Turbulence Structure of Marine Stratocumulus

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ABSTRACT

Aircraft measurements are analysed from the “First Lagrangian” of the Atlantic Stratocumulus Transition Experiment (ASTEX) from south east of the Azores Islands. In this experiment, Lagrangian strategy was used and the marine air mass, that advected southward, was followed during 12 to 14 June 1992. During the experiment, the stratocumulus clouds transitioned into thin and broken stratocumulus with cumulus cloud penetrating from below.

To characterise the vertical structure in the marine boundary layer the buoyancy fluxes, the variances, the turbulent kinetic energy, the momentum fluxes and humidity fluxes were examined. The buoyancy flux profiles were used to discover the decoupling of the stratocumulus and the sub-cloud layer. Turbulence analysis for all five flights shows that the cloud layer were decoupled from the underlying layer. In the cloud layer the buoyancy production due to longwave radiative cooling at cloud top, was the main source for driving the turbulence. In the sub-cloud layer, the variances indicate wind shear to be the main generator of turbulence for the first two days. Then, as sea surface temperature increases, buoyancy produced turbulence was more pronounced. The u-, v- and w-spectra and cospectra of wθ and uw give insight into the typical eddy sizes, and how the peak wavelengths vary with height. The peak wavelengths in sub-cloud and cloud layer were larger than layer depth and u- and v-spectral peak wavelengths often larger than the peak wavelength from w-spectra. While peak wavelengths in the sub-cloud layer vary with the height above the surface, they are approximately invariant with height in the cloud layer.
Stratocumulus clouds are classified as low level clouds, which means that their cloud base lying below 2000 m. The word stratus is Latin for “layer” and cumulus is Latin for “heap”. Figure 1 and 2 shows stratocumulus clouds were the sky is almost covered by a layer composed of large and rounded elements. The cloud elements are arranged almost regularly and nearly parallel and there is probably wind shear that makes the cloud pattern as long rows, especially seen in figure 2. Figure 3 and 4 shows stratocumulus clouds at sunset and the more patchy structure made by convection. These stratocumulus clouds at sunset often built up from cumulus clouds, which no longer grow vertically but spread out in the horizontal. These figures illustrate continental stratocumulus clouds while in this paper the data are taken from marine stratocumulus clouds.
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1. Introduction

Clouds have a large effect on the Earth climate. The Earth’s lower atmosphere is often cloudy and at any instant clouds overlie about half the planet’s surface (McIlveen 1992). Clouds vary in thickness from a few tens of meters to the full depth of the troposphere and are variable in both time and space. Stratocumulus cloud sheet tends to form where the ocean is cold compared to the lower atmosphere and in areas where large-scale subsidence forms a capping inversion. These clouds are classified as low-level clouds, which are horizontally extensive sheets of clouds. If the cloud base is not too high, drizzle or snow at the surface can result (McIlween 1992). Stratocumulus is very common over the cooler parts of oceans bordering to subtropical west coasts of continents and in the Arctic basin during boreal summer (Nucciarone and Young 1991).

Boundary layers with stratocumulus clouds are important for the Earth’s energy balance and thus also for the climate. Thick and low clouds are efficient reflectors of incoming shortwave radiation from the sun. Comparing to cloud-free areas, marine stratocumulus clouds have a high albedo relative to the sea surface, which means that more shortwave radiation is reflected in the cloudy areas. Typical albedo for cloud-free area is 5-10% and with clouds, the albedo is often as high as 60-90%. In the infrared wavelengths, even a rather thin cloud acts as a blackbody. Due to the relatively low altitude of the cloud tops, the cloud-top temperature is not so different from the sea surface temperature. This implies, in contrast to for example cirrus clouds, that the longwave radiation budget can not compensate the change in the net radiation.

Through these important radiative exchanges, and because the marine stratocumulus clouds have large horizontal extents, they are of great interest for the global climate and in climate modelling. To parameterize boundary layer clouds, one needs to understand the process that generate, maintain and dissipate these clouds. Presently, these clouds are difficult to simulate in weather forecast and climate models, partly due to lacking knowledge of the processes that are important to cloud development (Duynkerke et al. 1995).

One process that is particularly important for the development of stratocumulus clouds and needs to be understood is turbulence and turbulent transport. This paper will focus on turbulent statistics calculated from aircraft measurements during the First Lagrangian of the Atlantic Stratocumulus Transition Experiment (ASTEX) (Albrecht et al. 1995). Variances and covariances are presented; the variances provide information about the turbulent energy and intensity while the covariances describes the kinematic turbulent fluxes. These are also given as spectra and cospectra. The turbulent statistics in the stratocumulus clouds and sub-cloud layer are studied to infer whether these layers are coupled or decoupled. Turbulent structure and decoupling during ASTEX is also analysed in deRoode and Duynkerke (1997).

Section 2 provides a background about boundary layer processes and section 3 provides some theory about energy spectra and scaling. Section 4 describes the ASTEX experiment and in section 5 the treatment of the data is described. In section 6 the results are presented, divided into turbulence statistics, spectral analysis and length-scales. Conclusions are found in section 7.
2. Background

The troposphere extends from the Earth surface up to an average altitude of 11 km. The lowest part of the troposphere, which is influenced by the Earth surface, is defined as the boundary layer (Stull 1988). The boundary layer thickness is variable in both time and space, ranging from a few tens of meters to O(1 km). It has a diurnal variation, which is also different over a land or a sea surface. This due to the differences in the effective heat capacities of the surfaces, leading to different temperature variations near the surface.

The boundary layer can be divided into a few sub-layers. The lowest 10% of the boundary layer is called the surface layer. Within the surface layer, turbulent fluxes and stresses vary by less than 10% in magnitude. This layer is thus sometimes referred to as the “constant-flux layer”. Nearest to the ground, there is an even thinner layer called the micro layer. This layer has a very small vertical extent, of about a few centimetres, and here the molecular transport dominates over turbulent transport. Above the surface boundary layer, there is a transition zone in which the influence of the surface weakens with increasing height. This layer is sometimes called the Ekman layer. Sometimes during daytime, particularly over land, a convective mixed layer builds up from surface layer and up to the capping inversion at the boundary layer top. Often at night, the low-level air layer becomes statically stable. Calm winds and very little turbulence characterise the stable boundary layer. Above the boundary layer top, the rest of the troposphere is called the free atmosphere. In the free atmosphere, the friction from the earth surface does not influence the flow directly any more and the free atmosphere shows little diurnal variation.

In the boundary layer the most important transport process is turbulence. Turbulence transport moisture, heat, momentum and pollutants, especially in the vertical. One of the driving forces for turbulence in the boundary layer is buoyancy. Thermals of warm air rise because this air is less dense than the surrounding air. Turbulence can also be produced by shear, arising when wind speed and/or direction vary with height. The turbulence tends to reduce the gradients, the shear, in the flow.

2.1 The mixed layer

If the boundary layer is convectively driven and topped by a capping inversion, turbulence can mix heat, moisture etc efficiently within the entire boundary layer. The turbulence then tends to distribute heat, moisture and momentum uniformly in the vertical, which leads to a well-mixed layer. This kind of boundary layer is common over land during days when the surface heating is strong and also over oceans when the sea surface temperature warms up the air near the sea surface, e.g. in cold-air advection.

2.2 Marine stratocumulus clouds

The lack of understanding of the governing physical processes makes marine stratocumulus clouds difficult to simulate in climate and weather forecast models. In most models, the turbulent fluxes are assumed to vary linearly with height and the scaling parameter is the distance from surface. However, within cloud-capped planetary
boundary layers (PBL) over the ocean, the turbulent structure is different compared to the classical PBL theory. Here turbulence is often caused by convective instability, due to the longwave radiative cooling at cloud top. The turbulent structure is then said to be “upside down”. Cold air at cloud top is descending, with compensating updrafts of warm air. For this case, the distance to surface can be unimportant which makes the turbulent parameterisation different when comparing with classical PBL theory.

The clouds and sub-cloud layers can also be either coupled or decoupled, which is also a factor that modellers have to consider. Turbulence in the cloud layer is produced by buoyancy and generally scales with the layer depth. During daytime when solar heating inside the cloud is large and fluxes from surface are small, a secondary inversion close to the cloud base can be generated. For this special case turbulence inside the cloud is only governed by in-cloud processes and the clouds can not ”feel” what is happening in the sub-cloud layer. In this case, the distance down to surface is unimportant and the two layers have turbulence that evolves in their own separate ways.

An indicator for decoupling is a local minimum in the turbulent fluxes at or near the cloud base. To decide if there exists a decoupling the buoyancy flux profile is useful. If there is a local minimum in the turbulent kinetic energy (TKE) profile, this means that the turbulence is damped here, which is also an indicator of decoupling. The variances give information about the turbulent energy and intensity. When studying the variation of the vertical velocity variance with height there are two maximums, one in each layer, if the layers are decoupled.

Due to the turbulent kinetic energy in the boundary layer, air can be mixed from above the inversion into the boundary layer; a process called entrainment (Duynkerke et al. 1995). Dry and relatively warmer air then mixes with the in-cloud air and when the dry air becomes saturated it releases latent heat. Turbulent fluxes from the sub-cloud layer are also a source of entrainment, at cloud base, with release of latent heat. The combination of radiative cooling and entrainment at cloud top drives the convection in the cloud layer.

As the decoupling reduces the height of the layer in contact with the surface, the sub-cloud layer is moistened more efficiently. Eventually, condensation may then occur at the top of this layer. If the inversion at the top of the sub-cloud layer is potentially unstable, cumulus clouds formed here may grow vertically and also penetrate all the way to the stratocumulus top. These turrets of clouds may transport heat and moisture very effectively, but their effect is difficult to estimate experimentally, since they cover a very small surface area, and are difficult to sample properly.

3. Theory

3.1 Energy spectra

Energy spectra are convenient when studying the contribution to the total turbulent kinetic energy and the partitioning between the different components in horizontal and vertical from eddies with all kind of sizes. Energy spectra can also be used to study how eddies with different sizes contribute to variations in temperature or humidity. Spectra show the distribution of energy or variance with respect to frequency.
When turbulence is measured, the turbulent eddies are usually converted to frequency, where the highest frequency represents the smallest eddies, the lower the frequency is the larger eddies are.

The energy spectrum is divided into three ranges. The highest frequencies are in the dissipation range. In the other end of the spectrum, usually called the energy containing range, are the lowest frequencies. Buoyancy or shear produces the energy in this range. The area in between, where energy is neither produced nor dissipated, is called inertial subrange. The energy is here transported by an eddy cascade down to smaller scales into the dissipation range. In the dissipation range the energy in the turbulent eddies is converted into internal energy, i.e. heat, by doing work against viscous stresses (Kaimal and Finnigan 1994).

Since energy is neither produced nor dissipated in the inertial subrange, the transfer of energy from the energy containing range down to the dissipation range, is only controlled by the dissipation, $\varepsilon$. Therefore the dissipation is the only independent parameter in inertial subrange, besides the frequency. According to Kolmogorovs (who conceived the idea of an inertial subrange) similarity theory, the energy in the inertial subrange is a function only of $\varepsilon$ [m$^2$s$^{-3}$]. From dimensional analysis the energy spectrum here is

$$E_a(\kappa) = \alpha \varepsilon \kappa^{-5/3}$$

(3.1)

where $\alpha$ is Kolmogorov constant, which is approximately 0.55, $\kappa$ is the wavenumber and subscript “a” represents one of the three velocity components, i.e. $u$, $v$ and $w$.

To apply this to measurements it is useful to express the energy spectrum in frequency, $f$, instead of wavenumber, $\kappa$. To convert between frequency scales and spatial scales, Taylor’s hypothesis is used where it is assumed that turbulent eddies advects with the mean wind:

$$\kappa = \frac{2\pi f}{u}$$

(3.2)

The velocity, $u$, is the translation speed of the eddies, normally the wind speed. For aircraft measurements, the speed of the aircraft is much larger than the wind speed. The aircraft travels through “frozen” eddies and the aircraft true airspeed, $u_a$, is used instead.

The TKE is

$$\overline{\varepsilon} = 0.5 \cdot \left( \overline{u'^2} + \overline{v'^2} + \overline{w'^2} \right)$$

(3.3)

where for example the horizontal velocity component $u$ can, in spectral analysis, be described as:

$$\int_0^\infty E_a(\kappa)d\kappa = \overline{u'^2} = \int_0^\infty S_1(f)df$$

(3.4)
Taylor’s hypothesis gives

$$\kappa \cdot E_u(\kappa) = f \cdot S_u(f)$$  \hspace{1cm} (3.5)

There are similar expressions for the v- and w-component of the wind.

If it is assumed that there exist an inertial subrange in the velocity spectra, and that the slope in the u-, v- and w-spectra is $-2/3$, the dissipation can be expressed as follows:

$$J_S(f) = \alpha e^{\frac{3}{2} \kappa^{\frac{1}{2}}}$$

$$J_{S_u}(f) = \alpha e^{\frac{3}{2} \left( \frac{2\pi}{u} \right)^{\frac{1}{2}}} f^{-\frac{3}{2}}$$

$$f^{\frac{5}{2}} [J_{S_u}(f)] = \alpha e^{\frac{3}{2} \left( \frac{2\pi}{u} \right)^{\frac{1}{2}}}$$

$$\Rightarrow \varepsilon = \frac{2\pi}{u} \alpha^{\frac{3}{2}} f [J_{S_u}(f)]^{\frac{1}{2}}$$  \hspace{1cm} (3.6)

In this study, this expression is used for calculating the dissipation.

### 3.2 Scaling

#### 3.2.1 Turbulent fluxes and variances

Expressions that are made dimensionless through division by scaling parameters are called normalised. The variance gives information about the turbulent energy and intensity while the covariance describes the kinematic turbulent fluxes. For evaluation of the scaling parameters, i.e. the friction velocity, $u_*$, convective velocity scale, $w_*$, and the temperature and humidity scales, $\theta_*$ and $q_*$, extrapolation of the fluxes was used. For the sub-cloud layer, the fluxes were extrapolated down to sea surface while for the cloud layer extrapolation up to cloud top was used. The subscript “t” refers to cloud top and “s” refers to the surface. This scaling for decoupled cloud layers was used in Tjernström and Rogers (1996). $\theta_*$ is the virtual potential temperature, defined as $\theta_* = \theta (1 + 0.61 q_{sat} - q_i)$ and $\Theta_0$ is a constant reference temperature. The height from the sea surface is scaled with the height of the sub-cloud layer, $z_i$. In the cloud layer the cloud thickness, $\Delta z$, is used as the scaling height. All vertical axes are then scaled to a thickness of unity. The figures, with scaled fluxes and variances, 6.1 to 6.7 do not show the few flight legs that gave values outside the presentation.

The scaling in the sub-cloud layer is summarised as:

$$u_*^2 = -\overline{u'w'}$$ \hspace{1cm} (3.7)

$$w_* = \left[ \frac{g}{\Theta_0} \frac{(w' \theta_v')}{z_i} \right]^{1/3}$$  \hspace{1cm} (3.8)

5
\[ \theta_s = \frac{\langle w' \theta' \rangle}{w_*} \]  
(3.9)

\[ q_s = \frac{\langle w' q' \rangle}{w_*} \]  
(3.10)

while the in-cloud scaling is:

\[ w_* = \left[ \frac{g}{\Theta_0} \langle w' \theta' \rangle \Delta z \right]^{-1/3} \]  
(3.11)

\[ \theta_* = \frac{\langle w' \theta' \rangle}{w_*} \]  
(3.12)

\[ q_* = \frac{\langle w' q' \rangle}{w_*} \]  
(3.13)

### 3.2.2 Spectra and cospectra

Spectra are here normalised in two different ways, using mixed-layer scaling (Kaimal et al. 1976) or surface-layer scaling (Kaimal et al. 1972). In the inertial subrange the energy transport down to dissipation is the only controlling parameter, as described above. This means that normalised spectra from the different flight legs should collapse into approximately the same line in the inertial subrange in both cases.

Dissipation has dimensions of \((\text{velocity})^3\) divided by length, so that normalising for the mixed layer gives:

\[ \frac{\varepsilon}{(w_*)^3} z_i = \frac{\varepsilon}{\Theta_0} \frac{z_i}{\langle w' \theta' \rangle_{z_i}} = \frac{\varepsilon}{\Theta_0} \frac{\langle w' \theta' \rangle}{\langle w' \theta' \rangle_{z_i}} = \psi_{\varepsilon} \]  
(3.14)

\[ \psi_{\varepsilon} \] represents the dimensionless mixed layer dissipation rate of turbulent kinetic energy, which is the ratio of dissipation and buoyant production. The frequency is made nondimensional by:

\[ n = \frac{f z_i}{u_a} \]  
(3.15)

so that the three velocity spectra in the mixed layer are normalised with \(w_*\) and \(\psi_{\varepsilon}\):

\[ \frac{f S_*(f)}{w_*^2 \psi_{\varepsilon}^{2/3}} = \frac{\alpha}{(2\pi)^{2/3}} n^{-2/3} \]  
(3.16)
Surface-layer velocity spectra are similarly normalised by $u_*$. The corresponding non-dimensional dissipation rate for surface layer is $\psi$ and scales the energy content. The non-dimensional frequency in surface-layer scaling is $\eta = fz / u$, where $z$ is the altitude of the flight legs. Together this causes the spectra in the inertial subrange to collapse on to the same line.

Cospectra in the cloud layer are normalised with mixed-layer scaling, so that $w_*$ and $\theta_*$ are used to make cospectra of $uw$ and $w\theta$ non-dimensional. Cospectra of $uw$ and $w\theta$ in the sub-cloud layer are nondimensionalized using $u_*$ and the corresponding $\theta_*$. The normalised frequency is the same as for the spectra.

4. The experiment

Aircraft measurements from the "First Lagrangian" of the Atlantic Stratocumulus Transition Experiment (ASTEX) are used in this study. The goal of ASTEX was to characterise the evolution of cloudiness and the vertical structure in a marine boundary layer (MBL) as it moves over a warmer surface (Bretherton and Pincus 1995). ASTEX examined the evolution of clouds, dynamical and vertical structure in a marine boundary layer air mass (Bretherton and Pincus 1995).

The ASTEX First Lagrangian data-area was in the Atlantic, south-east of the Azores Islands at about 28° - 40° N, 23° - 28° W. The data was collected during five research flights, by the British Meteorological Research Flight (MRF) C-130 and by the National Center for Atmospheric Research’s (NCAR) Electra aircraft. The measurements were collected with a Lagrangian strategy. That is, to follow approximately the same air mass when it moves with the mean boundary layer wind while measuring the physical processes and the evolution of cloudiness.

The First Lagrangian took place in a clean marine air mass from 1719 UTC 12 of June to 1302 UTC 14 of June 1992. For the first two days, the wind was about 10 m/s from north but for the third day the winds were calmer, at about 5 m/s and from northeast. The air mass moved over a sea surface where the temperature increased by 4 degrees, from 17°C to 21°C (Bretherton et al. 1995).

Trajectory and cloudiness forecasts were used to decide when to start collecting data by the aircrafts. The aircrafts were coordinated to fly and collect data continuously in time. However, during the last day there was a 14-hour gap, due to poor visibility at the airport on the island of Santa Maria in the Azores, where the aircrafts were stationed. When collecting data, the MRF C-130 flew crosswind legs while the NCAR Electra flew both crosswind and along-wind legs of about 60 km length. The aircraft flew well below cloud base, during daytime at 30-50 m and at night 150 m, slightly beneath the stratocumulus cloud base, inside the cloud sheet and above the clouds. Two research ships were also used during the campaign, measuring sea surface temperature and chemicals and made soundings. Balloons and satellites also made observations.
4.1 Cloud evolution

The air mass that was followed during the First Lagrangian had been a part of a marine air mass moving eastward across the North Atlantic. For the entire period of the First Lagrangian a solid stratocumulus layer was advected and the air mass moved southward (Bretherton and Pincus 1995). During the first day, the cloud top and cloud base of the solid stratocumulus sheet was at approximately constant altitudes. During the second day, the entire layer rapidly deepened while the cloud thickness remained approximately constant. Cumulus clouds then started to form in the sub-cloud layer, rising up into the stratocumulus cloud, which became thinner. The cumulus activity increased during the day and some cumulus clouds even reached the inversion. During the third day, cumulus clouds were still penetrating into thin and broken stratocumulus cloud layer.

Figure 4.1. The cloud evolution during the First Lagrangian of the ASTEX was between 1719 UTC 12 June 1992 and 1302 UTC 14 June 1992. The horizontal distance between flight 1 and 5 is approximately 1300 km. The grey scale in the lowest horizontal bar represents changing sea surface temperature, with increasing values from left to right. After de Roode and Duynkerke (1997).

5. The data

The data available from the two aircrafts contains the basic meteorological variables. There are two sets of data: one high rate data set, at 32 Hz from MRF C-130 and at 20 Hz from NCAR Electra, and one slow rate data set at 1 Hz from both aircrafts. In each of the five flights, there were 16 to 25 flight legs with a typical aircraft sampling length of about 60 km. The aircraft measurements were taken at different heights in cloud layer, in sub-cloud layer and over the stratocumulus clouds.

The very first step was to detrend the data by high-pass filtering to eliminate any trend and low-frequency noise. This was done at a low frequency, determined from the length of each flight leg. Next the data was high-pass filtered again, this time at a higher frequency, to approximately separate turbulence from mesoscale or other atmospheric low-frequency variability. To estimate this so called cut-off frequency, subjective study of the spectra was used. All flight legs from the same flight was fil-
tered using the same cut-off frequency. A problem in choosing this filter frequency is
the risk that it removes important information in the low frequency range. The lack of
a clear gap in most of these spectra makes the choice somewhat subjective. The cut-
off frequencies range from 0.007 Hz to 0.04 Hz for different flights. This data pro-
cessing was made using the Fast Fourier Transform (FFT), which is very useful for
treating this kind of data. FFT convert a noisy original signal to an applicable data set
divided in frequency. The output from the FFT is the variance, covariance, spectra
and cospectra used in this study.

The TKE dissipation was estimated from the velocity spectrum, using the assumption
of an inertial subrange, see above. The peak wavelengths were taken from the spectral
and cospectral peaks.

6 Results

6.1 Turbulence statistics

Prior to the scaling of the turbulence statistics, the unscaled fluxes for the individual
flights were examined to determine the layer structure. This was done in order to see
if there was any substantial decoupling between the sub- and in-cloud layers. The tur-
bulent fluxes were primarily used for this; if the buoyancy flux profile featured a local
minimum at or close to the cloud base, this was used as an indicator for decoupling.
Four of the five flights have a buoyancy flux profile with a clear minimum at or near
cloud base. For the only night flight, the buoyancy flux has a minimum well below
the cloud base. The last flight also has a somewhat different buoyancy-flux profile,
where the buoyancy fluxes were very small inside the cloud.

The normalised data is thus divided into a sub-cloud layer and an in-cloud layer. As
the in-cloud turbulence is mainly driven by longwave radiative loss from the cloud
top, the turbulent characteristics are expected to be that of a well-mixed layer but
“upside down”. Turbulent statistics in the in-cloud layer is thus scaled using
convective scaling, as described above. The sub-cloud layer is normalised both using
surface-layer and mixed-layer scaling.

To determine the scaling parameters, the buoyancy fluxes and the humidity fluxes are
extrapolated to the sea surface for the sub-cloud layer and to the cloud top for in-
cloud layer. The vertical axis is scaled so that each layer has a thickness of unity. The
normalised buoyancy flux profiles from all the flights are shown in figure 6.1.

The scaled variances from all the five flights are shown in figures 6.2-6.4 for the w-, u
and v-components, respectively. In figure 6.5 the normalised TKE is shown. The
dashed lines in the plots of the sub-cloud variances are taken from Brost et al. (1982).
The dashed lines in the plots for the buoyancy fluxes and the momentum fluxes
represent the expected linear decrease in the fluxes away from the surface or the cloud
top in the sub- and in-cloud layers, respectively.
Figure 6.1. Normalised buoyancy flux for the cloud and the sub-cloud layer. Dashed line represents theoretical linear decrease. Plot symbols: ◦ flight 1, x flight 2, ● flight 3, + flight 4 and Δ flight 5.

Table 6.1. Scaling parameters and layer depth for the in-cloud and the sub-cloud layer.

<table>
<thead>
<tr>
<th>Flight</th>
<th>Flight 1</th>
<th>Flight 2</th>
<th>Flight 3</th>
<th>Flight 4</th>
<th>Flight 5</th>
</tr>
</thead>
<tbody>
<tr>
<td>In-cloud layer:</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>$w_0$ [m s$^{-1}$]</td>
<td>0.53</td>
<td>0.75</td>
<td>0.60</td>
<td>0.58</td>
<td>0.34</td>
</tr>
<tr>
<td>$\theta_0$ [K]</td>
<td>0.019</td>
<td>0.027</td>
<td>0.022</td>
<td>0.021</td>
<td>0.003</td>
</tr>
<tr>
<td>$q_0$ [g kg$^{-1}$]</td>
<td>0.028</td>
<td>0.003</td>
<td>0.017</td>
<td>0.069</td>
<td>0.003</td>
</tr>
<tr>
<td>$\Delta z$ [m]</td>
<td>440</td>
<td>605</td>
<td>470</td>
<td>470</td>
<td>1100</td>
</tr>
<tr>
<td>Sub-cloud layer:</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>$u_0$ [m s$^{-1}$]</td>
<td>0.20</td>
<td>0.32</td>
<td>0.30</td>
<td>0.28</td>
<td>0.11</td>
</tr>
<tr>
<td>$w_0$ [m s$^{-1}$]</td>
<td>0.17</td>
<td>0.30</td>
<td>0.31</td>
<td>0.40</td>
<td>0.26</td>
</tr>
<tr>
<td>$\theta_0$ [K]</td>
<td>0.003</td>
<td>0.017</td>
<td>0.001</td>
<td>0.008</td>
<td>0.004</td>
</tr>
<tr>
<td>$q_0$ [g kg$^{-1}$]</td>
<td>0.029</td>
<td>0.051</td>
<td>0.041</td>
<td>0.050</td>
<td>0.027</td>
</tr>
<tr>
<td>$z_0$ [m]</td>
<td>300</td>
<td>150</td>
<td>300</td>
<td>600</td>
<td>500</td>
</tr>
</tbody>
</table>

Flight 1
For flight 1, a minimum in the buoyancy flux near the cloud base is easily seen. The buoyancy flux then increase within the cloud, with maximum near cloud top (figure 6.1). The vertical velocity variance (figure 6.2) also seems to have a local minimum near the cloud base and then an in-cloud maximum approximately at same level as the maximum buoyancy flux. The TKE (see figure 6.5) is quite scattered, but it appears to
have two local maximum, one in the cloud layer and one in the sub-cloud layer. During this flight decoupling of the layers were assumed.

**Flight 2**

In flight 2, when comparing turbulent fluxes and the vertical velocity variance it is difficult to find a clear decoupling. One reason for this can be that this is a night flight. The vertical velocity variance reaches a maximum some distance below the cloud base. The buoyancy flux is positive in the cloud layer and further down, under the cloud base. Still, the two layers are assumed to be decoupled, because of the local minimum in the buoyancy flux.

![Figure 6.2. Normalised vertical velocity variance. Plot symbols; o flight 1, x flight 2, o flight 3, + flight 4 and Δ flight 5. Dashed line represents theoretical linear decrease from Brost et al. (1982).](image)

The TKE and the wind speed variances, $u'^2$ and $v'^2$, decreases with height all the way from the surface and up to the cloud top. The momentum flux, however, is approximately zero in the cloud, see figure 6.6. This case thus seems to have a mixture of properties, both in-cloud mixed layer properties, without much height dependence, and sub-cloud properties where turbulence decay away from the surface. Perhaps part of the turbulence in the cloud layer is generated by shear at the surface, causing the variances to continuously decay, while part of the sub-cloud turbulence is due to buoyant thermals generated by cloud-top cooling. The minimum in the buoyancy flux below the cloud base represents the height to which these thermals can be seen. The absence of solar radiation prevents the formation of an inversion at the cloud base,
thus inhibiting a clear division of the two layers. This results in a combination of bottom-up and top-down properties.

**Flight 3**
Also in the third flight, it is difficult to see a clear minimum in the vertical velocity variance near the cloud base, but it do increase inside the cloud. The buoyancy flux profile also increases inside the cloud. The momentum flux decreases with height in the sub-cloud layer up to the cloud base, but is very small and constant in the cloud, indicating that one can assume that the sub-cloud and cloud layers are decoupled.

**Flight 4**
For flight 4, two separate maximums in the vertical velocity variance were found, one in the sub-cloud and one in the cloud layer, suggesting that these two layers were decoupled. The buoyancy flux has a minimum at the cloud base and a maximum inside the cloud, which also indicates a decoupling. Furthermore, the TKE has also a minimum at the cloud base, which means that the turbulence is damped. If the cloud layer is decoupled from the sub-cloud layer it seems that turbulence generated from cloud top reach down to the cloud base.

![Figure 6.3. Normalised wind speed variance for the u-component. Plot symbols: o flight 1, x flight 2, ◊ flight 3, + flight 4 and Δ flight 5. Dashed line represents theoretically decrease from Brost et al. (1982).](image-url)
**Flight 5**

During flight 5, the character of the PBL has changed. The boundary layer is quite deep and the stratocumulus cover at the top is becoming thin and broken. In the sub-cloud layer, cumulus clouds dominate that penetrates through the entire layer and into the thin and broken stratocumulus clouds above (deRoode and Duynkerke 1997). Because of the highly temporal character of the cumulus clouds, it is difficult to find any common turbulent structures indicating decoupling. Still, decoupling is here assumed to be at the height were the fluxes of momentum, sensible heat and buoyancy have a local minimum, which is actually the stratocumulus cloud base.

![Figure 6.4](image.png)

**Figure 6.4.** Normalised wind speed variance for the v-component. Plot symbols; o flight 1, x flight 2, o flight 3, + flight 4 and Δ flight 5. Dashed line represents theoretical decrease from Brost et al. (1982).

In summary, the buoyancy flux in the sub-cloud layer decreases linearly with height, starting at a positive value near the surface, and becoming slightly negative near the cloud base. This behaviour is similar to a continental cloud free PBL. For flight 1, the buoyancy flux is negative throughout much of the sub-cloud layer, which may indicate a significant entrainment from above or more or less stable stratification. It could also be due to latent cooling as drizzle drops evaporate in the sub-cloud air. The buoyancy flux in the cloud layers is in general close to zero at the clouds base and then increases towards the cloud top. This confirms the theory about the cloud top
cooling being the driving mechanisms for turbulence generation in the cloud. The only night-time flight, flight 2, deviates somewhat in the sense that the thermals\(^1\) from the cloud top seems to penetrate below the cloud base (Nicholls 1989), but is otherwise similar to the main feature. The last flight does not follow this structure, which is probably due to the cloud development during the flight with many cumulus towers penetrating from the sub-cloud layer into the stratocumulus. These clouds can be very efficient in transporting heat and moisture, but due to their local character, they are difficult to sample with an aircraft.

The momentum fluxes in the cloud layer is small and constant with height which indicates clearly that the sub-cloud and in-cloud layers were decoupled. Above the sub-cloud layer the momentum fluxes are around zero for all flights. Momentum flux profiles in the sub-cloud layer behaves as expected, with a linear decrease from the surface in all flights.

The humidity flux also behaves as expected in a mixed layer through both layers, except during flight 4, see figure 6.7. In flight 4, the humidity flux increases in the cloud layer to the cloud top. This can have something to do with the increased cumulus activity during the flight. Cumulus clouds sometimes start to penetrate into the stratocumulus clouds, some cumulus reaching as high up as the stratocumulus cloud top.

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\(^1\) The word thermal is here, and onward, used to represent downward moving cold packets of air from the cloud top rather than the opposite which is more common.
The positive values near cloud top can be related to dry air being entrained down into the cloud layer.

The dashed lines in the plots of the scaled variances for the sub-cloud layer is a parameterisation from Brost et al. (1982). These lines fit this data well for the variance of the u- and w-components of the wind. An exception is flight 2, where the vertical velocity variance increases instead through the sub-cloud layer. This is probably related to the fact that this is a night-time flight. As discussed above, this means that there is no inversion at the cloud base, suppressing the vertical motion. The measured variance of the v-component is much larger than indicated by the dashed line. This indicates that the wind-direction shear in the sub-cloud layer is important, most so in the first three flights, which is before the sea surface temperature has started to increase substantially. As expected the variance is the smallest in the w-component, indicating wind shear to be the main generator of the turbulence. In the cloud layer the variance of three wind components were approximately of the same size, which is a clear indication that turbulence is here generated by buoyancy and then redistributed to the horizontal components.

\[ u'w'/w^2 \]

**Figure 6.6.** Normalised momentum flux. Plot symbols; o flight 1, x flight 2, ◊ flight 3, + flight 4 and Δ flight 5. Dashed line represents theoretically linear decrease.
Figure 6.7. Normalised humidity flux. Plot symbols; o flight 1, x flight 2, o flight 3, + flight 4 and Δ flight 5.

6.2 Spectra

The atmospheric turbulence spectra cover a wide range of frequencies, here from 10 s$^{-1}$ to 0.001 s$^{-1}$. Because of the wide range in frequency, the best way to present this is in a log-log representation. Focusing on the spectral peak, just multiply the spectral energy density $S(f)$ with frequency $f$, without losing information about higher frequencies. If the spectra are presented as log $[f \times S(f)]$ vs. log $f$, the area under the curve is, however, not proportional to the variance anymore. The high frequency end of spectra for flight 2 and 5 has a substantial increase at higher frequencies, which is not the case for the other flights. Both these flights were flown by the MRF C-130 and this may have something to do with the instrumentation and/or the sampling on this aircraft. This may perhaps be an effect of aliasing, where unresolved eddies creates an erroneous energy at the highest frequencies that are possible to represent correctly, at the Nyquist frequency.

The spectra from flight 5 are not shown because the results from this flight appear different. As discussed previously a possible reason can be the complicated cloud structure, with cumulus cloud penetrating into the stratocumulus clouds. These two kinds of clouds develop in their own way and at the same time interact, which make the normalisation troublesome. Scaling in the inertial subrange did not work out as expected for neither the sub-cloud nor the in-cloud layer since the $-2/3$-slope was not
fulfilled for all flight legs. The buoyancy flux differs comparing with the other four flights. The buoyancy flux was small from approximately 1000 m and up to cloud top and decreased in the cloud layer (figure 6.1).

The spectra in the cloud layer were normalised using mixed layer similarity. The spectral energy for the mixed layer were normalised by $w*$ and $\psi_1$ as described in the theory above. The frequency is normalised with layer thickness and by the aircraft true airspeed, $n = f \Delta z / u_a$, and equals unity at the spectral peak if the typical eddy sizes are equal to the layer depth. At the spectral peak, the wavelength $\lambda_m$, is representative to the eddy sizes with most energy. For the mixed layer it is expected that the most energetic eddies are related to wavelengths which correspond to the layer depth. As the scaling parameters $\Delta z$ and $w*$ are not varying with height in the mixed layer, the spectra are expected to be invariant with height (Kaimal and Finnigan 1994). In figure 6.8 this is illustrated using in-cloud layer flight legs for flights 1–4. The data is here averaged into ten height intervals, each 0.1$\Delta z$ thick, from the cloud base to the cloud top.

The normalisation in the sub-cloud layer was made in two ways, using surface-layer scaling and mixed-layer scaling. For the mixed-layer scaling the convective velocity-scale, $w*$, is used and the frequency is normalised with the layer depth, $z_i$. For the surface-layer scaling $u*$ is used while the frequency is normalised with the altitude of the flight legs from sea surface. The spectral energy normalised with $u*$ and $\phi_0$ is determined by trial and error to get the inertial subrange collapsed onto one curve. See figure 6.9 for surface-layer scaling and the mixed-layer scaling for flights 1–4 together. The sub-cloud spectra were averaged in the same way as for the in-cloud spectra.

The dashed lines in the figures represents flight legs which are near the sea surface or the cloud base, the sub-cloud the cloud layer, respectively. This is defined as flights that are within less than 20 % of $z_i$ for sub-cloud layer, and of $\Delta z$ in the cloud layer. For sub-cloud flight legs near cloud base or in-cloud flights near the cloud top dotted lines are used, also here this defined as being within 20 % from top of each layer respectively.

Figure 6.8 shows averaged values for the first four flights in the cloud layer. The mixed-layer scaling in the w-spectra indicates peak wavelengths at approximately 2$\Delta z$. Small peak wavelength were found close to cloud top, $(\lambda_m)_w = 0.6\Delta z$, and largest peak wavelength were found at the altitude interval (0.2-0.29)$\Delta z$ where $(\lambda_m)_w = 3.5\Delta z$. Nucciarone and Young (1991) found the smallest peak wavelength nearest cloud top for marine stratocumulus due to the capping inversion. One should also expect small peak wavelength close to the cloud base, but $(\lambda_m)_w = 2.0\Delta z$ is consistent when study each flight alone at the altitude nearest the cloud base. A problem with the averaging process here is the different numbers of flight legs over which the averaging is made. Some altitude intervals only average two flight legs. In the w-spectra, one level within the centre of the cloud has a peak wavelength that is smaller than the cloud layer depth; this happened to be from only two flight legs from the one night flight. If the averaging were made using five intervals instead of ten, each average would contain more flight legs as the intervals would be too large but some information on the vertical structure could also be lost this way.
In the cloud layer the u- and v-spectra show larger peak wavelengths than the w-spectra. The mixed-layer scaling for u-spectra seems to approximate \((\lambda_m)_u = 3\Delta z\) and more to be more scattered for the v-spectra which have a tendency of double peaks. Peak wavelengths for u- and v-spectra are assumed to be constant with height in the mixed layer, \((\lambda_m)_{u,v} = 1.5\Delta z\), (Kaimal and Finnigan 1994). Peak wavelengths in upper part of the cloud layer indicate an increase with maximum at cloud top, which was also found in Nicholls 1989. This local maximum can be explained by the capping inversion that suppresses vertical motions and causes horizontal motions to increase. Shear produced turbulence indicate a maximum in the horizontal velocity variance towards cloud top (Stull 1988).

Figure 6.9 shows averaged values from u-, v- and w-spectra in the sub-cloud layer, as previous without flight 5. As with the cloud layer, the sub-cloud layer indicates larger peak wavelengths for u- and v-spectra than in the w-spectra for both the surface-layer and mixed-layer scaling. The peak wavelength in v-spectra is scattered and difficult to
Figure 6.9. Average values over normalised u-, v- and w-spectra from the sub-cloud layer where flight legs are divided into ten intervals depending on altitude. The averaging made from flight legs from flight 1 to 4. The left frames present surface-layer scaling and the right frames mixed-layer scaling. Dotted lines represent the two intervals close to cloud base and the dashed lines represent the two lowest intervals close to the sea surface.

decide for both scaling processes. The peak wavelengths in w-spectra are approximately $3z_i$ for the surface-layer scaling. A slight decrease for both peak wavelengths close to cloud base and sea surface where $(\lambda_m)_w = 1.8z_i$ for the surface-layer scaling. Largest peak wavelength at altitude interval (0.2-0.29)z_i where $(\lambda_m)_w = 5.2z_i$. When comparing with the individual flights, flight 2 and 3 have maximum in peak wavelength at this altitude. The dotted spectral line that differs a lot comes from two flight legs from the night flight. For surface-layer scaling there are a division in the low-frequency area for the horizontal spectra, which may indicate that the scaling with z is
not useful for all frequencies. Kaimal (1978) found from experiments that horizontal velocity field follows different similarity laws in different regions of the spectral range. The scaling with mixed-layer theory did not work out well for the sub-cloud layer. But for the w-spectra without the two lowest intervals, the two dashed lines, the scaling in the inertial subrange is rather good. The w-spectra have smallest peak wavelength near the sea surface and then, over altitude 0.2z_i, are the peak wavelength approximately between 1z_i to 3z_i. The dotted line that differs in the inertial subrange for the three spectra is an averaging from two flight legs from the same flight, flight 1.

6.2.1 Eddy sizes and energy

All flight’s u-, v- and w-spectra are studied to get an idea of how the spectral peak is related to the sub-cloud or the cloud layer depth. The assumption of decoupling would give typical eddy sizes in w-spectra related to the layer depth.

*Flight 1*

For flight 1 the w-spectra for in-cloud layer shows that all of the flight legs has spectral peak at eddy sizes larger than the layer depth. Flight legs from the upper part of the cloud indicate the smallest eddy sizes with peak wavelengths, (λ_m)_w = 1.1Δz and in lower part of cloud (λ_m)_w = 1.1Δz to 2.8Δz. Spectral peak for u and v are mostly larger than for w-spectra and there is a tendency for spectral peak in v to be larger than in u-spectra. Otherwise no common structure between flight legs from same altitude in u- and v-spectra is found. Peak wavelength in the u-spectra is approximately 3Δz and for the v-spectra 4Δz.

![Graphs showing normalized spectra for flight 1 in the cloud layer.](image)

*Figure 6.10.* Normalised mixed layer u-, v- and w-spectra for flight 1 in the cloud layer.
In the sub-cloud layer in the w-spectra eddies at spectral peak with smallest wavelengths are found for flight legs from 0.17z_i. The peak wavelength ranges between (\lambda_m)_w = 0.9z_i and 1.7z_i for these legs. The ocean surface may act like a wall and the vertical motions of the eddies were damped. Otherwise the spectral peak in the w-spectra indicate eddies with local maximum at altitude 0.6z_i where (\lambda_m)_w = 3.2z_i. The u- and v-spectra indicate two peaks, especially for flight legs from 0.17z_i and 0.6z_i. Peak wavelength for the six flight legs from 0.17z_i is quite similar with most (\lambda_m)_u between 2z_i and 5z_i and (\lambda_m)_v between 1z_i and 4z_i. Peak wavelengths were even larger in the other flight legs at higher altitudes in the sub-cloud layer.

Figure 6.11. Normalised u-, v- and w-spectra for sub-cloud layer flight 1, surface-layer scaling. Dotted lines represent flight legs near the cloud base and dashed lines represent flight legs near the sea surface.
Flight 2
The night flight, flight 2, has the highest flight legs at altitude, 0.5Δz, in the cloud layer. Flight legs higher up were flown in a zigzag pattern through the cloud top, and could not be used for this analysis. Flight legs at lowest altitude in w-spectra have the largest peak wavelength, (λ_m)_w between 2Δz and 6Δz. The peak wavelength for the w-spectra then slightly decreases with altitude. Flight legs from 0.25Δz have (λ_m)_w between 2Δz and 4Δz and smallest peak wavelength for flight legs from 0.5Δz where (λ_m)_w = 0.5Δz to 1.5Δz. For this flight eddies are more extended in the horizontal and v-spectra indicate a larger peak wavelength than in the u-spectra. Several flight legs from the three different altitudes have (λ_m)_u = 1.5Δz and the v-spectra is more scattered, (λ_m)_v = 3.5Δz to 9.5Δz.

Figure 6.12. Normalised mixed layer u-, v- and w-spectra for flight 2 in the cloud layer. Dashed lines represent flight legs near the cloud base.
The sub-cloud w-spectra show that eddies were larger at altitude 0.25z\textsubscript{i} than for 0.7-0.8z\textsubscript{i}. The eddies in all flight legs were at spectral peak larger than layer depth and were at spectral peaks 10 times larger than the layer depth. Spectral peaks in u and v were also larger than layer depth and v-spectra indicate a larger peak wavelength than for u-spectra. The three flight legs from 0.7-0.8z\textsubscript{i} were very similar in u- and v-spectra, (λ\textsubscript{m})\textsubscript{u} = 7.0z\textsubscript{i} and (λ\textsubscript{m})\textsubscript{v} = 30z\textsubscript{i} and the three legs close to the sea surface have even larger peak wavelengths. When comparing between in-cloud and sub-cloud spectra, it is clear that there is more energy in the sub-cloud spectra. The spectral peaks for the sub-cloud layer was at lower frequencies than for the in-cloud layer, indicating larger eddies.

![Normalized spectra](image)

**Figure 6.13.** Normalised u-, v- and w-spectra for sub-cloud layer flight 2, surface-layer scaling. Dashed line represents flight legs near the sea surface.
**Flight 3**

Five of seven in-cloud flight legs in flight 3 for the w-spectra have typical eddy size at $(\lambda_m)_w = 3\Delta z$ and these legs are from different altitudes. In the u-spectra, there is a tendency for some flight legs to have double peaks. When looking at the peaks at the highest frequency they are between $(\lambda_m)_u = 1.8\Delta z$ and $(\lambda_m)_u = 2.9\Delta z$. The rest of the flight legs, close to the cloud base and near the cloud top shows no common structure, but peak wavelengths are larger than the layer depth. The v-spectra indicate larger eddies at spectral peak than the u-spectra.

![Graphs of u, v, and w spectra](https://via.placeholder.com/150)

**Figure 6.14.** Normalised mixed layer u-, v- and w-spectra for flight 3 in the cloud layer. Dotted line represents flight legs near the cloud top and dashed lines represent flight legs near the cloud base.
Flight legs in the sub-cloud layer in the w-spectra have peak wavelength that are larger than the layer depth, and the largest peak wavelength found at 0.25z_i, \((\lambda_m)_w\) varies from \((\lambda_m)_w = 15z_i\) at 0.25z_i to \((\lambda_m)_w = 2.5z_i\) in the middle of the sub-cloud layer. Good conclusions for the sub-cloud layer can be difficult because there are only flight legs at normalised heights \(z_i = 0.25\) and \(z_i = 0.4\). The u- and v-spectra indicate double peaks and especially for the v-spectra it is difficult to decide the peak wavelength. Peak wavelengths are larger in u- and v-spectra compared to w-spectra. Also for u and v, the peak wavelength is larger at altitude 0.25z_i than at 0.4z_i.

![Figure 6.15. Normalised u-, v- and w-spectra for sub-cloud layer flight 3, surface-layer scaling.](image-url)
Flight 4
Flight 4 has flight legs from four different altitudes inside the cloud layer. In w-spectra, the flight legs close to the cloud top and the cloud base indicate peak wavelength equal to the layer depth or even smaller, $(\lambda_m)_w = 0.4\Delta z$ for the smallest peak wavelength. Flight legs close to the cloud base have two peaks. In the middle of the cloud layer, the peak wavelength is larger than the layer depth, $(\lambda_m)_w = 1.5\Delta z$ to $2.0\Delta z$. Peak wavelength for the u- and v-spectra are quite similar with eddy sizes larger than the layer depth, $(\lambda_m)_u$ and $(\lambda_m)_v$ are approximately $3\Delta z$. It is difficult to find any common structure for the u- and v-spectra when comparing between different altitudes. Perhaps this has something to do with the fact that collecting of data takes time, and the boundary layer at different heights is not measured at the same time.

Figure 6.16. Normalised mixed layer u-, v- and w-spectra for flight 4 in the cloud layer. Dotted lines represent flight legs near the cloud top and dashed lines represent flight legs near the cloud base.
In the sub-cloud layer, w-spectra have spectral peak at frequencies related to eddy sizes larger than sub-cloud layer depth. Flight legs close to the surface have peak wavelength in w-spectra between $\lambda_m = 2z_i$ and $\lambda_m = 5z_i$. In the middle of the sub-cloud layer $\lambda_m = 3z_i$ to $5z_i$ and in upper part, at $0.8z_i$, $\lambda_m = 2z_i$ to $4z_i$. There seems to be larger peak wavelengths for the u- and v-spectra, compared to the w-spectra.

![Figure 6.17](image)

**Figure 6.17.** Normalised u-, v- and w-spectra for the sub-cloud layer flight 4, surface-layer scaling. Dashed lines represent flight legs near the sea surface.

### 6.2.2. Summary

**Cloud layer**

Typical eddy sizes for w-spectra in the cloud layer are mostly larger than the layer depth for all four flights. Flight legs with peak wavelength equal to or smaller than layer depth are found in the upper part of the cloud layer, close to the inversion. All the four flights that were analysed indicate that the horizontal u- and v-spectra at all levels have peak wavelengths that are larger than the peak wavelength in w-spectra. For flight 1, 2 and 3 the v-spectral peak seems to be larger also than that in the u-spectra. When scaled with mixed-layer scaling, most of the spectra collapse on top of each other and seems to be invariant with height. The spectral peaks in w- and u-spectra are generally within 1-3Δz, although in some cases the spectral peak is sub-
stantially larger. This indicates that mixed-layer scaling is appropriate and that cloud-top buoyancy is the main process that generates turbulence in the cloud. However, spectral peaks in the v-spectra seems to be substantially larger than those in both the u- and w-spectra, sometimes as large as $10\Delta z$. The reason for this is not clear, but could be associated with some mesoscale circulation.

**Sub-cloud layer**

Peak wavelengths in the w-spectra are larger than the layer depth for almost all flight legs in all flights. Peak wavelength are smaller than layer depth for w-spectra in flight 1 for some of the flight legs close to surface, possibly suppressed in the vertical due to the vicinity of the sea surface. The magnitude for the peak wavelength in u-, v- and w-spectra for the night flight differs from the other flights, and the peak wavelengths were largest in flight 2. The sub-cloud layers show that the peak wavelengths in u- and v-spectra are larger than in the w-spectra. Mixed-layer scaling does not collapse spectra from different flights and different heights well. This seems to invalidate the use of this scaling here. However, using surface-layer scaling, taking the height variation of the spectral peak into account, collapses the spectra only marginally better. In all likelihood, with the changing surface conditions during this Lagrangian experiment, no single scaling applies to all the spectra from start to finish.

Finally, some sub-cloud spectra also seem to feature double peaks to a larger extent than the in-cloud spectra. This may be interpreted two ways. It may be a consequence of flying through isolated cumulus clouds that develops below the clouds during the later parts of the Lagrangian. However, as these clouds also penetrate into the clouds (Wang and Lenschow 1995), one would perhaps expect this to show up also in the in-cloud spectra. The alternative explanation is that some of the most energetic thermals formed in the cloud layer actually makes it down into the sub-cloud layer, although the turbulent flux is zero or close to zero at a level in-between. This was also observed by Khalsa (1992) when analysing FIRE data from Californian stratus using conditional sampling. If this is indeed the case also here, this shows that the issue of defining decoupling is far from solved.

**6.2.3. Cospectra**

The normalization for cospectra is made similar as for spectra. In sub-cloud layer the friction velocity and temperature scale is used for normalization of cospectra. The normalization in the cloud layer is with matching vertical velocity scale and temperature scale. The dotted and dashed lines also made in the same way as for the spectra. In figure 6.18 flight 1-4 is represented together for both sub-cloud and in-cloud layer and the averaging made in the same way as for spectra.

The averaging for cospectra of wθ in the cloud layer have a tendency of linear increase of heat flux with maximum at cloud top and positive heat flux, see figure 6.18. The cospectra were quite noisy, but the eddy sizes larger than layer depth and the cospectra peak at the highest frequency are approximately $2\Delta z$. There seem to be a second peak at a lower frequency for peak wavelength approximately $10\Delta z$ with also positive heat flux.
The cospectra of $w\theta$ in the sub-cloud layer show that almost all the negative contribution to the flux is represented by the upper part of the layer. Also the cospectra of $w\theta$ have peak wavelength larger than layer depth. Peak wavelength approximately between $2z_i$ and $5z_i$. The peak at the lower frequency shows up here also with approximately the same wavelength, $10z_i$. This peak with downward heat flux is only seen in the heat flux from upper part of the sub-cloud layer.

In the sub-cloud layer the cospectra of $uw$ shows negative flux as expected. The most contributions come from the lower part of the sub-cloud layer. Peak wavelength larger than layer depth and approximately between $1.5z_i$ to $5z_i$. Also here at a lower frequency, where $(\lambda)_{uw}$ is approximately $10z_i$, there is a peak with downward flux through almost the whole sub-cloud layer. As expected there is little turbulent production from cospectra of $uw$ in the cloud layer when comparing with momentum flux profile, figure 6.6.
**Flight 1**

In the cloud layer, $\omega$ cospectra shows that the four flight legs from the same altitude, $0.3\Delta z$, are quite different. One of them has a negative heat flux. There is an indication that the heat flux has two peaks, one were the peak wavelength is approximately $2\Delta z$ and the other approximately $8\Delta z$. In upper part of the cloud layer the peak wavelength increases marginally, at highest measured altitude, $0.75\Delta z$, it was $(\lambda_m)_{\omega \theta} = 2.2\Delta z$.

The peak wavelength in the sub-cloud cospectra of uw quite equal for the three measured altitudes. For flight legs near the sea surface, $0.17z_i$, and at $0.6z_i$ the $(\lambda_m)_{uw}$ are approximately $2.5z_i$. Near the cloud top at $0.98z_i$ the two flight legs have $(\lambda_m)_{uw} = 3.0z_i$ and $(\lambda_m)_{uw} = 1.5z_i$.

In the sub-cloud layer for $\omega \theta$ cospectra, the fluxes were positive for some flight legs near the sea surface. When comparing with scaled buoyancy flux profile, see figure 6.1, the only positive fluxes also comes from lower part of the layer. The flight legs from upper part indicate negative fluxes. Peak wavelength for negative fluxes are approximately $3z_i$ with some scatter.

![Figure 6.19](image)

**Figure 6.19.** Normalised cospectra of uw and $\omega \theta$ for flight 1. Dashed lines in the sub-cloud spectra represent flight legs near the sea surface and dotted lines near the cloud base.
Flight 2

The wθ cospectra in the cloud layer show positive fluxes for all flight legs except one. Peak wavelength approximately the same for all flight legs, \((\lambda_m)_{wθ} = 9\Delta z\).

In the sub-cloud layer the peak wavelength for all flight legs in the uw cospectra were more than 10 times larger than the layer depth and peak wavelength seems to decrease with minimum near the cloud base. To decide values for peak wavelengths is rather difficult because of the large scatter.

The wθ cospectra in the sub-cloud layer also show large eddy sizes. Flight legs from the lower part of the sub-cloud layer indicate positive fluxes, while in the upper parts the fluxes are negative.

Figure 6.20. Normalised cospectra of uw and wθ for flight 2. Dashed lines represent flight legs near the sea surface for sub-cloud and near the cloud base for the cloud layer.
Flight 3
This flight has the largest peak wavelengths for the cospectra of \( w_0 \), in general, comparing to the other flights. The cospectra of \( w_0 \) in the cloud layer indicate positive fluxes for all flight legs except one. Peak wavelength for several flights legs from different altitudes are approximately 10 times larger than layer depth.

All the flight legs for the \( w_0 \) and \( u_w \) cospectra are from the lower part in sub-cloud layer and because of this it is difficult to find any common structures for the whole sub-cloud layer. The flight legs in cospectra of \( u_w \) have negative turbulent fluxes as expected. All legs show peak wavelength larger than layer depth, some even more than 10 times larger.

All flight legs in cospectra of \( w_0 \) in the sub-cloud shows positive fluxes and some have peak wavelength more than 10 times larger than the layer depth. When not taking account to the large fluctuation, at the lower frequencies, the peak wavelengths for the different flights were approximately \( 3z_i \).

![Cospectra of \( w_0 \) and \( u_w \) for Flight 3](image)

*Figure 6.21.* Normalised cospectra of \( u_w \) and \( w_0 \) for Flight 3. Dashed lines represent flight legs near the cloud base.
Flight 4

There is a tendency in the $\mathbf{w}\theta$ cospectra in the cloud layer that flight legs from upper part have positive fluxes and flight legs from lower part have negative fluxes. The peak wavelength in upper part are approximately $2\Delta z$. For the lower layers the peak wavelength varies between $(\lambda_m)_{w\theta}=3.0\Delta z$ and $(\lambda_m)_{w\theta}=6.0\Delta z$.

Peak wavelength in the sub-cloud layer for cospectra of $\mathbf{uw}$ indicate a minimum near cloud base. At the altitude $0.8z_i$ the peak wavelength were approximately equal to layer depth. The largest eddies were near the sea surface with eddy sizes 10 times layer depth.

The fluxes in the cospectra of $\mathbf{w}\theta$ from flight legs in the lower part of the sub-cloud were positive, while in the upper part, the fluxes were negative. The peak wavelength was approximately equal for all flight legs from the different altitudes, $(\lambda_m)_{w\theta}=3.0z_i$.

![Cospectra of $\mathbf{uw}$, flight 4, in-cloud](image1)

![Cospectra of $\mathbf{w}\theta$, flight 4, in-cloud](image2)

![Cospectra of $\mathbf{uw}$, flight 4, sub-cloud](image3)

![Cospectra of $\mathbf{w}\theta$, flight 4, sub-cloud](image4)

Figure 6.22. Normalised cospectra of $\mathbf{uw}$ and $\mathbf{w}\theta$ for flight 4. Dashed lines represent flight legs near the sea surface and near the cloud base for the cloud layer. Dotted lines represent flight legs near the cloud top.

6.2.4. Summary

All the cospectra, of both the momentum and the heat flux, were quite noisy which makes an interpretation difficult.
The cospectra of uw in both layers for all flights have very similar characteristics in general. There was little turbulent production from the momentum fluxes in the cloud layer, as expected when comparing with the momentum flux profiles. The flights show that the turbulent fluxes of momentum, cospectra of uw, in the sub-cloud layer decrease with height to a minimum near the cloud base. This is also expected, while the friction against surface decreases with height, see also the momentum flux profiles figure 6.6. The largest peak wavelengths were found in flight 2 for uw cospectra in the sub-cloud layer.

The cospectra of wθ in the sub-cloud layer follow the buoyancy flux profiles, figure 6.1. Turbulent fluxes in the sub-cloud cospectra of wθ shows an indication of positive values in the lower part and then negative values in the upper part. Also here in the sub-cloud layer, the largest peak wavelength in cospectra of wθ were found in flight 2. The turbulent fluxes in the cospectra of wθ in the cloud layer were in general positive for all flights. The largest peak wavelength were found in flight 3, where most peak wavelength were about 10 times larger than the layer depth.

### 6.3 Length scales

Length scales are often used for closure techniques in models. One of the length scales is the dissipation length scale, $l_\varepsilon$, which is derived from the parameterisation for the dissipation of TKE. The dissipation length scale is often used as a master length scale, which means that all other length scales in the closure problem is related to this master length scale. The dissipation length scale in this study is taken from Tjernström and Rogers (1996):

$$l_\varepsilon = C_\varepsilon \frac{e^3}{\varepsilon}$$

where $e^2$ is the TKE, $\varepsilon$ is the rate of dissipation of TKE and $C_\varepsilon$ is a dissipation closure constant, here 0.2. The dissipation length scales is expected to be related to the peak wavelengths from spectra. Large eddies contribute to more TKE and the spectral peak at higher energy. Through the cascade process and with homogeneous turbulent flow appears the dissipation at a higher level than if there where smaller turbulent eddies.

Normalization of the length scale is made in two ways: with the sub-cloud or cloud layer depth, respectively, or with the peak wavelength from w-spectra. See figure 6.23.

In the sub-cloud layer using the layer-depth scaling, the dissipation length scale increases with height over the lower half of the layer, as expected, asymptotically reaching a maximum in the upper half of the layer. Flight 2, the only night flight, is different since length scale here was larger than for all the other flights; this flight also has a large TKE. For the cloud layer, using the layer-depth scaling, the vertical shape is quite similar for all flights. The length scale is approximately constant with height or slightly decreases towards the cloud top. Flight 5 differs from the other flights, having lower values. From mixed layer theory, $l_\varepsilon$ is expected to be constant with height.
The normalization with peak wavelength in the sub-cloud layer indicates profiles with maximum near the cloud base. Probably a maximum near the sea surface for flights 1 and 5, which also scale with peak wavelength that decreases close to the sea surface. The normalised dissipation length scale seems to increase faster in the upper part of the sub-cloud layer than in the lower part. When comparing with peak wavelength from the first four flights they have a tendency of maximum inside the sub-cloud layer and minimum near the cloud base and the sea surface. In the upper part of the layer there is perhaps a relationship between the dissipation length scale and the spectral peak for the w-spectra. The peak wavelength decreases when the normalised dissipation length scale increases. In the cloud layer, the normalised dissipation length seems to be relatively constant with height, with a tendency for an increase towards the cloud top. The night flight, with no information in its upper part of the cloud layer, increases with maximum in the middle of the layer. However, the peak wavelength from the night flight decreases with altitude.

The two ways of scaling gave different vertical pattern for the cloud and the sub-cloud layer. This behaviour comes probably from that the peak wavelength and layer depth is related to different properties and their variation with height. The peak wavelengths are related to the energy production scale and vary more or less with height. While the layer depth is related to the boundary layer structure in the vertical and not varying with height, in the calculations, for each flight.

![Figure 6.23.](image)

Figure 6.23. The two left frames show dissipation length scale normalised with the peak wavelength from w-spectra. The two frames at right show dissipation length scale normalised with layer depth. Plot symbols; o flight 1, x flight 2, ø flight 3, + flight 4 and Δ flight 5.
7. Conclusions

The turbulent structure of marine stratocumulus clouds has been analysed. The data are from aircraft measurements taken during the First Lagrangian of the Atlantic Stratocumulus Transition Experiment during 12 to 14 of June 1992. The air mass was followed southeast of the Azores Island for about a distance of 1300 km. The marine stratocumulus clouds are important for global climate and in climate modelling and the development of these clouds are therefore of great interest. On the first day of the First Lagrangian there was a solid stratocumulus cloud layer with some drizzle. The stratocumulus clouds moved southward and at late afternoon the second day cumulus clouds started to penetrate into the stratocumulus cloud. The air mass moved over warmer sea surface and the last day more cumulus clouds developed and the stratocumulus clouds were thin and broken.

Longwave radiative cooling from cloud top mainly drove the turbulent structures in the cloud layer. This means that the turbulence derive its energy from thermals started at the cloud top, which gives an upside down profile for the buoyancy fluxes comparing to classical PBL. The buoyancy flux profile indicate a maximum in the cloud layer near the cloud top and a minimum near the cloud base for all five flights except for the last flight. The last flight has probably a more complex turbulent structure because of the cumulus clouds that penetrate into the stratocumulus clouds. In the sub-cloud layer the buoyancy flux profile shows a decrease with a minimum near the cloud base as similar for a convective mixed layer. The night flight was different with a minimum in the buoyancy flux further down under the cloud base. Since the sub-cloud layer and cloud layer derive its turbulent energy from different sources, they are decoupled. The decoupling would probably give zero or a minimum in the buoyancy flux profile close to the cloud base which is the case here. The decoupling is also confirmed by the vertical velocity variance, the TKE and the momentum flux profiles. A minimum near the cloud base in the vertical velocity and the TKE profiles indicates damped motions and the two-layered structure caused by decoupling. This was the case for flight 1, 3 and 4. The night flight, flight 2, seems to have a mixture of both in-cloud and sub-cloud properties. For the night flight the TKE and $u'^2$ and $v'^2$ decreases from the sea surface up to the cloud top indicating shear produced turbulence, while the buoyancy flux profile indicate thermals generated by cloud-top cooling reaching below the cloud base. For the last flight, with its complex cloud structure, decoupling was assumed due to decreasing momentum flux, sensible flux and buoyancy flux profiles from surface up to the cloud base. The momentum flux profile in the sub-cloud layer decreases linearly as expected for all flights, while it in the cloud layer is small and constant with height.

In the scaled variances for the sub-cloud layer the dashed lines represents parameterisation from Brost et al. (1982). These lines fit this data well with the exception for the v-component. The v-component is much larger than the dashed line indicating that the wind-direction shear in the sub-cloud layer is important. The variance is smallest for the w-component indicating that wind shear is the main generator of the turbulence in the sub-cloud layer. The three components of the variance for the cloud-layer were approximately of the same size. This indicates that turbulence is generated by buoyancy in the cloud layer.
The spectral analysis is used to study eddy sizes and energy for the different flights in the cloud layer and the sub-cloud layer. The spectral peak represents eddies with most energy and the corresponding wavelength should be related to the layer depth. The spectra from flight 5 are not shown because the normalisation appeared troublesome, probably because of the complicated cloud structure.

The spectra for the cloud layer are normalised with mixed layer theory from Kaimal et al. (1976). The use of mixed-layer scaling shows that the spectra collapse on top of each other. The spectral peak show that the eddies were mostly larger than the cloud layer depth for all four flights. The w-spectra near the cloud top and the cloud base indicate smallest peak wavelength and a maximum of peak wavelength inside the cloud for flight 1 and 4. The vertical motions seem to be suppressed near the cloud top due to the capping inversion and close to cloud base. Flight 2 has largest peak wavelength for w-spectra in the lowest part of cloud layer and then decreasing with higher altitude. This structure may be due to the interaction between the shear and buoyancy produced turbulence. Approximately constant peak wavelength with height for flight 3 with peak wavelength larger than layer depth, \(\lambda_m = 3\Delta z\). Otherwise the spectral peaks in w- and u-spectra are generally within 1-3\Delta z. The v-spectral peaks seems to be larger than those in both w- and u-spectra and can have something to do with some mesoscale circulation.

The sub-cloud spectra are normalised with both mixed-layer scaling and surface-layer scaling after Kaimal et al. (1972). The scaling with mixed layer theory does not work out well because of the poor collapse onto one curve in the inertial sub-range. However, the peak wavelengths are larger than sub-cloud layer depth and the u- and v-spectra shows larger \(\lambda_m\) than the w-spectra. The surface-layer scaling, where the height variation is taken into account, collapses the spectra better in the inertial subrange. Also for the sub-cloud layer the peak wavelengths are larger than the layer depth for almost all flight legs and the peak wavelength in u- and v- spectra larger than w-spectra. Maximum peak wavelengths in the w-spectra is found between altitudes 0.25\Delta z and 0.6\Delta z for all four flights. The night flight differs from the other flight with largest peak wavelength in u-, v- and w-spectra. The spectral peaks in w-spectra for flight 1 and 4 are generally between 1\Delta z and 5\Delta z, while flight 2 and 3 shows generally larger peak wavelength.

The sub-cloud spectra, and for some extent in the cloud layer, show double peaks in the u- and v-spectra. This feature may need some further studies. However, it can be due to the developing cumulus clouds in the later parts of the Lagrangian. But then it would appear also in the cloud spectra because the cumulus clouds penetrate into the stratuscumulus clouds. Another explanation is that thermals formed in the cloud layer transports down into the sub-cloud layer without seeing anything in the turbulent fluxes, which shows up zero or close to zero in the level between the cloud and sub-cloud layer (Khalsa 1992). This means that more studies are needed about the decoupling. Another problem shows up in the two horizontal u- and v-spectra with the cut-off frequencies. For some flight legs, the cut-off frequency seems to be too tight for the low frequencies in the spectra.

The cospectra of uw in the cloud layer shows little turbulent production from momentum fluxes, for all four flights. This behaviour is also seen in the vertical profiles of momentum fluxes where the fluxes are small and constant. The cospectra of uw in
the sub-cloud layer decreases with a minimum close to the cloud base. This behaviour is also expected since the friction against the surface decay with height. In the sub-cloud layer the cospectra of uw is quite noisy when looking at each flight. The cospectra of uw are also negative for all flights and peak wavelength larger than layer depth. The averaged cospectra of uw indicate a second peak, approximately 10Δz, at lower frequency with also negative flux and this feature comes probably from a meso-scale circulation. The turbulent fluxes in cospectra of wθ follows the characteristics from the buoyancy flux profile. The cospectra of wθ in the cloud layer shows in general positive fluxes, if there are negative fluxes it comes from flight legs in the lower part of the cloud layer, and peak wavelength larger than layer depth. The peak wavelength variation in the cloud layer is different between the four flights and it is hard to find common structures. However, flight 1, 2 and 3 have approximately constant peak wavelength through each cloud layer and flight 4 has the largest peak wavelength in general, comparing to the other flights. For flight 4, the largest peak wavelength is found in the lower part of the cloud layer. When the cospectra of wθ is averaged with all four flights there is a tendency of linear increase for the turbulent production with a maximum near the cloud top. Also the averaged cospectra of wθ indicate a second peak at a lower frequency, approximately 10Δz. The sub-cloud layer for cospectra of wθ indicate for flight 1, 2 and 4 positive fluxes near the sea surface, in the upper part negative contributions. Flight 3 has positive fluxes but there are also flight legs only from the lower half of the layer. The peak wavelength in the cospectra of wθ in the sub-cloud layer is larger than layer depth for all flight legs and in general, the largest peak wavelength was found in flight 2. The averaged cospectra of wθ have peak wavelengths approximately between 2z_1 and 5z_1. Also here there is a second peak at a lower frequency, approximately 10z_1. This peak shows up only in the heat flux from the upper part of the sub-cloud layer.

The dissipation length scale is derived from the parameterisation of the dissipation of TKE and is often used as a master length scale in models. This length scale is normalised with two methods, with the sub-cloud or cloud layer depth, respectively, or with the peak wavelength from w-spectra. In the cloud layer with layer depth scaling, the vertical profile is similar for all flights with constant or with a slight decrease towards cloud top. Mixed layer theory indicates dissipation length scale to be constant with height. The vertical profile in the sub-cloud layer shows an increasing character with a maximum at the cloud base. The other normalisation with peak wavelength, in the cloud layer, shows a tendency of an increase towards the cloud top. The increase close to the cloud top is probably due to that some of the peak wavelength decreases near the cloud top. In the sub-cloud layer the normalised dissipation length scale have a maximum near the cloud base.

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References


