial_z\overline{\theta_l} = 0$). For a horizontally homogeneous flow ($\partial_t\overline{\theta_l} = \partial_z(\overline{w'\theta_l'} + F_{\theta_l})$), this implies a linear profile of the sum of diabatic and turbulent fluxes of conserved quantities ($\partial_z\partial_z(\overline{w'\theta_l'} + F_{\theta_l}) = 0$). Therefore, in the MLM, given a knowledge of the diabatic forcings, F, and turbulent fluxes at the flow boundaries, quasi-stationarity determines the profile of the turbulent fluxes of θ_l and q_t ($\overline{w'\theta_l'}$ and $\overline{w'q_t'}$).

Furthermore, because the state variables determine the buoyancy ($b \approx \theta_v' \,/\, \theta_0 \approx \alpha\, \theta_l' \,/\, \theta_0 + \beta\, q_t$), their turbulent fluxes also determine the turbulent buoyancy flux, $\mathcal{B}$. However, the presence of the condensate once again complicates the problem. In moist unsaturated air, the thermal effect on buoyancy is more important than the moisture effect, whereas in saturated air, the moisture becomes more important due to the release of latent heat by the phase change. In the expression for $\mathcal{B}$:

$$\mathcal{B} = \overline{w'b'} = g \begin{cases} \alpha_u(\overline{w'\theta_l'}/\theta_0) + \beta_u \overline{w'q_t'} & q_t < q_s, \\ \alpha_s(\overline{w'\theta_l'}/\theta_0) + \beta_s \overline{w'q_t'} & q_t \geq q_s, \end{cases} \tag{5}$$

these effects are isolated in the partial derivatives α_u, α_s, β_u, and β_s (where $\alpha = \frac{\partial \theta_v}{\partial \theta_l}$ and $\beta = \frac{\partial \theta_v}{\partial q_t}$), which are functions of state and can be determined analytically. For shallow flows, the partial derivatives can be approximated to be constant, and for $\theta_l = 288$ K and $q_t = 10$ g kg^{-1} they have the following values: $\alpha_u = 1.06$, $\alpha_s = 0.608$, $\beta_u = 0.49$, and $\beta_s = 3.3$ (Stevens, 2004). As the values for the partial derivatives show, the thermal effect on buoyancy in the cloud is reduced almost to the half of its subcloud value, whereas the moisture effect is increased almost an order of magnitude. Also in (5), g is the gravitational acceleration and θ_0 is the reference potential temperature.

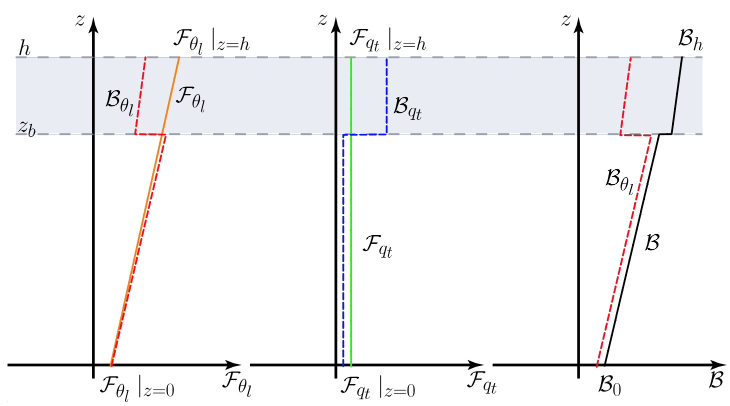

Figure 4: Cartoon of turbulent fluxes in the non-precipitating and weakly-entraining STBL. $\mathcal{F}_{\theta_l} = g\,\overline{w'\theta_l'} \,/\, \theta_0$ and $\mathcal{F}_{q_t} = g\,\overline{w'q_t'}$ are in the buoyancy units. $\mathcal{B}_{\theta_l}$ and $\mathcal{B}_{q_t}$ are the turbulent buoyancy fluxes associated with the fluxes of θ_l and q_t, respectively, while $\mathcal{B}$ is the sum of the two.

The effect of the cloud presence on the buoyancy flux profile in the weakly-entraining and non-drizzling STBL is illustrated in a cartoon in Fig. 4. As discussed above and depicted in the cartoon, the STBL in the quasi-steady state has a linear sum of turbulent and diabatic fluxes of the state variables. As the flow is mainly driven by the radiative cooling, the whole layer is cooling, while for the moisture content we have assumed stationarity, *i.e.*, a balance between the surface moistening and evaporative drying. The resulting buoyancy flux profile is therefore positive throughout the whole layer, with the jump at the cloud base:

$$\Delta_{z_b}\mathcal{B} = \mathcal{B}_{z_{b_+}} - \mathcal{B}_{z_{b_-}} = g[(\alpha_s - \alpha_u)\,\overline{w'\theta_l'}/\theta_0 + (\beta_s - \beta_u)\,\overline{w'q_t'}] = \\ = g[-0.452\,\overline{w'\theta_l'}/\theta_0 + 2.81\,\overline{w'q_t'}], \tag{6}$$

where $\mathcal{B}_{z_{b_+}}$ and $\mathcal{B}_{z_{b_-}}$ are the buoyancy fluxes just above and just below the cloud base, respectively, with z_b being the height of the cloud base. The buoyancy jump is related to the release of latent heat at the cloud base, and therefore tightly coupled to the upward moisture flux. This is also confirmed in Eq. (6), which implies that in the buoyancy driven STBL, for the buoyancy jump to be positive $\overline{w'q_t'}$ has to be positive as well.

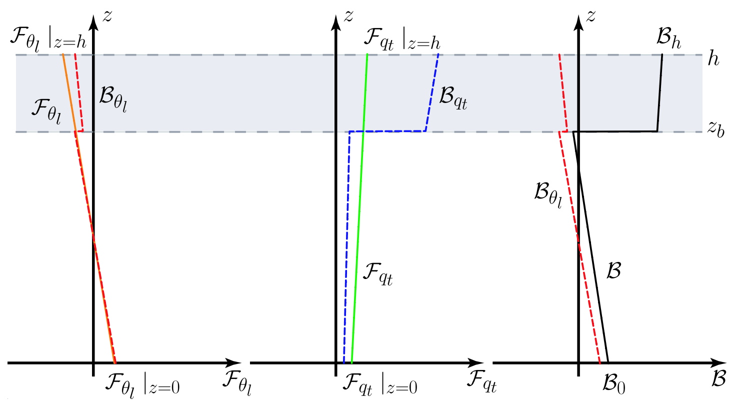

Figure 5: Cartoon of turbulent fluxes in the non-precipitating and strongly-entraining STBL. The symbols are as in Fig. 4.

If the entrainment is strong enough, the layer may actually warm. For the STBL to stay well mixed then, the buoyancy flux just below cloud base could become negative, as in Fig. 5. The negative buoyancy flux below the cloud base can be interpretated as a need for the cloud layer to do work on the subcloud layer to keep the well-mixed thermodynamic state and supply the moisture to the cloud. In these terms, the cloud has only a limited ability to do such work. This was once a motivation for another class of entrainment closure. These 'flux limiting' closures actually allow entrainment only until the ratio of vertically integrated net negative buoyancy flux to the vertically integrated positive buoyancy flux reaches some prescribed limit (Lilly, 1968; Schubert *et al.*, 1979).

2.3 Decoupling

One could ask what happens to the STBL when the rate of entrainment warming requires a negative buoyancy flux over a substantial height at the top of the subcloud layer, as in Fig. 6. This question has been addressed in numerous studies, which suggested that the STBL decouples and the Sc decks break up in this scenario. For instance, Turton and Nicholls (1987) and Bretherton and Wyant (1997), who studied the diurnal and deepening-warming decoupling, respectively, with mixed-layer models, suggested that if the negative buoyancy flux in the subcloud layer overcomes the cloud ability for downward mixing, Sc either thins (Turton and Nicholls, 1987) or transitions to Cu (Bretherton and Wyant, 1997). The difference in their

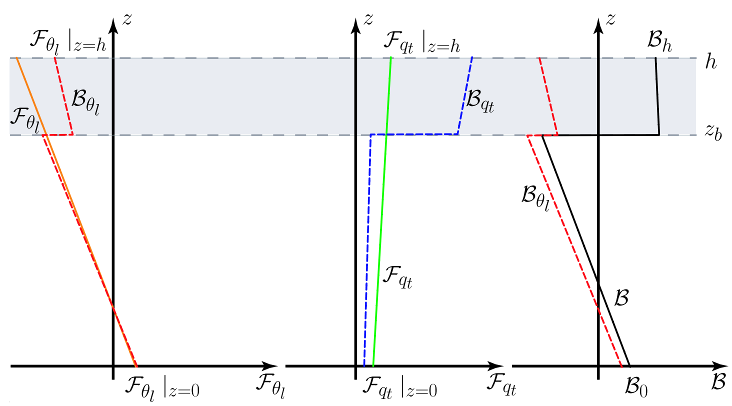

Figure 6: Cartoon of turbulent fluxes in the decoupled STBL. The symbols are as in Fig. 4.

results follows from the difference in the sources of decoupling. In the Turton and Nicholls (1987) study, the cloud top radiative cooling gets offset by the short-wave cloud warming and therefore the main source of turbulence gets reduced, while the entrainment continues, which causes the evaporation of the cloud. In the Bretherton and Wyant (1997) study, however, the surface fluxes, particularly the moisture flux, become additional sources that enhance the turbulent motion and therefore the entrainment, which then overcomes the radiative cooling and enhances the warming over the whole layer, providing the conditions for more vigorous Cu dynamics.

Although the sources of decoupling and the final state of the cloud are different in these studies, they both recognize the importance of the buoyancy flux and diagnose the decoupling in a similar manner. Turton and Nicholls (1987) introduced the threshold for the ratio of vertically integrated buoyancy flux in the subcloud layer to the vertically integrated buoyancy flux in the cloud layer to diagnose the separation of layers. They estimated the threshold to be -40%. Bretherton and Wyant (1997), though, modified the metric to the 'buoyancy integral ratio' (BIR), which is the negative ratio of vertically integrated negative buoyancy flux in the subcloud layer to the vertically integrated positive buoyancy flux throughout the rest of the boundary layer. The BIR threshold they used is 15%. However, an LES study by Stevens (2000) suggests that some degree of decoupling is present for any positive value of BIR if the decoupling is measured by the reduced variance of vertical velocity, or by the development of vertical gradients of scalars. Moreover, the study suggests that the mixed-layer models with values of BIR higher than 10% are ill defined, as that value denotes a pronounced development of a two layer structure. Stevens (2000) even suggests that this threshold value can be used in the general circulation models for distinguishing the boundary layer regimes between Sc-topped and Cu-coupled (including Cu-under-Sc) boundary layers. Regardless of the value of the threshold, this body of literature provides the basis for understanding Sc, decoupling and the transition to Cu. It also provides the framework for presenting and discussing results of future studies.

2.4 Drizzle

One of the processes thought to play an important role in the life cycle of Sc, but not presented in Fig. 3, is precipitation in the form of drizzle. Although it has been recognized that drizzle is abundant and that it affects cloud microphysics, cloud morphology, and radiative processes, as well as the stability of the STBL, drizzle has often been neglected or treated marginally by the modeling community. Therefore the question of how these effects relate and how drizzle interacts with other processes acting in the STBL is still open.

Studies that focus on the influence of drizzle emerged after the *in situ* measurements showed that drizzle is abundant and that the drizzle flux can be comparable to the turbulent fluxes of liquid water (Brost *et al.*, 1982; Nicholls, 1984). For instance, Paluch and Lenschow (1991) invoked drizzle in explaining Sc-to-trade-Cu transition, whereas Ackerman *et al.* (1993) related drizzle to the dissipation of Sc through the collapse of the STBL. Furthermore, Pincus and Baker (1994) argued that drizzle restricts the depth of Sc, while Stevens *et al.* (1998) showed that in the presence of drizzle Sc transform into Cu, but without an increase in the boundary-layer depth that is characteristic for a trade-Cu boundary layers. In addition to the different effects, each of these studies also offered different mechanisms through which drizzle acts upon the STBL.

A conceptual model of the drizzle-induced cloud transformation developed by Paluch and Lenschow (1991) implies that drizzle interacts with the STBL turbulence mainly by stabilizing the subcloud layer through moistening and cooling, and that further development of trade Cu is a consequence of the surface heating. Their study is based on *in situ* measurements, and thus would benefit from further studies using the experiments with controlled conditions, where the role of individual processes can be isolated.

The drizzle-related Sc dissipation proposed by Ackerman *et al.* (1993) involves the drizzle interaction with the cloud microphysical, and subsequently radiative, properties, which causes the stabilization of the cloud layer. By stabilizing the cloud layer, as Ackerman *et al.* (1993) argue, drizzle allows the subsiding air to push the capping inversion downward, leading to the collapse of the STBL. However, although Ackerman *et al.* (1993) used a very sophisticated cloud microphysical model, the dynamics of the STBL in their study was represented by a 1D turbulent closure model that does not resolve the dynamics of the boundary layer, and hence invites for further study using a model that would allow full development of the dynamics.

The drizzle-induced decrease of the cloud depth is argued by Pincus and Baker (1994) to be due to the decrease of the cloud-top height. As they argue, by affecting the energy budget drizzle reduces the entrainment rate and the STBL height. In their study Pincus and Baker (1994) used a mixed-layer model, which provides considerable insights, but also has some limitations. For instance the entrainment parameterization is still a topic of an ongoing research, and MLM results are sensitive to the choice of the entrainment parameterization (Stevens, 2002). Pincus and Baker (1994) used one of the 'flux limiting' schemes, and therefore prevented the model from developing the decoupled state, without understanding if it might happen in the nature. Nevertheless, their result that the cloud albedo susceptibility is affected by the propensity of cloud to drizzle is in agreement with observations, which leaves this problem open for further study, preferably including the full 3D STBL dynamics.

The drizzle-caused transition from Sc to Cu is shown by Stevens *et al.* (1998) to be mainly through the stabilizing effect of drizzle on the buoyancy flux. In their LES study Stevens *et al.* (1998) showed that, in addition to the cooling and moistening of the subcloud layer, drizzle significantly increases the buoyancy of the downdrafts within the cloud layer. Because this stabilizing effect on downdrafts is apparent only if the level of saturation of the parcels within the downdrafts is above the cloud base, they named it 'potential buoyancy'. Their result can also be interpreted as the decoupling, since the increase of the downdraft buoyancy suggests the decrease of the buoyancy flux below the mean cloud base. However, the question of the relative contribution of processes within the cloud versus subcloud layer remains open, as does the question of whether decoupling theory can be used to explain the results of the simulation.

A recent observational study of vanZanten *et al.* (2004) confirmed the consistency of the theoretical studies with the observations regarding the drizzle-induced cloud transformation. Their analysis of *in situ* and radar data showed that drizzle is prevalent, and that the heat flux equivalent to the drizzle rate is comparable to the cloud-top radiative forcing. Furthermore, vanZanten *et al.* (2004) observed that the time scale for drizzling regions is at least an hour, and that relatively rare, but intense, drizzle events contribute disproportionately to the overall distribution. Therefore they concluded that the observations imply that the greater the value of the mean drizzle rate, the more likely it is that drizzle covers small spatial areas, which is consistent with the idea of drizzle-induced cloud transformation.

An underlying idea of these studies is that drizzle can cause the decoupling of the well-mixed STBL, since drizzle represents an additional forcing of the STBL energetics. In the mixed-layer framework, the sum of the diabatic and turbulent fluxes of conserved quantities has a linear profile. Therefore, a presence of drizzle alters the turbulent buoyancy flux profile, through the effect on both θ_l and q_t turbulent fluxes. In addition to the positive jump in the buoyancy flux profile at cloud base, a region of buoyancy consumption of turbulence kinetic energy develops at the top of the subcloud layer (similar to Fig. 6). In other words, drizzle forces the STBL to do work to mix cloudy parcels downward and clear air upward so as to maintain a well mixed state. Decoupling is thought to occur when the necessary amount of work exceeds the amount that the STBL can provide.

These studies seem to disagree regarding the final effect of the drizzle-induced decoupling. Paluch and Lenschow (1991), Stevens *et al.* (1998) and vanZanten *et al.* (2004) argue for the cloud transformation, whereas Pincus and Baker (1994) and Ackerman *et al.* (1993) for cloud thinning and destruction. However, one could argue that the resulting arguments of the Ackerman *et al.* (1993) and Pincus and Baker (1994) studies can be due to the misinterpretation of the models with highly parameterized dynamics. Another study that supports the decoupling of the STBL leading to Cu formation is the study of Wang and Wang (1994). They argue that the moistening and cooling of subcloud layer is the most important effect, and that drizzle does not significantly reduce cloud liquid water content. Although they also used a 1D model, they could better discuss the distribution of the variables in addition to their mean values because they used a third order turbulence closure. Nevertheless, they could not reproduce 'heavy' drizzle, and the conclusions they made are for the 'weakly-drizzling' STBL.

As Turton and Nicholls (1987) and Bretherton and Wyant (1997) showed that the cloud warming due to solar radiation and an increase in the SST can cause the decoupling of the STBL, we would like to see if drizzle can be a source of decoupling as well. The above studies support the drizzle-induced decoupling, but the open questions are whether there is a threshold in the drizzle rate needed for the decoupling to occur, and if there is, how the other forcings influence this threshold. Furthermore, what particular aspect of drizzle induces the decoupling? Although the mixed layer theory and associated bulk energetics provide a framework for answering these questions, the LESs provide a way to check the theoretical predictions. Another question is the spatial distribution of drizzle. Is there a minimum area of drizzling STBL for decoupling? Furthermore, does the drizzling-induced decoupling contain the mechanism to spread out, or does it just get advected by the mean wind, or does the STBL have a mechanism to fill in the cloud layer that gets decoupled from the subcloud layer due to the effect of drizzle? These are the questions that we would like to address.

2.5 POCs

Compact regions of lower albedo embedded in the otherwise strongly reflecting Sc decks, seen on Figs. 1 and 2, namely POCs, are not just curious features that intrigue our imagination. Their appearance reduces the area coverage of Sc, and therefore reduces the cooling effect of Sc in the Earth radiative budget. Furthermore, as they seem to exist in the environmental conditions that are relatively similar to those that favor Sc, their appearance encourages asking what particular STBL attribute determines the cloud regime that develops. One could even ask if anthropogenic effects on atmospheric aerosol can play any role in their formation.

POCs were observed as soon as satellite imagery emerged, but there are no theoretical studies of them. The only study so far that addresses them by using the combination of the satellite, radar and *in situ* data is Stevens *et al.* (2004), which described them as spatially-compact, cellular-patterned, low-reflectivity regions embedded in the otherwise stratiform cloud fields. These observations suggest that drizzling areas of the sampled STBL are coinciding with the cell walls in the POCs, which invites theoretical study of the drizzling STBL using the models with domains large enough to allow the formation of POCs.

In addition to the visible images, POCs can also be detected from nocturnal satellite imagery by locating the small values of the difference between the 11 and 4 μm brightness temperature, which allows for the continuous detection of the evolution of POCs. Stevens *et al.* (2004) observed that POCs are coherent, long-lived (longer than 10 h) and advected by the mean boundary layer wind. Radar reflectivities and *in situ* data imply that the precipitation in the sampled STBL is localized in the walls of the open cellular convection and that the drizzle rate at the surface exceeds more than twice the surface evaporation, while the drizzle rate at the cloud base is twice to three times larger than the surface values. As this much precipitation can completely dry out the cloud in about 10 min, if not replenished, Stevens *et al.* (2004) concluded that POCs represent stable flow configurations that organize to maintain the moisture supply to the precipitating cell boundaries. However, a question of what makes them so long-lived, and self-sustaining remains open.

Another intriguing question that emerges from the Stevens *et al.* (2004) study is the air humidity in the vicinity of POCs. *In situ* data suggest that both the boundary layer and the free tropospheric air are moister in the vicinity of POCs. In the proposed study we cannot explore the source of the difference in the environmental conditions, but we would like to address its effect onto the STBL and look for the possible link between the surface and free tropospheric moisture.

3 LES

To conduct a comprehensive study of the interaction of drizzle with turbulence in the STBL one needs to employ a tool that fully resolves the flow characteristics in all three spatial dimensions, as well as in time. LES is thought to be a method that satisfies these requirements, because it resolves the energetics of the largest eddies in the flow, which are responsible for most of the transport of momentum, energy and mass. An example of successful application of LES is the Stevens *et al.* (1998) study, which suggested the drizzle-induced transition from Sc- to Cu-type dynamics. However, to expand our understanding of the necessary conditions and the mechanism of the transition, as well as the newly developed structures, there is a need to carry on with the implementation of the LES in the new studies. For instance, domain size needs to be increased to allow development of POCs, as the observations suggest a link between drizzle and POCs (Stevens *et al.*, 2004). The increase in the domain size raises new questions; *e.g.*, one has to develop ways for compensating between the increased computation time and need to have fine enough resolution to resolve the eddies that are in the inertial range, where the energy is only transported to the smaller, dissipation, scales.

LES solves the equations that describe the motion and the thermodynamic state of the fluid, as well as the evolution of the scalars of interest, on the scales larger than its grid-cell size. The equations for the fields resolved in the LES are derived by applying a low-pass filter to the original equation. The sub-filter scale (SFS) terms in these equations represent the effects of smaller scales onto the larger – that being the removal of energy – and they are either mimicked by using numerical schemes with sufficient numerical diffusion or damping, or accounted for by explicitly introducing physical assumptions, namely making SFS closures.

The particular LES that will be used in this study is the UCLA LES. Initially it will be used as a black box, with the exception of the implementation of the drizzle parameterization. However, in the course of the study, some additional improvement will occur when necessary. The UCLA LES solves the Ogura-Phillips anelastic equations that exclude sound waves, but represent all other types of motion in the atmosphere. In particular, it solves for the three components of velocity, $\bar{u}_i$, liquid-water potential temperature, $\bar{\theta}_l$, and total-water mixing ratio, $\bar{q}_t$:

$$\frac{\partial \bar{u}_i}{\partial t} = -\bar{u}_j \frac{\partial \bar{u}_i}{\partial x_j} - \frac{\partial \bar{p}}{\partial x_i} + \frac{g\bar{\theta}''_v}{\theta_0}\,\delta_{i3} + f_k(\bar{u}_j - u_{jg})\epsilon_{ijk} + \frac{1}{\rho_0}\,\frac{\partial(\rho_0\,\tau_{ij})}{\partial x_j}, \tag{7}$$

$$\frac{\partial \bar{\theta}_l}{\partial t} = -\bar{u}_j \frac{\partial\,\bar{\theta}_l}{\partial x_j} + \frac{1}{\rho_0}\,\frac{\partial(\rho_0\,\gamma_j)}{\partial x_j} + \frac{\partial F_{\theta_l}}{\partial x_j}\delta_{j3}, \tag{8}$$

$$\frac{\partial \bar{q}_t}{\partial t} = -\bar{u}_j \frac{\partial\,\bar{q}_t}{\partial x_j} + \frac{1}{\rho_0}\,\frac{\partial(\rho_0\,\upsilon_j)}{\partial x_j} + \frac{\partial F_{q_t}}{\partial x_j}\delta_{j3}, \tag{9}$$

while pressure, $\bar{p}$, is diagnosed at each time step by solving the Poisson equation:

$$\bigtriangledown^2 \bar{p} = \frac{\partial}{\partial x_i}\left(-\bar{u}_j \frac{\partial \bar{u}_i}{\partial x_j} + \frac{g\bar{\theta}''_v}{\theta_0}\,\delta_{i3} + f_k(\bar{u}_j - u_{jg})\epsilon_{ijk} + \frac{1}{\rho_0}\,\frac{\partial(\rho_0\,\tau_{ij})}{\partial x_j}\right), \tag{10}$$

which follows from the momentum and continuity ($\frac{\partial(\rho_0\,\bar{u}_j)}{\partial x_j} = 0$) equations. As discussed before, $\theta_v = \alpha\,\theta_l + \beta\,\theta_0\,q_t$ is virtual potential temperature that characterizes the buoyancy, where coefficients α and β take different values for saturated and unsaturated air. The reference state, given by $(p_0, \theta_0, \rho_0, u_g, v_g)$, is chosen to be in hydrostatic and geostrophic balances and to satisfy the ideal gas law for a dry atmosphere. Superscript $''$ denotes the thermodynamic perturbations from the reference state. The sub-filter scale contributions to the momentum ($\tau_{ij} = \nu_t\, S_{ij}$), liquid-water potential temperature ($\gamma_j = P_r^{-1}\, \nu_t\, \frac{\partial \bar{\theta}_l}{\partial x_j}$) and total-water mixing ratio ($\upsilon_j = P_r^{-1}\, \nu_t\, \frac{\partial \bar{q}_t}{\partial x_j}$) are parameterized in a manner described below. Diabatic fluxes F_{θ_l} and F_{q_t} will be parameterized as well. Simple models for radiation and drizzle will be introduced to reduce the computational costs and allow for the large computational domain. Diabatic fluxes will then be calculated from the radiative and drizzle fluxes, F_r and F_d respectively, as:

$$\frac{\partial F_{\theta_l}}{\partial z} = \frac{\partial F_r}{\partial z} - \frac{L}{c_p} \frac{\theta_l}{T} \frac{\partial F_d}{\partial z} \qquad (11)$$

and $F_{q_t} = F_d$.

To parameterize radiative forcing, a simple model of the net long-wave radiative flux will be used, which has been developed by Stevens *et al.* (2004) who showed that the model reasonably approximates radiative fluxes calculated by the δ-four stream radiative-transfer code developed by Fu and Liou (1993), which is computationally more expensive. It is a diagnostic model in which the three terms represent the effects of cloud top cooling, cloud base warming, and cooling in the free troposphere just above the cloud top:

$$F_r(x, y, z, t) = F_0 e^{-Q(z,\infty)} + F_1 e^{-Q(0,z)} + \rho_i c_p D \alpha_z \left[\frac{(z - z_i)^{4/3}}{4} + z_i (z - z_i)^{1/3} \right], \qquad (12)$$

where

$$Q(a, b) = \kappa \int_a^b \rho q_l dz. \qquad (13)$$

Also, ρ_i is the air density just below cloud top, D is the large scale divergence, and F_0, F_1, α_z and κ are the tuning parameters, adjusted so the simple model fits the profile from the full model for a given initial state. The horizontal and temporal dependence of F_r follows from the spatio-temporal variability in both q_l and z_i (inversion height).

To parameterize drizzle, another simple model will be implemented. This model is prognostic, but simple enough to have a minimal contribution to the computational expenses. In the model, a drizzle rate, R, which is proportional to the drizzle flux, F_d, by a factor of 10^{-5}, is relaxed over some time τ to the equilibrium profile R_{eq}:

$$\frac{DR}{Dt} = \frac{R_{eq} - R}{\tau}, \qquad (14)$$

where $\frac{D}{Dt} = \frac{\partial}{\partial t} + \bar{u}_i \frac{\partial}{\partial x_i}$. The equilibrium drizzle rate can be represented by the shape function $f(z)$ that follows Stevens *et al.* (1998), and a value at the cloud base $R_{eq}|_{z=z_b}$:

$$R_{eq} = R_{eq}|_{z=z_b} \, f(z), \tag{15}$$

where $R_{eq}|_{z=z_b}$ can be estimated from the empirical relation that is motivated by the observations analyzed by vanZanten *et al.* (2004):

$$R_{eq}|_{z=z_b} = A \, \frac{H^3}{N}. \tag{16}$$

Here, A is the proportionality coefficient, H is the cloud depth and N is the cloud droplet number concentration. The proportionality coefficient depends on the time scale over which the relaxation occurs. The shorter the time scale the higher the value of A. Cloud depth can be diagnosed from the LES results, while the cloud droplet number can be assumed to be equal to the number of condensation nuclei, which will be prescribed.

We would like to stress that this drizzle parameterization is new and that in the case of it beeing too simplistic we plan to use other parameterizations that are relatively simple and have been shown to be successful, such as the bulk parameterization developed by Khairoutdinov (1998).

Sub-filter fluxes are modeled using the Smagorinsky-Lilly model, which is an eddy viscosity model that includes the effects of the stabilization by modifying the mixing length scale. In the model, the eddy viscosity, ν_t, is assumed to be proportional to the mixing length scale and to the characteristic turbulent velocity, which is estimated from the local strain rate, S_{ij}:

$$\nu_t = l_0^2 \sqrt{1 - P_r^{-1} R_i} \, \left(\frac{1}{2} S_{ij} S_{kl} \delta_{ik} \delta_{jl} \right). \tag{17}$$

Here l_0 is an isotropic mixing length scale, P_r is an eddy Prandtl number specified to be $\frac{1}{3}$, R_i is a gradient Richardson number and S_{ij} is in the traceless form:

$$S_{ij} = \left(\frac{\partial \bar{u}_i}{\partial x_j} + \frac{\partial \bar{u}_j}{\partial x_i} - \frac{1}{2} \delta_{ij} \frac{\partial \bar{u}_k}{\partial x_k} \right). \tag{18}$$

Because it has been shown that the LES is sensitive to the SFS representation at the inversion interface (Stevens *et al.*, 2004), this problem will be explored by either turning off the parameterization, or tuning its stability threshold.

The solver uses finite differences on a regular-horizontal and stretched-vertical mesh. We propose studies based on the horizontal spacing of 50 m in both directions, and the vertical spacing of 10 m near the surface and refined above (10% per layer) to obtain a fixed 5 m spacing about the inversion, above which the grid is stretched again. We will explore a variety of domain sizes, including those that allow for the organization of convection in more than one open cell. The code has recently been configured for the large horizontal domain sizes and a computation of 700×700×131 point mesh has been performed with a 70 m horizontal grid spacing, *i.e.*, 50 km scale, which is substantially larger than in the Stevens *et al.* (1998) study. The top of the domain is at 1470 m, and a sponge layer, that prevents the spurious influence of the upper boundary, occupies the upper five levels. The side boundary conditions are cyclic, while the bottom surface has prescribed SST from which the surface fluxes are calculated. The momentum terms are time-stepped using a leap-frog scheme with an Asselin filter for the damping of the computational mode. Momentum advection is computed using fourth-order centered differences. Scalar terms are time-stepped using a forward scheme staggered with respect to the time-levels of the momentum terms, so that the advecting winds correspond to the mid-point times. Scalar advection is TVD (total variation diminishing) and uses the MC flux-limiters. The code is parallelized using MPI with a 1D decomposition. The initial calculations with the small domain will be performed on the Intel Pentium 4, CPU 2 GHz, while the calculations with the large domain will be performed on many processors of an IBM-SP4.

4 Work Plan

As the current literature suggests that drizzle induces a change of the circulation into the Cu-type, we would like to explore what about drizzle is critical to dynamic evolution. We propose to use an LES with a domain large enough to capture the development of several Cu. That would provide the unbiased statistics (which might have been the limitation in Stevens *et al.*, 1998), and also the possibility for the development of the open cell circulation. The study will be based on numerous LESs, which will be described in this section.

4.1 LES validation

The environmental forcing and initial conditions in this study will be initially based on the results of the analysis of the observations from the second research flight (RF02) of DYCOMS-II. These data are being analyzed by M. C. van Zanten and B. Stevens and will also be used as the basis for the eighth GCSS LES intercomparison. GCSS stands for GEWEX (Global

Energy and Water Cycle Experiment) Cloud System Studies, and the GCSS Boundary Layer Cloud working group aims to improve physical parameterizations of clouds, other boundary layer processes, and their interactions. Previous intercomparison studies have examined many aspects of cloud topped boundary layers including the nocturnal non-precipitating STBL (Stevens *et al.*, 2004). The eighth LES intercomparison explore drizzling nocturnal STBL, and we plan to participate.

The control experiment in this study will be the simulation of the non-drizzling STBL. This will be performed by simply excluding the drizzle parameterization and therefore preventing drizzle formation. The horizontal distribution of the environmental conditions will be uniform. From the control experiment we will learn about the profiles and horizontal distribution of the thermodynamic variables and vertical velocity, as well as their fluxes and higher order statistics. We expect that these will show a well mixed STBL with homogeneous distribution of up- and downdrafts and with uniform cloud depth. These results will further be compared with the drizzling cases.

The first drizzling case will also have horizontally uniform environmental conditions, but will use a parameterized drizzle rate. We would like to use the simple drizzle parameterization described in section 3, which provides the most easily controlled drizzle rate. With this approach we would like to avoid expensive microphysical calculations and to concentrate on the turbulent processes, which are important for the change in the circulation type. However, if this approach proves too simplistic, we will explore the use of other bulk drizzle schemes, for instance Khairoutdinov (1998).

The resulting profiles and horizontal distributions of the thermodynamic variables and vertical velocity, as well as their fluxes and higher order statistics will be analyzed. This analysis will provide information about the time and space scales present in the drizzling STBL, which will be necessary for our further analysis discussed later. The observations suggest that the temperature and moisture are anti-correlated on scales $\sim$10 km (Paluch and Lenschow, 1991), while the clouds persist for a much longer time than is needed for drizzle to completely dry out the cloud (vanZanten *et al.*, 2004). These together suggest a need for analyzing the scales developed within the LES of drizzling STBL and a minimal domain size required for the representation of the drizzling STBL.

The question of the vertical resolution and the representation of the SFS processes will also be addressed in these simulations. As the LES comparison study by Stevens *et al.* (2004)

showed, the results vary depending on the SFS parameterizations, especially at the inversion interface, where they affect the mixing of the STBL air with the free-tropospheric air and can cause spurious entrainment. We expect that this will be an issue in this study too, as the stronger entrainment may reduce the cloud propensity to drizzle. Therefore, the initial phase of the study will also include experiments that address these issues.

4.2 Drizzle

After establishing an optimal domain size we can investigate the effect drizzle has on the STBL by comparing the drizzling with the control run. Based on the current literature, we expect that the STBL will decouple as drizzle rates increase, preventing the profiles of θ_l and q_t from being as well mixed as in the control run. Furthermore, the variance of the vertical velocity can be expected to be smaller and with two local maxima – one within the cloud and one in the subcloud layer. Another variable expected to be different is the skewness of the vertical velocity, since the distribution of the up- and downdrafts would not be horizontally homogeneous as in the control run. However, one could also expect that the drizzle does not change the canonical view of the STBL as a well mixed layer. The result would depend on the other forcings and on how much forcing drizzle provides. The mixed layer theory can provide the estimate of the dependence of the drizzle rate decoupling threshold on the strength of the other STBL forcings, while these simulations can provide a way to evaluate the theoretical predictions. An additional interesting question would also be the sustainability of drizzle both in the coupled and decoupled state.

These simulations will be a valuable contribution to the study of the Sc to trade-Cu transition, as they will allow us to see if the cloud propensity to drizzle can be an additional control factor for the transition, as is suggested by Bretherton and Wyant (1997), or if it can be the cause of transition, as suggested by Paluch and Lenschow (1991).

With established drizzle rate decoupling threshold we can explore what particular aspect of drizzle is critical for the change in the dynamics. The simulations that will address this question will be run without the drizzle parameterization, but with the included additional sources of the warming and drying that will mimic the drizzle effects. We would like to explore the warming and drying effects separately, as well as in various combinations of their values in the cloud and subcloud layer. We would use various profiles that would differ in the shape to represent the peculiarity of the effects if the forcing is concentrated in particular sublayers, or is spread over the whole STBL height. This could also resolve the question of the importance of cloud warming/drying versus subcloud cooling/moistening. However, as drizzle has nonuniform horizontal distribution, we would like to explore the importance of that aspect of drizzle as well. The simulations that would address this question would have included the above profiles only where the cloud depth exceeds some critical value.

4.3 POCs

Further in the study, we would like to explore the question related to the non-uniformity of the environmental conditions, for instance by running the experiments with horizontally varying large scale forcings (*e.g.*, moisture content in the overlying free troposphere, that affects the radiative forcings as well as the buoyancy jump at the inversion interface; SST that affects the surface latent and sensible heat fluxes; and large-scale divergence that regulates the depth of the STBL and therefore the cloud depth) and cloud propensity to drizzle. Each of these variables will be changed in the individual experiments. The results of such experiments can provide insight into the response of the drizzling STBL due non-uniformity of the environmental conditions and we can learn if the drizzle can cause compact regions of open cells in the otherwise homogeneous Sc decks. Furthermore, we can investigate whether POCs induce large scale circulations that help maintain them.

5 Future Work

Recognizing the importance of Sc for climate and possible Sc effects on climate change (Slingo, 1990) there are attempts in our community to improve the representation of these clouds in the GCMs (General Circulation Models). An example of a novel effort in this respect is MMAP (Multi-scale Modeling of Atmospheric Processes, Randall *et al.*, 2004). In this approach, the idea of parameterizing the boundary layer processes is to implement numerous coarse resolution (∼100 m in the horizontal and ∼25 m in the vertical) LESs or 2D ERMs (Eddy Resolving Models) per GCM grid cells. In order to do such an implementation, there is a need to evaluate the results of such coarse and 2D models. As the future work we would like to participate in MMAP and study the representation of drizzle effects on the STBL in the coarse LES and 2D ERM. More precisely, for the evaluation we would follow the approach of Moeng *et al.* (2004) in comparing 3D LES and 2D ERM. The results from the proposed LES study of the effects of drizzle on the STBL would be a basis for this future work, as there are no current results on this topic.

6 Timeline

In defining a timeline the governing idea is scheduling time for answering the questions we have raised. I would like to graduate in Fall 2006 and we believe that these two years will provide enough time to successfully complete the proposed project.

Prior to addressing the questions, we need to spend some time on learning how the model works and comparing it with other similar models. Therefore we propose the initial phase of the study to end in April 2005, when there is a meeting of GCSS 8 group. The sub-phases will include: learning about the UCLA LES; running non-drizzling LESs on variety of domains to benchmark them on RF01; implementing drizzle parameterizations; running drizzling LESs on variety of domains to benchmark them on RF02 and to establish the minimal scale for the representation of drizzle in the LES.

The second phase is proposed to last from April till October 2005. In this period we would like to explore the environmental conditions that lead to the decoupling and also what particular aspect of drizzle is critical for the dynamics evolution. The LES would be run with drizzle parameterization on a domain size established in the first phase to check the mixed-layer theory predictions about the circulation change. For the decoupling runs, we would like to explore what about drizzle induces the decoupling by removing the drizzle parameterization and implementing the variety of heating and moistening profiles. Furthermore, the importance of the non-uniformity of the drizzle onto the dynamics would be explored by applying the heating and moistening profiles horizontally uniform over the whole domain, and including the dependence on the cloud depth.

The third phase would explore the non-homogeneity in the environmental conditions and the ability of POCs to maintain themselves. In this phase the LESs will be on the large horizontal domain (at least 20×20 km^2), and we propose that this phase lasts from October 2005 till April 2006.

Depending on the progress, we would like to participate in the MMAP and explore the drizzle representation in coarse LESs and 2D ERMs. We expect that from April 2006 till October 2006 we will be able to do that part of the study. However, we also leave the possibility that this part of the project exceeds designated time for the graduation, in which case it could form the basis of a proposal for post graduate work.

References

Ackerman, A. S., O. B. Toon, and P. V. Hobbs, 1993: Dissipation of Marine Stratiform Clouds and Collapse of the Marine Boundary Layer Due to the Depletion of Cloud Condensation Nuclei by Clouds. *Science,* **262**, 226–229.

Agee, E. M., T. S. Chen, and K. E. Dowell, 1973: A Review of Mesoscale Cellular Convection. *Bull. Amer. Meteor. Soc.,* **54**, 1004–1012.

Albrecht, B. A., 1989: Aerosols, Cloud microphysics, and Fractional Cloudiness. *Science,* **245**, 1227–1230.

Bretherton, C. S. and M. C. Wyant, 1997: Moisture Transport, Lower-Tropospheric Stability, and Decoupling of Cloud-Topped Boundary Layers. *J. Atmos. Sci.,* **54**, 148–167.

Brost, R. A., J. C. Wyngaard, and D. H. Lenschow, 1982: Marine Stratocumulus Layers. Part II: Turbulence Budgets. *J. Atmos. Sci.,* **39**, 818–836.

Fu, Q. and K. N. Liou, 1993: Parameterization of the radiative properties of cirrus clouds. *J. Atmos. Sci.,* **50**, 2008–2025.

Garay, M. J., R. Davies, C. Averill, and J. A. Westphal, 2004: Actinoform clouds: Overlooked Examples of Cloud Self-Organization at the Mesoscale. *Bull. Amer. Meteor. Soc.,* **85**, 1585–1594.

Khairoutdinov, M., 1998: *Large-Eddy Simulation of Stratocumulus-Topped Boundary Layer With an Explicit and a New Bulk Microphysics Scheme.* Ph. D. thesis, University of Oklahoma.

Klein, S. A. and D. L. Hartmann, 1993: The Seasonal Cycle of Low Stratiform Clouds. *J. Climate,* **6**, 1587–1606.

Klein, S. A., D. L. Hartmann, and J. R. Norris, 1995: On the relationships among low-cloud structure, sea surface temperature, and atmospheric circulation in the summertime northeast pacific. *J. Climate,* **8**, 1140–1155.

Lilly, D. K., 1968: Models of cloud-topped mixed layers under a strong inversion. *Quart. J. R. Met. Soc.,* **94**, 3353–3361.

Moeng, C.-H., J. C. McWilliams, R. Rotunno, and P. P. S. J. Weil, 2004: Investigating 2d modeling of atmospheric convection in the pbl. *J. Atmos. Sci.,* **61**, 889–903.

Nicholls, S., 1984: The dynamics of stratocumulus: aircraft observations and comparisons with a mixed layer model. *Quart. J. R. Met. Soc.,* **110**, 783–820.

Paluch, I. R. and D. H. Lenschow, 1991: Stratiform Cloud Formation in the Marine Boundary Layer. *J. Atmos. Sci.,* **48**, 2141–2158.

Pincus, F. and M. B. Baker, 1994: Effect of precipitation on the albedo susceptibility of clouds in the marine boundary layer. *Nature,* **372**, 250–252.

Randall, D. A., A. S. Denning, J. Helly, C.-H. Moeng, and W. Schubert, 2004: Center for Multi-Scale Modeling of Atmospheric Processes. Proposal 0338050, National Science Foundation.

Schubert, W. H., J. S. Wakefield, E. J. Steiner, and S. K. Cox, 1979: Marine Stratomulus Convection. Part I: Governing Equations and Horizontally Homogeneous Solutions. *J. Atmos. Sci.,* **36**, 1286–1307.

Slingo, A., 1990: Sensitivity of the Earth's radiation budget to changes in low clouds. *Nature,* **343**, 49–51.

Stevens, B., 2000: Cloud transition and decoupling in shear-free stratocumulus-topped boundary layers. *Geophys. Res. Lett.,* **27**, 2557–2560.

Stevens, B., 2002: Entrainment in stratocumulus-topped mixed layers. *Quart. J. R. Met. Soc.,* **128**, 2663–2690.

Stevens, B., 2004: Atmospheric Moist Convection. *Annu. Rev. Earth Planet. Sci*. submitted.

Stevens, B., W. R. Cotton, G. Feingold, and C.-H. Moeng, 1998: Large-Eddy Simulations of Strongly Precipitating, Shallow, Stratocumulus-Topped Boundary Layer. *J. Atmos. Sci.,* **55**, 3616–3638.

Stevens, B., D. H. Lenschow, G. Vali, H. Gerber, A. Bandy, B. Blomquist, J. L. Brenguier, C. S. Bretherton, F. Burnet, T. Campos, S. Chai, I. Faloona, D. Friesen, S. Haimov, K. Laursen, D. K. Lilly, S. M. Loehrer, S. P. Malinowski, P. Szymon, B. Morely, M. D. Petters, D. C. Rogers, L. Russell, V. Savic-Jovcic, J. R. Snider, D. Straub, M. J. Szumowski, H. Takagi,

D. C. Thorton, M. Tschudi, C. Twohy, M. Wetzel, and M. C. van Zanten, 2003: Dynamics and Chemistry of Marine Stratocumulus - DYCOMS-II. *Bull. Amer. Meteor. Soc.,* **84**, 579–593.

Stevens, B., C.-H. Moeng, A. S. Ackerman, C. S. Bretherton, A. Chlond, S. de Roode, J. Edwards, J.-C. Golaz, H. Jiang, M. Khairoutdinov, M. P. Kirkpatrick, D. C. Lewellen, A. Lock, F. Muller, D. E. Stevens, E. Whelan, and P. Zhu, 2004: Evaluation of large-eddy simulations via observations of nocturnal marine stratocumulus. *Mon. Wea. Rev.*. submitted.

Stevens, B., G. Vali, K. Comstock, R. Wood, M. C. vanZanten, P. H. Austin, C. S. Bretherton, and D. H. Lenschow, 2004: Pockets of Open Cells (POCs) and Drizzle in Marine Stratocumulus. *Bull. Amer. Meteor. Soc.*. submitted.

Turton, J. D. and S. Nicholls, 1987: A study of the diurnal variation of stratocumulus using a multiple mixed layer model. *Quart. J. R. Met. Soc.,* **113**, 969–1009.

vanZanten, M. C., B. Stevens, G. Vali, and D. H. Lenschow, 2004: Observations of drizzle in nocturnal marine stratocumulus. *J. Atmos. Sci.*. accepted.

Wang, S. and Q. Wang, 1994: Roles of Drizzle in a One-Dimensional Third-Order Turbulence Closure Model of the Nocturnal Stratus-Topped Marine Boundary Layer. *J. Atmos. Sci.,* **51**, 1559–1576.

Wyant, M. C., C. S. Bretherton, H. A. Rand, and D. E. Stevens, 1997: Numerical Simulations and a Conceptual Model of the Stratocumulus to Trade Cumulus Transition. *J. Atmos. Sci.,* **54**, 168–192.