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Characteristics of Drizzle under Stratocumulus using Cloud Doppler Radars

Virendra Ghate

University of Miami, vghate@rsmas.miami.edu

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UNIVERSITY OF MIAMI

CHARACTERISTICS OF DRIZZLE UNDER STRATOCUMULUS USING CLOUD DOPPLER RADARS

By
Virendra Prakash Ghate

A THESIS

Submitted to the Faculty of the University of Miami in partial fulfillment of the requirements for the degree of Master of Science

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the requirements for the degree of
Master of Science

CHARACTERISTICS OF DRIZZLE UNDER STRATOCUMULUS USING CLOUD DOPPLER RADARS

Virendra Prakash Ghate

Approved:

Dr. Bruce A. Albrecht
Committee Chair
Professor of Meteorology

Dr. Steven G. Ullmann
Dean of the Graduate School

Dr. Paquita Zuidema
Assistant Professor of Meteorology

Dr. Pavlos Kollias
Associate Scientist
Brookhaven National Laboratory
Marine stratocumulus clouds cover extensive areas of the subtropical oceans and greatly influence Earth's radiation by strongly reflecting the incoming solar radiation. The most climatologically pronounced stratus regime is located in the South-East Pacific. Drizzle is one of the several physical processes that affects the lifecycle and evolution of marine stratus by depleting the cloud liquid water and by stabilizing the marine boundary layer through evaporative cooling.

In this study we use ship-borne radar observations from two innovative research cloud radars — a Millimeter Cloud Radar (MMCR) (λ=8 mm) and Frequency Modulated Continuous Wave (FMCW) radar (λ=3 mm) to study the fallout of drizzle in the sub-cloud layer. Radar inter-comparison is used to perform calibration and quality control of the FMCW radar. The FMCW observations suffer no saturation and provide profiles of radar Doppler moments from the ship level to the cloud base. A lognormal drizzle drop size distribution is assumed and the parameters (N₀, r₀ and σₓ) are retrieved using the observed radar reflectivity and mean Doppler velocity profiles. The retrieved parameters are used to extract bulk parameters of the drizzle size distribution such as liquid water content, total number of droplets and rainfall rates at various heights within the sub-cloud layer.
(typically from 50-500 m). It is demonstrated that a simple evaporation model can be used to constrain the inversion from radar observables to drizzle size distribution parameters. The model output showed that the drizzle DSD is truncated at lower end due to the rapid evaporation of smaller drops and the logarithmic width of drizzle DSD is lower than the typically prescribed value of 0.35.
ACKNOWLEDGEMENT

Many people have been instrumental in the completion of this study. First of all I will like to thank my advisor Dr. Bruce Albrecht, for giving me the opportunity to work with him and guiding me throughout the course of this study. The guidance by Dr. Pavlos Kollias in radar engineering, and ship based data collection techniques was impeccable. The suggestions given by Dr. Paquita Zuidema were very helpful. Without the help of other UMRMG members Ieng Jo, Efthymios Serpetzoglou, Shaunna Vargas, Thomas Snowdon and Daniel Voss, this study would not have been completed. The inspiration and constant support by my parents and my brother, during the period of the study was the biggest motivation for me to complete it. Last but not the least; I will like to thank all of my friends in USA and in India who made the stay and the study much more fun than I expected.
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Chapter 1 – Introduction

1.1 Motivation

Marine stratocumulus clouds are shallow boundary layer clouds mainly observed at low levels over the eastern side of the subtropical oceans. These regions are characterized by cold surface waters and warm dry air subsiding aloft that favor the creation of a sharp temperature and moisture inversion. This inversion caps the Marine Atmospheric Boundary Layer (MABL) and traps the clouds at its top (Klein and Hartman 1993). These clouds are not only important due to their vast extent, but also due to their strong influence on the radiation budget. They have high albedo that significantly decreases the amount of solar radiation reaching the ocean’s surface, while their low altitude results in a small temperature difference between the cloud-top and the ocean surface resulting in little change in the long wave radiation emitted back to space. Hence, these clouds have cooling effect on the sea surface. Randall (1984) has shown that a 4% increase in the area covered by stratocumulus would result in atmospheric cooling that could offset global warming due to a doubling of the CO$_2$ content of the atmosphere. Although the role of stratocumulus clouds in affecting the radiation budget was recognized earlier, the growing need of their more accurate representation in the Global Climate Models (GCMs) has engaged many scientists in the pursuit of a better understanding of their radiative, microphysical and dynamical properties, the thermodynamic structure of the MABL, and the climatological variability of the respective areas (e.g. Stevens et al 2003). A full
understanding of the properties of these clouds will be vital for estimating global energy budgets and to develop realistic numerical weather and climate model.

The southeastern Pacific stratocumulus region reaches close to the equator, and extends 1500 km offshore all the way south to central Chile almost year-round (Klein and Hartmann, 1993). In addition to its large latitudinal extent, its interaction with El Nino Southern Oscillation (ENSO) and the special morphology of the South American continent also contributes to the unique character and high importance of the Southeastern Pacific stratocumulus regime (Li and Philander, 1996).

Observations and modeling studies in the past have shown that drizzle under these clouds is important principally because it is involved in determining the cloud lifetime and evolution. The impact of drizzle on MABL has been subject of several observation and modeling studies (e.g. Brost et al 1982; Nicholls 1987; Albrecht 1989, 1993; Wang and Wang 1994; Feingold et al 1996; Stevens et al 1998 etc.). These studies suggest that drizzle may have an important dynamical role in regulating the MABL structure. The above studies have shown that the chief dynamical effect is due to its differential vertical heating. It causes a net latent heating in the cloud layer by depleting the liquid water and causes an evaporative cooling in the sub-cloud layer. The drizzle induced stabilization of the MABL may induce its decoupling. On the other hand, heavy drizzling can deplete the cloud Liquid Water Content (LWC) and weaken the cloud structure leading to broken stratus clouds (or Pockets of Open Cells (POCs) as called by Stevens et al., 2003) or even complete cloud dissipation. On the other hand, under light
drizzle conditions cooling is confined to region just below the cloud base that destabilizes the sub-cloud layer resulting in a more vigorous circulation and a better mixed boundary layer (Feingold et al., 1996; Stevens et al., 1998). This in turn can increase the turbulence intensity, which can lead to drizzle drop formation by collision-coalescence, thus providing a feedback to more drizzle formation that might lead to heavy drizzle and cloud dissipation by promoting a drying of the cloud layer.

In summary, numerous studies have examined the production of drizzle and its effect on stratocumulus clouds and planetary boundary layer. It is clear that the presence of drizzle in stratus MABL significantly impacts the evolution of these clouds and the MABL structure itself. However, drizzle is a complicated phenomenon that includes many scales. Many of the hypotheses are based on models that have ad-hoc drizzle parameterization. Hence, good drizzle parameterizations are needed in order to represent these types of clouds in GCMs.

1.2 Scientific Objectives

In this study observations made during Tropical Atmosphere Ocean (TAO) and Stratus04 scientific cruise conducted in 2004 are used to characterize the drizzle under stratocumulus. Observations used in this study were made from two Cloud Doppler radars--the National Oceanic and Atmospheric Administration (NOAA) Earth System Research Laboratory (ESRL)’s K-band Millimeter Cloud Radar (MMCR) and Center for Interdisciplinary Remotely-Piloted Aircraft Studies
(CIRPAS)'s W-band Frequency Modulated Continuous Wave (FMCW) radar, form the base of this study. A detailed description of these instruments is given in the following chapters. The main objectives of this study can be summarized as follows:

1) To analyze the ESRL-MMCR observations from its two modes of operation to determine mode suitable for drizzle studies.

2) To calibrate the FMCW radar and test its use for drizzle detection.

3) To develop R-Z relationship based on the Drizzle Drop Size Distribution (DSD) model described by Frisch et al. 1995.

4) To develop an evaporation model for sub-saturated conditions below the Lifting Condensation Level (LCL) for constraining the Frisch model retrieved DSD to better fit the evaporative conditions.
Chapter 2 - Data Collection & Instrumentation

2.1 Cruise description

In November 2004, the NOAA research vessel Ronald H. Brown (RHB) conducted a regular buoy maintenance cruise in central and southeast Pacific. The Physical Science Division (PSD) of the NOAA-ESRL together with the University of Miami Radar Meteorology Group (UMRMG) conducted joint measurements of MABL, stratocumulus clouds, boundary layer thermodynamic structure, surface fluxes and meteorology during this cruise. The first leg of the cruise was named the Tropical Atmosphere Ocean (TAO) and the second one as Stratus04.

For the TAO cruise, the RHB left Panama City, Panama on 24 October 2004 and went south-westward to reach the southernmost buoy on the 95° W longitude. Then the ship overhauled all the buoys on that longitude up to the northernmost buoy located at 12° N. The same was done while going south along the 110° W longitude. After overhauling the last buoy at 8° S and 110° W the ship headed South-Eastward towards the port of Arica, Chile and ended the TAO cruise on 27 November 2004.
Figure 2.1: Ship Track of Research Vessel Ronald H. Brown for Tropical Atmosphere Ocean TAO (solid) and Stratus04 (dashed) cruises conducted in the northern Hemisphere fall of 2004.

For Stratus04 cruise, the ship departed Arica, Chile on 5 December 2004. Then it followed the same easterly path towards 20° S and 85° W like that of the earlier cruises like EPIC (East Pacific Investigation of Climate) 2001 and PACS (Pan American Climate Studies) 2003. During this cruise the WHOI (Woods Hole Oceanographic Institution) Ocean Reference Station (WORS) located at 20° S and 85° W was overhauled. For the purpose of comparison of the ship based instruments with the buoy instruments, the ship was stationed at this location for 5 days from 11 December 2004 (Julian Day 346) to 16 December 2004 (Julian
Day 351). The cruise ended on 23 December 2004 after a south eastward route at the southern port of Valparaiso, Chile. Fig. 2.1 shows the ship track of both TAO and Stratus04 cruises.

2.2 Instrumentation

There were several instruments deployed onboard Research Vessel (R/V) RHB during both the cruises. Some of these include the MMCR, FMCW radar, scanning C-band radar, wind profiler, laser ceilometer, Micro-Wave Radiometer (MWR), Flux suite and upper-air soundings. Fig. 3.1 shows the instrument setup for the cruise. A brief description of the instruments used in this study is given in following sections.
2.2.1 MMCR

The 34.86 GHz (8.6 mm) Doppler pulse cloud radar known as MMCR from ESRL was operated by UMRMG during both the cruises. It is a ultra-high sensitivity Doppler radar with a peak transmitting power of 100 W. The antenna is 1.8 m in diameter and has a tilted flat radome. The beam is circular with a width of 0.3 deg. During both the cruises, the radar was set to run in two modes: Boundary Layer (BL) mode and Precipitation (PR) mode. To do so, it was transmitting pulse widths of 300 µs and 600 µs alternatively to give the two
modes of operation. For the purpose of data archiving and self-calibration the radar was not collecting data for approximately two minutes at the end of every half hour and every top of the hour. The BL mode was specifically designed for studying Boundary Layer Clouds, while the PR mode was set for obtaining true reflectivity during precipitation events. The radar has a single antenna Transmitter/Receiver (T/R) system. Since there is a switching between the T/R which is practically not instantaneous, the radar is not able to collect any data for the first 114 m above the antenna. This first 114 m zone is referred as “Dead Zone” in the rest of the literature. The radar is very well calibrated to the accuracy of ±1 db. The operating parameters of the radar are given in the Table 2.1.

TABLE 2.1 Operating parameters of MMCR during TAO and Stratus04 cruise conducted in 2004. # More on this in chapter 3

<table>
<thead>
<tr>
<th>Parameter</th>
<th>MMCR-BL</th>
<th>MMCR-PR</th>
</tr>
</thead>
<tbody>
<tr>
<td>Frequency</td>
<td>34.86 GHz</td>
<td>34.86 GHz</td>
</tr>
<tr>
<td>Pulse Width</td>
<td>300 µs</td>
<td>600 µs</td>
</tr>
<tr>
<td>Inter Pulse Period</td>
<td>68 µs</td>
<td>106 µs</td>
</tr>
<tr>
<td>Maximum Range</td>
<td>5019 m</td>
<td>15054 m</td>
</tr>
<tr>
<td>Vertical Resolution</td>
<td>45 m</td>
<td>90 m</td>
</tr>
<tr>
<td>Number Spectra Averaged</td>
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<td>9</td>
</tr>
<tr>
<td>Number of FFT points</td>
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<td>128</td>
</tr>
<tr>
<td>Time Resolution</td>
<td>8 s</td>
<td>8 s</td>
</tr>
<tr>
<td>Sensitivity at 5 km</td>
<td>-47 dBZ</td>
<td>-42 dBZ</td>
</tr>
<tr>
<td>Suffers Saturation#</td>
<td>Yes</td>
<td>No</td>
</tr>
</tbody>
</table>
2.2.2 FMCW Radar

The other radar onboard the RHB during both cruises was the CIRPAS’s 94.8 GHz (3.16 mm) FMCW compact airborne cloud Doppler radar, which is developed by ProSensing Inc. For the first time this radar was used for ship based cloud studies, unlike the flight based studies for which the radar was designed. Unlike the MMCR, which is a pulsed radar, this radar is a continuous wave radar. It is light weight radar (about 37 kg) and a transmit power of only 300 mW. The radar has separate transmitter and receiver antennae, both of them 30 cm in diameter and of dual cassegrain design. The power consumption of the system is low 2.5 A at 28 VDC. The operating characteristics of the radar are given in Table 2.2

Table 2.2 Operating Characteristics of FMCW radar during TAO and Stratus04 cruises. * denotes values for TAO cruise while ** values for Stratus04 cruise. All the values are same unless otherwise stated. # More on this in chapter 3.

<table>
<thead>
<tr>
<th>Parameter</th>
<th>Value</th>
</tr>
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<tbody>
<tr>
<td>Center Frequency</td>
<td>94.8 GHz</td>
</tr>
<tr>
<td>Sweep Frequency</td>
<td>10 KHz</td>
</tr>
<tr>
<td>Number of FFT points</td>
<td>128</td>
</tr>
<tr>
<td>Number of Averages</td>
<td>200</td>
</tr>
<tr>
<td>Vertical Resolution</td>
<td>5m* and 10 m**</td>
</tr>
<tr>
<td>Range</td>
<td>1250 m* and 2500 m**</td>
</tr>
<tr>
<td>Time Resolution</td>
<td>1.3 sec</td>
</tr>
<tr>
<td>Sensitivity at 1 km</td>
<td>-10 dBZ</td>
</tr>
<tr>
<td>Suffers Saturation*</td>
<td>No</td>
</tr>
</tbody>
</table>
One of the several features of this radar that separates it from the MMCR is the absence of a dead zone. This is due to the separate transmitter and receiver system. The absence of a dead zone makes the radar ideal for the study of precipitation and evaporation at lower levels i.e. in the bottom 500 m of the MABL.

The radar went through a change of the Analog to Digital (A/D) board and a change of the receiver system just before the TAO cruise. Hence, the radar was not calibrated and just the radar receiver power was recorded during both the cruises. Both the radars were connected to the same GPS system to maintain the same time stamps for future comparisons.

### 2.2.3 Other Instruments

Other remote sensing instruments on the ship included a scanning C-band radar, a 915 MHz wind profiler, a laser ceilometer and a microwave radiometer. The laser ceilometer present onboard was a Vaisala model CT-25K type ceilometer. It was running with a temporal resolution of 15 sec and spatial resolution of 30 m during both cruises. The ceilometer was recording the first three cloud base heights and the backscatter coefficient. The three-channel Microwave Radiometer operated at 20.6, 31.6 and 90 GHz frequency. From the MWR the column integrated liquid and vapor amounts were calculated. Present also on the ship were various instruments incorporated in the flux suite. From these instruments the sensible heat, latent heat and momentum fluxes along with
broadband radiative fluxes were calculated. The flux suite also included instruments for measuring wind speed, Sea Surface Temperature (SST), Ship heading, ship location and other standard meteorological parameters. Radiosondes of Vaisala, the RS-92 type were used during both the cruises. Soundings were launched after every 6 hours from 1 Nov to 20 Nov for TAO and after every 4 hours from 6 Dec to 22 Dec for Stratus04 cruise.

2.3 Data Archiving and Processing techniques

Since many instruments were continuously operating during the cruises, data archiving and processing was a challenging task. The MMCR was collecting the Doppler Spectra from which the first three Doppler moments of the Spectra i.e. Reflectivity, Doppler Velocity and Doppler Spectrum Width were calculated. All the data was stored in netcdf format. The FMCW radar was recording the Doppler Spectra in a binary data (*.dat) format. From these data files the Doppler moments were calculated and were saved in netcdf format for archiving. Since the radar was not calibrated at the time of the cruise, instead of reflectivity (dBZ), the radar power was saved in dBm. Data from all the other instruments viz. Ceilometer, Radiometer, Sounding and Flux were converted from binary data files to ASCII files. All the data were plotted and analyzed to synthesize observations from the instruments and the radars.
2.4 Technical Difficulties

The FMCW radar was used for the first time for a ship based cloud study experiment. The radar is designed for aircraft cloud studies, which usually last for a few hours. Due to continuous operation on the cruise, there was radar overheating that could have been fatal for the radar components. In order to prevent this overheating, the radar was turned off whenever the radar temperature went above 45°C. Hence, there are periods with no FMCW data that extended for a few hours throughout both the cruises.

The MMCR had a component failure on the 5 December 2004, which marks the start of the Stratus04 cruise. Hence, the more sensitive MMCR was not available during this cruise.

The laser ceilometer was operating under reduced sensitivity after day 6 of the Stratus04 cruise. A deterioration of the optical fiber that carries the signal to the detector prevented the instrument from detecting many clouds during daytime when sunlight may have contaminated the optical returns (Serpetzoglou, 2005). Fortunately, the cloud base height is accurate when clouds are detected. After careful analysis it was decided that the data between local time 8:00 to 16:00 was unusable.

2.5 Cruise conditions

As the objective of the study is to characterize drizzle under Stratus clouds, the focus is primarily on the stratus04 cruise. This section describes the general meteorological conditions for the Stratus04 cruise.
Since the MMCR was not working during this cruise, the cloud top was calculated from the 915 MHz wind profiler data. Fig. 2.3 shows the ceilometer detected cloud base (blue) and the wind-profiler derived cloud-top (black) for the entire cruise. It can be seen that there was a steady increase in cloud-top height and a corresponding increase in cloud-base height as the cruise progressed southwards. While the cloud-thickness remained constant throughout the cruise. The average cloud-base height throughout the cruise was 1.1 km with standard deviation of 185 m, while the average cloud-top height was 1.5 km with a standard deviation of 170 m. The corresponding cloud-thickness was 323 m with a standard deviation of 134 m. The ship was stationed at the WORS from 11 December 2004 (Julian Day 346) till the 16 December 2004 (Julian Day 351).
Figure 2.3 Ceilometer detected cloud base (blue) and wind-profiler derived cloud-top (black) for Stratus04 cruise. The average cloud base for the entire experiment was 1.104 km with a standard deviation of 185 m while the average cloud-top was 1.474 km with a standard deviation of 170 m.

Shown in Fig. 2.4 is the time-height section of the relative humidity from the soundings for the entire cruise. The Lifting Condensation Level (LCL) height calculated from the surface meteorological instruments using Bolton (1980) is shown in black dots while the wind-profiler derived cloud-top is also shown. It can be seen that, the wind-profiler derived cloud-top matches well with the top of the boundary layer indicated by the low-level moisture decrease observed
throughout the cruise. Hence the strong capping inversion acts like a rigid lid for the stratus clouds to arrest their vertical extent, thus the inversion base height derived from wind-profiler can be used as a proxy for cloud-top height.

An example of data from the FMCW radar is shown in Fig. 2.5. It shows a time-height mapping of the reflectivity captured during 15 December 2004. The event selected is between 0300 UTC till 1700 UTC, which corresponds to 2300
of 14 December 2004 till 1300 of 15 December 2004 local time. The cloud-top and ceilometer cloud base is also shown in black dots and black asterisk respectively. Several features are noticeable from the figure. Due to the lower sensitivity of the FMCW radar, it is not able to capture the thin stratus cloud, and is only able to detect drizzle that has a much higher reflectivity than the cloud. The drizzle does not necessarily start from the cloud top. The duration of drizzle events can vary substantially and a majority of the drizzle evaporates before reaching the surface.

Figure 2.5 Time-height section of reflectivity (dBZ) captured by FMCW radar on 15 December 2004. The ceilometer detected cloud-base is shown in black dots while the black * show the win-profiler derived cloud-top.
Previous studies have used different thresholds to distinguish between drizzle and cloud using the radar reflectivity. In this study, the threshold used by Frisch 1995 of -17 dBZ is used. Comstock et al. (2004) used a reflectivity threshold of -20 dBZ. VanZanten et al. (2005) who developed an R-Z relation using radar and in-situ Drop Size Distribution (DSD) measurements used 0.3 mm/day as a threshold for drizzle, which corresponds to reflectivity of -28.8 dBZ at cloud base using the R-Z relationship they developed. They also differentiate between light and heavy drizzle using a threshold of 1 mm/day, which corresponds to reflectivity of -6.4 dBZ at the cloud base. In this study we do not intend to do any fine-scale drizzle classification since the lower sensitivity of the radar, only allows heavy drizzle events to be captured by the radar.
Chapter 3 - Radar Analysis

3.1 MMCR modes

Here the two operational modes of the MMCR are analyzed to investigate their suitability for drizzle studies. Furthermore, using radar inter-comparison (MMCR versus 94-GHz FMCW) we calibrate the FMCW radar. In previous research cruises conducted in the South-East Pacific (e.g., EPIC, PACS03), the MMCR onboard the RHB was operating again with these two modes, however during both modes the MMCR was saturated in the presence of drizzle near the radar. This problem (saturation) was diagnosed during the PACS 2003 cruise and the radar hardware was modified during the precipitation (PR) mode to allow the radar receiver to stay below saturation in the presence of drizzle in the sub-cloud layer.

The performance of the new PR mode design was tested during the 2004 cruise and it is demonstrated in the following analysis. A 10.5 min drizzle event on the 24 Nov 2004 is chosen. The ship location was 10.67° S and 100.04° W during the observation period. The data chosen are from 0400 UTC, which is near midnight local time. The reflectivity time-height field captured by the two modes of MMCR during this time is shown in Fig 3.1
Figure 3.1 Time-height section of reflectivity captured by MMCR-BL (Top) and MMCR-PR (Bottom) on 24 November 2004.

The dead zone for both radar modes is 114 m although the first useful radar data are collected around 200 m because the lowest radar range gate occurs within the period that the radar is alternating from transmit to receive mode. The time-height fields of radar reflectivity from the two MMCR modes illustrate the finer vertical resolution of the BL mode and indicate the presence of higher radar echo intensities at the PR mode. At each range gate the reflectivity
field from each mode is averaged. The averaging was done in the units of the reflectivity factor $Z$ (mm$^6$ m$^{-3}$).

$$Z(\text{mm}^6\text{m}^{-3}) = 10^{(\text{dbz}/10)}$$  \hspace{1cm} (3.1)

Fig. 3.2 shows the mean reflectivity for both the modes in the upper panel and the maximum reflectivity in the lower panel.

Figure 3.2 Mean Reflectivity recorded by MMCR-PR (Red) and MMCR-BL (Blue) in the top panel. Maximum reflectivity recorded by both the modes for the same time is shown in bottom panel.
The results in Fig. 3.2 show that the mean reflectivity of PR mode is higher than that of the BL mode at all heights. This shows that the BL mode suffers from saturation in lower gates. The bottom panel shows the maximum observed reflectivity from each mode as a function of height. It can be seen that the maximum reflectivity of PR mode at all the heights is also higher than that of the BL mode. The difference in the reflectivity from the two modes of the MMCR is due to higher MMCR receiver dynamic range during the PR mode. During the signal receiving cycle, all the T/R switches remain closed in the BL mode to protect the receiver from the incoming scattered signal, while one of the switches remains open for the PR mode that results in a higher dynamic range for the PR mode than the BL mode. This made the precipitation mode 22-23 dB less sensitive but on the other hand extended the dynamic range of the MMCR by the same power amount.
In order to investigate the effect of MMCR saturation in previous field experiments conducted in the same area, the percent of PR mode returns that were detected above the BL mode saturation (BL\text{sat}) are calculated. The maximum reflectivity recorded by the BL mode for this event at a gate is treated as the saturation reflectivity for that gate. Also the percent of PR detection above the BL\text{sat} - 3 dBZ were calculated. The result is shown in Fig. 3.3. Less than 10 % of PR mode observations are above the BL\text{sat}, while more than 30% are above

Figure 3.3 Number of MMCR-PR samples detected above MMCR-BL saturation (Red) and above MMCR-BL saturation - 3 (Blue).
the BL_{\text{sat}} -3 dBZ threshold. This is a significant portion of the data set and could lead to bias results and analysis on drizzle intensity and evaporation in the sub-cloud layer.

The above analysis shows that the BL mode suffers saturation and fails to record the true reflectivity during precipitation events. Since the main objective of the study is to study drizzle characteristics at lower levels, it was decided not to use this mode for further study. Although not useful for this study, the BL mode is more sensitive than the PR mode and its noise floor is 15-20 db lower than the noise floor of the PR mode. This along with higher spatial resolution makes this mode useful for the study of clouds rather than precipitation. However, our focus here is to study the drizzle structure in the sub-cloud layer and thus we decided not to use the BL mode data.

3.2 FMCW calibration

The MMCR is a well calibrated radar within an accuracy of ±1 db and as shown in the previous section, the MMCR-PR profiles record the true reflectivity during drizzle events. Thus we used the MMCR to calibrate the FMCW radar. The precipitation event described above obtained on 24 November 2004 is used to calibrate the FMCW radar.

The FMCW radar was not calibrated during the cruises and hence, only the radar power was recorded in dBm. The radar reflectivity at each gate can be calculated from the radar power by using following equation.

\[
\text{Reflectivity} = 10 \times \log_{10}(\text{RadarPower}) + 20 \times \log_{10}(\text{Height}) + R_c \quad (3.1)
\]
where the Reflectivity is in dBZ, and the radar power is in db counts, height is in km and $R_c$ is the radar calibration constant that depends on the characteristics of the radar antenna, radar front-end and the particular setting of the radar Analog-to-Digital (A/D) receiver.

The event used for calibration is shown in Fig. 3.4. The top panel shows the reflectivity captured by the MMCR-PR while the bottom panel shows the range corrected radar power of FMCW radar for the same time. Hence shown in the bottom panel is the sum of first two terms on the Right Hand Side of Eq. (3.2) for the FMCW radar. To calculate $R_c$, the reflectivity of both the radars was averaged for the entire period and compared at 300 m.
Figure 3.4: Time-height section of reflectivity captured by MMCR-PR on 24 November 2004 is shown in the top panel, while bottom panel shows the time-height section of range corrected radar power captured by FMCW radar for the same period.

The height of 300 m was chosen to minimize attenuation in the FMCW radar returns, since, most attenuation at 94-GHz is expected above the cloud base and the average cloud base height for the above period was 400 m. The radar inter-comparison resulted in a radar calibration constant $R_c = -89$ dBZ for the FMCW radar. Hence the equation for the reflectivity from the FMCW radar becomes
\[ \text{Reflectivity} = 10 \times \log_{10}(\text{RadarPower}) + 20 \times \log_{10}(\text{Height}) - 89 \quad (3.2) \]

Using Eq. (3.2), the reflectivity for the FMCW radar was calculated for both the cruises. Fig. 3.5 shows the scatter plot between the reflectivity of MMCR-PR and FMCW radar at 300 m. The figure shows excellent correlation between the two radars indicating that the FMCW radar is calibrated and supporting the use of the radar constant of -89 for the entire period.

Figure 3.5 Scatter plot between the reflectivity captured by FMCW radar and the MMCR-PR at 300 m for the period shown in Figure 3.4.
The calibration of the FMCW radar revealed that the radar has a sensitivity of -13 dBZ at 300 m for a vertical resolution of 5 m. For coarser vertical resolutions the sensitivity will be higher. To further analyze the FMCW radar performance, the mean and maximum reflectivities for the event from FMCW radar and MMCR-PR were calculated and are shown in Figure 3.6. It can be seen that the FMCW radar suffers attenuation above the cloud base as its reflectivity decreases with height above 400 m. The sudden knot and gradual decrease thereafter in the FMCW reflectivity profile is due to the fact that the radar signal gets severely attenuated by sudden increase in the Liquid Water Content (LWC) from the low drizzle LWC (0.001 g/m³ to 0.1 g/m³) to high cloud LWC (0.1 g/m³ to 0.4 g/m³) above the cloud base. But the FMCW radar does not suffer from any form of saturation as it matches with the MMCR-PR reflectivity reasonably well below the 400 m level. Further, the maximum reflectivity recorded by the radar at higher levels is also comparable with the MMCR-PR.
Figure 3.6: Top panel shows the average reflectivity observed by MMCR-PR (red) and FMCW radar (black) for the period in discussion on 24 Nov 2004. The bottom panel shows the maximum reflectivity observed by both the radars during the same period.
Chapter 4 – Drizzle Characteristics

4.1 Background on Rain-rate Reflectivity (R-Z) relationship

Accurate rainfall measurement is of great importance for a series of important applications in atmospheric research ranging from the lifecycle of stratocumulus clouds to the latent heat release in the atmosphere. Rainfall rate varies from a few millimeters per day in the case of drizzle under Stratiform clouds to tens of centimeters per hour for deep convective clouds. Rain gages are used to measure surface rainfall rate. The tipping bucket rain-gage has typical accuracy of 0.24 mm/hr while an optical rain gage can measure rainfall rate as small as 0.1 mm/hr. Drizzle rainfall rates are generally less than a few mm day$^{-1}$, thus these instruments are of little use to provide quantitative measurements of surface precipitation.

Radars can probe clouds and precipitation over large areas and can provide with detail maps of radar reflectivity and other radar observables but radar reflectivity is not equal to rain rate. The power intercepted by radar antenna is proportional to the reflectivity, which is the integral of the backscatter cross-section $\sigma$, of hydrometeors per unit volume of air. If all the particles in the volume are less than 800 µm in diameter, so as to fall in the Rayleigh regime at $\lambda = 3$ mm, the reflectivity at a wavelength $\lambda$ can be calculated as

$$\eta = \int_{0}^{\infty} N(D)\sigma(D)dD$$

(4.1)

and
\[\sigma(D) = \frac{\pi^5 D^6 |K|^2}{\lambda^5}\]  \hspace{1cm} (4.2)

where \(K\) is the refractive index factor of the substance involved, which in our case is water. \(D\) is the diameter (mm) and \(N(D)\) (m\(^{-3}\)) is the size distribution. The radar reflectivity factor is \(Z\) (mm\(^6\) m\(^{-3}\)) is proportional to the sixth moment of the Drop Size Distribution (DSD) and is defined as

\[Z = \int_0^{D_{\text{max}}} D^6 N(D) dD\]  \hspace{1cm} (4.3)

so that

\[\eta = \frac{\pi^5}{\lambda^5} |K|^2 Z\]  \hspace{1cm} (4.4)

Hence, if the scattering medium is known, then the reflectivity only depends on the DSD. Rain Rate (R) is proportional to the fourth moment of the DSD and can be written as

\[R = 6\pi \times 10^{-4} \int_0^{D_{\text{max}}} D^3 v(D) N(D) dD\]  \hspace{1cm} (4.5)

where \(v(D)\) is the terminal fall velocity of falling raindrops.

Without information about droplet spectra, it is difficult to predict a direct relation between \(Z\) and \(R\). Several approaches have been used in the past to derive R-Z relationships: 1) \(R\) and \(Z\) calculated from assumed or measured droplet spectra 2) Comparison of radar and rain gage. 3) Measurement of \(Z\) converted via an assumed R-Z relationship to predict \(R\), then comparing the
predicted \( R \) with measured \( R \). In this study the first approach is applied and a lognormal drop size distribution is assumed.

Most previous studies yielded empirical relationships of the form \( Z=aR^b \), where \( a \) and \( b \) are constants. The coefficients, \( a \) and \( b \) have been found to vary widely for different rain types and locations. Table 4.1 lists some of the numerous \( Z-R \) relationships developed previously. Although traditionally the relation is expressed with \( Z \) as an unknown expressed as a function of \( R \), generally \( Z \) from a radar is used to get an estimate of rainfall rate. Hence, rather than \( Z=aR^b \), a relation \( R=aZ^n \) is developed here. The last three relationships were developed in recent field experiment and for the same form of precipitation considered in this work. Hence, more insight is given to them than others (Comstock et al. 2004, VanZanten et al. 2005, Sharon (personal communication)).

Table 4.1 Empirical relationships between reflectivity \( Z \) (\( \text{mm}^6 \text{ m}^{-3} \)) and Rain Rate \( R \) (\( \text{mm hr}^{-1} \)) * denotes that the Rain Rate is in (\( \text{mm day}^{-1} \)) otherwise in (\( \text{mm hr}^{-1} \)).

<table>
<thead>
<tr>
<th>Equation</th>
<th>Reference</th>
<th>Location</th>
<th>Remarks</th>
</tr>
</thead>
<tbody>
<tr>
<td>( Z=162R^{1.16} )</td>
<td>Atlas &amp; Chmela (1957)</td>
<td>Lexington, MA</td>
<td>Stratiform Rain</td>
</tr>
<tr>
<td>( Z=350R^{1.42} )</td>
<td>Atlas &amp; Chmela (1957)</td>
<td>Lexington, MA</td>
<td>Stratiform Rain</td>
</tr>
<tr>
<td>* ( Z=25R^{1.3} )</td>
<td>Comstock et al</td>
<td>South East Pacific</td>
<td>Stratiform Drizzle</td>
</tr>
<tr>
<td>* ( R=0.51Z^{0.34} )</td>
<td>VanZanten et al.</td>
<td>California Coast</td>
<td>Coastal Stratiform</td>
</tr>
<tr>
<td>* ( R=0.24Z^{0.62} )</td>
<td>Sharon (personal communication)</td>
<td>California Coast</td>
<td>Coastal Stratiform</td>
</tr>
</tbody>
</table>

The Comstock et al. (2004) analysis of the EPIC 2001 field experiment radar data proposes a \( Z-R \) relationships for the estimation of the drizzle rainfall rate at the cloud base height: \( Z=25R^{1.3} \) or \( R=0.08Z^{0.76} \). For developing this \( Z-R \)
relationship, MMCR observations were used (although as we pointed in the previous section the MMCR suffered from saturation in the presence of drizzle in the field experiments prior to the 2004 Stratus cruise); the Drop Size Distribution was calculated from the filter paper used during the cruise and from in-situ microphysical measurements from the flights conducted below northeast Atlantic drizzling stratocumulus and Stratus by the UK Met Office C-130 aircraft (Comstock et al. 2004). A reflectivity threshold of -20 dBZ was used to separate cloud and drizzle returns. Evaporation of falling drizzle was reported. To explore a differential Z-R at cloud base and at the surface, an evaporation-sedimentation model was applied and from the model a relation between Z at cloud base and R at the surface was determined. The developed Z-R relations were then applied to the reflectivity captured by the scanning C-band radar aboard the RHB to estimate, rainfall rates for a larger volume. The best estimate of the mean area-averaged rain rate during the chosen days in EPIC was estimated to be 0.7 mm day$^{-1}$ at cloud base, with a total uncertainty envelope of 0.2-2.3 mm day$^{-1}$. The rain rate at the surface was estimated as 0.2 mm day$^{-1}$, with an uncertainty range of 0.1-0.6 mm day$^{-1}$.

Data from flights during the Dynamics and Chemistry of Marine Stratocumulus (DYCOMS-II) field experiment obtained several hundred kilometers South-Southwest of San Diego, California were used to develop the Z-R relation proposed by VanZanten et al (2005). The instrumentation onboard the NCAR/NSF C-130 aircraft included: the 95 GHz Wyoming Cloud Radar and three instruments for estimating the DSD; one based upon the single particle scattering
and the other two on shadowing of light. R-Z relations at cloud base and at surface were proposed by VanZanten et al. (2005). Instead of a reflectivity threshold, a threshold of 0.3 mm day$^{-1}$ in rain-rate intensity was used to distinguish between cloud and drizzle. Distinction was also made between light and heavy drizzle with a threshold of 1 mm day$^{-1}$. A rain-rate of 1 mm day$^{-1}$ corresponds to reflectivity value of -6.4 dBZ at cloud base and 8.6 dBZ at the surface using their derived R-Z relationship. Drizzle evaporation of 60% to 100% is reported from cloud base to the surface. Drizzle rates at the surface measured from the in-situ instruments were always found to be less than 1 mm day$^{-1}$.

During the Drizzle and Entrainment Cloud Study (DECS), conducted in the summer of 1999 on and offshore of Monterey Bay, California (Sharon et al. 2005), a Forward Scatter Spectrometer Probe (FSSP) measured the size distribution of cloud droplets while a Cloud Imaging Probe (CIP) was used to measure distributions of drizzle size droplets and derive a Z-R relationship at the cloud base height.

Figure 4.1 shows the rain-rate in mm day$^{-1}$ for range of reflectivity -20 dBZ to 20 dBZ, using the R-Z relations described above. It can be seen that all three of them yield different rain-rate. It is worth noting that the three relations, although for the same form of precipitation and cloud type, were developed using different instruments at different time and locations.
4.2 Derivation of R-Z using Frisch et al. (1995)

To derive a R-Z relationship from the dataset used in our study, a technique proposed by Frisch et al. (1995) is used (hereafter referred as Frisch model). The retrieval technique uses the Doppler Spectral moments to derive the DSD parameters. The quantities of interest are modal radius ($r_0$ (µm)), total number of droplets ($N$ (m\(^{-3}\))), drizzle rain rate ($R$ (mm day\(^{-1}\))) and drizzle liquid water content ($LWC$ (g m\(^{-3}\))). A brief description of the technique is given in the following section.
4.2.1 Description of Frisch model

The Frisch model is based on the assumption that the average drizzle droplets size distribution (DSD) can be represented by a log-normal size distribution that has the form:

\[
n = \frac{N}{\sqrt{2\pi \sigma_x}} \exp \left[ \frac{(x - x_0)^2}{2\sigma_x^2} \right]
\]

(4.6)

where \( x = \ln(r) \) and \( x_0 = \ln(r_0) \), \( r_0 \) is the modal radius (\( \mu m \)) and \( \sigma_x \) is the logarithmic spread of the distribution and \( N \) is the total number of drops per m\(^3\). The droplet number density spectrum \( n \) (m\(^{-3}\) \( \mu m^{-1} \)) defines the number of droplets per m\(^3\) between radius \( r - \frac{dr}{2} \) and \( r + \frac{dr}{2} \) where \( dr \) is an infinitesimal increment in the drizzle radius. The unknown parameters in this distribution are: the total number of drops \( N \), the modal radius \( r_0 \) and the logarithmic width of the distribution \( \sigma_x \) and the idea is to use the first three Doppler moments (reflectivity, mean Doppler velocity and Doppler spectrum width) as input (known) parameters and derive the three parameter of the lognormal distribution Fig. 4.2 shows an example of a lognormal DSD.
Figure 4.2 Sample distribution function of drizzle droplets. The modal radius is 120 µm, total number of droplets 10000 and the standard deviation is 0.35 µm.

The $k^{th}$ moment of the distribution can be defined as

$$
\langle r^k \rangle = r_0^k \exp \left( \frac{k^2}{2\sigma_s^2} \right)
$$

Using the above relation, all the moments of the Doppler Spectrum can be expressed in terms of the DSD parameters. The radar reflectivity which is the sixth moment of DSD and zeroth moment of Doppler spectrum can be expressed as
where $Z$ is in the units of mm$^6$m$^{-3}$, $r$ is in mm and $N$ is in drops m$^{-3}$.

The radar scattering properties of the droplet distributions can be expressed in terms of the radar reflectivity $\eta$ and the incremental scattering contribution of each size droplet as estimated by the standard Rayleigh approximation (Battan 1973):

$$\frac{\partial \eta}{\partial r} = 4\pi (2\pi / \lambda)^4 |K|^2 r^6 n (r)$$

where $|K|^2 = 0.93$ is the refractive index factor for liquid water and $\lambda$ is the radar wavelength. Thus, $\eta$ is the integral of Eq. (4.9) over all droplets.

$$\eta = \int \frac{\partial \eta}{\partial r} \partial r = \pi^4 \lambda^{-4} |K|^2 Z$$

The fundamental measurement of a Doppler radar is the Doppler spectrum. The Doppler spectrum $\partial \eta / \partial V$ is defined as the backscatter intensity between Doppler velocities $V - \frac{dV}{2}$ and $V + \frac{dV}{2}$ or alternatively the reflectivity weighted velocity distribution. The Doppler spectrum is directly related to the backscatter distribution defined in Eq. (4.10) through the fall velocity dependence on radius.

$$\frac{\partial \eta}{\partial V} = \frac{\partial \eta}{\partial r} \frac{\partial r}{\partial V_f} = 4\pi (2\pi / \lambda)^4 |K|^2 r^6 \frac{\partial r}{\partial V_f} n (r)$$

The $k^{th}$ velocity moment is then expressed as

$$\langle V^k \rangle_{D} = \eta^{-1} \int V^k \frac{\partial \eta}{\partial V} dV = \eta^{-1} \int [V_f (r)]^k \frac{\partial \eta}{\partial r} dr$$
\[ \langle V^k \rangle_D = \frac{\left\langle r^6 \left[ V_f(r) \right]^k \right\rangle}{\left\langle r^6 \right\rangle} \]  

(4.12)

From the above it is clear that once we specify \( V_f \), the above equation can be used to get all the parameters for the DSD. In order to do that the definition of drizzle is restricted to drizzle size drops which range from 45 \( \mu \)m to 400 \( \mu \)m in radius. This also assumes that the drops fall in Rayleigh regime. Hence the droplet fall velocity for this range of droplets can be assumed to be (Gossard et al. 1990)

\[ r = a V_f + b \]  

(4.13)

where \( a = 1.2 \times 10^{-4} \) s and \( b = 1.0 \times 10^{-5} \) m. This approximation is reasonable for 45 \( \mu \)m < \( r < 400 \) \( \mu \)m and for 0.3 m/s < \( V_f < 3.0 \) m/s. Particles smaller that 45 \( \mu \)m are considered as “Cloud Drops” while particles larger than 400 \( \mu \)m are considered as “Rain Drops”. For the purpose of our study, we can get the modal radius from the last two equations as

\[ r_0 = (a \langle V \rangle_D + b) \exp \left( -\frac{13 \sigma_x^2}{2} \right) \]  

(4.14)

where \( r_0 \) is in meter while \( \langle V \rangle_D \) is in m s\(^{-1}\). The total number of droplets can be calculated from the Z as

\[ N = \frac{Z \exp(-18 \sigma_x^2)}{2^6 \tau^6} \]  

(4.15)

The lognormal width of the DSD \( \sigma_x \) was initial taken constant equal to 0.35 throughout the calculations in this study. Since, all the parameters of the DSD at
all levels are calculated, the Rain rate (mm day$^{-1}$) and Liquid Water Content (g m$^{-3}$) can also be calculated by using the following equations

$$LWC = \left(\frac{4\pi}{3}\right)\rho_w N_0^3 \exp\left(\frac{9\sigma^2}{2}\right)$$

(4.16)

$$R = 36 \times 10^5 \times \frac{LWC}{\rho_w} \left[\left(\langle V_d \rangle + \frac{b}{a}\right) \exp\left(-3\sigma_x^2\right) - \frac{b}{a}\right]$$

(4.17)

where $\rho_w$ is expressed in g m$^{-3}$.

### 4.2.2 Model application to selected cases

In order to further characterize drizzle, an R-Z relation is developed using the above-mentioned technique. In order to do this, eight separate events are chosen from the entire cruise. The details about the event are given in Table 4.2

<table>
<thead>
<tr>
<th>Date</th>
<th>Drizzle event</th>
<th>Time (UTC)</th>
<th>Duration (min)</th>
</tr>
</thead>
<tbody>
<tr>
<td>11 Dec 2004</td>
<td>1</td>
<td>5:00 – 5:39</td>
<td>39</td>
</tr>
<tr>
<td>15 Dec 2004</td>
<td>2</td>
<td>7:48 – 8:40</td>
<td>52</td>
</tr>
<tr>
<td>15 Dec 2004</td>
<td>3</td>
<td>13:42 – 14:18</td>
<td>36</td>
</tr>
<tr>
<td>15 Dec 2004</td>
<td>4</td>
<td>15:21 – 16:58</td>
<td>97</td>
</tr>
<tr>
<td>16 Dec 2004</td>
<td>5</td>
<td>5:31 – 6:15</td>
<td>44</td>
</tr>
<tr>
<td>16 Dec 2004</td>
<td>6</td>
<td>6:15 – 7:20</td>
<td>65</td>
</tr>
<tr>
<td>16 Dec 2004</td>
<td>7</td>
<td>10:49 – 11:43</td>
<td>54</td>
</tr>
<tr>
<td>19 Dec 2004</td>
<td>8</td>
<td>8:06 – 9:13</td>
<td>67</td>
</tr>
</tbody>
</table>

Assuming uniform (homogeneous) drizzle DSD during each event, the reflectivity and Doppler velocity is averaged for the entire period to get a single vertical profile of reflectivity and mean Doppler velocity. In order to differentiate
between cloud and drizzle, reflectivity threshold of -17 dBZ is used. Thus, only radar observations with reflectivity greater than –17 dBZ are used in the averaging. The averaging was done in reflectivity factor $Z$ ($\text{mm}^6\text{m}^{-3}$) rather than in reflectivity itself which is expressed in dBZ. While the averaging of Doppler velocity was done by weighting it by reflectivity, hence average Doppler velocity at any height was calculated as

$$\langle V_d \rangle_{\text{average}} = \frac{\sum Z \times V}{\sum Z}$$

(4.18)

The technique was applied to these averaged vertical profiles to get the drizzle DSD and its integral parameters such as LWC and R. Fig. 4.3 to Fig. 4.14 illustrate the selected events and the retrieved parameters. As the FMCW radar attenuates really fast above the cloud base, the model is only applied to the lower 500 m of the MABL as the average ceilometer derived cloud base for the Stratus04 cruise was 1104 m (Serpetzoglou 2005), by choosing the lower 500 m we make sure that all the samples are un-attenuated.

a) 11 December 2004

The first event chosen for analysis is a 39-minute drizzle event observed on 11 December 2004. The ship was stationed at the WORS (20° S 85° W) during this period. The event occurred between 0500 and 0539 Universal Time Constant (UTC) which corresponds to 0100 and 0139 Local Time (LT). It is a nighttime drizzle event. The Doppler Spectra moments captured by the FMCW radar during this time are shown in Fig. 4.3. The Doppler spectrum width is less
around 0.3 m/s with little variation both in time and height. This suggests that the width of the DSD is also narrow and does not change much in time and height. The patchiness seen in the Doppler velocity plot is due to the ship motion (pitch and roll), which moves the radar itself up and down adding a ship velocity component to the radar measured Doppler velocity. Averaging is done in the time domain to remove the ship motion contamination in the radar signal.

The mean Doppler velocity and radar reflectivity are average at all heights and are shown in Fig. 4.4. It can be inferred from the average profiles that there is a significant decrease in the reflectivity near the surface while there is little change in the Doppler velocity. Using these values of reflectivity and velocity, the modal radius \( r_0 \), number of droplets \( N \), drizzle rainfall-rate \( R \) and Liquid Water Content \( \text{LWC} \) were calculated at each height. These parameters are shown in Fig. 4.5. There is a significant decrease in the number of droplets, as the drizzle descends, but there is little change in the modal radius. The drizzle number concentration decreases rapidly between 500 m and 200 m from 15000 m\(^{-3}\) to 2500 m\(^{-3}\), but vary little below this level. This suggests that the evaporation occurs mainly between 500 m till 200 m. The vertical profiles of \( R \) and \( \text{LWC} \) also indicate the evaporation of drizzle as it falls down, since both of them decrease gradually. The rain-rate decreases from 4.5 mm day\(^{-1}\) at 500 m to 2 mm day\(^{-1}\) at 100 m. The LWC decreased from 0.056 g m\(^{-3}\) at 500 m to 0.008 g m\(^{-3}\) at 10 m. The rapid decrease above 200 m and slower decrease below can also be seen in the LWC profile.
Figure 4.3 Time-height sections of Doppler spectra moments for 11 December 2004. The top panel shows the reflectivity (dBZ), middle panel mean Doppler Velocity (ms$^{-1}$) and bottom panel Doppler spectrum width (ms$^{-1}$). The ceilometer detected cloud base during this event is plotted in the top panel with black dots.
Figure 4.4 Average Reflectivity (dBZ) and average reflectivity weighted mean Doppler velocity (ms$^{-1}$) for event 1 shown in Fig. 4.3.
Figure 4.5: Derived micro-physical parameters for event 1 using the Frisch model. Figure shows the modal radius, total number of droplets, LWC and Drizzle rate as a function of height clockwise from top left.
b) 15 December 2004

Drizzle events chosen for analysis from 15 December 2004 are shown in Fig. 4.6. The ship was stationed at the WORS (20° S 85° W) on this day. The periods chosen are bounded by red lines and are numbered as event 2, 3 and 4 respectively. The time-averaged profiles of reflectivity and Doppler velocity for these events are shown in Fig 4.7. It can be seen that although the events occur almost continuously with little time gap between them, there is considerable variation in the magnitude of these averaged profiles. The derived parameters after the application of Frisch model for these events are shown in Fig. 4.8. Several things can be observed from these profiles. The modal radius for all the events show little variation in height and in the magnitude and remain constant around 120 µm, while there is a considerable amount of variation in the total number of droplets. Also, it is worthwhile to note that event 3 which occurs in the middle is characterized by strong evaporative profiles of rain rate and liquid water content, while the events at the edges do not show strong evaporation.
Figure 4.6 Time-height sections of Doppler spectra moments for 11 December 2004. The top panel shows the reflectivity (dBZ), middle panel mean Doppler Velocity (ms$^{-1}$) and bottom panel Doppler spectrum width (ms$^{-1}$). The ceilometer detected cloud base during this event is plotted in the top panel in black dots. The regions between red lines are the chosen event 2, 3 and 4 respectively.
Figure 4.7 Average Reflectivity (dBZ) and average reflectivity weighted mean Doppler velocity (ms$^{-1}$) for event 2 (red), 3 (blue) and 4 (green).
Figure 4.8 Derived micro-physical parameters for event 2 (red), 3 (blue) and 4 (green) using the Frisch model. Figure shows the modal radius, total number of droplets, LWC and Drizzle rate as a function of height clockwise from top left.
16 December 2004

The drizzle events for 16 December 2004 are shown in Fig. 4.9. The individual events chosen for averaging and further analysis are shown between red lines. The events are numbered as event 5 thru 7 in the order. The ship was stationed at the WORS during this period. It can be seen that much of the drizzle evaporates before reaching the surface. The three events occurred at 0130, 0215 and 0650 local time respectively. Hence all of them can be regarded as night time events.

The averaged profiles of reflectivity and Doppler velocity for these events are shown in Fig 4.10. Event 6 is characterized by low values of reflectivity and Doppler velocity while event 5 and 6 show similar behavior as event 2 and 4. The derived parameters for the events are shown in Fig 4.11. The modal radiiuses for all the events remain constant throughout the 500 m and also show little variation in magnitude from each other. The exponential increase in the total number of droplets for event 6 below 120 m seems to be an artifact as no mechanism to cause such an increase at lower levels can be thought of. Similar to 15 Dec, the event which happened in the middle is characterized by strong evaporation while the edge events show little evaporation.
Figure 4.9 Time-height sections of Doppler spectra moments for 16 December 2004. The top panel shows the reflectivity (dBZ), middle panel mean Doppler Velocity (ms$^{-1}$) and bottom panel Doppler spectrum width (ms$^{-1}$). The ceilometer detected cloud base during this event is plotted in the top panel in black dots. The regions between red lines are the chosen event 5, 6 and 7 respectively.
Figure 4.10 Average Reflectivity (dBZ) and average reflectivity weighted mean Doppler velocity (ms\(^{-1}\)) for event 5 (red), 6 (blue) and 7 (green).
Figure 4.11 Derived micro-physical parameters for event 5 (red), 6 (blue) and 7 (green) using the Frisch model. Figure shows the modal radius, total number of droplets, LWC and Drizzle rate as a function of height clockwise from top left.
d) 19 December 2004

The last event chosen for analysis occurred on 19 December 2004 and the first three Doppler moments for the event are shown in Fig. 4.12. The 67 min event occurred when the ship was moving eastwards towards the port of Valparaiso, Chile at around 0400 local time. The averaged Reflectivity and Doppler velocity for this event is show in Fig. 4.13, while the derived parameters are shown in Fig. 4.14. This event also shows the artifact of exponential increase in total number of droplets at low levels as in event 6. Otherwise the event show similar behavior as any of the other previous events without strong evaporation.
Figure 4.12 Time-height sections of Doppler spectra moments for 19 December 2004. The top panel shows the reflectivity (dBZ), middle panel mean Doppler Velocity (ms\(^{-1}\)) and bottom panel Doppler spectrum width (ms\(^{-1}\)). The ceilometer detected cloud base during this event is plotted in the top panel in black dots.
Figure 4.13 Average Reflectivity (dBZ) and average reflectivity weighted mean Doppler velocity (ms\(^{-1}\)) for event 8.
Figure 4.14 Derived micro-physical parameters for event 8 using the Frisch model. Figure shows the modal radius, total number of droplets, LWC and Drizzle rate as a function of height clockwise from top left.
4.2.3 R-Z relation

To develop a reflectivity-rain rate relationship for Stratus04, the averaged values of these two at 500 m, 300 m and 20 m are separately plotted and the best fit lines for these were calculated. Fig. 4.15 shows the combined plot of all the points and the best fit line for each set of points. It can be seen that the slope of the best fit line increases with height. As the R-Z relations are developed in the form of \( R = aZ^n \), the values of parameters \( a \) and \( n \) were determined for the best fit lines and are tabulated in Table 4.3.
Table 4.3 Values of parameter $a$ and power $n$ in the R-Z relationship $R=aZ^n$ developed from Stratus04 data.

<table>
<thead>
<tr>
<th>Height</th>
<th>$a$</th>
<th>$n$</th>
</tr>
</thead>
<tbody>
<tr>
<td>500 m</td>
<td>1.2667</td>
<td>0.9317</td>
</tr>
<tr>
<td>300 m</td>
<td>1.2582</td>
<td>0.8263</td>
</tr>
<tr>
<td>20 m</td>
<td>1.1607</td>
<td>0.5537</td>
</tr>
</tbody>
</table>

For further comparison of the developed R-Z with others, the R-Z at 500 m is plotted with other R-Z developed for cloud-base values; Fig. 4.16 and R-Z at 20 m is plotted with others developed for surface; Fig. 4.17. It should be noted that the radar used in developing the R-Z relation is this study is W-band radar and except the study by VanZanten et al. (2005), none of the others have used W-band radar. It can be seen that the R-Z developed by Comstock et al. (2004) for C-band radar yields the lowest rainfall rate compared to any other R-Z for same reflectivity. While R-Z developed by VanZanten et al. (2005) yield heaviest rainfall rate for low reflectivity. The R-Z developed here yields high rainfall rates but has almost the same slope developed by Comstock et al. (2004). There are few R-Zs developed for surface values. The R-Z developed in this study and by VanZanten et al. (2005) is plotted in Fig. 4.17. It can be seen that the slope of the R-Z is considerably less than the one at the cloud base. It can also be noted that the rainfall rate for 20 dBZ at surface is little less than 11 mm day$^{-1}$, compared with 100 mm day$^{-1}$ at cloud base. Hence, the evaporation of drizzle is implicit in the R-Z relation.
Figure 4.16 Comparison of developed (blue) R-Z at 500 m with other R-Z developed previously; Comstock et al. 2004 (red), VanZanten et al. 2005 (black), Sharon (green) at cloud base for the same cloud type and precipitation.
4.2.4 Sensitivity of width of DSD to retrievals

Throughout the analysis presented here, a constant logarithmic width of the DSD of 0.35 was assumed, as proposed by Frisch et al. (1990). Fig 4.18 shows the retrievals for event 3 assuming different values of DSD width. It can be seen that in order to get the correct drizzle DSD and true value of the derived parameters, it is necessary to get the true value of the DSD width. On the other hand the retrieved profiles of rain rate and liquid water content for the events
described in section 4.2.2 advocate the fact that there is lot of evaporation associated with the falling drizzle in the lower half of the MABL as the drizzle encounters sub-saturated conditions. Hence, in order to retrieve the true parameters from the Frisch model, it is necessary to force the model to account for the sub-saturated conditions and hence evaporation. To address the issue of determining the correct width of drizzle DSD and incorporating drizzle evaporation in the Frisch model, a simple prognostic evaporation model is described in the next section.
Frisch derived parameters with $\sigma_x = 0.25$ (green), $\sigma_x = 0.30$ (blue) and $\sigma_x = 0.35$ (red) for event 3.

Figure 4.18 From top left, profiles of modal radius, total number of droplets, Liquid water Content and Rain rate for event 3 for values of logarithmic width of distribution 0.25, 0.30 and 0.35 in green, blue and red respectively.
4.3 Evaporation Model

As the falling drizzle encounters sub-saturated environment in the lower 500 m of the MABL, evaporation of drizzle is inevitable. The evaporation model described here provides an alternate way to study the boundary layer structure and also to test the Frisch model. Similar attempts made earlier (Wood 2005; Comstock et al. 2004) have used different setup and parameters. The model setup for this study is described in the following section.

4.3.1 Model setup

In a sub-saturated environment, raindrops evaporate by the process of diffusion of water vapor. The rate of diffusion of water vapor is proportional to the gradient of the mole fraction of the water vapor. Here, we use an approximation, commonly employed in cloud microphysics that takes the rate of diffusion to be proportional to the gradient of vapor density. The error incurred is small and justified in view of other approximations made (Rogers and Yau 1996). With this approximation the following equation is obtained (Rogers and Yau 1996) for the rate of change of mass drop by diffusion.

\[
\frac{dm}{dt} = 4\pi r D_v f_v \Delta \rho_v
\]  

(4.19)

where \(m\) is the mass of the raindrop with radius \(r\), \(t\) is the time, \(D_v\) is the diffusivity of water vapor in air, \(f_v\) is the ventilation coefficient for the diffusion of water vapor, and \(\Delta \rho_v\) is the difference of vapor density between the environment and the drop surface. After substituting the mass in terms of radius and density in the
above equation and multiplying the numerator and the denominator of the left hand side by \(dh\), we can write the above equation as below

\[
V_r \frac{dr}{dh} = \frac{1}{\rho_w} D_v f_v \Delta \rho_v
\]  

(4.20)

where the vertical co-ordinate \(h\) is measured downwards from a certain reference level, \(\rho_w\) is density of falling drop and \(V\) is the terminal fall velocity of the raindrop.

We now provide some insight to the variables in the above equation. The terminal fall velocity of the falling drizzle drop can be represented in terms of the radius of the drop (Gossard et al. 1990).

\[
V = \frac{r - b}{a}
\]  

(4.21)

where \(r\) is the radius of the falling drop in meters and \(a = 1.2 \times 10^{-4} \text{s}\), \(b = 1.0 \times 10^{-5} \text{m}\). The values of \(a\) and \(b\) are reasonable for the sizes of drizzle drops considered in the study. If the solute effect is neglected, the density of the falling drop can be regarded as water density, which is 1000 kg m\(^{-3}\).

From the Smithsonian Meteorological Tables (SMT), the Diffusivity of water vapor in air \(D\) at any given temperature and pressure can be calculated as

\[
D_v = 0.211 \times 10^{-4} \left( \frac{T}{T_0} \right)^{1.94} \left( \frac{P_0}{P} \right)
\]  

(4.22)

where \(T_0 = 273.15 \text{ K}\) and pressure \(P_0 = 1013.25 \text{ mb}\). We intend to apply the evaporation model to the lowest 500 m of the MABL, hence calculations of these parameters are performed in this domain only. If we consider the lapse rate to follow the dry-adiabatic lapse rate, then the vertical profile of \(D_v\) follows that
shown in Fig. 4.22. It can be seen that the variation of $D_v$ in the domain considered is $6.07 \times 10^{-7} \text{ m}^2/\text{s}$. Li and Srivastava (2001) have shown that $D_v$ varies slowly with height and can be well represented by its mid-level value to a high degree of accuracy. Calculation made using actual values varied by less than 3% with those obtained from using mid-level values. Hence, the value of $D_v$ at 250 m $2.42 \times 10^{-5}$ is used in this study.

Figure 4.19 Vertical profile of Diffusion of water vapor ($D_v$) in the air for the bottom 500 m of MABL assuming that the temperature profile follows the dry adiabatic lapse rate.
The ventilation coefficient $f_v$ is a function of diffusivity $D_v$ (m$^2$ s$^{-1}$), kinematic viscosity of air $\nu$ (m$^2$ s$^{-1}$), raindrop radius $r$ (m) and the terminal fall velocity of the raindrop $V$ (m s$^{-1}$). Pruppacher and Klett (1997) have proposed relations for ventilation coefficients depending on the value of a deciding parameter $N_{Sc}^{0.33}N_{Re}^{0.5}$, where $N_{Sc}$ is Schmidt number and $N_{Re}$ is the Reynolds number. For values of the deciding parameter less than 1.4 following relation is proposed.

\[
f_v = 1 + 0.108(\nu / D_v)^{0.33}(2Vr / \nu)^{0.5}
\] (4.23)

while for deciding parameter greater than 1.4 the relation is

\[
f_v = 0.78 + 0.308(\nu / D_v)^{0.33}(2Vr / \nu)^{0.5}
\] (4.24)

In the above equations, we can assume that the kinematic viscosity remains constant in our domain and assign its value to be $0.1529 \times 10^{-4}$ m$^2$/s, which corresponds to 1000 mb and 20°C from the SMT. The fall velocity $V$ and drop radius are related to each other by Eq. (4.21). The changes in $f_v$ over the range of drop radius considered here is shown in Fig. 4.23. It can be seen that $f_v$ changes from unity for drops smaller than 40 µm to little more than 4 for drop size of 400 µm. Calculations performed using the middle value of $f_v$ (Li & Srivastava 2001) overestimated the ventilation effects for smaller droplets, producing erroneous values of the final drop radius. Hence, $f_v$ was calculated for each droplet for its initial size at each level. Convergence studies for $f_v$ by reducing the grid size are not attempted at this point.
The difference between the vapor density of the environment and vapor density at the drop wall can be expressed as following after some simple calculations.

\[
\Delta \rho_v = \frac{e_s}{R_v T} (RH - 1)
\]

where \(e_s\) is the saturation vapor pressure at environmental temperature \(T\), \(R_v\) is the gas constant for water vapor and \(RH\) is the environmental Relative Humidity (RH). While deriving the above equation it was assumed that the temperature of
the drop and the environment is the same i.e. any form of latent heat is quickly
dissipated so as to have same temperature as of the environment. Also the vapor
pressure at the drop surface is equal to the saturation vapor pressure at that
temperature. Following the calculations from Bolton (1980) the saturation vapor
pressure can be calculated as follows

\[ e_s = 6.112 \times \exp \left( \frac{17.67 \times T}{T + 243.5} \right) \]  

(4.26)

where \( e_s \) is in milli-bars when \( T \) is expressed in °C. It can be assumed for the
temperature profile to follow the dry-diabatic lapse rate of 9.8 °C/km till the Lifting
Condensation Level (LCL) height. Hence, the temperature dependence of \( e_s \) can
be converted to height dependence as

\[ e_s = 611.2 \times \exp \left( \frac{17.67 \times (T_0 - 0.0098 \times h)}{T_0 - 0.0098 \times h + 243.5} \right) \]  

(4.27)

where \( T_0 \) is the temperature at the surface in °C and \( h \) is the height measured
upwards from surface in meters. The relative humidity vertical profile can be
assumed to vary from the surface value till 100% at the Lifting Condensation
Level (LCL). Hence, the relative humidity can be expressed as

\[ \text{RH}(h) = \left(1 - \frac{h}{h_{\text{LCL}}} \right) \times \text{RH}_0 + \frac{h}{h_{\text{LCL}}} \]  

(4.28)

where \( \text{RH}_0 \) is the relative humidity at the surface, \( h \) is the height in meter
measured upwards from the surface and \( h_{\text{LCL}} \) is the height of the LCL in meter.

The height of the LCL can be found out by following the dry adiabatic till the
temperature becomes equal to the temperature at LCL, which is given by Bolton (1980) as follows

\[ T_{LCL} = \frac{2840}{3.5 \times \log(T_0 + 273) - \log(e) - 4.805} + 55 - 273 \]  \hspace{1cm} (4.29)

and

\[ h_{LCL} = 15.5 + \frac{T_0 - T_{LCL}}{0.0098} \]  \hspace{1cm} (4.30)

where, \( e \) is the partial pressure due to water vapor at the surface in milli-bars, all temperatures are in °C and heights in meter. In order to calculate \( e \), we need the surface pressure which is assumed to be constant 1015 mb in this study. During the entire cruise the surface pressure varied between 1012 mb to 1019 mb with mean value of 1015 mb. Hence, the density difference between environment and the drop surface can be calculated by using Eq. (4.30) which only has surface temperature and humidity as the input parameters.

\[ \Delta \rho_v = \frac{611.2}{R_v} \exp \left( \frac{17.67 \times (T_0 - 0.0098 \times h)}{T_0 - 0.0098 \times h + 243.5} \right) \times \frac{1}{T_0 - 0.0098 \times h + 273} \times (RH_0 - 1) \times \left( 1 - \frac{h}{h_{LCL}} \right) \]  \hspace{1cm} (4.31)

By using the above formula, \( \Delta \rho_v \) was calculated using the mean surface temperature and relative humidity values of event 3 described earlier. The vertical profile of it is shown in Fig. 4.24. Although the actual curve shown in blue is a second degree function, it can be assumed to be linear for the first 500 m above the surface. The linear curve is shown by red dots. Hence \( \Delta \rho_v \) can be safely expressed as
\[ \Delta \rho_v = (\Delta \rho_v)_0 + m \times h \]  

(4.32)

where \((\Delta \rho_v)_0\) is the value of \(\Delta \rho_v\) at surface. Since in eq. (4.20), height is measured downwards from a certain reference level, for simulating drizzle, \(\Delta \rho_v\) is expressed as

\[ \Delta \rho_v = (\Delta \rho_v)_z + m \times h \]  

(4.33)

where \((\Delta \rho_v)_z\) is the value of \(\Delta \rho_v\) at a certain level \(z\) from which the height is measured positive downwards. When we substitute Eq. (4.21) and Eq. (4.33) in Eq. (4.20), it becomes

\[ \left( \frac{r - b}{a} \right) r \frac{dr}{dh} = \frac{1}{\rho_w} \frac{D_v f_v}{\left( (\Delta \rho_v)_z + m \times h \right)} \]  

(4.34)

The above equation can be used to calculate the variation of drizzle drop radius with height, with specified temperature and relative humidity at the surface.
Figure 4.21 Difference between environmental vapor density and drop wall vapor density is shown as a function of height. The blue curve shows the actual density difference which is a 2nd degree function while the red dots show the linear fit of the curve.

4.3.2 Application of evaporation model

Before applying the evaporation model to cases during the cruise, some simulations were performed to test the severity and consistency of the model in the domain of consideration. In this study the radius of drizzle drop is assumed to vary from 45 µm to 400 µm. Fig 4.22 shows the evolution of the radius of drops in this range when they are allowed to fall in a sub-saturated environment with LCL
height of 600 m and surface relative humidity of 80%. It can be seen that as the evaporation rate is inversely proportional to the drop size, the smaller droplets evaporate much faster than the bigger one and are less likely to make it to the surface, while the bigger drops experience little change in their size. In the simulated conditions considered here, all drops less than 250 µm radius at the 500 m evaporate completely before reaching the surface.

Figure 4.22 Simulation of change in drop radius for drizzle size drops in a sub-saturated environment with LCL at 600 m and surface relative humidity of 80%.

It can be concluded from the previous simulation that the smaller drops to evaporate much faster than the larger ones. To study the impact of this
phenomenon on the evolution of the shape of drizzle DSD, the evaporation model was initialized by the Frisch model retrieved lognormal DSD using logarithmic width of distribution as 0.35 at 350 m for event 3 described earlier. The resulting DSDs at various levels below are shown in Fig. 4.23. It can be seen that the distribution gets truncated at the lower end just 50 m after the start due to rapid evaporation of the smaller drops. Hence, the shape of the distribution changes from lognormal to truncated lognormal at smaller radii. Below that level, there is a decrease in the total number of drops while the shape of the distribution remains the same as truncated lognormal.
Figure 4.23 Evolution of drizzle Drop Size Distribution starting at 350 m, after every 50 m for event 3. The initial DSD (350 m) is lognormal in shape, while it gets truncated at lower end at levels below due to sub-saturated conditions.

From the above mentioned simulations, it can be concluded that evaporation has a stronger impact on the number of smaller drops rather than larger ones and faster evaporation of smaller drops than the bigger drops modifies the drizzle DSD from a lognormal shape to truncated lognormal shape. Hence, we can say that evaporation has a strong impact on the evolution of the drizzle DSD and for getting the true DSD we need to incorporate it in the retrieval model. For doing that a simple algorithm illustrated in Fig. 4.24 is presented. In the proposed technique, we started out with the averaged profiles of Reflectivity
and Doppler velocity for an event observed during the cruise. Frisch DSD retrieval model is then applied to these profiles with an initial guess of the logarithmic width of the distribution, which is 0.35 in our case. DSD parameters were then retrieved for the entire sub-cloud layer and then the evaporation model was initiated with DSD at a reference level. The evaporation model yielded DSDs and other parameters at lower levels were then compared with the observed and the Frisch model derived DSDs and parameters. At this point minimization theory was applied to find out the agreement between the parameters yielded by the two models. If the agreement was good then the solution was treated as final else the process was repeated by changing the initial guess of the logarithmic width of the distribution in the Frisch model.
Described below are the results obtained for event 3 and event 6 after the application of the above mentioned algorithm. The evaporation model was initialized with DSD derived from the Frisch model at a certain height and the final drop radius was calculated numerically at every 10 m below it. All the calculations were made explicitly using a bin size of 1 µm. After every 10 m, the DSD was re-arranged and other parameters were calculated. The ventilation
coefficient was calculated at each height for the starting drop radius. To take into account the localized increase in the relative humidity due to the evaporation of small drops, a surface relative humidity value of 10% higher than the actual was used.

The reflectivity, total number of droplets, Rain-rate and Liquid Water Content derived from the Frisch model and the evaporation model for case 3 are shown in Fig. 4.25. The model is applied from 350 m to 50 m. The two models agreed with each other when the Frisch model logarithmic width of DSD was set to 0.25. The evaporation model yielded a reflectivity 2 dBZ less than the observed reflectivity at the surface. Effects of this difference are carried through into the other comparisons. The figure also shows the Frisch model derived parameters for different values of logarithmic width. It can be seen that for higher values of DSD width, the derived parameters are also higher, suggesting underestimation of evaporation in these profiles. Shown in Fig. 4.26 is the DSD yielded by Frisch model and the evaporation model at 50 m level. As stated earlier the evaporation model derived DSD has a truncated lognormal shape while the Frisch model assumes a lognormal DSD. The Frisch model derived DSD is much tightly spaced while the evaporation model yielded distribution is sparse. The sparseness of the distribution is due to different evaporation rate for smaller and bigger drops. Also, it is worth noting that in-situ measurements of drizzle DSD obtained by VanZanten et al. (2005) and Wood (2005) have also yielded truncated lognormal drizzle DSD.
Figure 4.25 Comparison between Frisch model with $\sigma_x=0.25$ (red), $\sigma_x=0.30$ (blue), $\sigma_x=0.35$ (green) and evap model (dotted) comparisons for event 3 described in section 4.2.2. Profiles of reflectivity, total number of droplets, liquid water content and rain rate are shown clockwise from top left.
The algorithm was then applied to event 6 at 350 m level. The optimal width of the DSD was found to be 0.30 in this case. The resulting plots of comparison between the two models are shown in Fig 4.27. It can be seen that the reflectivity calculated from the evaporation model matches well with the observed reflectivity throughout the domain. The profiles of total number of drops, rain rate and LWC also comply with the evaporation mode. Fig. 4.28 shows the DSD from both the models at 150 m. As the width of the distribution for the Frisch model is greater than that in the event 3, it can be seen that the
drops are not tightly spaced. In this case also the evaporation model ends up with a truncated lognormal distribution rather than a lognormal as assumed by the Frisch model. While, this is similar to the analysis for case 3, it is worth noting that in this case the derived parameters from both the models agree closely, whereas for case 3 they did not.
Figure 4.27 Comparison between Frisch model with $\sigma_x = 0.30$ (red), $\sigma_x = 0.35$ (blue), $\sigma_x = 0.40$ (green) and evap model (dotted) comparisons for event 6 described in section 4.2.2. Profiles of reflectivity, total number of droplets, liquid water content and rain rate are shown clockwise from top left.
Figure 4.28 Drizzle Drop Size Distribution for event 6 at 150 m derived using Frisch model (red) with $\sigma_x=0.30$ and evaporation model (blue).
Chapter 5 – Results and Future work

5.1 Summary and Results

Due to its strong feedback with ENSO, vast extent and huge impact on the radiation budget (Li & Philander 1996), the southeast Pacific stratocumulus regime is one of the major features in global climate which needs special attention both regarding observations and modeling studies. Drizzle under the Stratocumulus clouds not only affects the water budget in the clouds but also has a significant impact on the thermodynamic structure of the MABL, making it one of the most important features in the understanding of these clouds and the climatology of the region. Drizzle has also been linked to the Stratocumulus cloud coverage and thereby affecting the radiation budget (Albrecht 1989). Thus further elucidation of role drizzle in Stratocumulus is important.

In this study an attempt is made to characterize drizzle using the observations from two cloud Doppler radars during the Stratus04 cruise conducted in the northern hemisphere fall of 2004. One of the radars is the MilliMeter Cloud Radar (MMCR) which was also used in the previous field studies in the region; EPIC 2001, PACS 2003 (Bretherton et al. 2001, Comstock et al. 2004, Kollias et al. 2004). The radar was running with two modes; Boundary Layer (BL) mode and Precipitation (PR) mode. Comparison of these two modes suggested that the BL mode suffers from saturation during drizzle events while the PR mode does not. This confirms that MMCR-PR mode is a good tool for drizzle studies.
The other radar used in this study is CIRPAS’s Frequency Modulated Continuous Wave (FMCW) radar, which is primarily designed for aircraft cloud studies (Mead et al. 2003) and was used for the first time in a ship based cloud experiment. This radar benefits from the absence of “Dead Zone” along with finer spatial (10 m) and temporal (1.3 sec) resolution. The radar does not suffer from any form of saturation although suffers from attenuation above the cloud base.

The drizzle Drop Size Distribution (DSD) retrieval model described by Frisch et al. (1995) was then applied to the FMCW radar data from the lower 500 m of the boundary layer for specific cases. The derived parameters are modal radius, total number of droplets, Rain rate and Liquid Water Content (LWC) which indicate substantial evaporation of drizzle. A bi-level R-Z relationship was then developed from the retrievals. The resulting rain-rates for the chosen cases from the R-Z relation developed in this study, along with relations developed in the past (Comstock et al. 2004, VanZanten et al. 2005) are shown in Fig 5.1. The cooling induced due to the evaporation is shown in the bottom panel of the figure. It can be concluded from the figure that the rain-rates calculated from R-Z developed by VanZanten et al. (2005) are relatively high, while those by Comstock et al. (2004) are low compared to rain-rates calculated from relation developed in this study. It is also noteworthy that although the rain-rates vary a lot, the evaporative cooling rates inferred from the R-Z relationship developed here and by Comstock et al. (2004) are comparable.
Figure 5.1 Rain rate at cloud base (top panel) and at the surface (middle panel) for cases discussed in the literature from R-Z relations developed by Comstock et al. (2004) (red), VanZanten et al. (2005) (green) and Ghate et al. (black). The resulting cooling rates due to the evaporation are shown in the bottom panel.
The Frisch model assumes a lognormal DSD and a assumed value for the logarithmic width of distribution ($\sigma$), to which the derived parameters are sensitive. We therefore constructed a simple algorithm that iteratively solves for $\sigma$ by explicitly accounting for evaporation. The algorithm uses a simple prognostic evaporation model which is applicable to the conditions encountered during the cruise in the lower 500 m of the boundary layer and relies on Pruppacher and Klett (1997), Rogers and Yau (1996) and Li and Srivastava (2002) for numerical development. While the Frisch model assumes a constant logarithmic width value of 0.35, the iterative procedure yielded values of 0.30 and 0.25 for the two chosen cases. One of the cases ($\sigma = 0.30$) showed an excellent agreement between the two models while the other case ($\sigma = 0.25$) showed significant differences in the lower 150 m of the MABL where the difference could have been contributed by the surface processes.

Simulations from the evaporation model showed that drizzle size drops experience most of their evaporation in the lower half of the boundary layer. The rate of evaporation is inversely proportional to the droplet radius, so that the smaller droplets tend to evaporate much faster than the larger ones. Due to this phenomenon, the drizzle DSD changes in shape from the Frisch model assumed lognormal to a truncated lognormal with truncation at lower end. Aircraft observations (VanZanten et al. (2005), Wood (2005)) also support the finding of truncated lognormal drizzle DSD.
Hence the major findings of this study can be summarized as follows

- The Boundary Layer (BL) mode of MMCR saturates during the drizzle events failing to record the true reflectivity. This might have some implications on the datasets from the earlier cruises which used the same radar.

- It is necessary to take into account the sub-saturated environmental conditions below the cloud base as drizzle size droplets are severely prone to evaporation.

- The algorithm described in the literature can be successfully used to determine the true width of the drizzle DSD removing the need to assume a value of logarithmic width of the DSD within the Frisch model.

- The evaporation model is a good tool for studying drizzle evaporation and to determine meteorological parameter in the radar “Dead Zone”.

- Due to rapid evaporation of smaller droplets than the bigger drops, the lognormal drizzle DSD gets truncated at the lower end. This leads to a major change in the DSD few tens of meters below the initial height in the evaporation model.

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5.2 Future Work

In the view of the results and conclusion in this work, it can be seen that there is a need for future work for better understanding of the drizzle phenomena and
the boundary layer structure itself. As concluded in this study, evaporation of drizzle is an integral part of the boundary layer and lack of current instruments ability to measure the drizzle rate makes radar the only and the best tool to study the phenomena. Rain gages that generally have only point measurements are not able to capture the low rain rates associated with drizzle under stratus clouds. Hence, there is a need for developing surface drizzle gages that can sample a larger area with the sensitivity needed to detect low rainfall rates.

Simulations from the evaporation model suggested transition of drizzle DSD from lognormal to truncated lognormal in few tens of meters below the cloud base. To further evaluate and study this phenomenon, the evaporation model can be applied to in-situ DSD measurements (e.g. from DECS-99, Sharon et al. 2006) at the cloud base and then compared with measurements at certain level below it.

As the reflectivity of radar is proportional to the $6^{th}$ power of drop diameter, it is not the best parameter for retrieving meteorological parameters. The reflectivity of a single drop of 100 µm diameter is exactly equal to reflectivity of one million drops of 10 µm diameter, while the later contains much more water than the earlier. To avoid this paradox, dual wavelength observations of drizzle should be made since the radar attenuation is directly proportional to the liquid water content.

A combination of millimeter wavelength radars (W, K or X-band) along with in-situ measurements of cloud microphysics and aerosols from aircraft penetration in the context of a large field experiment (VOCALS 2007) in the future can
improve our understanding of marine stratus. Currently NOAA-ESRL is developing a high sensitivity 94 GHz cloud Doppler radar that is motion compensated for ship based observations. Use of this radar in a large field experiment like VOCALS 2007 can lead to generation of a new dataset and better understanding of the stratus MABL and the meteorology of the Southeast Pacific region in general.

Other major future work might be the application of the Frisch DSD retrieval technique and the evaporation model along with the algorithm described in the literature, to the datasets collected earlier in that region; EPIC 2001, PACS 2003, Stratus 2005. This will lead to not only a better understanding of the southeast Pacific region but also will help in testing the models consistency.
BIBLIOGRAPHY


