Lecture 15. Subtropical stratocumulus-capped boundary layers

In this lecture…

- Physical processes and their impact on Sc boundary layer structure

Introduction

Clear turbulent boundary layers over land are usually driven mainly by surface heat fluxes or drag. Stratocumulus-capped boundary layers (SCBLs) are more complicated (Fig. 15.1; **note all figures refer to the associated slides**). The cloud usually forms because turbulence lifts moist air from near the surface up to its condensation level. The cloud plays an active role in maintaining the turbulence and building a sharp, strong capping inversion. Radiative cooling at cloud top and heating within the cloud, as well as latent heating due to condensation or evaporation of cloud and drizzle all have strong feedbacks on the boundary-layer structure and turbulence. The strong capping inversion inhibits turbulent mixing or entrainment of the warmer and drier overlying air into the SCBL. This keeps the boundary layer cool and moist, helping the cloud persist. A strong capping inversion goes with more lower tropospheric stability, and also keeps the boundary layer moist and cloud-capped. This is a major reason for the observed correlation between lower tropospheric stability and stratus cloudiness.

Moist-conserved variables

In the study of MBLs, it is often useful to work with moist-conserved variables preserved during adiabatic changes including phase changes between vapor and liquid, e. g. the total water mixing ratio $q_t = q_v + q_l$ (sum of vapor and liquid water). Moist-conserved temperature-like variables include the equivalent potential temperature $\theta_e \approx \theta \exp(Lq_v/c_pT_{LCL})$ (see Bolton 1980 for a more accurate definition) and the liquid water potential temperature $\theta_l = \theta \exp(-Lq_l/c_pT_{LCL})$. If the parcel vertical displacement is nearly hydrostatic (a good approximation for the MBL), one can instead use simpler moist-conserved variables, the moist static energy $h = c_pT + gz + Lq_v$, or the liquid-water static energy $s_l = c_pT + gz - Lq_l$. All four of these choices are commonly used in studies of SCBLs.

As an air parcel rises moist-adiabatically above its lifted condensation level (LCL), it becomes cloudy and condenses liquid water at a rate $(dq_l/dz)_{\mathrm{ma}} = -(dq^*/dz)_{\mathrm{ma}} \approx 2$ g kg^{-1} km^{-1} for thermodynamic conditions typical for Sc (cloud base temperature of 285 K, cloud base pressure of 950 hPa). Here, 'ma' stands for 'moist adiabatic'.

Radiation

The SCBL interacts strongly with longwave and shortwave radiation, and radiative cooling is the main driver of turbulent convection in subtropical SCBLs. Clouds as little as 50 m thick efficiently absorb and emit longwave radiation. Although the clouds mainly scatter sunlight, they also absorb a little of it. The upper left figure in Fig. 15.2 shows a comparison of measurements and radiative transfer model calculations for a thick summertime North Sea stratocumulus cloud around noon. The symbols S and L refer to shortwave and longwave radiation, and arrows indicate upward and downward fluxes. About 2/3 of the incident sunlight is reflected, but about 60 W m^{-2} (6%) is absorbed in the cloud. Upwelling longwave radiation emitted from the warm cloud top is almost 100 W m^{-2} larger than downwelling longwave radiation emitted by the dry and mostly colder overlying atmosphere. Within the cloud, the photon path is short and the net longwave flux is small, while below cloud base, there is a net upward longwave flux of about 10 W m^{-2} because the SST slightly exceeds the cloud base temperature.

Combining longwave and shortwave fluxes, we get the net upward radiative flux during the middle of the day. From just the longwave flux, we get the net upward radiative flux at night. (middle figures). The dashed line in the night-time panel shows the daily-averaged net upward radiative flux. Vertical convergence or divergence of the net radiative flux implies radiative heating or cooling, respectively. During the night, the flux profile implies slight radiative warming near cloud base and strong cooling in the 50 m below cloud top, with a net 60 W m^{-2} longwave cooling integrated over the SCBL in the case shown. The strong cloud-top cooling favors vigorous convection and often generates a mixed-layer structure. During the daytime, the additional absorption of sunlight warms most of the cloud layer, but the strong longwave cooling still dominates at the cloud top. The 60 W m^{-2} of solar absorption roughly cancel the SCBL longwave cooling, so the effect of radiation at noon is only to destabilize the cloud layer, not the entire boundary layer. This favors daytime decoupling of SCBLs – surface heat fluxes cause convection near the sea-surface, and radiation drives convection within the stratocumulus cloud layer.

Averaged over the whole diurnal cycle, the net longwave cooling of the SCBL is roughly 3-4 times as large as net solar heating, and radiation is strongly destabilizing the SCBL by cooling its top. The diurnal cycle of SCBL radiative energy loss is shown at lower right, where it is also compared to a typical value of surface sensible heat flux over the subtropical oceans. This plot suggests that in subtropical SCBLs, radiation is more important to the energy budget and generation of turbulence than is the surface heat flux. The strong radiative cooling also helps maintain the sharp 5-10 K inversions that usually top such boundary layers.

Theoretical studies of cloud-topped mixed layers sometimes treat the radiative flux divergence as concentrated entirely at the cloud top, and often specify it as an external parameter $\Delta R_N \approx 50$ W m^{-2} (rightmost profile). In reality, ΔR_N is strongly dependent on above-SCBL humidity, cloud and temperature, as well as cloud-top temperature and insolation. In particular, ΔR_N is largest under a clear, dry atmospheric column.

Mixed-layer structure

Especially at night and in the early morning, SCBLs commonly exhibit <u>mixed-layer</u> structure in which moist-conserved thermodynamic variables and the horizontal velocity components are approximately uniform with height. This is a sign of strong vertical mixing by turbulent eddies extending from the surface all the way to the cloud-top. In SCBLs, the eddy updrafts and downdrafts are typically on the order of 1 m s^{-1}.

Fig. 15.3 shows an example from the DYCOMS-II field experiment off the California coast in July 2001 (satellite picture and map showing flight track and sea-surface temperature are at upper right). Several aircraft profiles through the same cloud layer a few tens of km apart are shown, all flown at night. One sees well-mixed structure in q_t and s_l, and the linear rise of q_l with height above cloud base. We also see the strong (10 K), sharp inversion separating the cool marine airmass from the much warmer and very dry air above, which has slowly subsided from much higher in the troposphere in the descending branch of the Hadley circulation.

Decoupled structure and the diurnal cycle

The large impact of solar absorption on the diurnal cycle of STBLs was first discussed in detail by Nicholls (1984, *QJRMS*, **110**, 783-820). The 1987 FIRE-MSC experiment documented the diurnal cycle in Sc 100 km off the California coast. Considerable cloud thinning (a factor of four decrease in liquid water path) and a reduction in cloud albedo from 50% to 30% is observed during the morning (Fig. 15.4). This is not simply cloud 'burning off' due to warming of cloudy air by the absorbed sunlight, resulting in evaporation of the cloud. Instead it is mainly driven by diurnal decoupling and the associated diurnal cycle of BL turbulent structure.

Fig. 15.5 shows 'decoupled' vertical structure in θ_e, q_t, and the wind components. This is also commonly seen in SCBLs, especially during mid-day – these aircraft profiles were made near noon in North Atlantic summer stratocumulus. The SCBL is separated into two mixed layers, one starting at the surface, and one extending down from the cloud layer, with a stratified layer in between. In this middle layer, there is little turbulence (visible in the slide as less fine-scale vertical variability). 'Scud' clouds can sometimes.form at the top of the surface mixed layer. Given long enough, these clouds can develop into cumulus convection, leading to a 'cumulus-coupled' SCBL in which cumulus convection fluxes moisture from the lower to the upper mixed layer.

Fig. 15.6 shows mean daytime and nighttime vertical velocity variance profiles measured by a tethered balloon at San Nicolas Island off the CA coast during FIRE-MSC.during about three weeks of June 1987. During the night, the vertical velocity variance is maximum in the middle of the layer where large-eddy up and downdrafts are maximum, while during the day separate surface layer and cloud layer maxima exist, indicative of two convective layers separated by a stable layer with gravity wave activity and perhaps intermittent turbulence.

The connection between decoupling and the diurnal cycle of Sc thickness is important to understand. After morning decoupling occurs, the Sc mixed layer dries due to entrainment, so the Sc base steadily rises and the Sc may partly or fully dissipate. The surface mixed layer moistens due to surface fluxes, which is what allows the scud or Cu to develop at its top. Late in the afternoon, when shortwave heating diminishes, the upper mixed layer begins to cool from longwave emission. As it cools, its turbulent eddies penetrate further down into the transition layer. During the evening the Sc mixed layer can 'reconnect' with the surface mixed layer, allowing moisture to efficiently flux into the Sc layer and rapidly deepen the he Sc as a single mixed layer is reestablished.

Fig. 15.7 shows a 6-day time series of radiosonde profiles from the October 2001 EPIC research cruise into the SE Pacific stratocumulus region. These nicely show a pronounced diurnal cycle. The difference between cloud base and near-surface LCL (measured by the ship at a height of 15 m above sea level) is a good measure of decoupling. It would be zero in an ideal mixed layer, in which the near-surface air had exactly the same properties as cloud base air. This is never seen, because in the surface layer (lowest 5-10% of the BL) there is a 'log-layer' in which air properties transition from those in the bulk of the boundary layer and the saturated air in a mm thick skin next to the sea-surface.) However, smaller values (less than 10-15% of the cloud base height) indicate a mixed layer, and larger values (more than 250 m) indicate a more decoupled boundary layer in which the surface air is distinctly moister than that in the cloud layer. This measure shows mixed-layer structure at night and slightly decoupled structure during the day (noon local time = 18 UTC) as well as during periods of drizzle. Also note that the inversion rises at night and falls during the day. One contributor to this is stronger entrainment at night. Another, specific to this location, is a diurnal oscillation in mean subsidence associated with a gravity wave that is forced by daytime heating on the Andean slopes and then propagates offshore (Garreaud and Munoz 1984).

Precipitation

Because stratocumulus are thin and rely on the surface for their supply of liquid water, they can be sensitive to even a little precipitation. Precipitation in stratocumulus can be somewhat artificially divided into droplet sedimentation and drizzle. Sedimentation is the slow settling of 'cloud' droplets less than 20 μm in radius. It occurs only within the cloud, but can result in downward water fluxes of several mm/day, which proves important for the water budget of the upper part of the cloud layer. Drizzle is the settling of larger drops created by collision-coalescence processes, and tends to be dominated by drops 100 μm and larger. Drizzle tends to maximize near cloud base, and rapidly evaporates below the cloud. Light drizzle is sometimes observed in shallow cloud-topped boundary layers, especially when aerosol concentrations are low or the cloud is thick (which is most common in the night and early morning).

Fig. 15.8 shows typical profiles of sedimentation and drizzle to the downward precipitation flux in a moderately drizzling Sc, corresponding to cloud base precipitation of 2 mm day^{-1} and surface precipitation of 0.25 mm day^{-1}. Sedimentation removes liquid water from the top of the cloud, forcing turbulence to lift it up again. This decreases entrainment (see Bretherton et al. 2007 for a detailed explanation) and tends to reduce turbulence in the cloud layer. Drizzle causes net condensation and latent heating in the cloud layer and evaporation and cooling of the subcloud layer, stabilizing the BL to convection. Often, drizzling shallow Sc layers are observed to have some stratification of potential temperature and mixing ratio, and cloud cover may be less homogeneous. Both sedimentation and drizzle are much larger when aerosol (and hence cloud droplet) concentrations are low. Thus, these processes are important to understanding the effects on anthropogenic aerosols on SCBLs and climate.

The bottom of Fig. 15.8 shows a 6-day time-height section of mm-wavelength upward-pointing radar returns from SE Pacific stratocumulus during the EPIC cruise. Reflectivities less than -10 dBZ correspond to nearly non-drizzling cloud; stronger reflectivities indicate drizzle. When the drizzle is weak, it all evaporates near cloud base; when the drizzle is strong it gets down to the surface. A strong diurnal cycle of drizzle is evident, connected to night-time cloud thickening.

Entrainment

Entrainment is the incorporation of filaments or blobs of overlying non-turbulent air into the SCBL by turbulent eddies. Entrainment occurs in a thin entrainment zone near the cloud-top. Over boundary layer updrafts, the entrainment zone is thin (as little as 1 m thick), and it is thicker (up to 100 m) in downdraft regions, especially if the inversion is weak or there is a lot of wind shear at the inversion. The physical mechanisms are somewhat complicated and the cloud itself affects the entrainment process through evaporative cooling –we'll discuss this more later when we talk about entrainment closures. What is clear is that entrainment is faster if the turbulence is stronger or the overlying inversion is weaker. For now, we simply define the entrainment rate w_e, which is the rate at which overlying air is incorporated into the SCBL. For subtropical SCBLs, w_e is usually only a few mm s^{-1}.

Consider a variable F with no sources or sinks in a thin entrainment zone, and a typical value F^- below the entrainment zone and F^+ above the entrainment zone (Fig. 15.9, top right). The flux - w_eF^+ of F through the top of entrainment zone must balance the flux of F through the bottom of the entrainment zone (which has a mean component - w_eF^- and a turbulent component). We deduce that a turbulent entrainment flux

$$\left.\overline{w'F'}\right|_e = -w_e\Delta F\ ,\ \Delta F = F^+ - F^- \qquad (15.1)$$

is needed to mix the entrained air into the SCBL.

Using (15.1), we can deduce entrainment from aircraft measurements of the below-inversion flux and cross-inversion jump of suitable variables. Total water, ozone, and DMS have been successfully used for this purpose. Alternatively, we can derive a heat, moisture or mass budget for the entire SCBL, deduce the entrainment flux by measuring all other terms in the budget, and then apply the flux-jump approach. Fig. 15.9 shows a nice comparison of these approaches using airborne measurements during DYCOMS-II (Stevens et al. 2003). Fig. 15.10 shows another example of the budget approach using ship-based measurements, in which we see reasonable consistency between the diurnal cycle of entrainment deduced from heat, moisture and mass budgets during a 6-day period in SE Pacific stratocumulus (Caldwell and Bretherton 2005). This approach works because entrainment is a dominant term in all three budgets.

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