Aerosol-Cloud-Precipitation Interactions in the Trade Wind Boundary Layer

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UNIVERSITY OF MIAMI

AEROSOL-CLOUD-PRECIPITATION INTERACTIONS IN THE TRADE WIND BOUNDARY LAYER

By

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AEROSOL-CLOUD-PRECIPITATION INTERACTIONS IN THE TRADE WIND BOUNDARY LAYER

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This dissertation includes an overview of aerosol, cloud, and precipitation properties associated with shallow marine cumulus clouds observed during the Barbados Aerosol Cloud Experiment (BACEX, March-April 2010) and a discussion of their interactions. The principal observing platform for the experiment was the Cooperative Institute for Remotely Piloted Aircraft Studies (CIRPAS) Twin Otter (TO) research aircraft that was equipped with aerosol, cloud, and precipitation probes, standard meteorological instruments, and a up-looking cloud radar.

The temporal variations and vertical distributions of aerosols observed on the 15 flights show a wide range of aerosol conditions that include the most intense African dust event observed at the Barbados surface site during all of 2010. An average CCN varied from 50 cm\(^{-3}\) to 800 cm\(^{-3}\) at super-saturation of 0.6 %, for example. The 10-day backward trajectories show that three distinctive air masses (originality of air mass as well as the vertical structure) dominate over the Eastern Caribbean (e.g., typical maritime air mass, Saharan Air Layer (SAL), Middle latitude dry air) with characteristic aerosol vertical structures.

Many clouds in various phases of growth during BACEX are sampled. The maximum cloud depth observed is about less than 3 km and in most of the clouds is less
than 1 km. Two types of precipitation features were observed for the shallow marine cumulus clouds with different impacts on boundary layer. In one, precipitation shafts are observed to emanate from the cloud base with evaporation in the sub-cloud layer (stabilize the sub-cloud layer). In the other, precipitation shafts emanate mainly near the cloud top on the downshear side of the cloud and evaporate in the cloud layer, leading to destabilizing the cloud layer and providing moisture to the layer. Only 42-44% of clouds sampled were purely non-precipitating throughout the clouds; the remainder of the clouds showed precipitation somewhere in the cloud, predominantly closer to the cloud top.

The relationship between aerosol (Na), cloud droplets (Nd), and precipitation rates (R) is addressed to explore aerosol-cloud-precipitation interactions. A robust increase in Nd with increase in aerosol concentrations is documented. Further, a strong linear relation between sub-cloud CCN and cloud-base Nd is observed in updrafts. The sensitivity of Nd to changes in vertical velocity perturbations w′ (i.e., dlnNd/dlnw′), is greater in the regimes of high aerosol concentrations, suggesting a slight increase in updrafts (or w′) in polluted conditions can lead to greater increases in Nd. Suppression of precipitation with aerosol is a common feature during BACEX. To quantify this decrease of precipitation toward higher aerosol concentration, the sensitivity of precipitation to changes in aerosol (i.e., precipitation susceptibility S₀) is examined. S₀ exhibits three regimes and peaks at intermediate range of cloud thickness. Further, the removal of Nd, due to the rain (wet scavenging), makes susceptibility stronger overall.

In addition to the aerosol feeding clouds from the sub-cloud layer, small cumuli can alter the aerosol properties of their immediate environment through cloud and precipitation processes. In the warm cumuli studied, the depletion of aerosols near cloud
field (so-called cloud halos/shell regimes) are notable, and the reduction of aerosols is more significant in precipitating clouds compared with non-and/or light-precipitating clouds.

The modification of boundary layer aerosol by cloud processes is also explored. The comparisons of the thermodynamic structures observed over Africa with those at Barbados indicate that layers below the SAL are moistened by surface fluxes and convective processes as the air masses are advected across the Atlantic over 7-10 days.
This dissertation is dedicated to my parents and my advisor.
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1. General

Many people contributed to the completion of this work and I would like to acknowledge their efforts and support.

First and foremost, I would like to thank my advisor, Dr. Bruce Albrecht. This work would not have been possible without him. His endless patience, generosity, kind guidance and encouragement were fundamental to the completion of this work. He was not only a great advisor but also a father figure to me. I was fortunate to participate in many field projects. It was a great opportunity to get to know scientists not only from my discipline, but also from other fields. I am also thankful for his encouragement to pursue all of my research interests.

I am deeply grateful to my committee members for their unconditional support, ability to stimulate students, constant encouragement, patience and in-depth knowledge. They are the best committee members a student could ever have. They are amazing mentors, human beings and role models. This work was strengthened by their diverse insight and professional expertise.

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Thanks are due to Dr. Judd Welton and staff, and Dr. Joseph M. Prospero and staff for establishing and maintaining the Ragged Point MPLNET, AERONET sites used in this investigation. Figure 5.7 is obtained from MPLNET (http://mplnet.gsfc.nasa.gov/). Figure 5.2 is downloaded from the MODIS satellite web site (http://modis-atmos.gsfc.nasa.gov/IMAGES/index.html). Sounding data in Fig. 5.3 were obtained from the University of Wyoming’s online Upper Air Data (http://weather.uwyo.edu/upperair/sounding.html).

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# TABLE OF CONTENTS

List of Figures ............................................................................................................. xi

List of Tables ............................................................................................................. xxii

List of Equations ......................................................................................................... xxiii

Chapter 1: Introduction .............................................................................................. 1
  1.1 Challenges in our current understanding of the aerosol-cloud-
  precipitation interactions over trade cumuli regime ............................ 1
    1.1.1 Trade cumuli ................................................................. 1
    1.1.2 Aerosol-cloud-precipitation interactions ............................. 2
    1.1.3 Aerosols over the trade cumuli regime in tropical Atlantic Ocean
    .............................................................................................................. 4
  1.2 Scientific objectives ..................................................................................... 5

Chapter 2: Data and Observing Techniques ............................................................. 9
  2.1 Experiment description ........................................................................... 9
  2.2 Data and instruments ......................................................................... 11
    2.2.1 Aircraft data ............................................................................. 11
    2.2.2 95-GHz FMCW Doppler airborne radar .................................. 13
    2.2.3 Chaff tracer .............................................................................. 14
    2.2.4 Ragged Point aerosol measurements ......................................... 14
    2.2.5 Sounding data and back trajectories ......................................... 15
    2.2.6 AERONET Sun photometer and MPLNET .............................. 16
  2.3 Observing techniques-Calibration of radar reflectivity ......................... 16
  2.4 Observing techniques-Mie technique .................................................. 22
    2.4.1 Background .............................................................................. 22
    2.4.2 Methodology ............................................................................ 24
      2.4.2.1 Mie scattering ....................................................................... 24
      2.4.2.2 Platform motion correction and vertical air motion retrieval 
      .............................................................................................................. 25
    2.4.3 Validation of the technique ...................................................... 30
    2.4.4 Summary .................................................................................. 39

Chapter 3: Aerosol Variation and Cloud Properties over Caribbean Sea ............ 41
  3.1 Motivation .............................................................................................. 41
  3.2 Cloud and aerosol size and distribution variables ............................... 42
    3.2.1 Cloud droplet size and distribution ......................................... 42
    3.2.2 Aerosol particle size and distribution ....................................... 42
    3.2.3 Rainfall (Precipitation) rate ....................................................... 43
  3.3 Large-scale conditions ........................................................................... 43
  3.4 Back trajectories ............................................................................... 47
  3.5 Aerosol properties ............................................................................. 48
Chapter 3: Aerosol-Cloud-Precipitation Interaction

3.5 Aerosol-Cloud-Precipitation Interaction

3.5.1 Ragged Point aerosol measurement .............................................................. 48
3.5.2 Vertical and temporal variation ................................................................... 50
3.5.3 Aerosol particle size distribution ................................................................. 52
3.5.4 Cloud and precipitation properties ............................................................... 56
3.5.5 Representative of thermodynamic structures ........................................... 70
3.5.6 Discussion and summary ............................................................................ 74

Chapter 4: Aerosol-Cloud-Precipitation Interaction

4.1 Aerosol-cloud interaction .............................................................................. 78
4.1.1 Aerosol effects on cloud properties ............................................................ 78
4.1.2 Aerosol effects on precipitation ................................................................. 89
4.2 Aerosol-cloud interactions: Entrainment and flows in and around small
    cumulus clouds ................................................................................................. 94
4.2.1 Background ............................................................................................... 94
4.2.2 Chaff experiment ....................................................................................... 96
4.2.3 Cloud thermodynamics and dynamics ..................................................... 98
4.2.4 Radar analysis-Chaff returns ................................................................. 102
4.2.5 Discussion and summary ......................................................................... 106
4.3 Modification of aerosols near cloud field (cloud effects on aerosols)
    ......................................................................................................................... 109
4.4 Summary ....................................................................................................... 112

Chapter 5: Vertical Structure of Aerosols, Temperature and Moisture Associated with an
Intense African Dust Event Observed over the Eastern Caribbean

5.1 Motivation and background ......................................................................... 116
5.2 Between source (Africa) and site (Barbados) ............................................. 118
5.3 Observations ................................................................................................. 123
    5.3.1 Flight paths .............................................................................................. 123
    5.3.2 Thermodynamic and aerosol profiles ................................................... 125
    5.3.3 Vertical profiles of various aerosol characterizations .............................. 129
    5.3.4 MPL Aerosol returns ............................................................................ 132
5.4 Discussion of transports and processes ......................................................... 133
    5.4.1 Mixing diagram ..................................................................................... 133
    5.4.2 Size and composition ........................................................................... 137
    5.4.3 Changes in particle size distribution .................................................... 146
5.5 Discussion and summary ............................................................................. 148

Chapter 6: Summary, Conclusions and Future Work

6.1 Summary ....................................................................................................... 152
6.2 Implication to model parameterization ......................................................... 164
    6.2.1 First-indirect effect ............................................................................... 164
    6.2.2 Second-indirect effect .......................................................................... 165
6.3 Future work ................................................................................................. 166

Appendix A ............................................................................................................. 170
<table>
<thead>
<tr>
<th>Section</th>
<th>Page</th>
</tr>
</thead>
<tbody>
<tr>
<td>Appendix B</td>
<td>171</td>
</tr>
<tr>
<td>Appendix C</td>
<td>173</td>
</tr>
<tr>
<td>References</td>
<td>175</td>
</tr>
</tbody>
</table>
LIST OF FIGURES

Figure 2.1: Location of data collection. Flight domain is shown as square (2° in longitude; 1.5° in latitude) in left panel. Images from maps.google.com. Courtesy by Bruce A. Albrecht. ................................................................. 9

Figure 2.2: (a) Cloud, aerosol, precipitation probe sensors attached beneath the right wing of the CIRPAS Twin Otter research aircraft. (b) Cloud radar is mounted on top of the radar. (c) Chaff fibers used in this study for 95-GHz radar (unit of ruler in cm), (d) chaff dispenser attached beneath the left wing of the CIRPAS TO research aircraft. ........... 10

Figure 2.3: (a) Time series of range corrected power with height, and (b) that measured from the lowest level of radar (black) and the 6th moments of drop size distribution estimated from probe (magenta) for precipitating cloud sampled on 22 March 2010. 18

Figure 2.4: Scatter diagrams of (a) range corrected power measured from the lowest level of cloud radar (y-axis) and radar reflectivity (the 6th moments of drop size distribution) estimated from probe (x-axis) for the same cloud shown in Fig. 2.3. The difference between range corrected radar power and probe reflectivity is shown as numerical number in Fig. 2.4(b). The difference here indicates term D in Eq. (2.3). Radar reflectivity, after subtracting the difference between them, is shown in x-axis in Fig. 2.4(b). The dashed blue line in Fig. 2.4(b) shows the one to one line. ...................... 19

Figure 2.5: (a,c) Time series of range corrected power measured from the lowest level of radar (black) and the 6th moments of drop size distribution estimated from probe (magenta) for lightly precipitating cloud sampled on (a) 23 March, and (c) 30 March, 2010. (b,d) Scatter diagrams of radar reflectivity (y-axis) and probe reflectivity (x-axis). The difference between range corrected radar power and probe reflectivity is shown as numerical number in Fig. 2.5(b,d). The blue dashed line shows the one to one line. .......................... 20

Figure 2.6: Same as Fig. 2.5 except for non-precipitating cloud sampled on 7 April 2010. Probe radar reflectivity is estimated from CAS forward scattering data. ................. 20

Figure 2.7: An example of time series of radar reflectivity with height for precipitating cloud sampled on 22 March 2010. ................................................................. 21

Figure 2.8: Example of radar Doppler spectrum observed at a certain range (e.g., 156 m) from the aircraft. (a) The raw recorded radar Doppler spectrum with the observed location of the first Mie minimum peak $V_{D,1}$ indicated by the vertical dashed line and the expected location of the first Mie minimum peak $V_{f,1}$ in the absence of platform motion and air motion indicated by the vertical solid line. (b) The raw radar Doppler spectrum shifted to zero contribution from the platform and the air motion. The observed locations
of the first and second Mie maxima are denoted as upward arrows (right to left) in Fig. 2.8b.

Figure 2.9: A time-height cross-section of (a) radar reflectivity and (b) motion corrected Doppler velocity (+: upward) in the precipitating cloud on 5 April 2010 during BACEX from 16:16:48 to 16:18:36 (UTC). The reported height is Above the airborne Radar Level (ARL). Zero height corresponds to 768 m above sea level. The dashed line indicates a specific time for the Doppler spectra in Fig. 2.10.

Figure 2.10: Doppler power spectra observed on 16:17:45 UTC 5 April 2010, which is denoted as a dashed line in Fig. 2.9. Vertical profiles of (a) uncorrected and (b) corrected for platform motion Doppler spectra. The vertical lines in (a) and (b) correspond to the $V_{f1}$ and $V_{D,p}$. (c) Vertical profiles of peaks and valleys of motion corrected Doppler spectra, corresponding to Fig. 2.10b. Red (magenta) circles and blue crosses indicate 1$^{st}$ (2$^{nd}$) maximum and 1$^{st}$ minimum location in the Doppler spectra due to the Mie oscillation. Blue asterisks represent peaks due to the cloud droplets. An example of radar Doppler spectrum observed at 156 m is shown in Fig. 2.8.

Figure 2.11: Vertical air motions with height, retrieved from the Mie technique (circles) and obtained from the Doppler velocity of cloud droplets (air tracer technique, asterisks) on 16:17:45 UTC (hh:mm:ss) on 5 April 2010, denoted as dashed line in Fig. 2.9.

Figure 2.12: A time-height cross-section of (a) radar reflectivity and (b) motion corrected Doppler velocity (+: upward) above the aircraft/radar level (ARL) in the precipitating cloud on 5 April 2010 during BACEX from 16:13:48 to 16:15:36 UTC (a time duration of 108 seconds). Dotted lines denote a specific time periods when larger particles are observed in the Doppler spectra. (A)-(E) are used in Fig. 2.13.

Figure 2.13: Time-height cross-sections of Doppler power spectra for the cloud in Fig. 2.12 (for the 30 seconds period from 16:14:24 to 16:14:53 UTC, 1-second interval; approximately from A to E in Fig. 2.12) on 5 April, 2010. The vertical dashed line indicates the location of zero velocity. Details of (A)-(E) are described in the text.

Figure 2.14: (a) Time-height cross-section of vertical air motions and (b) vertical velocity measured from the aircraft motion sensing and inertial navigation system (INS) at a cloud penetration height (~770 m). The vertical air motions in Fig. 2.14a are retrieved from the Mie technique for the periods between dashed lines, which is enclosed by dashed line in Fig. 2.12b on 5 April 2010 and from the cloud droplets’ spectral peak for the rest of the periods.

Figure 3.1: Time-height cross-section of (a) relative humidity (%) (b) potential temperature (K) (c) wind speed (m s$^{-1}$) and (d) wind direction (degree) obtained from soundings launched at Grantley Adams airport in Barbados at 12:00 UTC from 14 March
to 16 April, 2010. The first (19 March) to the last (11 April) flights are denoted as vertical black solid lines. Periods of strong dust events (31 March-5 April) are also denoted as vertical dashed lines. Lifting Condensation Level (LCL) and 0 °C isotherm are laid in Fig. 3.1(a) and (b) as black lines connected with circle symbols. The primary and secondary inversion heights are shown as square and cross symbols, respectively. Sounding data were obtained from the University of Wyoming’s online Upper air Air Data (http://weather.uwyo.edu/upperair/sounding.html).

Figure 3.2: Profiles of potential temperature, Θ (left), water vapor mixing ratio (middle) and aerosol concentrations obtained from PCASP (cm⁻³; right) during aircraft’s ascents and/or descents. The mean profiles of each variable are denoted as black dots. Color bar indicates the number of research flights (RF #), shown in Table 2.1. The sounding shown here is arbitrary and is denoted in Table 2.1 inside parenthesis.

Figure 3.3: The 10-day back trajectories arriving at 500-m in the middle of the BACEX flight domain. Dates for each back-trajectory are shown accordingly.

Figure 3.4: Dust concentrations recorded at the Barbados Ragged Point surface site (13.2 °N, 59.5 °W) during 2010. Dust surface concentrations (red) over the period of BACEX (3/19-4/11) are shown in lower panel along with sea salt surface concentrations (blue) in the left axis, level 2 Aerosol Optical Depth (AOD) at 500 nm wavelength (magenta) and Angstrom exponent at wavelengths ranging from 440 to 675 nm (grey) from AERONET in the right axis. Dust and sea salt surface concentration data are provided by Dr. Joseph M. Prospero of the University of Miami.

Figure 3.5: (a) Temporal variation of aerosols at sub-cloud layer and (b) vertical distribution of aerosol obtained from aircraft. CCN (s=0.3 %) is overlaid on 23 March for vertical profiles since no PCASP is available on the day. Sub-cloud aerosol concentrations (Fig. 3.5a) are obtained from the level leg run at sub-cloud layer, and vertical profiles of PCASP aerosol concentrations (Fig. 3.5b) were obtained during aircraft’s ascents and/or descents.

Figure 3.6: Daily averaged aerosol particle size distributions (PSDs) ranging from 0.1 µm to 2.5 µm obtained from the PCASP. Color bar indicates the research flight number (RF #), shown in Table 2.1. PSDs from the odd (even) RF numbers are shown as solid (dashed) lines. PSDs estimated between RF07 and RF10 (3/29, 3/30, 3/31, 4/1) are denoted as bold lines. PSD of RF01 (19 March) is not shown due to instrument malfunction on that day.

Figure 3.7: Daily averaged aerosol particle size distributions (PSDs) for the sub-cloud level flights. PSDs obtained from PCASP and CAS probes are combined to attain PSDs ranging from 0.1 µm to 30 µm. Color bar indicates the research flight number (RF #), shown in Table 2.1. PSDs from the odd (even) number of RF are shown as solid (dashed)
lines. PSDs between RF07 (29 March) and RF10 (1 April) are denoted as bold lines. PSD of RF01 (19 March) and RF03 (23 March) are not shown due to the instrument malfunction (RF01) and the absence of PCASP data (RF03) for the days. The scale of Fig. 3.6 is shown as a box (dotted) in Fig. 3.7 in upper-left corner.

Figure 3.8: Time-height cross section of reflectivity on (a) 22 March, (b) 24 March, (c) 29 March and (d) 11 April, 2010 from the cloud-base level flight during 5-minute periods (about 18 km in horizontal extent) at an air speed of about 60 m s\(^{-1}\). Data were sampled from (a-b) precipitating and (c-d) non-precipitating clouds.

Figure 3.9: MODIS satellite images on (a) 22 March, (b) 24 March, (c) 29 March and (d) 30 March, 2010 over the ocean near Barbados. The flight domains are shown as red dotted boxes. The outer box indicates an average flight domain during BACEX. The flight domain of the particular day is overlaid as an inner box, if satellite image is obtained during the flight periods. The numerical number shown at the lower-right side of the figure indicates Julian day with UTC (e.g., 088.1730 indicates Julian day 088, 1730 UTC). Images were downloaded from the MODIS website (http://modis-atmos.gsfc.nasa.gov/IMAGES/index.html).

Figure 3.10: Normalized velocity-reflectivity number frequency distribution on each day during BACEX. Intervals of 2-dBz, and 0.1 ms\(^{-1}\) are used to obtain the frequency distribution. Color bar is shown in upper right corner. Reflectivity of -20 dBz and Doppler velocity of 0 m s\(^{-1}\) are denoted by dotted line. No clouds are observed on 1-2 April during the cloud-base level flights. Time and periods of each cloud-base flights are listed in Table 3.1.

Figure 3.11: Reflectivity and velocity distributions estimated from an average of all individual days (12 cases in Fig. 3.10), (b) using three precipitating clouds (clouds sampled on 3/22, 3/24, and 3/30), (c) using 11 days except for clouds on 22 March, which sampled the strongest precipitating clouds, and (d) from non-precipitating and/or lightly precipitating clouds (remaining 9 days in Fig. 3.10).

Figure 3.12: Frequency distribution of reflectivity and velocity with heights, by composite (a-b) all individual days (12 cases in Fig. 3.10), and (c-d) 9 days excluding three major precipitating clouds events sampled on 22, 24, and 30 March, 2010. Intervals of 30-m (vertical), 2-dBz, and 0.1 ms\(^{-1}\) are used to obtain the frequency distribution.

Figure 3.13: Normalized number of samples (divided by the number of data sampled) with height for all sampled clouds (gray) and for precipitating clouds (black). Precipitating clouds are defined as data points with Z > -20 dBz and vertical velocity < 0 m s\(^{-1}\). No precipitating clouds are observed on 25, 26, 29, 31 March and 11 April.
Figure 3.14: The individual cloud is identified by threshold of Gerber Liquid water content greater than 0.02 g m\(^{-3}\) and duration of data recording (lasts longer than 3s). Precipitating cloud is classified by thresholds of precipitation liquid water content (CIP volume \(\times\) density of water) great than 0.1 g m\(^{-3}\).

Figure 3.15: (a) Percentage of precipitating clouds estimated from all clouds sampled. Precipitating clouds are defined as clouds with precipitation liquid water content (PLWC) larger than 0.1 g m\(^{-3}\) (light-grey solid) and 1 g m\(^{-3}\) (medium- grey dashed). The CIP probe volume concentration (cm\(^3\) m\(^{-3}\)) is multiplied by density of water to obtain PLWC. (b) Percentage of precipitating clouds estimated from all clouds sampled (grey; column-averaged), and from clouds sampled during the cloud-base flights (black; cloud-base) with a threshold of PLWC > 0.1 g m\(^{-3}\). (c) Flight-averaged (grey) and cloud-base (black) precipitation rate (mm day\(^{-1}\)).

Figure 3.16: (a) Cloud water and (b) droplet number concentration \(N_d\) in cloud core (\(w > 1\) m s\(^{-1}\)) sampled by the Twin Otter during BACEX. Non-precipitating samples (CIP volume < 0.01) are used to estimate \(N_d\) in Fig. 3.16b Mean (minimum and maximum) values of LCL are denoted by denoted as dashed (dotted) lines.

Figure 3.17: Profiles of (a) potential temperature, \(\Theta\), water vapor mixing ratio obtained from BACEX (black) with \(\pm 1\sigma\) (grey), BOMEX (blue), RICO (red) and ATEX (magenta) field campaigns. Data of BOMEX, RICO, and ATEX are obtained from GCSS (GEWEX Cloud System Study) boundary layer cloud homepage. BACEX profiles are obtained from all data sampled during the experiment.

Figure 4.1: Daily averaged (a) cloud tops (CT, black solid line connected with filled circles), bases (CB, gray solid line connected with filled circles) and thickness H (dashed line connected with open circles) obtained from radar measurements during cloud-base flights; and (b) aerosol concentrations obtained from PCASP during sub-cloud flights with research flight number (RF #). RF # is listed in Table 3.1.

Figure 4.2: Profiles of aerosols (PCASP, cm\(^{-3}\), temperatures (T, °C), potential temperatures (\(\Theta, K\)), and mixing ratios (g kg\(^{-1}\)) for pristine (black, 3/29) and dusty (red, 4/1; grey, 4/2) days. Soundings are obtained during the 2\(^{nd}\) ascents of aircrafts on the day for 29 March and 1 April and during the first ascent of an aircraft for 2 April.

Figure 4.3: Profiles of (a) potential temperatures (\(\Theta, K\)) and (b) aerosols (PCASP, cm\(^{-3}\)) for three days (3/30, 3/31, and 4/5) that have similar aerosol concentrations in the sub-cloud layer, and for 25 March that has a thick aerosol layer above inversion, and for 29-30 March that have typical aerosol profiles in marine boundary layer. Cloud depths (from base to top) are denoted as vertical bars accordingly. Soundings are obtained during the 2\(^{nd}\) ascents of aircrafts on the day. Cloud bases and tops are obtained from the cloud-radar measurements during cloud-base flights on the days.
Figure 4.4: Relations between aerosol number concentrations, \( N_a \) (here CCN \( s=0.3 \% \)) and (a, c ,e) cloud number concentrations, \( N_d \) and (b, d, f) mean effective size of cloud droplets, \( D_e \): (a) flight averaged \( N_a \) vs. flight averaged \( N_d \), (b) flight-averaged \( N_a \) vs. flight-averaged \( D_e \); (c) sub-cloud \( N_a \) and flight-averaged \( N_d \), (d) sub-cloud \( N_a \) and flight-averaged \( D_e \), (e) sub-cloud \( N_a \) vs. cloud-base \( N_d \), (f) sub-cloud \( N_a \) and cloud-base \( D_e \). Colors in Fig. 4.4 (b, d, and f) indicate the daily mean cloud thickness, estimated from cloud-base flights. ..............................................

Figure 4.5: CCN in sub-cloud layer and \( N_d \) just above cloud base for different thresholds of updraft: (a) \( w' > 0 \), (b) \( w' > 0.5 \), and (c) \( w' > 1 \text{ m s}^{-1} \). ..............................................

Figure 4.6: The \( N_d \) versus \( w' \) just above cloud base for (a) polluted (5 April) and (b) pristine (10 April) environments. Green dots in Fig. 4.6a and Fig. 4.6b indicate all the data points for the day. Cloud-base \( N_d \) is averaged in 20 \( \text{cm}^3 \)-intervals and shown as black dots. The mean values of sub-cloud CCN (0.3 % and 0.6 %) are shown as horizontal lines. The mean profiles of \( N_d \) vs. \( w' \) for two pristine (29 March, 10 April) and dusty (31 March and 5 April) cases are shown in Fig. 4.6c with slopes \( (dN_d/dw') \) estimated for the regime of \( w' > 0 \). ..............................................

Figure 4.7: (a) The distribution of \( k \) during BACEX. (b) Same as Fig. 4.6(c) except for log-value. The slopes are shown as numerical numbers in the figure. ..................

Figure 4.8: Scatter diagrams of aerosol number concentrations (cm\(^{-3}\)) and rainfall rates (mm hr\(^{-1}\)) in and near the clouds during BACEX; (a) Flight-averaged CCN and rainfall rate and (b) sub-cloud CCN versus flight-averaged rainfall rate (mm hr\(^{-1}\)). Rainfall rate is calculated by Eq. (3.2) with CIP probe data. Colors indicate cloud thickness. Note that no CCN data was available on 11 April. ..............................................

Figure 4.9: Precipitation susceptibility \( S_0 \) as a function of cloud thickness, \( H \), estimated from cloud-base radar measurements (time resolution of 1-sec, height resolution of 24-m used). All clouds (12 cases; dashed) sampled during the BACEX and data excluding two major precipitating clouds (22 March and 30 March, solid) in Fig. 3.10, are shown to examine the scavenging effect. ..............................................

Figure 4.10: Cloud thickness distributions during BACEX. Cloud thicknesses are estimated from the cloud radar (1-s and 24-m resolutions) during the cloud-base level flights. Colors indicate rainfall rates (mm day\(^{-1}\)). Dates are shown adjacent to dashed lines. ..............................................

Figure 4.11: (a) A flight pattern and (b) corresponding altitude during the chaff experiment. Colors indicate time elapse (earlier time is denoted as a darker color). The spatial locations where the chaff is released are shown as cross symbols in Fig. 4.11a (black-upwind, gray-cross wind) and the target cloud is located in the middle of the cross
symbols. The duration of the cloud penetration is denoted as vertical dotted lines in Fig. 4.11b. Flight paths, after LEG 3, are not shown in Fig. 4.11a.

Figure 4.12: Photos taken towards the west, approximately a) 80 s, b) 20 s and c) 10 s prior to the first cloud penetration (LEG 1). The airplane penetrated the cloud from east to west direction during LEG1 and the background wind is into the direction of photo (easterly).

Figure 4.13: (a, e) Time series of vertical velocity ($w$, m s$^{-1}$; black), virtual temperature ($T_v$, °C; magenta) and liquid water content (LWC) multiplied by 10 (g m$^{-3}$, blue) for the cloud penetration on LEG 1 and LEG 2. Time-height cross sections of (b, f) reflectivity, (c, g) mean Doppler velocity and (d, h) spectrum width of Doppler spectrum, for the chaffed cloud on LEG 1 (left) and LEG 2 (right). The cloud penetrations were made at constant heights (along the wind-LEG 1, cross the wind-LEG 2) and the average penetration heights are shown with level legs. The aircraft observations were made at an air speed of about 60 m s$^{-1}$ and thus cloud widths are approximately 120 m and 480 m for LEG 1 and LEG 2. Note that reversed time axes are used on LEG 1 to match the penetration direction with other LEGs. Vertical dashed lines in Fig. 4.13f are used to show Doppler spectra in Fig. 4.15. East on the right (LEG 1); Northeast on the right (LEG 2); Upshear side on the right (LEG 1).

Figure 4.14: Same as Fig. 4.13 except for LEG 3 (left) and LEG 4 (right). 1-Hz resolution of LWC added by an average cloud penetration height, obtained from Gerber sensor, is overlaid in the reflectivity field. East and upshear side on the right (LEG 3, LEG 4).

Figure 4.15: Time-height cross-sections of Doppler power spectra for the chaffed cloud, through LEGs 2, 3, and 4. Radar reflectivity, Doppler velocities, and spectrum width corresponding to each LEG are shown in Fig. 4.13 and Fig. 4.14 with vertical dashed lines accordingly. Details of (A)-(F) for each leg are given in the context. Time resolution of Doppler spectra is 3-Hz. Note that C1, C2, C3 on LEG 3 are denoted as one vertical dashed line in Fig. 4.14 since, time resolution of radar moments shown in Fig. 4.14 is 1-Hz. Same as in A1 and A2 on LEG 4.

Figure 4.16: Time series of virtual potential temperature (red) and liquid water contents (black) for LEG 1 and LEG 3. The environmental virtual potential temperatures obtained from three soundings during the flights are overlaid (horizontal dashed lines). Note that reversed time axis is used on LEG 1.

Figure 4.17: Schematic of entrainment process and incloud flow pattern for the shallow marine cumulus cloud, overlaid in the radar reflectivity field observed on Leg 3. Cloud boundaries are shown as blue solid lines. Chaff movements and in-cloud flow patterns are denoted as black solid/dashed lines, and details of the numerical numbers are described in the text. Background winds are denoted as arrows to show the upshear/downshear
direction. 'E' indicates east, while 'W' indicates west direction. A 1-Hz resolution liquid water contents (LWC) are added as black solid line. Detrained cloud elements are seen above the main cloud as two turrets toward downshear side of the cloud.

Figure 4.18: Time series of aerosols (CCN 0.3% sampled at sub-cloud layer; red), LWC (green) and vertical velocity (w, black) sampled during cloud-base level flight for the cloud in Fig. 2.9 (Mie cloud in the text).

Figure 4.19: Same as Fig. 4.18 except data were sampled during LEGs 1-3 for the cloud in Fig. 4.12 (Chaff cloud).

Figure 4.20: Time Series of sub-cloud aerosols (CCN, 0.6%; black), cloud droplets ($N_d$) and drizzle number concentrations on 24 March 2010, showing depletion of CCNs where drizzle (shown as magenta) and/or cloud (shown as cyan dots) present. Aerosols, cloud droplets and drizzle number concentrations are measured by CCN counter, CAS and CIP probes, respectively.

Figure 5.1: The 13-day back trajectory arriving at 500 m in the middle of the BACEX flight domain for 17 UTC 1 April (magenta) and 2 April (cyan) 2010. Trajectories of each day at 00 UTC are denoted as cross symbols and several dates of back-trajectory are shown accordingly. Locations of sounding stations at Africa are added as cross.

Figure 5.2: MODIS satellite images on (a) 29 March, 2010 over the Atlantic and on (b) 22 March, 2010 over western Africa. Images were downloaded from the MODIS website (http://modis-atmos.gsfc.nasa.gov/IMAGES/index.html).

Figure 5.3: Profiles of (a) potential temperature and (b) mixing ratio on 21 March 09:00 UTC at GOTT, 22 March 12:00 UTC at GOOY, in Africa, and on 1-2 April 12:00 UTC at Barbados. Profiles of wind direction on 21 March 09:00 UTC at GOTT, 22 March 12:00 UTC at GOOY in Africa are shown in Fig. 5.3(c). SAL observed at Barbados is denoted as arrows.

Figure 5.4: (a, c) Flight paths, (b, d) time series of flight altitudes on (a, b) 1 April and (c, d) 2 April, 2010. Numerical numbers and colors are assigned to the individual level runs accordingly. Dashed lines in Fig. 5.4(a) and Fig. 5.4(c) represent the first (up1) and second (up2) ascent soundings while solid lines indicate data from level runs (leg 1 through leg 7). The starting point of each leg is denoted as cross symbol in Fig. 5.4(a) and Fig. 5.4(c). The Barbados Ragged Point surface site is located at 13.2 °N, 59.5 °W.

Figure 5.5: Profiles of (left) potential temperature $\theta$, (middle) water vapor mixing ratio $r$, and (right) PCASP aerosol concentrations from aircraft during ascents. Data (10 Hz) from each level run are denoted by black bold colors and numbered accordingly. Thin black dots represent data obtained during descents between two consecutive legs. For 1
April 2010 (upper panel); Leg 1: top of the haze layer, Leg 2: middle of Saharan air layer (SAL), Leg 3-Leg 5: intermediate layer (IL), Leg 6-7: sub-cloud layer (SCL). For 2 April, 2010 (lower panel); Leg 1: middle of Saharan air layer (SAL), Leg 2-Leg4: intermediate layer (IL), leg 5-7: sub-cloud layer (SCL). The layers of Saharan air, intermediate and sub-cloud are labeled in Fig. 5.5(a) on the first sounding from 1 April to illustrate the layers.

Figure 5.6: Profiles of various aerosols from various probes for (a, c) first (up1) and (b, d) second (up2) sounding on (a-b) 1 April and (c-d) 2 April 2010. CN (green), PCASP (black), CCN (0.3 %; blue) and CCN (0.6 %; red) are shown. Average heights of individual levels (~10 minute) runs are denoted as thick gray horizontal lines (top down: leg 1 to leg 7). CCN (0.3 %) indicates cloud condensation nuclei (CCN) activated at super-saturation of 0.3 %. Same notation for CCN (0.6 %). Information of data acquisition time and location is shown in Fig. 5.4.

Figure 5.7: Time-height cross-section of NRB (Normalized Relative Backscatter) obtained from Micro-Pulse LIDAR (MPL) at Ragged point in Barbados during 1-2 April 2010. Figures were obtained from MPLNET (http://mplnet.gsfc.nasa.gov/). The LIDAR is blocked from the sun for a 2-hour period centered on around local noon and indicated by the dark areas in this figure. Local time=UTC-5.

Figure 5.8: Mixing diagram of CCN (0.6 %) per mass (µg⁻¹) and mixing ratio (g kg⁻¹) on (a) 1 April and (b) 2 April, 2010. Black squares are obtained from the ascent environmental soundings and colored dots are obtained from each level run. (a): leg 1(top of the SAL), leg 2 (SAL), legs 3-5 (IL), legs 6-7 (SCL), (b): leg 1 (SAL), legs 2-4 (IL), legs 5-7 (SCL).

Figure 5.9: Mixing diagram of aerosol concentrations (µg⁻¹) for CCNs and CN versus mixing ratio (g kg⁻¹) for each level run, for (a-e) 1 April, and for (f-i) 2 April, 2010.

Figure 5.10: Aerosol particle size distributions for (a) 1 April 2010 from top of the SAL to sub-cloud layer and for (b) 2 April 2010 from SAL to sub-cloud layer.

Figure 5.11: Mean profiles of CCNs (blue; CCN 0.3 %, red; CCN 0.6 %) at each level leg for (a) 1 April and (b) 2 April, 2010. The difference of number concentrations between CN and PCASP at each level is also shown (black squares). Horizontal bar indicates ±1σ from the mean value at each leg.

Figure 5.12: Profiles of (a, c) potential temperature (black solid line), CCNs (0.3 %, blue; 0.6 %, red), and (b, d) ultrafine aerosol concentrations (red, 3 < D < 15 nm; blue, 3 < D < 10 nm; and green solid lines, 3 < D < 15 nm) for the second soundings on (a) 1 and (b) 2 April, 2012. Average heights of individual level-runs are denoted as thick gray horizontal lines (top down: leg 1 to leg 7) in Fig. 5.12(a) and Fig. 5.12(c).
Figure 5.13: Aerosol particle size distributions calculated from the layers of A, B and C in Fig. 5.12, from the second soundings of (a) 1 April 2010 and for (b) 2 April 2010. The layer A represents a layer above SAL, layer B indicate the very top of the SAL (Fig. 5.13a) and a transit layer from the very top of the SAL to a layer above the SAL (Fig. 5.13b), and a layer C indicates the middle of the SAL. The locations of three layers are shown in Figure 5.12.

Figure 5.14: Time series of (a) aerosol number concentrations (CCN, 0.6 % per mass) and (b) mixing ratio sampled on leg 3 (intermediate layer, IL) on 1 April for dry (magenta), moist (cyan), and transit (grey dots) intermediate layers (ILs). A mixing diagram of aerosol concentrations (µg⁻¹) versus mixing ratio (g kg⁻¹) for these three air masses is shown in Fig. 5.14(c).

Figure 5.15: Aerosol particle size distributions \( \frac{dN}{d(logD)} \) in upper, \( \frac{dV}{d(logD)} \) in lower panels) on 1 April, 2010 obtained from SAL, three intermediate layers (moist IL, dry IL and transit IL), and sub-cloud layer.

Figure 6.1: Profiles of aerosol number concentrations obtained from PCASP. Left: Typical maritime vertical structure; Middle: vertical structure associated with middle latitude dry air; Right: aerosol vertical structure during African dust events. Color bar indicates the number of research flights (RF #), shown in Table 2.1.

Figure 6.2: A cloud photo that shows two types of precipitation; Precipitation shafts emanate near the cloud top on the downshear side of the cloud; Precipitation shafts emanating near the cloud base. The photo was taken by Bruce Albrecht from the balcony of his office at Rosenstiel School of Marine and Atmospheric Science (RSMAS), Miami. Photo: Courtesy Bruce A. Albrecht.

Figure 6.3: The sensitivity of \( N_d \) to changes in vertical velocity perturbations \( w' \) \( (dlnN_d/dlnw') \) for two dusty (4/5-red and 3/31-magenta), and two pristine (4/10-cyan and 3/29-blue) conditions. The slope shows 0.2 and 0.1 for dusty and pristine conditions, respectively.

Figure 6.4: Precipitation susceptibility, \( S_0 \) with fixed cloud thickness from BACEX (upper panel), and from models and satellite observations (lower panel).

Figure 6.5: (a) Vertical profile of mixing ratio, and (b-c) mixing diagram of aerosol concentrations (PCASP, µg⁻¹) and mixing ratios (g kg⁻¹) obtained during the second ascent sounding on 2 April, 2010. Colors correspond to the heights. Aerosols and mixing ratios, sampled in the dry layer, right below the SAL in Fig. 6.5(a), are overlaid in Fig. 6.5(c) as black cross-symbols. The upper and lower boundary of the dry layer is denoted as horizontal dashed lines in Fig. 6.5(a).
Figure B1: Number of samples with height for all sampled clouds (gray) and for precipitating clouds (black and red). Precipitating clouds are defined as data points with $Z > -20$ dBz and vertical velocity $< 0$ m s$^{-1}$ (black), and are defined as data points with $Z > -20$ dBz (red).
LIST OF TABLES

Table 2.1: Flight list. Dates that passive tracers released are also listed. ....................... 11

Table 2.2: Characteristics of instruments used in Barbados Aerosol Cloud Experiment. .................................................................................................................. 12

Table 2.3: Characteristics of a FMCW radar. ...................................................................... 13

Table 3.1: Cloud-base level leg used in radar analysis for cloud statistics. ....................... 56

Table 5.1: The critical diameters (D_c) and kappa (κ) at each level for 1-2 April, 2010, with values of k from Twomey (1959) and number concentration of CN-PCASP. ...... 145

Table A1: Table of acronyms and symbols used in this study. ........................................ 170
LIST OF EQUATIONS

Equation 1.1. .......................................................................................................................... 3
Equation 2.1. .......................................................................................................................... 17
Equation 2.2. .......................................................................................................................... 17
Equation 2.3. .......................................................................................................................... 17
Equation 2.4. .......................................................................................................................... 25
Equation 2.5. .......................................................................................................................... 25
Equation 2.6. .......................................................................................................................... 26
Equation 2.7. .......................................................................................................................... 28
Equation 2.8. .......................................................................................................................... 29
Equation 3.1. .......................................................................................................................... 42
Equation 3.2. .......................................................................................................................... 43
Equation 3.3. .......................................................................................................................... 71
Equation 3.4. .......................................................................................................................... 71
Equation 3.5. .......................................................................................................................... 72
Equation 3.6. .......................................................................................................................... 72
Equation 4.1. .......................................................................................................................... 88
Equation 4.2. .......................................................................................................................... 88
Equation 4.3. .......................................................................................................................... 88
Equation 4.4. .......................................................................................................................... 91
Equation 4.5. .......................................................................................................................... 94
Equation 5.1. .......................................................................................................................... 122
Chapter 1: Introduction

1.1 Challenges in our current understanding of the aerosol-cloud-precipitation interactions over trade cumuli regime

1.1.1 Trade cumuli

Trade cumulus clouds are relatively less investigated compared with the other major oceanic warm cloud systems-- stratocumulus (Sc)--, because they do not have as significant impact on the radiative fields as Sc clouds do. However, small marine cumulus clouds are frequently observed over the Earth’s oceans and are by far the most common type of cloud in the world (Norris, 1998). Although the fractional cloudiness associated with these cumulus clouds is typically 15 to 25 %, the extensive areas that they cover make their radiative impacts an important factor in the climate system. In addition, the smallest clouds have the largest contribution to the mass flux transport (Nitta, 1975) and cloud fraction (Rodts et al., 2003) from cumulus. Further, they are a part of the feeder system for the deep convection in tropics and are critical to the energy and moisture budget of the trade-wind boundary layer. Besides, these clouds give the largest uncertainty in tropical cloud feedback in climate system (Bony and Dufresne, 2005). The importance of trade-wind cumuli to the climate system has motivated the development and execution of field campaigns over this regime (e.g., ATEX, BOMEX, and RICO). The Atlantic Trade winds EXperiment (ATEX) studied the development of the boundary layer in the trade winds near ITCZ (5-17 °N, 30-45 °W), from 6-21 February, 1969 (e.g., Augstein et al., 1974). The Barbados Oceanographic and Meteorological Experiment (BOMEX) took place east of Barbados (mainly) and regions between 7-18 °N and 52-57 °W during May and July of 1969, and focused on quantifying heat, moisture, and momentum exchanges between the tropical ocean and atmosphere.
The recent Rain in Cumulus over the Ocean (RICO) field campaign took place in the trades over the western Atlantic during November 2004 and January 2005 (8 weeks), and focused on the processes related to the formations of rain in shallow cumuli (Rauber et al., 2007; http://www.eol.ucar.edu/projects/rico/).

1.1.2 Aerosol-cloud-precipitation interactions

Aerosols have a significant impact on the Earth’s energy budget directly by scattering and absorbing solar radiation. They mainly consist of non-absorbing materials which scatter solar radiation back to space, leading to a negative radiative forcing (cooling of the climate system). They also can contain absorbing materials such as black carbon (soot) from the biomass burning (e.g., Ackerman et al., 2000; Koren et al., 2004) and dust from arid regions, leading to positive radiative forcing locally that partly offsets aerosol scattering effects (IPCC, 2001).

Some aerosols serve as cloud condensation nuclei (CCN), and thus lead to an increase in the cloud droplet number concentration \( N_d \) and a decrease in the size of cloud droplets for a fixed liquid water path (LWP) and, in turn, an increase in the scattering to the space yields to increase cloud albedo (Twomey, 1977; first indirect effect; cooling the climate system). The more numerous small size droplets under high aerosol concentrations are difficult to grow into the larger drops, as their fall velocities are similar. Consequently, under these conditions a decrease in drop coalescence efficiency inhibits precipitation, and therefore may increase liquid water path and cloud fraction, and possibly increase cloud lifetime (Albrecht, 1989; second indirect effect).
Cloud-aerosol interactions are considered to be one of the most important, but least known forcing factors in the climate system (IPCC, 2007). There are considerable disagreements regarding not only the magnitude but also direction of aerosol perturbations on cloud properties and associated feedbacks. For instance, although there are agreements on the suppression of precipitation with aerosols in warm boundary layer, considerable disagreements are found in cloud life time effects (Stevens and Feingold, 2009). Jiang et al (2006) showed an unchanged or even decreased cloud lifetime with aerosol concentrations varying from clean to polluted conditions for shallow cumulus clouds. Cloud fraction also shows two counter responses with aerosols: as aerosol increases cloud fraction increases in some studies (e.g. Norris, 2001; Kaufman et al., 2005), but cloud fraction decreased in others (Ackerman et al., 2000; Feingold et al., 2005; Jiang and Feingold, 2006; Xue and Feingold, 2006).

In the climate models, two general approaches have been used to relate changes in cloud droplet number concentrations ($N_d$) and aerosol concentrations ($N_a$). One is to use an empirical expression and/or equation between cloud droplet concentrations and aerosol concentrations (e.g., Boucher and Lohmann 1995) as

$$N_a = A(N_a)^b$$  \hspace{1cm} \text{(Equation 1.1)}

,where $N_a$ is an aerosol mass concentration of a specified aerosol. Mass of SO$_4^{2-}$ is the most commonly used in the model, as the characteristics of SO$_4^{2-}$, such as distribution and emission etc are known better than others. The problem of this approach is that the empirical relationship derived from a certain region may not appropriate for other region and thus, may not be representative globally. For example, empirical equation derived
from highly polluted area such as megacity may not proper to be applied to the biomass region.

The second method to relate changes in \( N_d \) to changes in \( N_a \) is based on a prognostic parameterization of the cloud droplet formation process (e.g., Abdul-Razzak and Ghan 2000). To treat cloud droplet formation accurately, the aerosol number concentration, its chemical composition and the vertical velocity on the cloud scale need to be known. The Abdul-Razzak and Ghan (2000) cloud droplet parameterization leads a stronger forcing in the major biomass burning regions in Africa and South America than does the Boucher and Lohmann (1995) parameterization (Penner et al., 2006).

1.1.3 Aerosols over the trade cumuli regime in tropical Atlantic Ocean

It is common to observe the westward movement of dust-laden air from Africa over the tropical Atlantic Ocean under the influence of the trade winds (Carlson and Prospero, 1972; Dunion, 2011). The individual SAL outbreaks can cover an area of the Atlantic as large as the 48 contiguous United States and can migrate to the eastern coast of Florida (Dunion and Velden, 2004). The large amount of dust in the SAL substantially impacts the radiative balance of the earth-atmosphere system-directly through scattering and absorption (e.g., sea surface temperature decreases during dust outbreak; Lau and Kim, 2007) and indirectly by modifying cloud microphysics acting as effective ice nuclei (e.g., DeMott et al., 2003; Sassen et al., 2003). Further, the dry, dusty SAL may weaken the strength of tropical cyclones (Dunion and Velden, 2004; Evan et al., 2006; Wu, 2007) and suppress convection in the interior of the SAL (Wong and Dessler, 2005). The vertical structure of the absorbing aerosol layer (e.g., dust) also can be important to
aerosol-cloud-precipitation interactions since clouds can be suppressed or enhanced depending on the relative location of the absorbing aerosol layer (Hansen et al., 1997; Feingold et al., 2005; Johnson et al., 2004) and is important to the accurate calculations of radiative forcing (e.g., Ackerman and Chung, 1992).

1.2 Scientific objectives

Although, field campaigns over the trade-cumuli regime have been made (see 1.1.1), none focused specifically on the interactions between aerosol, cloud and precipitations over this regime. To better understand aerosol-cloud-precipitation interactions over this trade cumuli regime, Barbados Aerosol Cloud Experiment (BACEX) took place off the Caribbean island of Barbados, within the northeast trades of the eastern Atlantic, during mid March and April 2010. The marine environment in Barbados provides an excellent area to sample marine clouds in which there is strong propensity for shallow clouds to precipitate. This cloud regime is the most ubiquitous on the planet, and is founded much of the current discrepancies in the representation of cloud feedbacks in model-based climate change estimates. In the trade-winds over Barbados, African dust (so-called Saharan Air Layer) is the dominant aerosol components during certain seasons and is expected to provide further opportunities for cloud modification by aerosols. The main goal of this dissertation is to improve our understanding in aerosol-cloud processes in trade-wind cumuli regime, and thus, to improve and/or to provide a basis for evaluating and improving the parameterization of cloud-aerosol-precipitation interactions in numerical models. In addition, this work is intended to better document the vertical
structure of the Saharan Air Layer (SAL) and provide standards for interpreting and comparing satellite data; further, to characterize the clouds properties in this regime.

This work examines several issues related to aerosol-cloud-precipitation interactions in shallow (cloud depths less than about 2-3 km) and warm (no ice) cumulus clouds. The entrainment process and flows in and around small cumulus clouds are also examined to illustrate the interactions between cloud and environmental air (e.g., aerosols) using chaff tracers. Modification of boundary layer aerosols by cloud process, associated with the intense African dust event, is discussed to address the clouds effects on aerosols.

The airborne cloud radar used in this study is an important tool to provide vertical structures of cloud and precipitation properties. To make full use of this tool, velocity correction and reflectivity calibrations are made and innovative techniques for making full use of the radar are developed. The specific scientific goals and questions to be addressed in this study are:

1) What are the overall characteristics of clouds, precipitation and aerosols over Caribbean Sea? (e.g., cloud tops, bases, reflectivity, vertical velocity, vertical structure of aerosols, so on)

2) How sensitive is cloud droplet concentration to changes in aerosol and changes in vertical velocity perturbation? \( \frac{d\ln N_d}{d\ln Na}, \frac{d\ln N_d}{d\ln w'} \)

3) How sensitive is the precipitation rate to changes in aerosol? (Precipitation susceptibility, \( S_0 \))

4) What is the effect of cloud on aerosol distribution near cloud field?

5) Can aerosol structures observed in the boundary layer be modified by cloud process?
The questions are addressed using observations collected during BACEX (March-April, 2010).

Chapter 2 describes the experiment, instruments and observing techniques (e.g., Mie technique) used for the remainder of the analyses shown in this dissertation.

In Chapter 3, the variations of aerosols, properties of clouds and precipitations, sampled during the experiments are studied. To characterize the clouds properties, such as precipitating or non-precipitating clouds, distributions of cloud tops and bases, reflectivity and Doppler velocity, so on, I heavily rely on radar measurements.

In Chapter 4, I focus on aerosols effects on clouds as well as cloud effects on aerosols in the near cloud field associated with entrainment and detrainment process. I will show that small cumuli can alter the aerosol properties of their immediate environment through cloud and precipitation processes, although the aerosols can affect cloud and precipitation processes and characteristics. The evaporation-mixing process associated with entrainment and detrainment can modify the near-field cloud environment and thus affect the subsequent evaporation and cloud life-cycle. Although the presence of cloud halos (regions of enhanced humidity in the vicinity of cumulus clouds; Radke and Hobbs, 1991; Perry and Hobbs, 1996; Lu et al., 2003; Laird, 2005) and shells surrounding clouds has been seen from aircraft observations and in Large Eddy Simulation (LES), the effects that these structures have on the near-field aerosol characteristics have not been examined. Further, direct observations of the mixing processes involved with entrainment and detrainment process in the vicinity of the cloud boundaries had not been available. Thus, in this chapter, the entrainment/detrainment process and flows in and around small cumulus clouds are also demonstrated directly.
with aids of chaff tracers to show the interactions between clouds and environmental air (i.e., aerosols).

In chapter 5, the clouds effects on aerosols in a larger scale--i.e., modification of boundary layer aerosols by clouds and cloud process--are examined. This is explored by studying the vertical structure of aerosols, temperature and moisture associated with an intense African dust event during the Barbados Aerosol Cloud Experiment. Chapter 6 concludes with summary of the conclusions reached during this work.
Chapter 2: Data and Observing Techniques

This chapter describes the experiment location, duration, instruments used for data collection and observing techniques used for data analysis. Section 2.1 provides details about the experiment location and period, and Section 2.2 describes data and instruments used in this study. The observing techniques are presented in Sections 2.3.

2.1 Experiment description

The Barbados Aerosol Cloud Experiment (BACEX 2010) took place, off the Caribbean island of Barbados during mid March to April, to study cloud, aerosol, and precipitation interactions in addition to the effects of aerosol variability on cloud properties and life-time. Operations were centered at around the domain of the east to northeast side of Barbados Ragged point surface site (13.2 °N, 59.5 °W). The experiment location is shown in Fig. 2.1.

Figure 2.1: Location of data collection. Flight domain is shown as square (2° in longitude; 1.5° in latitude) in left panel. Images from maps.google.com. Courtesy by Bruce A. Albrecht.
The principal observing platform for the experiment was the Collaborative Institute for Remotely-Piloted Aircraft Studies (CIRPAS) Twin Otter (TO) research aircraft that was equipped with aerosol, cloud, and precipitation probes (Fig. 2.2a) and standard meteorological instruments for observing the mean and turbulent thermodynamic and wind structures. To study the vertical structures of updrafts and downdrafts in clouds, 95-GHz cloud radar (designed and fabricated by ProSensing Inc. for CIRPAS) was mounted on top of the aircraft fuselage (Fig. 2.2b; see 2.2.2). In addition that the feasibility of using a passive tracer--radar chaff--was explored on some of the flights during the experiment to study the entrainment and transport processes in small cumuli. The general information for each flight is summarized in Table 2.1.
**Table 2.1:** Flight list. Dates that passive tracers released are also listed.

<table>
<thead>
<tr>
<th>Flight</th>
<th>Date</th>
<th>Time (UTC*)</th>
<th>number of soundings</th>
<th>Note</th>
</tr>
</thead>
<tbody>
<tr>
<td>RF01</td>
<td>19 Mar</td>
<td>15:14 - 16:40</td>
<td>2(2)</td>
<td>Spuriously high CAS $N_d$ on this flight</td>
</tr>
<tr>
<td>RF02</td>
<td>22 Mar</td>
<td>15:01 - 16:28</td>
<td>2(2)</td>
<td>-</td>
</tr>
<tr>
<td>RF03</td>
<td>23 Mar</td>
<td>14:28 - 18:20</td>
<td>4(3)</td>
<td>No PCASP data available, cloud chaff, clear air chaff</td>
</tr>
<tr>
<td>RF04</td>
<td>24 Mar</td>
<td>14:50 - 18:29</td>
<td>4(2)</td>
<td>Cloud chaff, clear air chaff</td>
</tr>
<tr>
<td>RF05</td>
<td>25 Mar</td>
<td>14:50 - 17:52</td>
<td>4(2)</td>
<td>Clear air chaff</td>
</tr>
<tr>
<td>RF06</td>
<td>26 Mar</td>
<td>14:45 - 16:04</td>
<td>2(2)</td>
<td>Cloud chaff</td>
</tr>
<tr>
<td>RF07</td>
<td>29 Mar</td>
<td>15:03 - 19:02</td>
<td>4(2)</td>
<td>Clouds chaff, clear air chaff</td>
</tr>
<tr>
<td>RF08</td>
<td>30 Mar</td>
<td>14:40 - 18:12</td>
<td>6(2)</td>
<td>Strong downdraft from cloud outflow</td>
</tr>
<tr>
<td>RF09</td>
<td>31 Mar</td>
<td>14:40 - 18:10</td>
<td>4(1)</td>
<td></td>
</tr>
<tr>
<td>RF10</td>
<td>1 Apr</td>
<td>15:14 - 18:18</td>
<td>3(2)</td>
<td>Intense African dust</td>
</tr>
<tr>
<td>RF11</td>
<td>2 Apr</td>
<td>14:40 - 17:18</td>
<td>3(2)</td>
<td>Intense African dust, clear air chaff</td>
</tr>
<tr>
<td>RF12</td>
<td>5 Apr</td>
<td>14:48 - 18:29</td>
<td>1(1)</td>
<td>Mie signature, water spout</td>
</tr>
<tr>
<td>RF13</td>
<td>7 Apr</td>
<td>14:33 - 17:17</td>
<td>3(1)</td>
<td>-</td>
</tr>
<tr>
<td>RF14</td>
<td>10 Apr</td>
<td>14:36 - 17:19</td>
<td>3(1)</td>
<td>-</td>
</tr>
<tr>
<td>RF15</td>
<td>11 Apr</td>
<td>14:36 - 18:09</td>
<td>3(1)</td>
<td>No CCN data available</td>
</tr>
</tbody>
</table>

*Local time: UTC-5

* The total number of soundings is included take-off and landing soundings. The number of the $n^{th}$ sounding used in Fig. 3.2 is shown inside parenthesis.

### 2.2 Data and instruments

#### 2.2.1 Aircraft data

The CIRPAS Twin Otter research aircraft made 15 flights upstream from Ragged Point (13.2 °N, 59.5 °W), eastern shore of Barbados where surface aerosol measurements were made, (see 2.2.4: Ragged Point Aerosol measurements), from 19 March to 11 April. Each flight had a duration of 3-4 hours and included at least one sounding excluding take off and landing soundings, and several horizontal legs from near the ocean surface at levels as low as 30 m and to above the trade inversion (~3 km). The aircraft was equipped with aerosol, cloud and precipitation probes (Fig. 2.2a) and standard meteorological instruments. The standard meteorological variables such as temperature, mixing ratio, liquid water content, winds are obtained at a frequency of 10-Hz and 1-Hz from PVM-
The aerosol data included aerosol number concentrations, \( N_a \), from a Passive Cavity Aerosol Spectrometer Probe (PCASP), a PMS cloud condensation nuclei (CCN) from a CCN spectrometer and cloud nuclei from the Condensation Particle Counters (CPCs). PCASP gives aerosol size distributions in the range of 0.1-2.5 \( \mu \text{m} \) over 20 bins. The dual-chamber PMS CCN spectrometer is set to give CCN concentrations at 0.3 % and 0.6 % super-saturations. Condensation nuclei (CN) measurements were made with CPCs with small-size cutoffs at 3 nm, 10 nm and 15 nm. The cloud and precipitation data included cloud droplets number concentrations, \( N_d \), from Cloud Aerosol Spectrometer (CAS) in the range of 0.6-60 \( \mu \text{m} \) over 20 bins, and drizzle water contents from the cloud imaging probe (CIP) in the range of 25-1550 \( \mu \text{m} \) over 62 bins. Descriptions of the instruments used on the TO research aircraft are given in Table 2.2. Acronyms and symbols used in this study are listed in Table A1 in the appendix.

**Table 2.2:** Characteristics of instruments used in Barbados Aerosol Cloud Experiment.

<table>
<thead>
<tr>
<th>Instrument</th>
<th>Observations/Purpose</th>
</tr>
</thead>
<tbody>
<tr>
<td>Standard meteorological instrument</td>
<td>Winds, temperature, dew-point, cloud liquid water content, surface temperature, etc</td>
</tr>
<tr>
<td>95 GHz Frequency Modulated Continuous Wave (FMCW) Doppler radar (upward vertically pointing mode)</td>
<td>Doppler spectrum; Cloud properties, in-cloud turbulence</td>
</tr>
<tr>
<td>CPCs</td>
<td>Ultrafine aerosols</td>
</tr>
<tr>
<td>Passive Cavity Aerosol Spectrometer Probe (PCASP)</td>
<td>Aerosols 0.1 - 2.5 ( \mu \text{m} ), 20 bins</td>
</tr>
<tr>
<td>Cloud Aerosol Spectrometer (CAS)</td>
<td>Aerosols and Clouds 0.6 - 60 ( \mu \text{m} ), 20 bins</td>
</tr>
<tr>
<td>Cloud Imaging Probe (CIP)</td>
<td>Drizzle 25 - 1550 ( \mu \text{m} ), 62 bins</td>
</tr>
<tr>
<td>CCN-200</td>
<td>CCN (super-saturation at 0.3 %, 0.6 %)</td>
</tr>
</tbody>
</table>
2.2.2 95-GHz FMCW Doppler airborne radar

The airborne radar used in this study, a solid-state Frequency Modulated Continuous Wave (FMCW) 95 GHz Doppler radar (Mead et al., 2003), is mounted on the CIRPAS research aircraft Twin Otter, and operated in an up-looking (zenith viewing) mode (perpendicular to the aircraft centerline; beam width 0.7°, Fig. 2.2b). The radar data were obtained at a sampling rate of 3-Hz (temporal resolution) with range gates at a 24 m spatial resolution and a velocity resolution of 0.16 m s\(^{-1}\) with a dead zone of less than 50 m, and thus, allows radar observation in close proximity to the in situ probe turbulence and microphysical measurements. The radar characteristics are summarized in Table 2.3.

**Table 2.3:** Characteristics of a FMCW radar.

<table>
<thead>
<tr>
<th>Parameter</th>
<th>Value</th>
</tr>
</thead>
<tbody>
<tr>
<td>Center frequency (GHz)</td>
<td>94.8</td>
</tr>
<tr>
<td>Peak transmit power (dBm)</td>
<td>30</td>
</tr>
<tr>
<td>minimum detectable sensitivity at 1 km (dBz)</td>
<td>-27</td>
</tr>
<tr>
<td>PRF (Hz)</td>
<td>39.78138</td>
</tr>
<tr>
<td>Maximum range (m)</td>
<td>5237</td>
</tr>
<tr>
<td>Maximum velocity (m s(^{-1}))</td>
<td>20</td>
</tr>
<tr>
<td>Range Resolution (m)</td>
<td>24</td>
</tr>
<tr>
<td>Receiver noise figure (dB)</td>
<td>7.0</td>
</tr>
<tr>
<td>Antenna Diameter (cm)</td>
<td>30 (12 inches)</td>
</tr>
<tr>
<td>Antenna Gain (dB)</td>
<td>46</td>
</tr>
<tr>
<td>Antenna Beam width (degrees)</td>
<td>0.7</td>
</tr>
</tbody>
</table>
| Radar beam orientation                                 | Up-looking  
(perpendicular to the aircraft centerline) |
2.2.3 Chaff tracer

The entrainment process and flows in and around the small cumulus cloud are studied by using a passive tracer (radar chaff) in chapter 4. The radar chaff elements used for this experiment are pre-cut metallic coated fibers, which are cut to a length of about 1/4 of the radar wavelength (typical size ~ 0.75 mm, see Fig. 2.2c) to maximize radar returns. The chaff fibers are packed in fiber tubes (14 cm length × 3.5 cm diameter), and are mounted in a dispenser, which holds a maximum of 24 tubes, mounted beneath the wing of the aircraft (Fig. 2.2d). Chaff from each of the tubes is dispersed into the air stream by a small explosive charge that is detonated on command. In this study the chaff was dispersed near the cloud tops and edges of a small cumulus cloud. Since the terminal velocity of the fibers is about 2 cm s⁻¹, they effectively track air motions. After chaff fibers were dispensed, the aircraft made penetrations of the cloud at lower levels, to observe the chaff signals above with the radar.

2.2.4 Ragged Point aerosol measurements

Aerosol measurements are made at a site located on the edge of a 30 m high bluff on the easternmost coast of Barbados. Daily aerosol samples are collected at the top of a 17 m high tower using a high-volume (nominal 1 m³ min⁻¹) filter sampling system. To minimize impacts from local sources, sampling is electronically controlled so as to sample winds that blow from the sea (i.e., within the sector extending from 335° to 130°) at wind speeds greater than 1 m s⁻¹. Filters are changed on a daily schedule and periodically returned to Miami where the soluble components are extracted with water (Li-Jones et al., 1998). The extracts were analyzed for major soluble inorganic ions: Na⁺
by flame atomic absorption; Cl, NO₃, SO₄ by suppressed ion chromatography (Savoie et al., 1989). Sea-salt aerosol concentrations are calculated from the Na⁺ concentration based on the weight ratio of sea-salt to Na, 3.256. The filters are then placed in a furnace at 500 °C for 14 hours; the filter ash is weighed and, after corrections for filter blank, the residue is ascribed to mineral dust (Prospero and Lamb, 2003; Trapp et al., 2010). The standard error in the mineral aerosol concentration is about ±10 % for concentrations greater than about 1 µg m⁻³.

The Ragged Point site supports a NASA AERONET sun photometer and a NASA MPLNET micro-pulse lidar. The photometer measures aerosol optical depth (AOD) at eight wavelengths (340, 380, 440, 500, 675, 870, 1020, and 1640 nm) and the Angstrom parameter in the wavelength interval of 440-675 nm (http://aeronet.gsfc.nasa.gov/new_web/units.html). We used level 2.0 data that were cloud screened and quality assured. Vertical structures of aerosols and clouds at Ragged Point were obtained from a micro-pulse lidar (MPL) system at Barbados (http://mplnet.gsfc.nasa.gov/cgi-bin/Mplnet/site_page_direct.cgi/allsite=Ragged_Point).

2.2.5 Sounding data and back trajectories

To characterize boundary layer (BL) structures of Africa and Barbados sites, daily rawinsonde observations from Grantley Adams in Barbados (13.06 °N, 59.48 °W, WMO ID: 78954), and GOOY (14.73 °N, 17.50 °W, WMO ID: 61641) and GOTT (13.76 °N, 13.68 °W, WMO ID: 61687) from western Africa are used. Sounding data were obtained from the University of Wyoming’s online upper-air data (http://weather.uwyo.edu/upperair/sounding.html). The history of air masses sampled by
the aircraft is estimated using the Hybrid Single Particle Lagrangian Integrated Trajectory (HYSPLIT) Model from the average location (13.2 °N, 59 °W) of the flight domains. NCEP Global Data Assimilation System (GDAS) data are used as input meteorological gridded data for each flight. The 315-hr (~ 13 day) and 10-day backward trajectories, arriving at Barbados at 500 m, are calculated to give a general sense of the origin of the air masses sampled on the aircraft missions.

### 2.2.6 AERONET Sun photometer and MPLNET

AERONET Sun photometers measured spectral aerosol optical thickness (AOT) at eight wavelengths (340, 380, 440, 500, 675, 870, 1020, and 1640 nm) and water vapor at wavelength 940 nm (http://aeronet.gsfc.nasa.gov/new_web/units.html) at Ragged point (13.2 °N, 59.5 °W) in Barbados. Level 2.0 data that were cloud screened and quality assured are used in this study. Vertical structures of aerosols and clouds at Ragged Point were obtained from a micro-pulse lidar (MPL) system at Barbados (http://mplnet.gsfc.nasa.gov/).

### 2.3 Observing techniques-Calibration of radar reflectivity

The airborne cloud radar used in this study is an important tool to provide vertical structures of cloud and precipitation properties. To make full use of this tool, velocity corrections to correct the aircraft motion as well as reflectivity calibration are made. In this section, calibration of radar reflectivity is described. The velocity correction procedure is addressed in Section 2.4.
Strength of radar signal (i.e., power) received in the radar receiver is expressed by

\[ P_r = \frac{C \eta}{r^2} \]  

(Equation 2.1)

where \( r \) is distance between radar (receiver) and target (e.g., cloud droplets), \( \eta \) is radar reflectivity factor \((\eta=\Sigma D^6\text{, in units of } \text{mm}^6 \text{ m}^{-3})\), and \( C \) indicates radar constant. The radar constant \( C \) is determined by radar characteristics such as radar wavelength, antenna size so on, and is provided by a radar manufacturer. Otherwise, it can be attained by comparing reflectivity measured by radar with the 6th moments of drop size distribution (i.e., radar reflectivity) estimated from probe data. Equation (2.1) can also be expressed by the logarithmic form as

\[ 10\log(P_r) = 10\log(C) + 10\log(\eta) - 20\log(r) \]  

(Equation 2.2)

and is rearranged as

\[ 10\log(\eta) = 10\log(P_r) + 20\log(r) - 10\log(C) \]  

(Equation 2.3)

Here term A represents radar reflectivity (hereafter \( Z \), 6th moments of DSD, in units of dBz), where \( \eta \) indicates a radar reflectivity factor, \( \eta = \int N(D)D^6dD \) (in units of \( \text{mm}^6 \text{ m}^{-3} \)). Term B indicates received power. Term C is related to range (distance) effect on received power of the radar; the signal received at closer distance would give the stronger signal. The combination of term B and C indicates range corrected power, but correction for term D is still needed to obtain the real strength (i.e., reflectivity) of target. Time series of range corrected power with height, and that measured from the lowest
level of the cloud radar (black) together with the 6th moments of drop size distribution estimated from probe data (magenta) are shown in Fig. 2.3 for the cloud sampled on 22 March.

![Image of radar data](image_url)

**Figure 2.3:** (a) Time series of range corrected power with height, and (b) that measured from the lowest level of radar (black) and the 6th moments of drop size distribution estimated from probe (magenta) for precipitating cloud sampled on 22 March 2010.

Term D in Eq. (2.3), included radar constant C, is determined by comparing range corrected power measured from radar at the lowest level (terms B+C) with radar reflectivity estimated from 6th moments of drop size distribution of probe data. The scatter diagrams of reflectivity, estimated from the probe (x-axis) and measured from the lowest level of the radar (y-axis) are shown in Fig. 2.4 for the same cloud shown in Fig. 2.3. The difference between range corrected power from the radar and reflectivity estimated from the probe is shown as numerical number in (b). Radar reflectivity after subtracting the difference (in this case, 30 dBz) is shown in y-axis in Fig. 2.4b. The
dashed blue line in Fig. 2.4b shows one to one line. The difference here indicates the value of term D in Eq. (2.3).

**Figure 2.4:** Scatter diagrams of (a) range corrected power measured from the lowest level of cloud radar (y-axis) and radar reflectivity (the 6th moments of drop size distribution) estimated from probe (x-axis) for the same cloud shown in Fig. 2.3. The difference between range corrected radar power and probe reflectivity is shown as numerical number in Fig. 2.4(b). The difference here indicates term D in Eq. (2.3). Radar reflectivity, after subtracting the difference between them, is shown in x-axis in Fig. 2.4(b). The dashed blue line in Fig. 2.4(b) shows the one to one line.

The difference between range corrected power and probe reflectivity is about 30 dBz, on average, in Fig. 2.4 for the heavily precipitating clouds. The short wavelength radar, such as cloud radar, in the rain conditions, may suffer from attenuation, and thus, it is possible to give a different value for the term D from that obtained from non-precipitating cloud. Further, Rayleigh scattering \(Z\) is proportional to \(D^6\) may not valid for heavier rain where larger drops (e.g., \(D > \sim 1.1 \text{ mm}\) for 95-GHz radar) exist. To get the radar constant that can be generally applied to the clouds sampled during the BACEX, time series of range corrected radar power, probe reflectivity, along with scatter diagrams of reflectivity with an average difference between them are shown in Fig. 2.5 and 2.6 for lightly precipitating, and non-precipitating clouds.
Figure 2.5: (a,c) Time series of range corrected power measured from the lowest level of radar (black) and the 6th moments of drop size distribution estimated from probe (magenta) for lightly precipitating cloud sampled on (a) 23 March, and (c) 30 March, 2010. (b,d) Scatter diagrams of radar reflectivity (y-axis) and probe reflectivity (x-axis). The difference between range corrected radar power and probe reflectivity is shown as numerical number in Fig. 2.5(b,d). The blue dashed line shows the one to one line.

Figure 2.6: Same as Fig. 2.5 except for non-precipitating cloud sampled on 7 April 2010. Probe radar reflectivity is estimated from CAS forward scattering data.
The difference between range corrected radar power and probe reflectivity is about 33 and 34 dBz (Fig. 2.5) in the lightly precipitating clouds. Reflectivity pattern estimated from the CIP probe is closer to reflectivity measured by the radar in the precipitating clouds. On the other hand, reflectivity pattern estimated from the CAS probe is closer to the reflectivity pattern measured by the radar in case of non-precipitating clouds. Thus, reflectivity estimated from CAS probe (forward scattering) are shown in Fig. 2.6. Further, difference between probe (CIP or CAS) reflectivity and range corrected radar power is smaller in precipitating clouds due to the attenuation; radar underestimates the real strength of the signal due to the attenuation during the precipitation. For example, the difference between range corrected radar power and probe reflectivity (i.e., term D in Eq. (2.3)) is about 29-31 dBz in the heavily precipitating clouds; it is about 33-34 dBz in the lightly precipitating clouds, and about 43 dBz in non-precipitating clouds. Consequently, the difference of 38 dBz is used for term D in Eq. (2.3), in this study, by considering the preponderance of lightly and non-precipitating clouds during the BACEX. Time series of radar reflectivity for the same cloud shown in Fig. 2.3, sampled for the precipitating cloud, and are shown in Fig. 2.7.

**Figure 2.7:** An example of time series of radar reflectivity with height for precipitating cloud sampled on 22 March 2010.
Here, radar reflectivity is calibrated by the effects of distance (term C) and radar constant (term D) of 38 dBz, that is generally applied to data during BACEX in Eq. (2.3).

2.4 Observing techniques-Mie technique

The cloud radar used in this study is mounted on top of the aircraft in a vertically pointing mode, so that the observed vertical component of the Doppler velocity is contaminated by the aircraft’s motion when the aircraft is not flying straight and level. This section describes the processes of platform motion correction to correct aircraft motion and techniques to retrieve vertical air motion using Mie-scattering.

2.4.1 Background

Measurements of vertical air motion and microphysics are essential for improving our understanding on the interaction of dynamics and microphysics in convective clouds. Aircraft penetrations in clouds and precipitation offer advantageous direct (in-situ) methods for measuring the vertical air motion and microphysics in cumulus clouds. However, they are limited to point measurements (1-D at flight level) and in areas that can be safely penetrated by aircraft. Measurements of vertical air motion in precipitating clouds using Doppler radars have been attempted since the 1960s (e.g., Probert-Jones and Harper, 1961; Doviak and Zrnic, 1993). Airborne Doppler radar systems (e.g., Heymsfield et al., 1996; French et al., 1999) extend aircraft measurements to 2-D and offering a more comprehensive characterization of the cloud and precipitation structure. However the retrieval of air motion and microphysics beyond the flight level is subject to several assumptions with respect to the Particle Size Distribution (PSD, e.g., Kollias et al., 2002).
From the ground, the absence of platform motion offers the opportunity of retrieving the vertical air motion with certain Doppler radar systems without relying on assumptions about the PSD. During the 1980s, wind profilers (Doppler radars operating at VHF and UHF) were adopted to extract the vertical air motion using Bragg scattering (e.g., Wakasugi et al., 1986; Gossard, 1988; Rogers et al., 1993; May and Rajopadhyaya, 1996; Rajopadhyaya et al., 1998). However, the deployment of a wind profiler from an aircraft is not practical due to typical antenna size requirements. During the past decade, the vertical air motion in precipitation has been retrieved with short wavelength (e.g., \(\lambda=3\) mm, 95-GHz frequency) radars, based on the fundamental physics (Kollias et al., 2002; Kollias et al., 2003; Kollias et al., 2007; Giangrande et al., 2010), using Mie scattering theory as first proposed by Lhermitte (1988). In a nutshell, the novel vertical air motion technique is based on the identification of spectral signatures (mainly the first minima of the backscattering cross-section) in the recorded Doppler spectra that occur at a particular particle diameter and their exploitation for estimating the Doppler shift due to the vertical air motion. Contrary to wind profilers, 95-GHz Doppler radars are highly portable and are available on moving platforms (e.g., Li et al., 2004). However, the platform motion can smear Mie scattering signatures in airborne Doppler radar spectra and such observations have not being reported.

Here, unique airborne 95-GHz Doppler radar observations from shallow marine cumulus clouds are presented. The observations clearly demonstrate that the Mie scattering signatures are available from airborne 95-GHz Doppler radar systems and thus offer the opportunity of extending the vertical air motion retrievals beyond the flight level. In addition to the radar Doppler spectra from precipitation, Doppler spectra from cloud
droplets are also observed in the same volume. This offers the opportunity to validate the Mie scattering-based air motion retrievals with the air motion retrievals from the cloud droplets’ spectral peaks.

2.4.2 Methodology

2.4.2.1 Mie scattering

Light scattering by spheres with size parameter \( x = \pi D/\lambda \ll 1 \), where \( \lambda \) is the wavelength, is described by the Rayleigh scattering approximation and the backscattering cross-section \( \sigma_b \left( \text{mm}^2 \right) \) is proportional to the sixth power of the particle diameter, \( D \). The monotonic increase of \( \sigma_b \) with particle size is disrupted in the Mie scattering regime (~ \( 0.1 < x < 100 \)), where, \( \sigma_b \) resonates with successive maxima and minima. Assuming spherically shaped raindrops, the characteristic raindrop sizes, where maxima and minima occur, are well described by Mie-theory (Mie, 1908); e.g., the locations of the first Mie maximum, first Mie minimum, and the second Mie maximum occur at around \( D \sim 1.15 \) mm, \( D \sim 1.65 \) mm and \( D \sim 2.25 \) mm (Giangrande et al., 2010). The use of an oblate spheroid model for large raindrops and T-matrix scattering theory (e.g., Mishchenko et al., 1996) provides slightly different solutions for the maxima and minima \( \sigma_b \) at 95 GHz. Nevertheless, the Mie solution to scattering by sphere is used here since the key research findings of this work are not altered.

The \( \sigma_b \) resonances have been frequently observed in the 95-GHz radar Doppler spectrum and used as a reference for the retrieval of the vertical air motion (e.g., Kollias et al., 2002; Giangrande et al., 2010). For a 95-GHz radar with a wavelength of 3.2 mm,
the first Mie minimum occurs at a raindrop diameter of 1.65 mm, which translates to a raindrop terminal fall velocity of 5.8 m s\(^{-1}\) based on the well-known Gunn and Kinzer (1949; hereafter GK49) terminal velocity data using the Beard (1985) fit:

\[
V_i (cm/s) = \exp(5.984 + 0.8515x - 0.1554x^2 - 0.03274x^3) \quad \text{(Equation 2.4)}
\]

where \(x = \ln[D(\text{mm})]\). This fit provides velocities, with an \textit{rms} deviation from those of GK49, of less than 1 \% between D=0.5 and 6 mm. Since the GK49 measurements were made at sea level for still air conditions, a density correction is applied to the fall velocity for the observations made aloft (at altitude z) by using the Beard (1985) formulation,

\[
V_f = V_z = \left(\frac{\rho_o}{\rho_z}\right)^m V_0 , \quad \text{(Equation 2.5)}
\]

where \(\rho_z\) is the air density at altitude (z) and \(\rho_o\) is the density at z=0 for standard conditions. Here, \(\rho_o=1.194 \, \text{kg m}^{-3}\). The coefficient \(m\) is a function of the raindrop diameter of interest: \(m = 0.375 + 0.025 \, \text{D [mm]}\). Hereafter, we will call the \(V_f\) as a density corrected fall velocity. \(V_f\) in Eq. (2.5) is \(v_{f,1}\) (will be seen later) when \(V_0\) of \(D_{1M}=1.65\) mm is used, where \(D_{1M}\) is the location of the first Mie minimum in terms of particle size.

\subsection*{2.4.2.2 Platform motion correction and vertical air motion retrieval}

The FMCW radar is mounted on top of the aircraft in an upward vertically pointing beam mode (perpendicular to the aircraft centerline). Therefore, the observed vertical component of the Doppler velocity is contaminated by the aircraft’s motion when the aircraft is not flying straight and level (Heymsfield, 1989; Lee et al., 1994; Leon and
Vali, 1998). However, the observed radar Doppler velocities can be corrected for the aircraft’s motions as recorded by the aircraft’s inertial navigation system (INS). The largest contributor to the vertical component of the Doppler velocity owing to the aircraft motion is pitch angle (Heymsfield, 1989). The effects of roll angle can be ignored in case that the aircraft performs the level leg, which the cloud penetrations are usually performed, as they are small compared with the effects of pitch angle (as will be shown later in this section). The radial Doppler velocity bias $V_{D,p}$, introduced to the 95-GHz FMCW due to the platform motion, involves the pitch angle, speed of aircraft and the vertical motion of the aircraft and can be expressed as

$$V_{D,p} = V_p \sin \alpha - W_p \cos \alpha$$

(Equation 2.6)

where $\alpha$ is the aircraft pitch angle, $W_p$ is the aircraft’s physical up and down motion velocity, and $V_p$ is the true speed of the airplane through the air ($G_S = V_p/\text{heading vector} + \text{drift vector due to wind}$, where $G_S$ is speed over ground). Here, positive (+) indicates radial velocities away from the platform (i.e., upward motion). In addition to the effects of the pitch angle and the aircraft’s physical up and down motion on the radial component of vertical velocity, the roll angle $\phi$ can also affect the observed vertical velocity. However, the effect is insignificant. For example, the roll angle is approximately $-0.98^\circ$ on average for the case studied, while the pitch angle is $3.6^\circ$ on average. So even with a cross-wind component of 10 m s$^{-1}$, a 1$^\circ$ variation in the roll would result in a 0.2 m s$^{-1}$ error in the retrieved vertical velocity. This error could be removed, since the aircraft roll and the cross-wind component can be obtained at the level of the aircraft observations. In the cases shown here the level legs are flown either
upwind or downwind. Thus in this case, the Doppler velocity corrections due to the roll angle will be small, since the air flow perpendicular to the aircraft fuselage is small, and since the radar is pointing upward, roll contributions are scaled by the sin of the roll angle. In the case studied here, the impact of the average roll angle on the observed vertical velocity component is estimated to be -0.02 m s\(^{-1}\), compared with 1.0 m s\(^{-1}\) from the pitch angle for the case studied (e.g., cloud in Fig. 2.12) for 1 m s\(^{-1}\) of wind, for example, and thus, is not included in Eq. (2.6) for the first-order correction (see Heymsfield, 1989, Eq. (6), for a comprehensive treatment of the pitch and roll contributions to the vertical velocity component).

**Figure 2.8:** Example of radar Doppler spectrum observed at a certain range (e.g., 156 m) from the aircraft. (a) The raw recorded radar Doppler spectrum with the observed location of the first Mie minimum peak \(V_{D,1}\) indicated by the vertical dashed line and the expected location of the first Mie minimum peak \(V_{f,1}\) in the absence of platform motion and air motion indicated by the vertical solid line. (b) The raw radar Doppler spectrum shifted to zero contribution from the platform and the air motion. The observed locations of the first and second Mie maxima are denoted as upward arrows (right to left) in Fig. 2.8b.
Doppler velocity $V_{D,1}$ of the first Mie minimum in each recorded radar Doppler spectrum is examined to identify the Mie features. An example of the spectrum from a shallow precipitating cumulus cloud is shown in Fig 2.8. In this case, the observed Doppler power spectrum shows three spectral peaks of Doppler velocity (Fig. 2.8). Negative Doppler velocities indicate motions towards the radar; thus, the largest precipitation particles in the radar sampling volume are on the left side of the Doppler spectrum. The locations of peaks (and valleys) in this study are determined using 2nd order Savitzky-Golay filtering (Savitzky and Golay, 1964; Steinier et al., 1972) for the power density (dB/ms$^{-1}$) or power (dB) as a function of velocity.

The first Mie minimum $V_{D,1}$ is observed at -0.2 m s$^{-1}$ (dashed line) in the raw recorded radar Doppler spectrum (Fig. 2.8a) and its theoretical location $V_{f,1}$ in the absence of platform or air motion is indicated by the solid line. Then, the raw recorded radar Doppler spectrum in Fig. 2.8a is shifted by the magnitude of $V_{D,P}$ to apply the platform motion (refer to the locations of $V_{D,1}$ in Fig. 2.8a and in Fig. 2.8b). The distance between $V_{D,1}$ in Fig. 2.8a and in Fig. 2.8b corresponds to the magnitude of $V_{D,P}$. The velocity Doppler shifts $V_{DS}$ between the expected ($V_{f,1}$) and observed ($V_{D,1}$) can be explained by the projection of the vertical air motion $w_a$ to the radar radial axis and $V_{D,P}$ and can be expressed as:

$$V_{DS} = V_{D,1} - V_{f,1} = (w_a \cos \alpha + V_{f,1} - W_p \cos \alpha + V_p \sin \alpha) - V_{f,1} = w_a \cos \alpha + V_{D,P} \quad (\text{Equation 2.7})$$

where $V_{f,1} = v_{f,1} \times \cos \alpha$, $v_{f,1}$ is a density corrected fall velocity for the first Mie minimum location (refer to Eq. (2.5)). For ground-based systems, if the first Mie minimum is found at its theoretical location, this indicates the absence of vertical air motion (e.g., Kollias et
For the airborne 95-GHz radar system, this implies that the radial air motion compensates the platform radial motion component. In summary, the vertical air motion $w_a$ is retrieved using the following expression:

$$w_a = \frac{V_D - V_{DP}}{\cos \alpha}$$  \hspace{1cm} \text{(Equation 2.8)}

Uncertainties related to the retrieval of $w_a$ include: an inaccurate positioning of the first Mie minimum location ($V_{D1}$, $D_{IM}$ in terms of velocity and diameter, respectively) in the Doppler spectrum, which are related to the performance of peak picking algorithm and resolution of the velocity bins (0.157 m s$^{-1}$ in this study); uncertainty due to the oblate rather than spherical shape of raindrops. However, oblate spheroid effects are minimal for the rain observed. Nevertheless, including these effects gives approximately a 0.15 m s$^{-1}$ difference in $v_{f1}$ estimates, and 0.3 m s$^{-1}$ difference in the $w_a$ estimates, when $D_{IM} = 1.65$ ($D_{IM}$ with spherical shape of raindrop) and/or $D_{IM} = 1.71$ mm ($D_{IM}$ with oblate shape of raindrop) is used as the position of the first Mie minimum. Uncertainty due to the regression fit to Gunn and Kinzer dataset in Eq. (2.4) and uncertainty due to using the mean density profile are negligible.

Further velocity corrections, as a function of height above the aircraft, would be needed if the Mie resonance occurs in a strongly sheared environment. In a case where the wind increases with height in the direction of the aircraft motion at a rate of 10 m s$^{-1}$ per km, at 1 km a 3.6° pitch angle would bias the vertical velocity by 6 m s$^{-1}$ ($10 \text{ m s}^{-1} \times \sin (3.6^\circ)$). For a similar shear in the cross-wind component, but a 1° change in the roll would give a 0.2 m s$^{-1}$ change in the retrieved vertical velocity. In the case presented here, the shear in the cloud layer is about 6 m s$^{-1}$ km$^{-1}$ and the flights were flown either up or
down wind. Thus the cross-wind component is small, and the uncertainty in the retrieved velocity 1 km above the aircraft will be about 0.4 m s\(^{-1}\) due to the pitch.

### 2.4.3 Validation of the technique

An example of the zeroth-moment (radar reflectivity) and the first-moment (mean Doppler velocity that were corrected for platform motion) of the 95-GHz FMCW Doppler spectrum from a relatively shallow precipitating cumulus cloud (observed about 80 km NE of Barbados), are shown in Fig. 2.9. The aircraft observations were made near the cloud base at an air speed of about 58 m s\(^{-1}\) and an altitude of 768 m above sea level. The aircraft within this 6 km long racetrack intercepted several clouds. Our analysis focuses on the shallow convective cloud observed between 40-70 seconds (after 16:16:48 UTC) in Fig. 2.9.

The average cloud top is 0.9 km above the aircraft flight level (~1.67 km MSL). Despite its shallow nature, this cloud produces strong radar reflectivity (e.g., maximum reflectivity of -8.5 dBz) due to the presence of large precipitating size particles. The precipitation particles exceed 1.65 mm in sizes that give the Mie resonance in the Doppler spectra, shown later in Fig. 2.10. Detrained cloud elements into the inversion layer that caps the convection are also seen above 0.9 km (ARL; Above Radar Level) earlier in the penetration (~40-50 s) with tilted features due to the strong wind shear across the inversion layer. The Doppler velocity (Fig. 2.9b) shows mostly downward motion associated with the precipitation shaft. However, alternating up and downward motions inside the cloud are noticeable, which implies the possibility of some re-circulation within the cloud that may help the growth of larger droplets. But,
interpretation of the mean Doppler velocities is not straightforward due to the presence of embedded vertical air motion in the presence of precipitation. Newly growing clouds without precipitation are also shown in Fig. 2.9 (echoes at around 10 s; 20-30 s and 80-90 s) and exhibit relatively weak returns with strong updrafts.

**Figure 2.9:** A time-height cross-section of (a) radar reflectivity and (b) motion corrected Doppler velocity (+: upward) in the precipitating cloud on 5 April 2010 during BACEX from 16:16:48 to 16:18:36 (UTC). The reported height is Above the airborne Radar Level (ARL). Zero height corresponds to 768 m above sea level. The dashed line indicates a specific time for the Doppler spectra in Fig. 2.10.

The aircraft motion correction procedure and the retrieval of the radial air motion is described below (Fig. 2.10). An example of uncorrected for platform motion 95-GHz FMCW radar Doppler spectrogram is shown in Fig. 2.10a; its corrected version in Fig. 2.10b and the locations of the Mie maxima and minima in Fig. 2.10c for a single profile recorded on 16:17:45 UTC 5 April 2010.
Figure 2.10: Doppler power spectra observed on 16:17:45 UTC 5 April 2010, which is denoted as a dashed line in Fig. 2.9. Vertical profiles of (a) uncorrected and (b) corrected for platform motion Doppler spectra. The vertical lines in (a) and (b) correspond to the $V_{f1}$ and $V_{D,p}$. (c) Vertical profiles of peaks and valleys of motion corrected Doppler spectra, corresponding to Fig. 2.10b. Red (magenta) circles and blue crosses indicate 1st (2nd) maximum and 1st minimum location in the Doppler spectra due to the Mie oscillation. Blue asterisks represent peaks due to the cloud droplets. An example of radar Doppler spectrum observed at 156 m is shown in Fig. 2.8.

The Mie maxima (denoted by red/yellow colors in Fig. 2.10b and circles in Fig. 2.10c) are separated by approximately 3 m s$^{-1}$ (centers on around -3 m s$^{-1}$ and -6 m s$^{-1}$ at radar level) and those peaks are due to the Mie scattering mechanism with differences between each peak being consistent with the predicted value: $\sim$2.5-3 m s$^{-1}$ of spacing between the 1st and 2nd Mie maximum; $\sim$1.5-2 m s$^{-1}$ spacing between 1st maximum and 1st minimum; $\sim$1.2 m s$^{-1}$ spacing between 1st minimum and 2nd maximum (Lhermitte, 1988; Giangrande et al., 2010). The peaks in Fig. 2.10c, denoted by blue asterisks, are from the cloud droplets’ return. Vertical air motion, as a function of height, retrieved by the Mie technique is shown in Fig. 2.11 (as circles).
Figure 2.11: Vertical air motions with height, retrieved from the Mie technique (circles) and obtained from the Doppler velocity of cloud droplets (air tracer technique, asterisks) on 16:17:45 UTC (hh:mm:ss) on 5 April 2010, denoted as dashed line in Fig. 2.9.

The profile of vertical air motion retrieved using the Mie signatures in the radar Doppler spectra can be compared directly with the vertical air motion estimates that we can deduce from the location of the cloud droplets’ spectral peak (small peaks on the right side of the recorded Doppler spectra, asterisks in Fig. 2.10c). The location of the spectral peaks due to cloud droplets are considered to track the vertical air motion, since these droplets have a negligible fall velocity (0.03 m s$^{-1}$ for D=10 μm, for example). This is the first time (to our knowledge) that separate peaks due to Rayleigh scattering from cloud droplets and due to Mie scattering from raindrops are observed and documented in the same radar Doppler spectra. This vertical air motion agrees well with those retrieved from the Mie technique (circles in Fig. 2.11). The magnitude of an updraft in this cumulus cloud (the echoes extended about 1 km above the aircraft, see Fig. 2.9) increases
with height from \( \sim 1 \text{ m s}^{-1} \) at the level of aircraft to \( \sim 5 \text{ m s}^{-1} \) at 500 m above the radar. The difference between vertical air motions obtained from the Mie technique and cloud droplet returns is less than 0.1-0.2 m s\(^{-1}\) with a correlation of 0.997 between them. But any uncertainties in this estimate due to aircraft motion corrections (\( \sim 3 \text{ m s}^{-1} \) in this example) affect both the Mie and the direct cloud droplets’ retrievals.

The presence of Mie scattering signatures and/or cloud droplets’ spectral peaks in the recorded Doppler spectra can be used to retrieve the vertical air motion structure in shallow precipitating cumulus clouds. This is illustrated below using observations of a different cumulus cloud observed on the same flight (5 April, 2010).

\[ \text{Figure 2.12: A time-height cross-section of (a) radar reflectivity and (b) motion corrected Doppler velocity (+: upward) above the aircraft/radar level (ARL) in the precipitating cloud on 5 April 2010 during BACEX from 16:13:48 to 16:15:36 UTC (a time duration of 108 seconds). Dotted lines denote a specific time periods when larger particles are observed in the Doppler spectra. (A)-(E) are used in Fig. 2.13.} \]
Figure 2.12 shows the time-height cross-section of radar reflectivity and mean Doppler velocity that has been corrected for platform motion. This precipitating cumulus cloud was observed at about 7 km NE of the cloud shown in Fig. 2.9. The aircraft observations were made near the cloud base (along the wind direction; downwind), at a mean air speed of about 59 m s\(^{-1}\) (thus, the horizontal size of the cloud ~2.7 km) and at an altitude of 770 m above sea level (cloud top heights ~1700 m MSL). This cloud shows a similar cloud structure to the cloud in Fig. 2.9 in terms of the reflectivity, Doppler velocity and cloud top height. The period enclosed by dashed lines indicates the area where Mie scattering signatures are observed in the recorded radar Doppler spectra. In Fig. 2.12a, the highest reflectivity is associated with precipitation in the cloud. In the Doppler velocity field (Fig. 2.12b), downward fall velocity is dominant overall, but a strong updraft is found along the downwind side of cloud edges near the cloud top. The platform motion correction in this case is 3.3±0.7 m s\(^{-1}\).

The evolution of the cloud droplets and precipitating size drop are shown in the Doppler power spectra in Fig. 2.13 for the cloud observed in Fig. 2.12. At the cloud edges (16:14:24-16:14:26, including point A in Fig. 2.13), positive vertical air velocities from 0-2 m s\(^{-1}\) are observed. Further into the cloud and near the cloud base (up to point B1, 16:14:27-16:14:35, Fig. 2.13), two separate Doppler spectra peaks are observed: one spectral peak from cloud droplets (right side of the recorded radar Doppler spectra) positioned around 2 m s\(^{-1}\) (suggesting the presence of an equal magnitude of updraft) and another spectra peak from larger, precipitating size particles centered around -2.5 m s\(^{-1}\). The actual fall velocities of the raindrops are ~ -4--4.5 m s\(^{-1}\) given that these drops are embedded in an updraft of approximately ~ 2 m s\(^{-1}\). Between 16:14:37-16:14:42 (points
B3 to C), three distinctive returns are observed: $v \sim +1.5 \text{ m s}^{-1}$, $v \sim -3.5 \text{ m s}^{-1}$, and $v \sim -5.5 \text{ m s}^{-1}$, indicating signals from cloud droplets, moderate sized raindrops, and larger raindrops (at least $D > 1.65 \text{ mm}$), respectively. The separation between the cloud droplets and moderate sized raindrops is induced by a particle size gap, while Mie scattering induces the separation between the two large raindrop related peaks.

![Figure 2.13: Time-height cross-sections of Doppler power spectra for the cloud in Fig. 2.12 (for the 30 seconds period from 16:14:24 to 16:14:53 UTC, 1-second interval; approximately from A to E in Fig. 2.12) on 5 April, 2010. The vertical dashed line indicates the location of zero velocity. Details of (A)-(E) are described in the text.](image)

**Figure 2.13:** Time-height cross-sections of Doppler power spectra for the cloud in Fig. 2.12 (for the 30 seconds period from 16:14:24 to 16:14:53 UTC, 1-second interval; approximately from A to E in Fig. 2.12) on 5 April, 2010. The vertical dashed line indicates the location of zero velocity. Details of (A)-(E) are described in the text.

Doppler spectra of the first three seconds when the larger raindrops ($D > 1.65 \text{ mm}$) first appear in the velocity-height plots are shown in B1, B2 and B3 (16:14:35-16:14:37). At point B1, returns from two Doppler velocities are shown: Returns from cloud droplets
(\(v \sim +2 \text{ m s}^{-1}\)) and moderate sizes of raindrops (\(v \sim -3 \text{ m s}^{-1}\)). The actual fall velocity of this raindrop is approximately \(~ -5 \text{ m s}^{-1}\), in turn, equaling to droplets with \(D \sim 1.4 \text{ mm}\).

At point B2, a hint of the 1\(^{\text{st}}\) Mie minimum emerges on the Doppler velocity-height plot at around \(v \sim -4.6 \text{ m s}^{-1}\). Then, at point B3 in Fig. 2.13, the well-defined first Mie minimum and the 2\(^{\text{nd}}\) Mie maximum are observed at around \(v \sim -4.6 \text{ m s}^{-1}\) and \(v \sim -6 \text{ m s}^{-1}\), respectively due to the presence of larger raindrops. The air velocity, shown as the most right peak, is approximately \(\sim +1.3 \text{ m s}^{-1}\). As a result, the actual fall velocities of these raindrops are \(\sim -5.9 \text{ m s}^{-1}\) (\(D \sim 1.65 \text{ mm}\)) and \(\sim -7.3 \text{ m s}^{-1}\) (\(D \sim 2.4 \text{ mm}\)), respectively.

The appearance of the three distinctive returns, which come from the cloud droplets, moderate sizes of raindrops, and raindrops larger than \(D > 1.65 \text{ mm}\), continues to point C. At point D (outside of the cloud core), radar returns mainly come from the negative velocities in the Doppler spectra; centered at around \(\sim -1 \text{ m s}^{-1}\) and \(-3 \text{ m s}^{-1}\), indicating 1 m s\(^{-1}\) of downdraft (air motion) and 2 m s\(^{-1}\) of fall velocity from raindrops of \(D \sim 0.5 \text{ mm}\).

Beyond point D (closer to outside of cloud edges), downward air motion (\(\sim 1 \text{ m s}^{-1}\)) is found without any larger cloud droplets and/or precipitating size drops. The wiggling features of the Doppler power spectra reflect the turbulent condition of the cloud.

Overall, as the aircraft flew into the heavy precipitation regime, larger raindrops were detected (A→C) and vise versa. For example, cloud droplets with updrafts (single peak) were detected at point A. But smaller/moderate raindrop sizes with cloud droplets (double peaks) were observed outside of the cloud core. In the cloud core, raindrops of \(D > 1.65 \text{ mm}\) started to emerge (triple peaks) in the Doppler spectra (e.g., between B3 and C).

As the aircraft flew out of the cloud core, again, smaller/moderate raindrop sizes with
cloud droplets (two peaks) were observed. Outside of the cloud edges, cloud droplets (single peak) with downdrafts were observed.

Figure 2.14: (a) Time-height cross-section of vertical air motions and (b) vertical velocity measured from the aircraft motion sensing and inertial navigation system (INS) at a cloud penetration height (~770 m). The vertical air motions in Fig. 2.14a are retrieved from the Mie technique for the periods between dashed lines, which is enclosed by dashed line in Fig. 2.12b on 5 April 2010 and from the cloud droplets’ spectral peak for the rest of the periods.

The retrieved vertical air motions from the Mie scattering signatures in the radar Doppler spectra are also compared to those obtained from aircraft motion sensing and inertial navigation system (INS) for the precipitating cloud studied (Fig. 2.14). The Mie technique is used for the period where larger particles are observed (periods between dashed lines in Fig. 2.14a). For the rest of the period, peaks from cloud droplets are used. Fig. 2.14 shows that an updraft ($w_u$) of 0.5-2.5 m s$^{-1}$ is embedded in the cloud within the downdraft regions observed in the area between dashed lines in Fig. 2.12. This feature is also shown from the Doppler spectra in Fig. 2.13 where there are returns from cloud droplets at around 0.5-2.5 m s$^{-1}$ between 16:14:29 and 16:14:47. Weak downdrafts were
found at the edges of the clouds compared with stronger updrafts found at the interior of the cloud. Overall, vertical air motions, measured from the radar ($w=1.3\pm0.5$ m s\(^{-1}\), on average), showed slightly stronger updraft velocities ($\sim 0.5$-1 m s\(^{-1}\)) compared with those obtained from the aircraft’s navigation system ($w=0.6\pm0.6$ m s\(^{-1}\), on average). These differences may be due to a vertical air motion increasing with height, but this should not result in large differences since the lowest radar samples are about 50 m above the aircraft.

### 2.4.4 Summary

At the radar wavelength of 3.2 mm (W-band, frequency of 95-GHz), the raindrop backscatter cross section varies between successive maxima and minima as a function of the raindrop diameter (D) that are well described by Mie theory. The first Mie minimum of the backscattering cross section occurs at $D = 1.65$ mm. The location of the first minimum in the recorded radar Doppler spectrum has been used successfully in the past with ground-based radars to retrieve the vertical air motion. Here, the first application of this technique to airborne W-band radar Doppler spectra during observations of warm precipitating cumulus clouds over the eastern Caribbean is presented.

Moreover, a separate spectral peak due to the cloud droplets is also observed in the same radar Doppler spectra that contain Mie signatures. This feature has not been observed from ground-based radars and provides an independent validation of the Mie scattering-based vertical air motion retrieval using the cloud droplets as a tracer of vertical air motion. This is the first demonstration of such Rayleigh and Mie scattering
signatures in the radar Doppler spectra. The application of the Mie technique, using an airborne W-band system, can lead to the new opportunities in cloud and precipitation research.

The unique data set shown in this study show the first-ever confirmation of the vertical velocity retrieval from the Mie maxima and minima with that from the cloud droplet returns. Possible air-borne applications of this technique include the mapping of vertical air velocity fields in hurricane and tropical storm rain bands and in mid-latitude warm cloud systems. Although not shown here, the technique also provides valuable information for mapping the relative raindrop size distribution (as a radar disdrometer) in clouds where the Mie maxima and minima are observed (Giangrande et al., 2010).
Chapter 3: Aerosol Variation and Cloud Properties over Caribbean Sea

3.1 Motivation

The island of Barbados - located on the eastern edge of the Caribbean Sea - typically experiences northeast trade winds and periodic African dust outbreaks. Trade cumulus clouds are the most ubiquitous on the planet (Norris, 1998), and is thought to give the largest uncertainty in tropical cloud feedback in climate system (Bony and Dufresne, 2005). In the trades over Barbados, African dust (so called Saharan Air Layer, SAL) is the dominant aerosol constituent during certain seasons. To better understand the interactions between aerosols, clouds, and precipitation, the overall distribution and characteristics of cloud and aerosol itself should be known. However, little is known about the distributions of cloud intensity (reflectivity), vertical velocity, cloud- thickness (tops and bases). Further, vertical structures of aerosols (e.g., Reid et al., 2002) are rarely documented (The vertical structure of SAL and its effects on atmospheric vertical thermodynamic structure are documented in chapter 5 in detail). This chapter documents variability of aerosol and cloud over Barbados and provides statistics of clouds (e.g., cloud thickness, tops and bases, and distribution of reflectivity and vertical velocity of clouds) to provide a basis for evaluating and improving numerical models in the parameterization of cloud-aerosol-precipitation interactions. The large-scale conditions during BACEX and back-trajectory of air masses arriving at Barbados are described in 3.3 and 3.4, respectively. Aerosol properties are reviewed in 3.5, followed by cloud and precipitation properties in 3.6. The chapter is summarized in 3.7.
3.2 Cloud and aerosol size and distribution variables

To discuss cloud and aerosol properties, it is useful to define several variables that used in this chapter and throughout the dissertation.

### 3.2.1 Cloud droplet size and distribution

Cloud droplet size (effective diameter) in this study is calculated as followings

\[
D_e = \frac{\int N(D)D^3dD}{\int N(D)D^2dD}
\]

(Equation 3.1)

where D is the bin-mean diameter of the probe. The drop size distributions \(N(D)\) were obtained by combining data obtained from the cloud probe (CAS forward scattering) and precipitation probe (CIP) to include cloud to drizzle drops. The first two channels (or bins) from CIP overlap with those from the last two bins of CAS, and have poor accuracy compared with those of CAS, therefore, the first two CIP size-bins were discarded when the data is combined (personal communication with Haflidi H. Jonsson from Naval Postgraduate School).

### 3.2.2 Aerosol particle size and distribution

Aerosol particle size \((D_a)\) and their size distributions (PSDs) in the sub-cloud layer are estimated by combining PSDs from the PCASP and CAS probes to give the full size and distribution information for the range from 0.1 µm to 60 µm. The first ten bins (0.6-2.2 µm) from CAS (0.6-60 µm, 20 bins) overlap with those from PCASP (0.1-2.5µm, 20 bins), and have poor accuracy compared with those of PCASP (in particular, the first
nine channels). Therefore, when the data is combined, the first nine CAS and the last PCASP size-bin were discarded (i.e., PCASP-up to channel 19; CAS-from channel 10 and up). Otherwise, Aerosol particle size and their distributions are estimated from PCASP with the exclusion of any measurements within clouds. Aerosol particle size ($D_a$) is calculated with the same form as Eq. (3.1).

### 3.2.3 Rainfall (Precipitation) rate

Rainfall rate (mm hr$^{-1}$) is estimated by using Drop size distributions from the CIP probe data as

$$R = \frac{\pi}{6} \int_{0}^{\infty} N(D)D^2 u(D) dD$$

(Equation 3.2)

following Rogers and Yau (1989), where $u(D)$ is the fall speed of particle of size $D$. Here, three fall speed formulations are used; One, $u = k_1 r^2$ is used for cloud droplets up to 30 µm radius with $k_1 \approx 1.19 \times 10^6$ cm$^{-1}$ s$^{-1}$. Two, the fall speed in the radius range $0.6 \text{ mm} < r < 2 \text{ mm}$, $u = k_2 r^{1/2}$ is used with $k_2 \approx 2.01 \times 10^3$ cm$^{1/2}$ s$^{-1}$. Three, in the intermediate size range, $40 \mu m < r < 0.6 \text{ mm}$, $u = k_3 r$ with $k_3 \approx 8 \times 10^3$ s$^{-1}$ is used.

### 3.3 Large-scale conditions

General features of the large-scale conditions over the study area are shown by time-height sections of humidity, temperature and winds from the Barbados soundings for the period of 14 March to 16 April, 2010 (Fig. 3.1). During the experiment, the LCL was lower than 1 km (~747 m on average) and the 0 °C isotherm appeared near 5 km (500-600 hPa), indicating that clouds of interest are warm (liquid) clouds. Inversion
height here was defined as the level below 6 km where an increase in the vertical virtual potential temperature is greatest over a 5 m interval. The inversion height increases from about 1.5 km to 3.7 km from 18 March to 25 March and then decreases to a minimum (~1000 m) on 3-4 April as dry air intrudes into the lower atmosphere. After 5 April, the inversion increases and the lower-troposphere stability (LTS) weaken (not shown). The appearance of multiple inversion heights was common during the experiment and the trade inversion is not clearly defined as in stratocumulus cases. The variations in inversion height agree with humidity, temperature and wind patterns: the lower inversions are observed during the dust outbreak, while the relatively higher inversions appear after the event. Overall, inversion heights are observed between 1.5 and 3 km in most cases, but sometimes it reaches 4-5 km when a large-scale disturbance influences the study area.

Large-scale humidity conditions over the study area are illustrated in Fig. 3.1a. A massive dry air intrusion into the layer below 2 km occurs during the periods of 31 March – 5 April and prior to 22 March. Further, an elongated feature of dry air (RH < 10 %) on 30 March, which sampled organized clusters of deeper convection with a strong outflow, is notable. On 5 April, a sharp transition from a dry to a moist zone through the entire lower atmosphere is also distinctive. Temperature fields during the experiment are shown as time-series of potential temperature in Fig. 3.1b. For the periods of pre-dust outbreaks, inversion heights appear in the ranges of potential temperatures between 305 K and 310 K. On the other hand, they are primarily observed in the range of 315-320 K, near the isotherm during the post-dust periods. The sub-cloud layer (below the LCL) is relatively well-mixed, showing a constant $\theta \sim 300$ K.
Figure 3.1: Time-height cross-section of (a) relative humidity (%) (b) potential temperature (K) (c) wind speed (m s\(^{-1}\)) and (d) wind direction (degree) obtained from soundings launched at Grantley Adams airport in Barbados at 12:00 UTC from 14 March to 16 April, 2010. The first (19 March) to the last (11 April) flights are denoted as vertical black solid lines. Periods of strong dust events (31 March - 5 April) are also denoted as vertical dashed lines. Lifting Condensation Level (LCL) and 0 °C isotherm are laid in Fig. 3.1(a) and (b) as black lines connected with circle symbols. The primary and secondary inversion heights are shown as square and cross symbols, respectively. Sounding data were obtained from the University of Wyoming’s online Upper Air Data (http://weather.uwyo.edu/upperair/sounding.html).

Large-scale winds conditions are shown in Fig. 3.1c-d. Easterly winds dominate throughout the atmosphere during the pre-dust outbreak periods (Fig. 3.1d). The height of the easterly winds lowers with the onset of dust outbreaks on 31 March then reaches a minimum height on 3-4 April when the inversion heights are the lowest and the air-mass driest. After 4 April, the region of easterlies ascends and shows a maximum height on 9-
10 April at around 4.5 km when the lower atmosphere is under the coldest (not shown) and the most humid conditions. The axes of the weakest winds are located above the inversion height prior to the dust event (e.g., prior to 31 March). However, the axes of the weakest winds descend and significantly weaker winds are noticeable below the inversion heights during the periods of dust outbreaks (Fig. 3.1c). In addition, strong westerly winds (> 15 m s\(^{-1}\)) are observed during 22-24 March from about 2 km to 5 km.

The overall atmospheric conditions and the variability observed during the flights are shown in Fig. 3.2 with the vertical profiles of potential temperature, mixing ratio and aerosols. Potential temperature (Fig. 3.2a) varies from 298 K to 312 K with poorly defined inversion heights in most flights. However, strong inversion layers are observed in RF #10 (yellow, corresponds to 1 April) and RF #11 (2 April), near 500-600 m and around 1500-1700 m, characterized by potential temperature jumps, and significant reductions in mixing ratios. Further, the overall concentrations are relatively high compared with other days. RF #1 (19 March) also shows a strong inversion near 1700 m, along with a significant reduction in water vapor mixing ratio near the layer. In contrast, profiles form RFs #13, #14, and #15 show monotonic patterns; potential temperature and mixing ratio decrease with height without any significant inversion or dry layers. Accumulation mode aerosol concentrations are relatively low on these days, with concentrations of less than 250 cm\(^{-3}\) below 1000 m, and a decrease with height to values close to 0 cm\(^{-3}\) above 1500 m.

The mean profile of PCASP decreases slightly with height from about 250 cm\(^{-3}\) near the surface, to 100 cm\(^{-3}\) near 2700 m. The structure shown in Fig. 3.2(a-b) indicates that the mean profiles of potential temperature and mixing ratio are reasonably well
mixed in the layer from the surface to about 500 m (sub-cloud layer). Above 500 m, potential temperature increases slightly with height from 300 K to 312 K. Mixing ratio decreases with height ranging from 17 g kg$^{-1}$ to 5 g kg$^{-1}$.

**Figure 3.2**: Profiles of potential temperature, $\Theta$ (left), water vapor mixing ratio (middle) and aerosol concentrations obtained from PCASP (cm$^{-3}$; right) during aircraft’s ascents and/or descents. The mean profiles of each variable are denoted as black dots. Color bar indicates the number of research flights (RF #), shown in Table 2.1. The sounding shown here is arbitrary and is denoted in Table 2.1 inside parenthesis.

### 3.4 Back trajectories

The 10-day backward trajectories of the observed air masses arriving at 500 m in the middle of the flight domain are shown in Fig. 3.3. Air masses within the boundary layer originated mainly from three regimes as suggested by Dunion (2011); air masses arriving at the flight domain during 30 March to 5 April, and on 19 March, which correspond to the periods of dry air intrusion into the lower troposphere (Fig. 3.1a), are traced back to the Africa. On the other hand, air masses arriving at Barbados between on 23 March and 26 March originated from further North (middle latitudes). For the rest of
the flights (e.g., 3/22, 3/29, 4/10, 4/11), air masses arriving at Barbados originated from the North Atlantic and were over the ocean for at least 10 days.

![10 day back-trajectory at 500 m](image)

**Figure 3.3:** The 10-day back trajectories arriving at 500-m in the middle of the BACEX flight domain. Dates for each back-trajectory are shown accordingly.

### 3.5 Aerosol properties

#### 3.5.1 Ragged Point aerosol measurement

The island of Barbados is located on the eastern side of the Caribbean Sea, within the northeast trades of the eastern Atlantic, and thus under the influence of periodic African dust outbreaks throughout the year. Dust concentrations recorded at the Barbados Ragged Point surface site during 2010 are shown in Fig. 3.4. Dust and sea salt surface concentrations over the period of BACEX (3/19-4/11) are shown in lower panel, along with AERONET level 2 Aerosol Optical Depth (AOD) at 550 nm wavelength and Angstrom exponent at wavelength 440-675 nm.
Figure 3.4: Dust concentrations recorded at the Barbados Ragged Point surface site (13.2 °N, 59.5 °W) during 2010. Dust surface concentrations (red) over the period of BACEX (3/19-4/11) are shown in lower panel along with sea salt surface concentrations (blue) in the left axis, level 2 Aerosol Optical Depth (AOD) at 500 nm wavelength (magenta) and Angstrom exponent at wavelengths ranging from 440 to 675 nm (grey) from AERONET in the right axis. Dust and sea salt surface concentration data are provided by Dr. Joseph M. Prospero of the University of Miami.

During the BACEX period (Fig. 3.4b), Ragged Point dust concentrations remained lower than 10 µg m⁻³ for a 9-day period, prior to 29 March, and then rapidly increased to a maximum on 1-2 April with concentrations exceeded 150 µg m⁻³ (44.5 µg m⁻³ on average, ranging from 0.7 µg m⁻³ on 28 March to 155.1 µg m⁻³ on 2 April). Throughout the BACEX period, sea salt surface concentrations ranged from 0.5 µg m⁻³ on 8 April to 72 µg m⁻³ on 11 April (on average, 20.5 µg m⁻³). Aerosol optical depth (AOD) fluctuates around 0.1 during the non-dusty period then increases rapidly from 29
March. For 1 April, the AOD is observed to be about 0.6 at a wavelength of 500 nm. The peaks in AOD at around on 25-26 March are associated with high concentrations of aerosol layers observed aloft during the period (will be shown in Fig. 3.5). Angstrom exponents are less than 0.2 during the dust outbreaks and reach a minimum of 0.1 on 1 April, supporting the prevalence of dust during the period.

### 3.5.2 Vertical and temporal variation

The aerosol distributions observed on 15 research flights (from 3/19 – 4/11) just upwind of Barbados during the Barbados Aerosol Cloud Experiment (BACEX 2010) are shown in Fig. 3.5; aerosols sampled at sub-cloud layer (Fig. 3.5a), and their vertical profiles in the trade-wind boundary layer (Fig. 3.5b). Temporal variations of sub-cloud aerosols during the experiment are shown in Fig. 3.5a. The mean values of CN (black), PCASP (blue) and CCN (activated at super-saturation of 0.6 %, hereafter $s=0.6\ %$, red) with their standard deviations are shown with vertical error bars. Mean value of CCN activated at super-saturation of 0.3 % are added (magenta square symbols).

Overall, CCN concentrations follow PCASP patterns reasonably well. Aerosol concentrations show a constant increasing trend after 29 March and peak around 1-5 April for CN, PCASP and CCN observations, consistent with the trend of dust surface concentrations in Fig. 3.4b. High aerosol concentrations on 23 March in the sub-cloud layer are notable, and the aerosol traces back to the U.S. continent (see Fig. 3.3). The mean (black solid line) and individual (colored lines) profiles of PCASP aerosol concentration are offset by 400 cm$^{-3}$ intervals for each flight (Fig. 3.5b) with each vertical dotted line representing a new axis to indicate aerosol concentrations for the day in
question. For example, PCASP concentrations on 29 March are nearly constant (~ 200 cm$^{-3}$) below 1000 m, and gradually decrease with height reaching ~ 0 cm$^{-3}$ around 2000 m; PCASP concentrations on 5 April are nearly constant (~ 300 cm$^{-3}$) below 600 m, and gradually increase with height and peak (~ 600 cm$^{-3}$) around 2000 m, then decrease with height reaching ~ 200 cm$^{-3}$ around 2300 m. PCASP aerosol concentrations on 23 March were not available, thus CCN activated at a super-saturation of 0.3 % were overlaid in Fig. 3.5b to give a sense of the vertical structure of aerosol concentration on the day.

**Figure 3.5:** (a) Temporal variation of aerosols at sub-cloud layer and (b) vertical distribution of aerosol obtained from aircraft. CCN ($s=0.3\%$) is overlaid on 23 March for vertical profiles since no PCASP is available on the day. Sub-cloud aerosol concentrations (Fig. 3.5a) are obtained from the level leg run at sub-cloud layer, and vertical profiles of PCASP aerosol concentrations (Fig. 3.5b) were obtained during aircraft’s ascents and/or descents.
First, the variety of vertical structures is of interest; aerosol concentrations decrease as heights increase on 22, 29, 30 March, 7, 10, 11 April with a maximum in the sub-cloud layer. The observations on 22 and 29 March and 10 and 11 April show pristine atmospheric conditions with maximum CCN concentrations of less than 200 cm\(^{-3}\) in the sub-cloud layer and close to 0 cm\(^{-3}\) above the inversion (see Fig. 3.1 for the inversion heights). These air masses appear to originate from over the Atlantic (Fig. 3.3). March 30 and 7 April correspond to transit periods from pristine to dusty, and dusty to pristine, respectively. There are a couple of days when high aerosol concentrations are observed above the inversion (e.g., 25 and 26 March) and low dust concentrations are recorded at the surface (Fig. 3.4b). However, both produce high valued of AOD in Fig. 3.4b, as AOD reflects the column integrated value. Air masses on these days (25-26 March) originate from middle latitudes (Fig. 3.3). During the period between 31 March and 2 April, significant aerosol variations occur through the entire layer below 3 km. PCASP peaks shown near 300-400 m on 30 March in Fig. 3.5b, are affected by strong downdrafts from the outflow of deep cloud clusters and are consistent with dry air descent (Fig. 3.1a), implying deep cloud outflows brought aerosols down to the sub-cloud layer.

### 3.5.3 Aerosol particle size distribution

Daily flight-averaged (or column-averaged) aerosol particle size distributions are obtained from the PCASP when no liquid water is detected (Fig. 3.6). Particle Size Distributions (PSDs) on RF02 (3/22; deep blue dashed line) show the lowest aerosol concentrations in the range of D < 1.7 µm. On the other hand, PSDs on RF10 (4/1; yellow dashed line) show the largest aerosol concentrations in the range of D > 0.4 µm.
The envelope between the largest and the lowest aerosol concentrations covers one order of magnitude, with the largest difference at around $D=8 \, \mu m$. PSDs on RF02 (3/22), RF07 (3/29), RF14 (4/10), and RF15 (4/11) show lower aerosol concentrations than those on other days and are consistent with clean conditions (PCASP < 200-300 cm$^{-3}$) in Fig. 3.5.

**Figure 3.6:** Daily averaged aerosol particle size distributions (PSDs) ranging from 0.1 µm to 2.5 µm obtained from the PCASP. Color bar indicates the research flight number (RF #), shown in Table 2.1. PSDs from the odd (even) RF numbers are shown as solid (dashed) lines. PSDs estimated between RF07 and RF10 (3/29, 3/30, 3/31, 4/1) are denoted as bold lines. PSD of RF01 (19 March) is not shown due to instrument malfunction on that day.

In contrast, PSDs between RF09 and RF 12 (3/31-4/5) show high $dN/d(\log D)$, and are consistent with polluted conditions shown in Fig. 3.5. PSDs obtained from RF07 (3/29; green bold-solid line) and RF08 (3/30; green bold-dashed line) have similar concentrations for the smaller size of aerosol (e.g., $D < 0.25 \, \mu m$), but the difference gets larger, with more abundant larger particles observed on 30 March. These two days (3/29, 3/30) have similar vertical structures in aerosol concentrations in Fig. 3.5b, with higher
concentrations on 30 March. On 31 March (RF09), when the African dust prevails throughout the boundary layer, PSDs over all sizes are increased compared with those from one day before (RF08, 3/30); and \(dN/d\log D\) peaks on 2 April. It is remarkable that the shapes of size distribution are so consistent from one day to the next, in spite of the huge difference in total mass loading, and origin of the aerosol. Aerosol on 25-26 March was confined above the inversion, and thus, PSDs reflect the lack of larger particles and the existence of abundant fine particles on the days.

To extend the information on the particle size distribution to the larger sizes, sub-cloud aerosol particle size distributions are estimated by combining data from PCASP with CAS sampled for the sub-cloud level flights (Fig. 3.7).

**Figure 3.7:** Daily averaged aerosol particle size distributions (PSDs) for the sub-cloud level flights. PSDs obtained from PCASP and CAS probes are combined to attain PSDs ranging from 0.1 µm to 30 µm. Color bar indicates the research flight number (RF #), shown in Table 2.1. PSDs from the odd (even) number of RF are shown as solid (dashed) lines. PSDs between RF07 (29 March) and RF10 (1 April) are denoted as bold lines. PSD of RF01 (19 March) and RF03 (23 March) are not shown due to the instrument malfunction (RF01) and the absence of PCASP data (RF03) for the days. The scale of Fig. 3.6 is shown as a box (dotted) in Fig. 3.7 in upper-left corner.
Two populations of PSDs are distinct from the sub-cloud level flights; PSDs from dusty days (flights on RF09, RF10, RF11 and RF12) have significantly more number and volume between 0.5 and 10 µm than the other sub-cloud level flights. For RF08 (30 March), a column-averaged PSD appears between clean and polluted conditions when the entire lower atmosphere is considered (a column-averaged PSD in Fig. 3.6), while aerosol concentrations in the sub-cloud layer (Fig. 3.7) show values as high as those of the African dust outbreaks. This is because a strong downdraft from the outflow of deep cloud cluster on the day brought aerosol down to the sub-cloud layer as shown in Fig. 3.5b near 300-400 m. The other PSDs are associated with non-African dust periods and show relatively low aerosol concentrations over all ranges of sizes.

The number (volume) concentrations of aerosol, for the African dust periods (RF09-RF12), decrease (increase) monotonically in the range of 0.8-10 µm. For the regimes of large particles (e.g., D > 10 µm), concentrations decrease rapidly in both number and volume, and are consistent with PSDs of non-dusty periods. PSDs from non-dusty periods show maxima near 7-10 µm and 10-20 µm in both number and volume, indicating the possible existence of giant nuclei during non-dusty periods in the sub-cloud layer. The peak near 2-3 µm, especially for the non-dusty days, is due to the discrepancy between PCASP and CAS in the transit regime. The PCASP (0.1 - 2.5 µm) dries the particles before measuring them while CAS (0.6 - 60 µm) sizes them in ambient conditions; thus, there can be discrepancies between the two values especially when relative humidity is high. Nevertheless, two size distributions are well agreed at the interface. The spread of aerosol PSD in the largest particle size is about 1-order of
magnitude; for example, it ranges from $10^2$ in volume number concentrations ($dV/d\log D$) on 22 and 24 March to $10^1$ in $dV/d\log D$ on 29 March and 1 April.

### 3.6 Cloud and precipitation properties

On each day during BACEX, TO flights were made about 50-100 km upwind of Ragged Point and included several consistent level legs, from near the surface to the trade-wind inversion height; e.g., levels of (1) below the cloud (sub-cloud level (leg) flight), (2) near the cloud base (cloud-base level (leg) flight), (3) near the cloud-top (cloud-top level (leg) flight). During the experiment, small cumulus clouds were observed on most of the days, but relatively deep cloud clusters (heights to about 2.5-3 km) were sampled on only a couple of days (e.g., 22 March, 24 March, and 30 March), and they show different characteristics from the small cumulus clouds.

**Table 3.1:** Cloud-base level leg used in radar analysis for cloud statistics.

<table>
<thead>
<tr>
<th>RF #</th>
<th>Date</th>
<th>Time (UTC*)</th>
<th>Flight height</th>
<th>Note</th>
</tr>
</thead>
<tbody>
<tr>
<td>RF01</td>
<td>19 Mar</td>
<td>-</td>
<td>-</td>
<td>-</td>
</tr>
<tr>
<td>RF02</td>
<td>22 Mar</td>
<td>15:52:48-16:10:48</td>
<td>1035 m</td>
<td>-</td>
</tr>
<tr>
<td>RF03</td>
<td>23 Mar</td>
<td>17:04:48-17:24:36</td>
<td>1065 m</td>
<td></td>
</tr>
<tr>
<td>RF04</td>
<td>24 Mar</td>
<td>17:01:48-17:24:36</td>
<td>525 m</td>
<td>Below anvil cloud</td>
</tr>
<tr>
<td>RF05</td>
<td>25 Mar</td>
<td>15:31:12-16:03:00</td>
<td>795 m</td>
<td>Non-precipitating clouds</td>
</tr>
<tr>
<td>RF06</td>
<td>26 Mar</td>
<td>15:27:54-15:36:00</td>
<td>1005 m</td>
<td>Non-precipitating clouds</td>
</tr>
<tr>
<td>RF07</td>
<td>29 Mar</td>
<td>17:06:00-17:18:36</td>
<td>885 m</td>
<td>Non-precipitating clouds</td>
</tr>
<tr>
<td>RF08</td>
<td>30 Mar</td>
<td>17:36:00-17:49:48</td>
<td>405 m</td>
<td>sub-cloud leg</td>
</tr>
<tr>
<td>RF09</td>
<td>31 Mar</td>
<td>16:53:24-17°07:48</td>
<td>705 m</td>
<td>Non-precipitating clouds</td>
</tr>
<tr>
<td>RF10</td>
<td>1 Apr</td>
<td>-</td>
<td>-</td>
<td>No clouds observed</td>
</tr>
<tr>
<td>RF11</td>
<td>2 Apr</td>
<td>-</td>
<td>-</td>
<td>No descent clouds are observed</td>
</tr>
<tr>
<td>RF12</td>
<td>5 Apr</td>
<td>16:09:00-16:24:36</td>
<td>825 m</td>
<td></td>
</tr>
<tr>
<td>RF13</td>
<td>7 Apr</td>
<td>16:32:46-16:43:12</td>
<td>735 m</td>
<td>-</td>
</tr>
<tr>
<td>RF14</td>
<td>10 Apr</td>
<td>16:19:30-16:28:12</td>
<td>1005 m</td>
<td>-</td>
</tr>
<tr>
<td>RF15</td>
<td>11 Apr</td>
<td>16:02:24-16:21:00</td>
<td>795 m</td>
<td>Non-precipitating clouds</td>
</tr>
</tbody>
</table>

*Local time: UTC-5*
To obtain a general sense of the characteristics of small oceanic cumulus cloud, such as distributions of cloud tops and bases, cloud fields sampled from the cloud radar during the cloud-base level flights are used. For 30 March, clouds sampled from the sub-cloud level flights are used for the radar analysis, since this day does not have a sufficiently long cloud-base level leg to use for the analysis. Date, time periods, average heights of flights used for this cloud-base flight for the radar analysis are summarized in Table 3.1, and examples of time-height cross-sections of radar reflectivity $Z$ for the 5-minute periods are shown in Fig. 3.8.

Figure 3.8: Time-height cross section of reflectivity on (a) 22 March, (b) 24 March, (c) 29 March and (d) 11 April, 2010 from the cloud-base level flight during 5-minute periods (about 18 km in horizontal extent) at an air speed of about 60 m s$^{-1}$. Data were sampled from (a-b) precipitating and (c-d) non-precipitating clouds.
Clouds on 22 and 24 March (Figs. 3.8a-b), are precipitating clouds characterized by organized features, significant sizes (horizontal and vertical) and strong reflectivity. On 29 March and 10 April (Figs 3.8c-d), typical marine shallow cumulus clouds are sampled, and reflectivity ranges from -40 dBz to -20 dBz, significantly weaker compared with those of precipitating clouds (Fig. 3.8a-b). The horizontal widths of the precipitating clouds are approximately 4-7 km (1-2 min) and the thickness of the cloud is about 1-2 km. On the other hand, the horizontal widths of the non-precipitating clouds (Fig. 3.8c-d) are less than 1 km and cloud depths are less than 500 m.

Precipitating clouds, compared with non-precipitating clouds, are characterized by organized features. Satellite imageries taken on 22, 24, 30 and 29 March are shown in Fig. 3.9, as examples of showing the difference of cloud organizations between precipitating and non-precipitating clouds, respectively. Images were downloaded from website and were modified (http://modis-atmos.gsfc.nasa.gov/IMAGES/index.html). The clouds on 22 March are comprised of fairly strong convection surrounded by lots of cloudiness, which is formed by outflow from the convective clouds. The appearance of clouds organized around the arc-shaped outflow boundaries from the (earlier) convection is also evident on 24 and 30 March in Fig. 3.9b and Fig. 3.9d. Convection associated with these features often reached cloud heights of about 2-3 km (e.g., Fig. 3.8a-b).

On the other hand, a typical fair weather cumulus clouds were sampled during the flight on 29 March (Fig. 3.9c). The size of clouds is significantly smaller than that observed from precipitating clouds in Fig. 3.9(a, b, d), and the clouds observed within the flight domain do not have neither well-organized convection nor outflow features. The
shallow convection here with no measurable precipitation showed unorganized feature and commonly showed cloud thickness of less than 500 m.

Figure 3.9: MODIS satellite images on (a) 22 March (b) 24 March (c) 29 March and (d) 30 March, 2010 over the ocean near Barbados. The flight domains are shown as red dotted boxes. The outer box indicates an average flight domain during BACEX. The flight domain of the particular day is overlaid as an inner box, if satellite image is obtained during the flight periods. The numerical number shown at the lower-right side of the figure indicates Julian day with UTC (e.g., 088.1730 indicates Julian day 088, 1730 UTC). Images were downloaded from the MODIS website (http://modis-atmos.gsfc.nasa.gov/IMAGES/index.html).
What ranges of radar reflectivity and vertical velocities are most commonly observed in these clouds? To address this question, the frequency distributions of radar reflectivity versus Doppler velocity are shown in Fig. 3.10. Radar reflectivity and velocity are estimated from the cloud radar during the cloud-base flight on the individual day. Here, upward motion (positive Doppler velocity) indicates updrafts. Note that the droplets’ vertical velocity obtained by a vertically pointing radar beam is the sum of the terminal velocity of the hydrometeor (e.g., raindrop) and the vertical air speed (updrafts and downdrafts). Upward or downward air velocity is embedded in downward motion in the Doppler velocity measurements. In contrast, upward velocities indicate updrafts in the Doppler velocity measurements.

**Figure 3.10:** Normalized velocity-reflectivity number frequency distribution on each day during BACEX. Intervals of 2-dBz, and 0.1 ms\(^{-1}\) are used to obtain the frequency distribution. Color bar is shown in upper right corner. Reflectivity of -20 dBz and Doppler velocity of 0 m s\(^{-1}\) are denoted by dotted line. No clouds are observed on 1-2 April during the cloud-base level flights. Time and periods of each cloud-base flights are listed in Table 3.1.
Clouds sampled on 25 March, 26 March, 29 March, 31 March, 10 April and 11 April show similar patterns in reflectivity and velocity; frequency distributions are horizontally oriented on the velocity-reflectivity diagram, with a wide range of Doppler velocities (-5 ~ 3 ms\(^{-1}\)) and narrow range of reflectivity (less than about -20 dBz). In contrast, clouds sampled on 22 March, 24 March, 30 March and 7 April show similar patterns; data is vertically oriented on the velocity-reflectivity diagram with wider range of reflectivity and relatively narrow range of negative Doppler velocities. Finally, clouds sampled on 23 March and 5 April show mix of both types of distributions. Clouds with exclusively weak radar reflectivity (e.g., clouds on 3/25, 3/26, 3/29, 3/31, 4/11) are associated with non-precipitating clouds, while clouds with strong reflectivity with negative Doppler velocity (e.g., clouds on 3/22) are mainly associated with precipitation. Clouds sampled on other days (e.g., 3/23, 3/24, 4/5, 4/7, 4/10) are characterized by light precipitation. Hereafter, these clouds are referred to as “non-precipitating clouds”. On three days in Fig. 3.10 (3/22, 3/24, and 3/30), the maximum frequency of reflectivity and Doppler velocity occurs in the range of \( Z > -20 \) dBz and velocity \(< 0 \) m s\(^{-1}\), and these three clouds are clearly precipitating. Hereafter, clouds sampled on these three days are referred to as “precipitating clouds”.

The composite of all the reflectivity and velocity during BACEX obtained from cloud-base level flights are shown in Fig. 3.11a. Clouds with reflectivity of -20 dBz to +5 dBz and Doppler velocity of -2 ms\(^{-1}\) to +1 ms\(^{-1}\) are sampled most frequently (number of samples > 300). However, the reflectivity and velocity sampled from the three precipitating clouds shown in Fig. 3.11b show similar structures to those sampled over the entire BACEX period (Fig. 3.11a), indicating that data in Fig. 3.11a are strongly
influenced by the precipitating clouds observed on these three flights. The reflectivity-velocity distribution estimated from excluding the strongest precipitating cloud sampled on 22 March (Fig. 3.11c) shows two populations of reflectivity and velocity: (1) reflectivity weaker than approximately -30 dBz with Doppler velocity ranging from -4 ms\(^{-1}\) to +3 ms\(^{-1}\) (horizontally oriented pattern in Fig. 3.11c, with a narrow range of reflectivity and broad range of velocity); (2) reflectivity ranging from -30 dBz to 5 dBz and Doppler velocity ranging from -2 ms\(^{-1}\) ~ -4 ms\(^{-1}\) (vertically oriented pattern in Fig. 3.11c with a broad range of reflectivity and a narrow range of velocity).

**Figure 3.11:** Reflectivity and velocity distributions estimated from an average of all individual days (12 cases in Fig. 3.10), (b) using three precipitating clouds (clouds sampled on 3/22, 3/24, and 3/30), (c) using 11 days except for clouds on 22 March, which sampled the strongest precipitating clouds, and (d) from non-precipitating and/or lightly precipitating clouds (remaining 9 days in Fig. 3.10).
The distribution of reflectivity and velocity that excludes the three precipitating cloud cases (Fig. 3.11d), highlights in the regimes of narrow range of reflectivity with broad velocity ranges (horizontally oriented pattern in Fig. 3.11d), indicating the regimes are associated with non-precipitating clouds. It also shows the other regime of reflectivity and velocity (vertically elongated pattern in velocity-Z diagram) that indicates the presence of (light) precipitation in these non-precipitating dominating shallow cumulus clouds. The cloud thicknesses are about 1300 m (Fig. 3.12c-d).

**Figure 3.12:** Frequency distribution of reflectivity and velocity with heights, by composite (a-b) all individual days (12 cases in Fig. 3.10), and (c-d) 9 days excluding three major precipitating clouds events sampled on 22, 24, and 30 March, 2010. Intervals of 30-m (vertical), 2-dBz, and 0.1 ms\(^{-1}\) are used to obtain the frequency distribution.
The vertical distributions of the clouds (Z and Doppler velocity) are shown in Fig. 3.12. The frequency distributions are estimated from cloud fields obtained from cloud radar during the cloud-base level flights with 1-s and 30-m resolutions. The vertical distribution of Z and Doppler velocity is estimated from using all the data are shown in Fig. 3.12a-b, while those estimated exclusively from non-precipitating and lightly precipitating clouds are shown in Fig. 3.12c-d to show the vertical distribution of clouds.

Two dominant populations of reflectivity are found from the vertical distribution of reflectivity, estimated using all the data shown in Fig. 3.12a: (1) at ~ -35 dBz and ~ 1 km (number of samples ~ 200); (2) -20 dBz < Z < 5 dBz in a height range of about 1000 to 2300 m. For the velocity distribution composited from all available data (Fig. 3.12b), Doppler velocities of about -2 ms\(^{-1}\) to +1 ms\(^{-1}\) are most frequently sampled from 1000 to 2300 m. All clouds (precipitating, non-precipitating, and lightly precipitating clouds) sampled during the BACEX show the lowest cloud bases and highest tops at about 400 m and 2700 m, respectively, indicating a maximum depth of the clouds of about 2300 m. To examine vertical distributions of non-precipitating clouds during BACEX, reflectivity and velocity frequency distributions of the 9 days that exclude the three precipitating clouds (3/22, 3/24, and 3/30) are shown in Fig. 3.12(c-d). During BACEX, reflectivity of ~ -35 dBz and Doppler velocities of ± 2 ms\(^{-1}\) (bimodal) between 600 m and 1300 m are the most frequently observed in these non-precipitating and/or lightly precipitating clouds. Cloud bases and tops for the non-precipitating clouds are about 700 m and 2000 m, respectively, indicating a cloud thickness of about 1300 m.

To examine the differences in vertical sampling statistics between precipitating (e.g., Fig. 3.11b) and non-precipitating clouds (e.g., Fig. 3.11d), the normalized number
of data points with height are shown in Fig. 3.13. Clouds sampled by cloud radar (e.g., Fig. 3.8) during the cloud-base level flights are used. The number of data points is summed up within 30-m vertical intervals from cloud-base to 3 km at 1-s resolution, then it is divided by the maximum number of data points observed on the day to obtain the normalized number of samples shown on the x-axis for the 12 cases studied. We assume that a precipitating point is a data point with $Z > -20$ dBz and Doppler velocity $< 0$ ms$^{-1}$ (downward motion). The data is averaged in 100 m vertical intervals to smooth out small variations.

Figure 3.13: Normalized number of samples (divided by the number of data sampled) with height for all sampled clouds (gray) and for precipitating clouds (black). Precipitating clouds are defined as data points with $Z > -20$ dBz and vertical velocity $< 0$ m s$^{-1}$. No precipitating clouds are observed on 25, 26, 29, 31 March and 11 April.
Two types of precipitation features are illustrated in Fig. 3.13. In one, precipitation shafts are observed mainly close to the cloud base (e.g., 3/22 and 3/30) with evaporation in the sub-cloud layer, especially when the clouds are deeper than the other lightly precipitating clouds. In the other, precipitation shafts emanate mainly near cloud top (e.g., 4/5) on the downshear side of the cloud (not shown) and evaporate in the cloud layer. This type of precipitating cloud is shallower than the first type of cloud, and is also sometimes accompanied by precipitation shafts emanating near cloud base (e.g., 3/23, 4/5, 4/7). Precipitation embedded in the updrafts is not detected by Doppler threshold adopted here (vertical velocity < 0 m s$^{-1}$). Precipitation features that were observed during BACEX without an updraft threshold are addressed in Fig. B1 in the appendix.

**Figure 3.14:** The individual cloud is identified by threshold of Gerber Liquid water content greater than 0.02 g m$^{-3}$ and duration of data recording (lasts longer than 3s). Precipitating cloud is classified by thresholds of precipitation liquid water content (CIP volume × density of water) great than 0.1 g m$^{-3}$.

During the flights, onboard scientists frequently observed precipitation on the window of the aircraft. Can we quantify how much and how frequently these clouds precipitate? To answer this, the daily percentage of precipitating clouds among the total
number of clouds is shown in Fig. 3.15. Rainfall rate (mm hr\(^{-1}\)) is calculated using Eq. (3.2) with CIP probe data. The percentage of precipitating clouds for a given day is estimated by the ratio of precipitating clouds to the total number of clouds sampled. A cloud is counted only if the Gerber liquid water content > 0.02 g m\(^{-3}\), and the data sample lasts longer than 3 seconds (~ 180 m width of clouds) to avoid counting small patches of clouds.

Figure 3.15: (a) Percentage of precipitating clouds estimated from all clouds sampled. Precipitating clouds are defined as clouds with precipitation liquid water content (PLWC) larger than 0.1 g m\(^{-3}\) (light-grey solid) and 1 g m\(^{-3}\) (medium-grey dashed). The CIP probe volume concentration (cm\(^{3}\) m\(^{-3}\)) is multiplied by density of water to obtain PLWC. (b) Percentage of precipitating clouds estimated from all clouds sampled (grey; column-averaged), and from clouds sampled during the cloud-base flights (black; cloud-base) with a threshold of PLWC > 0.1 g m\(^{-3}\). (c) Flight-averaged (grey) and cloud-base (black) precipitation rate (mm day\(^{-1}\)).
The cloud is classified as precipitating if the sum of the precipitation liquid water content, PLWC (CIP volume × density of water) for a given cloud is larger than 0.1 g m\(^{-3}\) (light-gray dashed). However the choice of the threshold is arbitrary and the percentage of precipitating clouds with PLWC larger than 1 g m\(^{-3}\) is overlaid in Fig. 3.15a (gray-dashed line).

The total number of clouds sampled on each day ranges from 50 to 200 (not shown here). However, the actual number of clouds occurring during the day can easily be less or more than the number of clouds sampled, since the aircraft sometimes penetrates the same cloud more than once (in this case, we may count the same cloud more than once through the multi-penetrations of the clouds) or avoids penetrating clouds due to strong updrafts or downdrafts (in this case, we sample a subset of clouds out there). Nevertheless, Fig. 3.15a shows that approximately 56 % of sampled clouds precipitate somewhere in the clouds when PLWC > 0.1 g m\(^{-3}\) is used, and thus about 44 % of clouds are purely non-precipitating clouds. Further, 41 % of sampled clouds are precipitating and 59 % of clouds are purely non-precipitating clouds, if PLWC > 1 g m\(^{-3}\) is used. It indicates that approximately 40-60 % of the clouds precipitate somewhere in the clouds and it is consistent with the percentage of the non-precipitating clouds in Fig. 3.13; no precipitating clouds were observed on five of the 12 flights (~ 42 % of clouds sampled are purely non-precipitating clouds). In Fig. 3.15b, the percentage of precipitating clouds sampled exclusively during the cloud-base flights (black) shows lower values compared with those of sampled during all of the flights, indicating a dominant mode of precipitation shafts emanating from the cloud top. Although more than about half of the clouds precipitate in Fig. 3.15a (PLWC > 0.1 g m\(^{-3}\)), rainfall rates in and around the cloud
during the BACEX (Fig. 3.15c) are far less than 10 mm day\(^{-1}\), except on 22 March (2.7 mm day\(^{-1}\), on average). Cloud-base rainfall rates on RF02 (3/22) are larger than those estimated from all the other flights. Rainfall rates on RF04 (3/24), RF08 (30 March) and RF13 (4/7) are similar. For the rest of the days, rainfall rates estimated from above cloud-base (gray) are larger than those at cloud-base.

There was no precipitation recorded at the surface weather station in Barbados during the campaign (no rain or trace recorded, http://www.wunderground.com/global/BR.html). The average precipitable water during BACEX, based on the soundings (Grantly Adams airport), was 4.1 cm (not shown) that about the same as observed in the Rain In shallow Cumulus over the Ocean (RICO) field campaign (Rauber et al., 2007).

Characteristics of cloud cores were defined using 10-m averaged liquid water mixing ratio and cloud droplet number concentrations \(N_d\) estimated from all data sampled during BACEX (Fig. 3.16). A cloud core here is defined by regimes of updrafts larger than 1 m s\(^{-1}\). An adiabatic liquid water mixing ratio \((w_i = \Gamma_i (z - z_B);\) Albrecht, 1990) is overlaid on Fig. 3.16a. Here, \(z\) indicates height; \(z_B\) is the height of cloud-base; \(\Gamma_i = -\frac{dw_i}{dz}\), and \(w_s\) is saturation mixing ratio. The 10-m vertically averaged liquid water mixing ratio for non-precipitating clouds (Fig. 3.16a) and \(N_d\) (Fig. 3.16b) are estimated from data with \(w > 1\) m s\(^{-1}\), Gerber LWC > 0.01 g m\(^{-3}\), and CIP volume number concentrations less than 0.01 cm\(^3\) cm\(^{-3}\) (PLWC = 10 g m\(^{-3}\)), since shattering of large drops can contaminate the measurements of \(N_d\) and large drops also tend to precipitate. Overall, clouds sampled below 1-km are mainly non-precipitating and close to adiabatic, while clouds observed above 2.2 km are mostly precipitating (Fig. 3.16a). Further, even the cloud cores are far
from adiabatic, in particular with increasing distance above cloud base (in good agreement with RICO; e.g., Rauber et al., 2007; Gerber et al., 2008, and continental cumulus; Lu et al., 2008). Cloud droplet number concentrations (Fig. 3.16b) vary from near 0 to 400 cm⁻³. The maximum \( N_d \) is at ~ 1200 m and clusters of cloud droplets are found near the mean LCL. Overall, \( N_d \) increases with height. The low \( N_d \) at high altitude (~ 2300 m) may be associated with entrainment mixing and wet scavenging.

**Figure 3.16:** (a) Cloud water and (b) droplet number concentration \( N_d \) in cloud core \((w > 1 \text{ m s}^{-1})\) sampled by the Twin Otter during BACEX. Non-precipitating samples (CIP volume < 0.01) are used to estimate \( N_d \) in Fig. 3.16b Mean (minimum and maximum) values of LCL are denoted by denoted as dashed (dotted) lines.

### 3.7 Representative of thermodynamic structures

How representative are the BACEX thermodynamic structures compared with those obtained from field campaigns in the cumulus regimes? To answer this, vertical profiles of potential temperature and mixing ratio obtained from BACEX, ATEX,
BOMEX and RICO field campaigns are compared. All data sampled (soundings + level leg flights) are used to attain the mean profiles of potential temperature and mixing ratio during BACEX. The data is averaged at every 20-m vertical interval first, and carried out 9-points moving average (87.5 m resolution) to smooth out small variations. Potential temperature and mixing ratio for RICO, ATEX, and BOMEX are obtained from GCSS (GEWEX Cloud System Study) boundary layer cloud homepage (http://www.knmi.nl/~siebesma/BLCWG/). The following equations and/or values of $q_t$ and $\theta_l$ are used for BOMEX, RICO and ATEX. Initially, it is assumed that there is no liquid water (i.e., $q_l = 0.0$), so that $\theta = \theta_l$ and $r_v = r_t$.

$q_t$ (g/kg):  
$$
\begin{align*}
0 < z < 520: & \quad 17.0 + (16.3 - 17.0) / (520) \times z \\
520 < z < 1480: & \quad 16.3 + (10.7 - 16.3) / (1480 - 520) \times (z - 520) \\
1480 < z < 2000: & \quad 10.7 + (4.2 - 10.7) / (2000 - 1480) \times (z - 1480) \\
z > 2000: & \quad 4.2 - 1.2E-3 \times (z - 2000)
\end{align*}
$$

$\theta_l$ (K):  
$$
\begin{align*}
0 < z < 520: & \quad 298.7 \\
520 < z < 1480: & \quad 298.7 + (302.4 - 298.7) / (1480 - 520) \times (z - 520) \\
1480 < z < 2000: & \quad 302.4 + (308.2 - 302.4) / (2000 - 1480) \times (z - 1480) \\
z > 2000: & \quad 308.2 + 3.65E-3 \times (z - 2000)
\end{align*}
$$

BOMEX (source: http://www.knmi.nl/~siebesma/gcss/bomexcomp.init.html)
ATEX (source: Table 1 from Stevens et al., 2000)

<table>
<thead>
<tr>
<th>z (m)</th>
<th>Θ (K)</th>
<th>Q (g kg⁻¹)</th>
<th>U₉ (m s⁻¹)</th>
<th>V₉ (m s⁻¹)</th>
</tr>
</thead>
<tbody>
<tr>
<td>0</td>
<td>295.750</td>
<td>13.00</td>
<td>-11.00</td>
<td>-2.00</td>
</tr>
<tr>
<td>150</td>
<td>295.750</td>
<td>12.50</td>
<td>-10.55</td>
<td>-1.90</td>
</tr>
<tr>
<td>700</td>
<td>295.750</td>
<td>12.50</td>
<td>-8.90</td>
<td>-1.10</td>
</tr>
<tr>
<td>750</td>
<td>296.125</td>
<td>11.50</td>
<td>-8.75</td>
<td>-1.00</td>
</tr>
<tr>
<td>1400</td>
<td>297.750</td>
<td>10.25</td>
<td>-6.80</td>
<td>-0.14</td>
</tr>
<tr>
<td>1650</td>
<td>306.750</td>
<td>4.50</td>
<td>-5.75</td>
<td>0.18</td>
</tr>
<tr>
<td>4000</td>
<td>314.975</td>
<td>4.50</td>
<td>1.00</td>
<td>2.75</td>
</tr>
</tbody>
</table>

$q_t$ (g/kg) :  

(Equation 3.5)

$0 < z < 740$: $16.0 + (13.8 – 16.0) / (740) \times z$

$740 < z < 3260$: $13.8 + (2.4 – 13.8) / (3260 - 740) \times (z - 740)$

$z > 3260$: $2.4 + (1.8 – 2.4) / (4000 - 3260) \times (z - 3260)$

$\theta_t$ (K) :

(Equation 3.6)

$0 < z < 740$: 297.9

$z > 740$: $297.9 + (317.0 – 297.9) / (4000 - 740) \times (z - 740)$

RICO (source: http://www.knmi.nl/samenw/rico/setup3d.html)

The profiles of potential temperature and mixing ratio of BACEX, ATEX, BOMEX and RICO field campaigns are shown in Fig. 3.17. Overall, the atmospheric condition during BACEX (March-April, 2010) shows warmer and moistener condition among others (ATEX: February 1969; BOMEX: May and July 1969; RICO: November 2004 – January 2005). BOMEX shows similar moisture conditions below inversion (~1500 m) but drier than BACEX by about 5 g kg⁻¹, above inversion. Potential temperature
during BOMEX is about 1 K cooler (warmer) than BACEX below (above) inversion. Potential temperature of BOMEX falls in the edge of ±1σ of θ of BACEX. RICO profile shows cooler and drier atmospheric conditions constantly throughout the entire boundary layer compared with those in BACEX; about 2 K cooler and 2 g kg⁻¹ drier (except between 1000 – 1300 m) than BACEX. ATEX profiles show the driest and coldest conditions among others. Potential temperature is about 4 K cooler than BACEX below inversion and about 1 K warmer (within the +1σ of θ ) above inversion (~ 1500 m). In addition, the atmospheric conditions during ATEX was about 4 g kg⁻¹ drier than those during BACEX.

Figure 3.17: Profiles of (a) potential temperature, Θ, water vapor mixing ratio obtained from BACEX (black) with ±1σ (grey), BOMEX (blue), RICO (red) and ATEX (magenta) field campaigns. Data of BOMEX, RICO, and ATEX are obtained from GCSS (GEWEX Cloud System Study) boundary layer cloud homepage. BACEX profiles are obtained from all data sampled during the experiment.
3.8 Discussion and summary

In this chapter, I have examined the variations of aerosol, cloud property over the Eastern Caribbean during Barbados Aerosol Cloud Experiment (BACEX), which took place from March 19 to April 11 2010 off the Caribbean island of Barbados. The temporal variations and vertical distributions of aerosols observed on the 15 flights made with the TO research aircraft show a wide range of aerosol conditions that include the most intense African dust outbreak (1-2 April) observed at the Barbados surface site during all of 2010. The intense African dust events concurred with a massive dry air intrusion into the lower atmosphere and very weak easterlies (Fig. 3.1). During the dust outbreak periods, clouds and precipitation were significantly suppressed. During the entire period of BACEX, the LCL was below 1 km (747 m on average, ranging from 440 to 955 m) and the 0 °C isotherm appears near 5 km (500-600 hPa) indicating that clouds of interest are warm clouds. Inversion heights are usually between 1.5-3 km, except when the large-scale disturbance dominates the system, and it reaches 4-5 km.

The 10-day backward trajectories of air masses arriving at 500 m in the middle of the flight domain show that three distinctive air masses dominate over the Eastern Caribbean (e.g., typical maritime air mass, Saharan Air mass, Middle latitude dry air, as shown in Dunion, 2011).

A variety of aerosol vertical structures is observed (Figs. 3.2c and 3.5b) and categorized by three distinct profiles: 1) aerosol concentrations decrease steadily with height, with its maximum concentrations below the trade-wind inversion < 250 cm\(^{-3}\), and are associated with a typical maritime air mass. 2) The second type of profile shows aerosol concentrations increasing with height, with a maximum concentration above the
inversion height; this profile is associated with air mass originated from middle latitude. 3) The third type of aerosol profile is associated with African dust events (e.g., 31 March to 5 April). High concentrations of aerosols are observed throughout the entire boundary layer with stratified aerosol structures.

The average precipitable water during BACEX was 4.1 cm that about the same as observed in the Rain In shallow Cumulus over the Ocean (RICO) field campaign (Rauber et al., 2007). The environment was dry and thus the precipitation did not reach surface in most days of the experiment (There was no precipitation detected at the surface weather station in Barbados during the campaign). The TO research aircraft was able to sample many clouds in various phases of growth during the BACEX. Data obtained from the vertically pointing cloud radar, mounted on top of the TO research aircraft, from the cloud-base level flight, provided the basis for the general characteristics of clouds sampled.

All clouds (precipitating, non-precipitating, and lightly precipitating clouds) sampled during the BACEX have cloud bases and tops at about 400 m and 2700 m, implying a maximum cloud depth of about 2300 m. However, it is shown that more than half of the clouds precipitate somewhere in the cloud (56 % on average in Fig. 3.15a) even though the precipitation in and near the cloud is less than 10 mm day$^{-1}$ as a whole, except on 22 March (Fig. 3.15c). In addition, as shown earlier (Rauber et al., 2007; Gerber et al., 2008) the cloud cores are far from adiabatic (Fig. 3.16a).

For the non-precipitating and/or lightly precipitating clouds, reflectivity of $\sim -35$ dBz and Doppler velocity of $\pm 2$ ms$^{-1}$ are the most frequently observed between 600 m
and 1300 m (Figs. 3.12c-d). Cloud bases and tops for these clouds are about 700 m and 2000 m, i.e., cloud thickness of about 1300 m.

Reflectivity and velocity distributions show two populations: (1) reflectivity weaker than approximately -30 dBz with Doppler velocity ranging from -4 ms\(^{-1}\) to +3 ms\(^{-1}\) (e.g., horizontally oriented pattern in Figs. 3.11c-d, with a narrow range of reflectivity and broad range of velocity); (2) reflectivity ranging from -30 dBz to 5 dBz and Doppler velocity ranging from -2 ms\(^{-1}\) ~ -4 ms\(^{-1}\) (vertically oriented pattern in Figs. 3.11c-d with a broad range of reflectivity and a narrow range of velocity). The horizontally oriented pattern is associated with non-precipitating clouds, while the vertically oriented pattern is attributed to precipitation.

During the BACEX, two types of precipitation features were observed (Fig. 3.13). In one, precipitation shafts are observed to emanate mainly from the cloud base with evaporation in the sub-cloud layer. In the other, precipitation shafts emanate mainly near cloud top on the downshear side of the cloud and evaporate in the cloud layer. This type of precipitating cloud is shallower than the first type of cloud, and also sometimes accompanies precipitation shafts emanating near cloud base. These two types of precipitation may have impacts on trade-wind boundary layer in different ways. For example, the cloud-base precipitation can stabilize the sub-cloud layer by the evaporation rain. As a result, the following cloud can grow only if it overcomes the stability. Thus, vertical velocities of sub-cloud eddies need to be strong enough to penetrate the inversion barrier if clouds are to form. On the other hand, when precipitation shafts emerge from cloud top and evaporate in the cloud layer would stabilize the cloud layer, but destabilize the atmosphere below the cloud layer. Consequently, small clouds could form relatively
easily near the LCL (i.e., cloud base). Further, the evaporation of precipitation in the cloud layer provides moistures to the near cloud field that may affect the moisture budget and increase cloud lifetime (e.g., Albrecht, 1981) of future clouds.
Chapter 4: Aerosol-Cloud-Precipitation Interaction

In Chapter 3, the variations of cumulus clouds and aerosols over the eastern Caribbean were examined. This chapter focuses on aerosol-cloud-precipitation interactions over this regime. Satellite based studies can be used to study aerosol indirect effects over a large geographical area with long time periods, but are known to suffer from retrieval bias (Loeb and Schuster, 2008) and clouds over this regime are not valid for the adiabatic assumption (Fig. 3.16a) that used in satellite studies; and the vertical distribution—a key component of aerosol indirect effect—is usually unknown. Thus, satellite data is not used in this work. Section 4.1 describes aerosol effects on cloud and precipitation. The effect of sub-cloud aerosol on the cloud thickness is first presented. The first aerosol indirect effect will be discussed by using relationships between aerosol concentrations ($N_a$) and cloud droplets number concentrations ($N_d$), and the sizes of effective droplets ($D_e$) with cloud-base and flight-averaged properties. The sensitivity of $N_d$ to vertical velocity perturbations ($w'$) is also introduced. The sensitivity of precipitation to changes in aerosol, so-called precipitation susceptibility (Feingold and Siebert, 2009) is also presented in this section. In Section 4.2, the interactions between clouds and environmental air (aerosols) are presented from the views of entrainment process and flows in and around small clouds. Section 4.3 shortly presents cloud effects on aerosols in the near-cloud field that connects to the Section 4.2.

4.1 Aerosol-cloud interaction

4.1.1 Aerosol effects on cloud properties

Aerosol may affect cloud properties by modulating cloud coverage and depth. To examine the effects of aerosols on cloud properties, in particular cloud thickness, cloud
tops (CT), thicknesses (H) and bases (CB) on research flights (RF #) during BACEX are shown in Fig. 4.1 together with sub-cloud aerosol number concentrations. The mean cloud properties (CT, H, and CB) are obtained from the cloud radar measurements during the cloud-base flights, and thus, flight levels are considered as cloud bases.

Figure 4.1: Daily averaged (a) cloud tops (CT, black solid line connected with filled circles), bases (CB, gray solid line connected with filled circles) and thickness H (dashed line connected with open circles) obtained from radar measurements during cloud-base flights; and (b) aerosol concentrations obtained from PCASP during sub-cloud flights with research flight number (RF #). RF # is listed in Table 3.1.

Cloud thickness (H) follows the pattern of cloud top heights fairly well in Fig. 4.1a; it ranges from 100 m on 26 March (RF06) to 2000 m on 30 March (RF08). Cloud top heights vary from 1-3 km, showing a similar pattern to that of the cloud thickness, since cloud-base heights vary around 500 m - 1000 m during the experiment. During the pre-dust outbreak periods (e.g., between RF06 and RF08; 3/26-3/30), cloud top heights and thickness increase as aerosol concentrations in the sub-cloud layer increase, in particular, between RF07 and RF08 (3/29 and 3/30). The vertical structures of aerosol on both days are similar (Fig. 3.5b), but show larger aerosol concentrations on 30 March
(RF08) when deeper clouds are sampled. The same patterns are observed during the periods of post-dust outbreak, between RF13 and RF15 (4/7-4/11); shallower clouds are observed when the trade-wind boundary layer is cleaner. Those five days (3/29, 3/30, 4/7, 4/10, and 4/11) show a typical aerosol vertical structure in the pristine marine boundary layer (aerosol concentrations of less than 200-300 cm$^{-3}$ decreasing with height monotonically) in Fig. 3.5b. In contrast, cloud thickness decreases at times closer to dust outbreaks (e.g., between RF08 and RF09; 3/30 and 3/31) and no clouds are observed during the middle of the dust outbreak (e.g., RF #10, 11 of 4/1-4/2).

Aerosols (dust in this study) may change the atmospheric thermodynamic structures (e.g., stabilize the atmosphere) and result in suppression of clouds. To examine any relations between aerosols and atmospheric thermodynamic structures, vertical structures of temperature, moisture and aerosol concentrations on pristine (29 March, RF07-dust free) and dusty (1 April, RF10) days are shown in Fig. 4.2.

![Figure 4.2: Profiles of aerosols (PCASP, cm$^{-3}$), temperatures (T, °C), potential temperatures (θ, K), and mixing ratios (g kg$^{-1}$) for pristine (black, 3/29) and dusty (red, 4/1; grey, 4/2) days. Soundings are obtained during the 2nd ascents of aircrafts on the day for 29 March and 1 April and during the first ascent of an aircraft for 2 April.](image)
Aerosols increase with height on 1 April (middle of the African dust outbreak). In contrast, aerosols decrease with height on 29 March (a typical pristine maritime environment). Potential temperatures, aerosols, and vapor mixing ratios are well-mixed in the sub-cloud layer (below 500 m). Further, the well-mixed thermodynamic and aerosol layer between 1750 m and 2500 m due to the Saharan air layer (SAL) is notable. A dusty day (4/1, red) shows a drier and warmer lower atmosphere than that of a pristine day (3/29) with more stable atmospheric structures (500-1800 m). The atmospheric temperature is 2-3 °C higher on 1 April than on 29 March in the layer where high concentrations of aerosols reside (e.g., between 500 m and 2500 m). In addition, potential temperature shows a stronger inversion than on the pristine day (29 March) within the layer, in particular near 500 m (transition layer) and below the SAL with an associated distinctive dry layer. Profiles on 2 April are overlaid on Fig. 4.2 to show the differences in thermodynamic and aerosol characteristics between two dusty days of 1 and 2 April. No clouds were observed on 1 April, over the course of the 3-hr flight. On the other hand, infrequent cumulus clouds, generally in small patches and almost all very optically thin, were sampled on 2 April during the flights near the bottom of the traditional cloud layer (~800 m). The transition layer is sufficiently strong to inhibit moist convection during most of the two flights. However, its strength is stronger on 1 April than that on 2 April; the strength of transition layer is about 20 K km$^{-1}$ on 1 April and 7 K km$^{-1}$ on 2 April. The details of vertical structure on the aerosol and thermodynamics for these two days of intense African dust events are discussed in Chapter 5.

Suppressed convection during the African dust outbreaks has been noticed in the early 1970s from large-scale tropical experiments such as BOMEX and GATE (e.g.,
Carlson and Prospero, 1972). Figure 4.2 shows that warmer and drier atmospheric conditions are observed during the periods of the African dust outbreak, with a more stable and stronger (in particular near 500 m) inversion layer. However, here, the effects of meteorology cannot be separated from the aerosol effects. Therefore, this change may be caused by either aerosols or meteorology.

In addition to the aerosol feeding clouds from the sub-cloud layer, the vertical structure of aerosols may play a role in cloud formation (Fig. 4.3). For instance, aerosol concentrations on 30 March (RF08), 31 March (RF09), and 5 April (RF12) are similar with ~300 cm$^{-3}$ in the sub-cloud layer (Fig. 4.1b). However the vertical structures of aerosols on these three days are quite different (Fig. 4.3); aerosol concentrations decrease with height on the 30 March with maximum of $N_a \sim 300$ cm$^{-3}$ in the sub-cloud layer (typical aerosol profile in the trade-wind boundary layer). In contrast, high concentrations of aerosol (~ 600 cm$^{-3}$) are observed throughout the entire boundary layer on 31 March (SAL profile) above 700 m. On 5 April, aerosols have similar concentrations as those in the sub-cloud layer (~300 cm$^{-3}$) until 1500 m, followed by a significant increase with height, with a maximum of ~700 cm$^{-3}$ slightly above the inversion. In Fig. 4.3, clouds extend to a height of around 2400 m on 30 March (thickness H ~ 2000 m); 2000 m on 5 April (H ~ 1000 m); and 1200 m on 31 March with H ~ 500 m. Further, when the high concentration of aerosol layer is deeper above the inversion (e.g., 25 March (RF05) in Fig. 4.3), clouds are suppressed. Clouds on 25 March extend to a height of around 1200-1300 m with a thickness H ~ 100 m, which is a minimum thickness during BACEX. It is possible that the thick-heavy-aerosol layer, confined above the inversion, on 25 March warms up the aerosol layer, which acts to stabilize the layer below the aerosol. The
potential temperature on 25 March changes from 300 K to 304 K, between 600 m and 1500 m (strength of stable layer is about 4 K km\(^{-1}\)). In Fig. 4.3(a), a strong inversion is observed where the aerosol concentrations increase significantly within a short distance, such as near 600-700 m and 1500 m on 31 March, and near 1600 m on 23 March. Here, the vertical structures of aerosol as well as its location (e.g., how high the aerosol layer locates with respect to cloud layer or inversion height), and depth of the aerosol layer residing above the inversion layer seems to play an important role on cloud development in Fig. 4.3. The stratification of aerosol structures on 31 March is possibly associated with cloud effects on aerosol. This modification of aerosol structure by cloud process is discussed later in Chapter 5.

**Figure 4.3:** Profiles of (a) potential temperatures (θ, K) and (b) aerosols (PCASP, cm\(^{-3}\)) for three days (3/30, 3/31, and 4/5) that have similar aerosol concentrations in the sub-cloud layer, and for 25 March that has a thick aerosol layer above inversion, and for 29-30 March that have typical aerosol profiles in marine boundary layer. Cloud depths (from base to top) are denoted as vertical bars accordingly. Soundings are obtained during the 2\(^{nd}\) ascents of aircrafts on the day. Cloud bases and tops are obtained from the cloud-radar measurements during cloud-base flights on the days.
The impact of aerosols on cloud droplet number concentrations and mean size of droplets are illustrated as scatter diagrams in Fig. 4.4. The average $N_a$ is estimated from data, from each flight obtained from all levels (flight-averaged CCN) and from the sub-cloud levels (sub-cloud CCN) for the day. The mean $N_d$ from all flight levels (e.g., $N_d$ in Fig. 4.4a and Fig. 4.4c) and those averaged for the cloud-base level flights (e.g., $N_d$ in Fig. 4.4e) are shown in Fig. 4.4 (a, c, e). Here, FA (flight-average) and CB (cloud-base) indicate an average value from all flight levels and from cloud-base level flights, respectively.

**Figure 4.4:** Relations between aerosol number concentrations, $N_a$ (here CCN $s=0.3$ %) and (a, c, e) cloud number concentrations, $N_d$ and (b, d, f) mean effective size of cloud droplets, $D_e$: (a) flight averaged $N_a$ vs. flight averaged $N_d$, (b) flight-averaged $N_a$ vs. flight-averaged $D_e$; (c) sub-cloud $N_a$ and flight-averaged $N_d$, (d) sub-cloud $N_a$ and flight-averaged $D_e$, (e) sub-cloud $N_a$ vs. cloud-base $N_d$, (f) sub-cloud $N_a$ and cloud-base $D_e$. Colors in Fig. 4.4 (b, d, and f) indicate the daily mean cloud thickness, estimated from cloud-base flights.
Overall, robust increases in $N_d$ with aerosols are shown in Fig. 4.4a, Fig. 4.4c, and Fig. 4.4e. However, a flight-averaged (FA) $N_d$ shows the least correlation with flight-averaged aerosol concentrations (Fig. 4.4a; $R=0.48$). On the other hand, sub-cloud aerosol concentrations fairly well represent the flight-averaged $N_d$ and cloud-base $N_d$, with correlations of $R=0.79$ and $R=0.71$, respectively. The highest correlation is found between sub-cloud aerosol concentrations and flight-averaged cloud droplet number concentration in Fig. 4.4c.

Relations between aerosols and effective sizes of droplets are shown in Fig. 4.4 (b, d, and f). Aerosols as well as liquid water contents affect the size of cloud droplets, and thus, data is colored by cloud thickness. In general, the average size of cloud droplets ($D_e$) tends to decrease as aerosol concentrations ($N_a$) increase. For instance, data from 23, 24 March and 5, 7, 10 April show similar cloud thickness (approximately 1000 m shown as greenish colors), and thus these dataset will be considered here; as $N_a$ increases from 24 March to 5 April, $D_e$ decreases from around 21 µm to 18 µm. However, $D_e$ tends to increase as $N_a$ increases under extremely pristine environments (e.g., 7, 11 April and 22, 30 March under CCN < 110 #/cc in Fig. 4.4b) and/or in case of thicker clouds (e.g., 22 and 30 March). Further, the inverse relationship between $D_e$ and $N_a$ is valid only for shallow clouds (e.g., clouds shallower than 500 m) when sub-cloud $N_a$ and cloud-base $D_e$ are considered (Fig. 4.4f), suggesting the possible roles of other cloud processes such as entrainment and precipitation.

Aerosol concentrations in the sub-cloud layer and cloud droplets concentrations in the cloud base are closely related (e.g., $R=0.71$ in Fig. 4.4e). The effects of sub-cloud aerosols on cloud properties is explored by examining the relationship between sub-cloud
CCN concentrations and $N_d$ observed just above cloud base, as a function of cloud updraft velocity (Fig. 4.5) for different updraft thresholds. Here, the vertical velocities ($w$) were averaged for the cloud-base level (~10-minute) legs and then the deviations from the mean were used to calculate $w'$. 

Figure 4.5: CCN in sub-cloud layer and $N_d$ just above cloud base for different thresholds of updraft: (a) $w' > 0$, (b) $w' > 0.5$, and (c) $w' > 1$ m s$^{-1}$.

Cloud-base $N_d$ is highly correlated to the sub-cloud CCN concentrations (Fig. 4.5), in particular, in the regimes of updrafts and/or positive perturbation of $w$ ($w' > 0$). In the case of $w > 1$ m s$^{-1}$ (Fig. 4.5c), 95 % of cloud-base $N_d$ variation can be explained by sub-cloud CCN. However, $D_e$ just above the cloud base is poorly related to the same parameter regardless of the thresholds (not shown here). To explore the sensitivity of $N_d$ to $w'$ just above cloud base, $N_d$, is shown as a function of $w'$ for clean and dusty conditions in Fig. 4.6.
Figure 4.6: The $N_d$ versus $w'$ just above cloud base for (a) polluted (5 April) and (b) pristine (10 April) environments. Green dots in Fig. 4.6a and Fig. 4.6b indicate all the data points for the day. Cloud-base $N_d$ is averaged in 20 cm$^{-3}$-intervals and shown as black dots. The mean values of sub-cloud CCN (0.3 % and 0.6 %) are shown as horizontal lines. The mean profiles of $N_d$ vs. $w'$ for two pristine (29 March, 10 April) and dusty (31 March and 5 April) cases are shown in Fig. 4.6c with slopes ($dN_d/dw'$) estimated for the regime of $w' > 0$.

Overall, cloud-base $N_d$ increases as $w'$ increases for both pristine and dusty cases. Sensitivity to $w'$, i.e., $dN_d/dw'$, is greater in polluted (or dustier) environments (e.g., 31 March and 5 April in Fig. 4.6c), suggesting a slight increase in positive $w'$ in polluted conditions results in a greater increase of $N_d$. The slopes of $dN_d/dw'$ for dusty conditions are about 122-123 (cm$^{-3}$/ms$^{-1}$), while those for pristine conditions are about 55, slightly less than half of that of dusty environments. Following Twomey (1959), Droplet concentration $N_d$ can be expressed by
where \( w \) is in (cm s\(^{-1}\)); \( C \) is a function of total aerosol number concentration; \( k \) is a parameter that depends on the air-mass type. Typical values of \( k \) are 0.5 and 0.75, respectively, for maritime and continental environments (e.g., Twomey, 1977). The number of particles (per unit volume) that are activated to become cloud droplets is expressed as a function of the super-saturation, \( s \)

\[
N_d = C s^k
\]

where \( s \) is the percent super-saturation (\( s = (S-1)*100\% \)); \( N_d \) is the number of nuclei per unit volume activated at super-saturations less than \( s \). The parameter \( k \) in Eq. (4.1) can be obtained from Eq. (4.2). The implication of the slope, \( dN_d/dw \)' in Fig. 4.6, may be used to evaluate the parameterization of Eq. (4.1). From Eq. (4.1), the sensitivity of \( N_d \) to changes in \( w \) is defined as,

\[
\frac{d \ln N_d}{d \ln w} \sim \frac{3k}{2(k+2)}
\]

and thus only depends on the parameter \( k \). However, \( k \) does not show significant differences between polluted and clean conditions (Fig. 4.7). The mean value of \( k \) and its standard deviation are shown in Fig. 4.7 with \( d\ln N_d/d\ln w \)' from the observational data shown in Fig. 4.6. The \( k \) value fluctuates around 0.5±0.1, and \( d\ln N_d/d\ln w \) from Eq. (4.3) is ~0.3. However, \( d\ln N_d/d\ln w \)' from the observational data of Fig. 4.7 gives a sensitivity of 0.1 for clean and 0.2 for dusty conditions. The sensitivity of \( N_d \) to \( w \) is overestimated from Twomey’s relationship and this is possibly due to the power-law size distributions used in Eq. (4.2). Aerosol size distributions are usually described by lognormal functions.
4.1.2 Aerosol effects on precipitation

Does aerosol play a role in regulating precipitation? Scatter diagrams of rainfall rates and CCN number concentrations are shown in Fig. 4.8. Flight-averaged rainfall rates and CCN are estimated from data from each flight from all levels for each day, and sub-cloud CCN are estimated from data from the sub-cloud level flights for each day. CIP probe data is used for the precipitation rate calculation. Data is colored by cloud thickness to capture the effects of aerosols on rainfall rate for similar macrophysics (i.e., cloud thickness).
Figure 4.8: Scatter diagrams of aerosol number concentrations (cm$^{-3}$) and rainfall rates (mm hr$^{-1}$) in and near the clouds during BACEX; (a) Flight-averaged CCN and rainfall rate and (b) sub-cloud CCN versus flight-averaged rainfall rate (mm hr$^{-1}$). Rainfall rate is calculated by Eq. (3.2) with CIP probe data. Colors indicate cloud thickness. Note that no CCN data was available on 11 April.

If we assume there is no sudden change of synoptic condition between two consecutive days that have a similar cloud thickness, and compare those two days (for example, between 23-24, 25-26 and 5-7 April, possibly), suppression of precipitation for the day with higher aerosol concentrations is notable in Fig. 4.8. Results are the same if we compare days with similar cloud thickness (e.g., days with $H \sim 1000$ m such as 23, 24 March, 5, 7, 10 April and; days with $H < 500$ m such as 25, 26, 31 March; days with $H \sim 2000$ m such as 22, 30 March). Regarding the cloud-base rainfall rates (not shown), aerosol-rainfall relations show the same pattern but are not as clear as flight-averaged rainfall rates. This probably is due to the precipitation properties—precipitation shafts emanating mainly near the cloud top and evaporating in the cloud layer, as observed in shallower clouds in Fig. 3.13. In short, towards thicker clouds (along the y-axis in Fig. 4.8), precipitation rate, R increases. Towards higher aerosols (along the x-axis), R decreases.
Precipitation susceptibility $S_0$—the sensitivity of precipitation to changes in aerosol—is used to quantify this decrease while minimizing the effects of macrophysics (Feingold and Siebert, 2009). It is defined as

$$S_0 = \left( -\frac{d \ln R}{d \ln N_d} \right)_{\text{macro}}$$

(Equation 4.4)

The subscript “macro” indicates that the susceptibility is calculated with fixed cloud macro-physical properties, such as fixed cloud thickness ($H$) or liquid water path (LWP). It is important to use fixed cloud macro-physical properties in $S_0$ calculation as aerosol concentrations tend to correlate with meteorological factors, making it difficult to separate the aerosol effects from macro-physical drivers. The logarithmic form of the equation reduces the sensitivity of $S_0$ to the measurement accuracy of $R$ and $N_d$, and the minus sign reflects that suppression of precipitation by aerosols. Precipitation susceptibility $S_0$ for $H$ held fixed are shown in Fig. 4.9. $H$ in this calculation is obtained from cloud radar measurements with resolutions of 1-s (in time) and 24-m (in vertical) during the cloud base level flights of each day. Precipitation rate is estimated from the CIP probe during the cloud-base level flight; and cloud-base $N_d$ is calculated from CAS (forward scattering) and CIP probes. The susceptibility is derived for eight bins of increasing cloud thickness $H$ from 0 to 1600 m with 200 m interval. Datasets from 10 of the 12 days (see Fig. 3.10), excluding two major precipitating clouds on 22 and 30 March as an attempt to account for sampling biases due to the rain effects on aerosols, i.e., wet scavenging, are used to calculate $S_0$ and compared with that for all 12 days to examine the effects of wet scavenging.

In Fig. 4.9, $S_0$ exhibits three regimes as discussed by Sorooshian et al. (2009) who used satellite measurements to estimate the susceptibility: (1) low cloud thickness, where
clouds do not precipitate, and are not sensitive to \( N_d \); (2) intermediate cloud depth, where precipitation begins to decrease due to the increased aerosols (and \( N_d \)); (3) high cloud thickness, where \( S_0 \) decreases as accretion, which is insensitive to \( N_d \), begins to dominate the collection process. \( S_0 \) peaks at intermediate values of cloud thickness (1000 m < \( H < 1200 \) m) with 10 days of clouds sampled excluding two major precipitating clouds (22 March and 30 March, see Fig. 3.10). The results are consistent with previous studies of warm cumulus clouds (e.g., Sorooshian et al., 2010; Jiang et al., 2010; Duong et al., 2011), reporting \( S_0 \) peaks at intermediate values of cloud LWP (~600-1300 gm\(^{-2}\)).

**Figure 4.9:** Precipitation susceptibility \( S_0 \) as a function of cloud thickness, \( H \), estimated from cloud-base radar measurements (time resolution of 1-sec, height resolution of 24-m used). All clouds (12 cases; dashed) sampled during the BACEX and data excluding two major precipitating clouds (22 March and 30 March, solid) in Fig. 3.10, are shown to examine the scavenging effect.

In contrast, Terai et al. (2012) showed that \( S_0 \) decreases as the cloud thickness increases in marine stratocumulus during the VOCALS Regional Experiment. They introduced precipitation susceptibility metrics \( S_R \) as the product of drizzle intensity (\( S_I \)) and drizzle fraction (\( S_f \)), and \( S_R \) decreases with increasing \( H \) mainly due to the \( S_f \) decrease, while \( S_I \) is unchanged.
In Fig. 4.9, removal of $N_d$ due to the rain (wet scavenging) makes susceptibility stronger overall (dashed line shows higher $S_0$ than that of solid line) and peaks at lower cloud thickness ($600 \text{ m} < H < 800 \text{ m}$) than that from lightly precipitating clouds ($1000 \text{ m} < H < 1200 \text{ m}$). Recently, Duong et al. (2011) also pointed out that wet scavenging leads to higher $S_0$. The reason that $S_0$ peaks near $H = 700 \text{ m}$ (primary) and $1300 \text{ m}$ (minor) is associated with heights that heavy precipitation are observed (Fig. 4.10).

**Figure 4.10:** Cloud thickness distributions during BACEX. Cloud thicknesses are estimated from the cloud radar (1-s and 24-m resolutions) during the cloud-base level flights. Colors indicate rainfall rates (mm day\(^{-1}\)). Dates are shown adjacent to dashed lines.

Heavy precipitation (e.g., $R > 1 \text{ mm day}^{-1}$) is mainly observed from thick clouds and dominates clouds sampled on 22 March, 30 March, and the upper parts of clouds sampled on 24 March. The lowest heights that precipitation are observed are approximately 700 m, consistent with the primary peak of cloud thickness in $S_0$, indicating the non-negligible effects of wet scavenging below the precipitation height.

Simple power law relationships have been used to relate the cloud base precipitation rate $R_{CB}$ to LWP (or H) and to $N_d$ (e.g., Geoffroy et al., 2008).
\[ R_{CB} = A H^a N_d^{-\beta} \]  

(Equation 4.5)

For clouds with the same H, the exponent \( \beta \) is equivalent to susceptibility \( S_0 \) in Eq. (4.4). For field studies of precipitating stratocumulus, the \( \beta \) value has been reported in the range of 0.8 to 1.75 for a fixed LWP (e.g., Pawlowska and Brenguier, 2003; Comstock et al., 2004; vanZanten and Stevens, 2005). However, the \( \beta \) value with cloud thickness has not been reported in these studies, primarily due to the measurements constraints. Recently, variations in \( \beta \) with cloud thickness are reported as from 0 to 3 for VOCALS Rex (Terai et al., 2012).

For studies of warm cumulus clouds, \( \beta \) varies from 0.5 to 1.1 with LWP, from the adiabatic parcel model (Feingold and Siebert, 2009) and \( \sim 1.1 \) from large eddy simulation (Jiang et al., 2010). However, precipitation susceptibility with either cloud thickness or with LWP has not been reported from field studies. Further, the \( \beta \) values reported in previous studies, which used satellite observations, are estimated from CCN proxies and thus direct comparison of \( \beta \) (magnitude) is not possible.

4.2 Aerosol-cloud interactions: Entrainment and flows in and around small cumulus clouds

4.2.1 Background

Entrainment can influence clouds and cloud processes in several ways: cloud lifetime, cloud radiative properties, electrification of clouds, and cloud chemistry. Entrainment and the resulting mixing processes (e.g., homogeneous and/or inhomogeneous process; Baker et al., 1980; Lehmann et al., 2009; Lu et al., 2011) play an important role on the evolution of a cloud droplet size distribution and the formation of precipitation.
Squires (1958) proposed that clouds are diluted from the top down; cloud air is mixed with environmental air at the cloud top and penetrates downward due to the evaporative cooling and diluted the cloud air below. This idea has played a central role in recent models of cumulus convection. Telford (1975) adapted this idea to a model and demonstrated that mixing in clouds is a buoyancy-driven process; entrainment occurred at the cloud top and mixed air is then carried down until it reaches a hydrostatic equilibrium with the environment. Later, Paluch (1979) showed that the entrained air in continental cumulus clouds in Colorado originated from a height close to the cloud top by using two conserved thermodynamic variables. A number of other studies have supported the idea of cloud-top mixing since then (Jensen et al., 1985; LaMontagne and Telford 1983; and many others) and it has been commonly accepted as entrainment/mixing process in cumulus clouds. However, other results that show the source of entrained air are close to (above or below) the height where observations were made has been reported (Gardiner and Rogers, 1987; Blyth and Raymond, 1988; Raga et al., 1990; Hicks et al., 1990; Taylor and Baker, 1991). Likewise, there are continuing debates on the origin of entrained air that revolve around cloud-top mixing versus lateral mixing (Blyth, 1993).

Further, the mechanics of how the environmental air is entrained into the cloud and the details of the mixing process are poorly observed. Blyth et al. (1988) suggested that entrainment occurs near the ascending cloud top as clouds grow, and the mixed parcels subsequently descend around the rising thermal (schematic model in Fig. 14 in their paper), showing that the source of entrained air was close to cloud top. On the other hand, Jonas (1990) used Paluch diagrams to show that the environmental air from near cloud top is transported to a lower level and then laterally entrained into the cloud.
Although, the Paluch diagrams provide an indirect method for studying mixing processes in and around clouds, more direct methods have been used; Stith (1992), for example, used sulfur hexafluoride (SF6) as a tracer to study entrainment mechanism. The gas was released from an aircraft that flew above the cloud top and it was detected by a second aircraft later. He observed that the maximum concentration of SFs at the edge of cloud in the downward moving air parcels. Earlier, Moninger and Kropfli (1987) proposed to use radar chaff tracers to study the entrainment of air into clouds. They used observations from a Ka band (35 GHz) radar with circular depolarization ratio (CDR) measurements and showed how radar returns from chaff could be distinguished from those of hydrometers. They provided a list of questions that could be addressed using chaff. These questions included: (1) “where does entrainment primarily occur in cumulus clouds?” (Lateral versus cloud top entrainment); (2) “are there specific entrainment levels?”; and (3) “once in the cloud, where does entrained air go?” This section describes entrainment and flows in and around small cumulus clouds, which are ultimately associated with interactions between clouds and environmental air (i.e., aerosols); here, aerosol-cloud interactions are represented by entrainment and detrainment process.

4.2.2 Chaff experiment

The cloud examined in this section, 29 March 2010, is sampled about 50-100 km east of Ragged Point. A bimodal distribution of cumulus clouds with cloud bases around 600 m was observed on this flight. The shallower clouds have tops at about ~1 km and deeper clouds have tops around 2 km. Many clouds observed on this flight have bubble-
like appearances with bubbles of cloudy air detraining from the cloud tops and evaporating.

Flight paths and time series of the altitudes flown on this day for the chaff experiment are shown in Fig. 4.11. Five sets of cloud penetrations were performed (top-down) and two sets of the level legs were made at the same altitude (Fig. 4.11b, e.g., LEGs 1 and 2, LEGs 4 and 5). Here, a level leg refers to a cloud penetration flight leg flown at a relatively constant height. Chaff was dispensed near the cloud top heights on LEG 1 along the wind and on LEG 2 cross the wind at the same level. After the chaff tracer was released, the aircraft made penetrations of the cloud at lower levels along LEG 3, LEG 4 and LEG 5- all flown along wind (Fig. 4.11b, shown as numerical labels for each level leg). The cloud had nearly disappeared on the fourth penetration (LEG 4) and thus, only the first four of the five penetrations of the cloud (LEGs 1 to 4) are discussed here.

**Figure 4.11:** (a) A flight pattern and (b) corresponding altitude during the chaff experiment. Colors indicate time elapse (earlier time is denoted as a darker color). The spatial locations where the chaff is released are shown as cross symbols in Fig. 4.11a (black-upwind, gray-cross wind) and the target cloud is located in the middle of the cross symbols. The duration of the cloud penetration is denoted as vertical dotted lines in Fig. 4.11b. Flight paths, after LEG 3, are not shown in Fig. 4.11a.
4.2.3 Cloud thermodynamics and dynamics

The cloud that was used for the chaff study is a shallow marine cumuli with a thickness and width of less than 1 km. The bubble-like features associated with this cloud are shown in the photos (Fig. 4.12). These photos were taken in a period approximately 10 s to 80 s before the first penetration of the cloud, from a camera looking forward out of the cockpit window. The photos clearly show the detrainment features that were observed during this flight.

![Figure 4.12: Photos taken towards the west, approximately a) 80 s, b) 20 s and c) 10 s prior to the first cloud penetration (LEG 1). The airplane penetrated the cloud from east to west direction during LEG1 and the background wind is into the direction of photo (easterly).](image)

The aircraft made penetrations of the cloud along wind on LEGs 1, 3 and 4, with LEGs 1 and 4 downwind and a LEG 3 upwind. In this study, the cloud penetration on LEG 3 is used as a reference; thus, reversed time axes were used for LEGs 1 and 4 in the all related time series shown in the figures, to provide a match of the cloud features in the spatial direction; east corresponds to the RHS and west to the LHS in the time-series figures (i.e. Figs. 4.13, 4.14). The background wind is from right to left (easterly) in the figure. Upshear is defined as the regions of $|\frac{du}{dz}| > 0$, where $u$ is horizontal components of the wind (Perry and Hobbs, 1996; Lu et al. 2003), and is defined as east-side of the cloud in this study (e.g., in Fig. 4.14).
The cloud structure near the top is obtained from the first two penetrations (LEG 1 and LEG 2) at heights of about 1400 m. The temperature, moisture and vertical velocity along the transit through the top of this cloud is shown in Fig. 4.13a and Fig. 4.13e. In addition, a detailed analysis of the radar returns observed on the cloud penetration is made to show the structure of the cloud sampled.

**Figure 4.13:** (a, e) Time series of vertical velocity ($w$, m s$^{-1}$; black), virtual temperature ($T_v$, °C; magenta) and liquid water content (LWC) multiplied by 10 (g m$^{-3}$, blue) for the cloud penetration on LEG 1 and LEG 2. Time-height cross sections of (b, f) reflectivity, (c, g) mean Doppler velocity and (d, h) spectrum width of Doppler spectrum, for the chaffed cloud on LEG 1 (left) and LEG 2 (right). The cloud penetrations were made at constant heights (along the wind-LEG 1, cross the wind-LEG 2) and the average penetration heights are shown with level legs. The aircraft observations were made at an air speed of about 60 m s$^{-1}$ and thus cloud widths are approximately 120 m and 480 m for LEG 1 and LEG 2. Note that reversed time axes are used on LEG 1 to match the penetration direction with other LEGs. Vertical dashed lines in Fig. 4.13f are used to show Doppler spectra in Fig. 4.15. East on the right (LEG 1); Northeast on the right (LEG 2); Upshear side on the right (LEG 1).
On the first penetration (Fig. 4.13a), the liquid water content (LWC) of the cloud reaches a maximum value of about 0.41 g m$^{-3}$. The aircraft measurements show that the regions of the highest LWC at around 16:01:50 – 16:01:52 UTC are in the updraft of \( w \sim 2 \text{ m s}^{-1} \). The strong downdrafts of about 3 m s$^{-1}$ are observed along the downshear edge of the cloud (~16:01:53 UTC). The in-cloud virtual temperatures within the high LWC areas are higher than the surrounding air and give positive buoyancy. The lower virtual temperatures, \( T_v \) (negative buoyancy) are found along the cloud edges, especially in the regions of the strongest descending motion at \( \sim 16:01:53 \text{ UTC} \) in the downshear side of the cloud. In addition, the updrafts are highly correlated with high LWC and positive buoyancy, implying that the cloud element penetrated at this stage (LEG 1) is actively growing. (e.g., Lehmann et al., 2009). Chaff is dispensed near the cloud top height on LEG 1 and thus, radar returns from the cloud at this stage, are hardly detected. Although the LEG 1 penetration was made at a height of about 1400 m where the LWC is about 0.41 g m$^{-3}$, the radar reflectivity indicates that tops extend to about 1.6 km (MSL).

The cloud structure on the 2nd penetration (LEG 2, Fig. 4.13e) also shows that the updraft is associated with high LWC (maximum of about 0.6 g m$^{-3}$) and positive buoyancy (associated with positive \( T_v \) perturbation), indicating the cloud core is in a growing stage. Again, strong negative perturbations of \( T_v \), presumably due to the evaporative cooling, and downdrafts are noticeable along the cloud edges (so-called cloud halos or cloud shell regions, which are regions of enhanced humidity in the vicinity of cumulus clouds; Radke and Hobbs, 1991; Perry and Hobbs, 1996; Lu et al., 2003; Laird, 2005; Heus and Jonker, 2008) in Fig. 4.13e as in Fig. 4.13a. These downdrafts (e.g., \( \sim 16:01:53 \text{ UTC} \) in Fig. 4.13a and both sides of cloud edges in Fig. 4.13e) extend
into the clear air surrounding the cloud. The radar reflectivity in Fig. 4.13f is about -26 dBz (yellowish colors) with significant updrafts (Fig. 4.13g; $w > 1-2 \text{ m s}^{-1}$) and spectrum widths (Fig. 4.13h) of 2-3 m s$^{-1}$ for the cloud, indicating the signals are mainly returned from growing cloud droplets. Cloud tops extend to about 1700 m.

The third penetration (Fig. 4.14a, about 2-3 minutes after the second penetration) is made about 300 m below the initial penetrations, and the LWC in the cloud reaches a maximum value of $\sim 0.3 \text{ g m}^{-3}$. The cloud at this stage shows mainly updrafts but shows downdrafts along the cloud edges with stronger downdrafts observed along the upshear
side of the cloud edges. No significant cooling is observed along the downshear side of the cloud edges. Rather, a wider area of higher $T_v$ is observed in the downshear side of the cloud. There still is good agreement between updrafts and LWCs, but no significant or even an opposite pattern between $T_v$ and, $w$ and LWC at this stage. The cloud at this stage has a cloud top of about 1400 m and two small turrets detached from the main cloud are also seen at around 1400-1600 m between 16:06:09-16:06:12. Reflectivities of $\sim$ -25 dBz (yellowish), along with updrafts of $\sim$ 2 m s$^{-1}$ and spectrum widths of $\sim$ 1-2 m s$^{-1}$ are observed at around 16:06:14 (shown as “E” in Fig. 4.14b). In this area, cloud droplets are growing as indicated by the enhanced reflectivity associated with an updraft and a relatively large spectrum width. Reflectivities, greater than -18 dBz (reddish colors) in Fig. 4.14b, are attributed to returns from chaff fibers as discussed in the next section.

The cloud observed on the fourth penetration (LEG 4, Fig 4.14e) shows a maximum LWC of 0.15 g m$^{-3}$ with a weak updraft ($\sim$ 0.7 m s$^{-1}$). The cloud is substantially weaker than when it observed on the previous legs and appears to be in inactive stage, and has almost dissipated. However, strong reflectivities (Fig. 4.14f) with relatively small spectrum width (Fig. 4.14h) are observed for the cloud. These returns are again attributed to chaff returns and will be discussed in the next section.

### 4.2.4 Radar analysis-Chaff returns

A detailed analysis of the radar returns from the chaff observed on the LEG 3 and LEG 4 is made to show the entrainment process and in-cloud flow patterns. Chaff are dispensed near the cloud top heights on LEG 1 and LEG 2, and chaff signals are detected on LEG 3 and LEG 4, about 300 m and 450 m below the initial penetrations.
Time-height cross-sections of reflectivity in LEG 3 (Fig. 4.14b) shows two areas of strong reflectivities (reddish colors): first, greater than -18 dBz reflectivity along the outside of the cloud edges in a vertically elongated pattern; and second, at the borders of the downshear side of the cloud (based on LWC) with a bubble like pattern. In the Doppler velocity field (Fig. 4.14c) downward motions are observed where the chaff signals are detected in the reflectivity field, indicating chaff descended to the lower heights, along the downshear side of the cloud edges through the downdraft. The downward motion along the cloud edge is consistent with vertical velocity obtained from aircraft’s inertial navigation system (INS) shown in Figs. 4.13a, 4.13e and Fig. 4.14a, and is in an area where no liquid water is observed. The spectrum width in the chaff is relatively small (~ < 0.5 m s\(^{-1}\)) in the area where the chaff signals are observed (e.g., > -18 dBz, reddish colors in Fig. 4.14b) compared with the cloud returns around 16:06:14 UTC on LEG 3 (shown as “E” in Fig. 4.14b). In contrast, high values of spectrum width are observed along the cloud edges and beneath the detached cloud turrets, suggesting active turbulent mixing or horizontal shear of the vertical velocity in these regions.

The bubble-like strong chaff reflectivity in Fig. 4.14b is located outside of the cloud based on LWC, but is located inside of the cloud based on the radar reflectivity (In Fig. 4.14b, cloud boundaries extend from 16:06:10 to 16:06:17 UTC at the flight level). The reasons that radar returns on LEG 3 (e.g., Fig. 4.14b) show inconsistent cloud boundaries are as follows: the lowest height that the radar acquires data is not exactly the same as the level where the LWC is measured: the radar is mounted on top of the aircraft and the Gerber LWC sensor is attached beneath the wing of the aircraft. In
addition that the lowest ~ 50 m is not detectable from the radar (radar dead zone of ~ 50 m). Further, the cloud studied here is tilted (westward) and consists of numerous bubbles.

The bubble-like strong chaff reflectivity in Fig. 4.14b is adjacent to the rising branch of reflectivity, highlighted as ~ -25 dBz (yellowish colors) along the upshear side of the cloud. As time proceeds the cloud appears to be filled with chaff as shown in Fig. 4.14f. The circulations inferred from the cloud and chaff returns indicate an updraft along the upshear side of the cloud overturns at cloud top and descends along the downshear side of cloud. As a result, the area of strong reflectivity in Fig. 4.14f shows a circular pattern (head part of an inversed P as described by Kitchen and Caughey, 1981) along the cloud edges. Spectrum widths within the chaff-filled cloud (Fig. 4.14h) show small values overall. In contrast, high spectrum widths along the lower parts of the cloud edges indicate turbulent mixing in these regions.

A detailed analysis of the entrainment process and comparisons of the radar signals from chaff and cloud droplets, observed during LEGs 2, 3 and 4 are shown in Fig. 4.15. The panel identifiers A-F in Fig. 4.15 are shown in Figs. 4.13 and 4.14 as vertical dashed lines. The cloud, and thus Doppler spectrum, observed on LEG 2, is not affected by chaff and thus shows returns from cloud droplets exclusively.

In Fig. 4.15 (first row), updrafts of about 0-3 m s⁻¹ are detected in most parts of the cloud area (A-E), while downdrafts are observed in the northeast side of the cloud in F at ~ 0.1 km above the cloud penetration height on LEG 2. Note that cloud penetration was made from southwest to northeast (see Fig. 4.11a). Doppler spectra, observed when the strong chaff signals were detected in the reflectivity field as a vertically elongated pattern (in Fig. 4.14b), clearly show the returns from the descending chaff in A on
16:06:09 (second row; LEG 3). This signal is compared with that of cloud droplets in E (second row; 16:06:15, LEG 3); the differences between strengths of the signal are noticeable.

Figure 4.15: Time-height cross-sections of Doppler power spectra for the chaffed cloud, through LEGs 2, 3, and 4. Radar reflectivity, Doppler velocities, and spectrum width corresponding to each LEG are shown in Fig. 4.13 and Fig. 4.14 with vertical dashed lines accordingly. Details of (A)-(F) for each leg are given in the context. Time resolution of Doppler spectra is 3-Hz. Note that C1, C2, C3 on LEG 3 are denoted as one vertical dashed line in Fig. 4.14 since, time resolution of radar moments shown in Fig. 4.14 is 1-Hz. Same as in A1 and A2 on LEG 4.

Snapshots of how chaff entrains into the main cloud are shown for LEG 3 through B to C3 in Fig. 4.15 (third row). At point B, downward motions dominate areas with chaff signals. However, from C1 to C3 on LEG 3, upward motions emerge from the
bottom (i.e., at the flight level) and chaff signals move upward. At point D2, the upper part of the chaff signals (slightly below 0.2 km), continue to move upward and connect to the rising branch of cloudy air in Fig. 4.14b.

Doppler spectra obtained from LEG 4 (fourth row) show signals exclusively from clouds in A1, and chaff in B and E. At points C and D (especially C), chaff signals are mainly observed below 0.1-0.2 km, but returns from cloud droplets are observed above 0.1-0.2 km. The radar detected the cloud above the flight level, although its base has been nearly dissipated. Through B and D on LEG 4, overturning circulations are observed above 0.1-1.2 km. Chaff signals appear to follow the overturning circulation and fill the entire cloud with chaff.

4.2.5 Discussion and summary

How far will the air parcel descend along the cloud halo? Will it reach the equilibrium temperature level or will it descend the cloud base by the conservation of mass? To answer this question a time series of virtual potential temperature at the level where the chaff was dispensed (LEG 1) and that at the level where chaff signals were detected (LEG 3) are examined in Fig. 4.16. Environmental virtual potential temperatures obtained from the three soundings on the day are overlaid as horizontal dashed lines.

The virtual potential temperatures are approximately equal to those of the environmental soundings, in the downshear side of the cloud edges (LHS of cloud in Fig. 4.16), where the chaff is entrained into the cloud on Leg 3. It shows that the air that is negatively buoyant near cloud top descends along the downshear side of the cloud halos. Temperature in the cloud halos is about 1.5- 2 °C lower than that of in-cloud (not shown),
so the unsaturated air in the halos descends dry adiabatically until it has the same
temperature to the environmental air. Though it is not shown here, the height is
approximately ~ 1000-1100 m from the vertical profile of potential temperature, similar
to the height inferred from the Paluch diagram.

![Figure 4.16](image)

**Figure 4.16:** Time series of virtual potential temperature (red) and liquid water contents (black) for LEG 1 and LEG 3. The environmental virtual potential temperatures obtained from three soundings during the flights are overlaid (horizontal dashed lines). Note that reversed time axis is used on LEG 1.

The entrainment process and the incloud flow patterns, superimposed on the cloud radar returns observed on Leg 3, are shown in Fig. 4.17 as a schematic.
Figure 4.17: Schematic of entrainment process and incloud flow pattern for the shallow marine cumulus cloud, overlaid in the radar reflectivity field observed on Leg 3. Cloud boundaries are shown as blue solid lines. Chaff movements and in-cloud flow patterns are denoted as black solid/dashed lines, and details of the numerical numbers are described in the text. Background winds are denoted as arrows to show the upshear/downshear direction. ‘E’ indicates east, while ‘W’ indicates west direction. A 1-Hz resolution liquid water contents (LWC) are added as black solid line. Detrained cloud elements are seen above the main cloud as two turrets toward downshear side of the cloud.

The chaff tracer facilitates the illustration of in-cloud flows as well as flows around the cloud; the environmental air descends to the equilibriun level, along the downshear side of cloud edges. A blob of chaff (bubble-like pattern) entrains laterally into the cloud in the cloud shell/halo regime (denoted as 2-3). Once entrained into the cloud, the air parcel moves horizontally to the upshear side of cloud edges (denoted as 4). On the upshear side of the cloud edge, air parcels move upward to the cloud top along the cloud edges through updrafts (denoted as 5). Finally, the air turns down toward the cloud base along the downshear side of cloud edges (denoted as 6). Detrained cloud elements are also seen above the main cloud, in the downshear side of the cloud. The incloud flow pattern shows inversed P letter as in Fig. 4 of Kitchen and Caughey (1981). Observations in non-cloudy air in this study, using chaff tracers, added new insight on small-scale
cloud circulations. However, the vertically pointing cloud radar, used in this study, is not easy to provide the time-space distribution of both chaff and cloud. Therefore, a scanning cloud radar with differential polarization capability would allow for an effective separation of the chaff and cloud returns, and a full view of the 3D evolution of the cloud and chaff.

4.3 Modification of aerosols near cloud field (cloud effects on aerosols)

In the regime of $w' < 0$ in Fig. 4.6, $N_d$ decreases as $-w'$ increases, and thus, this regime is associated with cloud dissipation. Further, downward motions were observed along the downshear side of cloud edges from chaff tracer experiment (e.g., Fig. 4.17). Thus, we may expect decrease of $N_d$ with $-w'$ in the cloud halos (shell) regime in near cloud field.

Time series of aerosol (CCN, 0.3%), cloud droplets number concentrations ($N_d$), liquid water content (LWC, g m$^{-3}$) and vertical velocity ($w$) near-cloud field are shown in Fig. 4.18 and Fig. 4.19. Clouds exampled here are clouds that used for vertical velocity retrieval in Chapter 2 (hereafter Mie cloud, same cloud as Fig. 2.9, shown in Fig. 4.18), and Clouds that used for chaff experiment (hereafter chaff cloud, same cloud as Fig. 4.13-4.14, shown in Fig. 4.19). The Mie cloud has a width of about 2 km, thickness of about 1 km (cloud tops ~ 1.6 km), while the chaff cloud has a width of about 500 m and cloud thickness of about 700 m (tops ~ 1.6 km).

Overall, for the Mie case (Fig. 4.18) updrafts are observed in cloud regimes at the aircraft level and downdrafts dominate along the cloud edges. The background CCN is ~ 200-250 (#/cc), but significant CCN depletion is found along the cloud edges (decrease
down to 50 cm$^3$). CCN depletion is more efficient in the regime of precipitation shafts emanating from the cloud base, shown as strong reflectivity with downward motion in radar measurements (e.g., around 16:17:42 rather than 16:17:24 in Fig. 4.18a; near dashed line in Fig. 2.9).

**Figure 4.18:** Time series of aerosols (CCN 0.3% sampled at sub-cloud layer; red), LWC (green) and vertical velocity (w, black) sampled during cloud-base level flight for the cloud in Fig. 2.9 (Mie cloud in the text).

For the chaff cloud case (Fig. 4.19), the background CCN is ~ 50-150 (#/cc), but depletion of CCN near cloud field is not as significant as in the Mie cloud in Fig. 4.18, though downdrafts are notable along the cloud edges; $w \sim -1$ m s$^{-1}$ for Mie cloud, while $w \sim -3$ m s$^{-1}$ for chaff cloud. This is possibly because the chaff cloud is smaller in size; and may not have strong circulation around the cloud as does Mie cloud; and, as a result, may have weaker effect on aerosol changes near cloud field than it does in the Mie cloud. Moreover, chaff cloud was purely non-precipitating cloud, while Mie cloud held large enough drops to give Mie resonance and precipitation.
Figure 4.19: Same as Fig. 4.18 except data were sampled during LEGs 1-3 for the cloud in Fig. 4.12 (Chaff cloud).

Precipitation shafts are observed to emanate from cloud base around 16:17:42, and it emanates near cloud top around 16:17:28 on the downshear side of the cloud with evaporation in the cloud layer. The largest CCN depletion in Fig. 4.18b is found around 16:17:42, when the precipitation shafts are observed at the flight level. The presence of precipitation, the size of cloud, and the amount of background CCN may be related to the efficiency of the CCN depletion. To examine the relations between CCN depletion and occurrence of precipitation, time series of aerosol concentrations on 24 March, which precipitated, is illustrated in Fig. 4.20 together with flight altitudes. Again, Fig. 4.20 shows that CCN depletions shown as black valleys concur with the presence of cloud and precipitation, indicating the presence of clouds and precipitation deplete CCN efficiently; but more efficiently with precipitation.
Figure 4.20: Time Series of sub-cloud aerosols (CCN, 0.6%; black), cloud droplets (N_d) and drizzle number concentrations on 24 March 2010, showing depletion of CCNs where drizzle (shown as magenta) and/or cloud (shown as cyan dots) present. Aerosols, cloud droplets and drizzle number concentrations are measured by CCN counter, CAS and CIP probes, respectively.

4.4 Summary

Data collected from the TO research aircraft equipped with cloud, aerosol and precipitation probes are used to examine the aerosol-cloud-precipitation interactions over Eastern Caribbean Sea. The 15-TO flights sampled many clouds in various phases of growth and a wide range of aerosol conditions including the most intense African dust events. During the periods of pre- and post-dust outbreak, under the typical maritime aerosol profiles, linear relations between sub-cloud aerosol concentrations and cloud thickness are found; deeper clouds are associated with high aerosol concentrations in the sub-cloud layer. On the other hand, clouds are significantly suppressed when high concentrations of aerosols prevail throughout the entire boundary layer (e.g., 31 March). Further, the vertical structures of aerosol as well as the depth of the aerosol layer resided aloft the inversion layer play important roles on the cloud formation and development.

The aerosol-cloud interactions are assessed by examining the relations between aerosol (N_a) and cloud droplets number concentrations (N_d), and the sizes of effective
droplets (\(D_e\)) with cloud-base and flight-averaged properties. A robust increase in \(N_d\) with aerosols is found (Fig. 4.4a, c, e): \(N_d\) averaged from the entire flights for a given day and \(N_d\) obtained from cloud base flights are fairly well represented by sub-cloud aerosol concentrations \((N_a)\), with \(R=0.79\) and \(R=0.71\), respectively. In general, the effective diameter of cloud droplets \(D_e\) tends to decreases as aerosol concentrations \((N_a)\) increase when data is stratified by cloud depth. However, the inverse relation between \(D_e\) and \(N_a\) is valid only for shallow clouds (e.g., clouds shallower than 500 m) when sub-cloud \(N_a\) and cloud-base \(D_e\) are considered (Fig. 4.4f), suggesting possible effects of cloud processes such as entrainment and precipitation on the cloud properties.

Cloud-base \(N_d\) is highly correlated to the sub-cloud CCN concentrations (Fig. 4.5), in particular, in the regimes of updrafts and/or positive perturbation of \(w\) \((w' > 0)\). For instance, 95% of cloud-base \(N_d\) variation can be explained by sub-cloud CCN for the regimes of \(w > 1\) m s\(^{-1}\). On the other hand, \(D_e\) in the cloud base is poorly related to the same parameter regardless of the thresholds (not shown here).

Cloud-base droplet concentrations increase as \(w'\) increases, and the sensitivity to \(w'\) \((dN_d/dw'\) as well as \(dlnN_d/dlnw'\)) is greater in the regimes of dustier environments (high aerosol concentrations, e.g., 31 March and 5 April in Fig. 4.6), suggesting a slight increase in positive \(w'\) in polluted conditions can make greater increase of \(N_d\). The sensitivity of \(N_d\) to changes in vertical velocity perturbations \(w'\) \((dlnN_d/dlnw')\) for the dusty and pristine conditions shows a slope of 0.2 and 0.1, respectively. Twomey’s formula shows a slope \((dlnN_d/dlnw')\) of 0.3, using the typical value of \(k \sim 0.5\), indicating the parameterization overestimates the sensitivity of \(N_d\) in changes in vertical velocity perturbations for shallow cumulus observed in the trade-wind boundary layer.
The suppression of precipitation toward higher aerosol concentrations is a general feature in data sampled during the BACEX. The sensitivity of precipitation to changes in aerosol, the so-called precipitation susceptibility $S_0$, is examined to quantify this suppression while minimizing the effects of macrophysics. Data from 10 of the 12 days (excluding two major precipitating clouds on 22 and 30 March as an attempt to account for sampling biases due to the rain effects on aerosols, i.e., wet scavenging) are used to calculate $S_0$ and compared with that from all 12 days. $S_0$ (Fig. 4.9) exhibits three regimes: (1) low cloud thickness, where clouds do not precipitate, and are not sensitive to $N_d$, $S_0 \sim 0$; (2) intermediate cloud depth, where precipitation begins to decrease due to the increased aerosols (and $N_d$), $S_0$ increases to 2 at $H \sim 1100$ m, where $S_0$ peaks; (3) high cloud thickness, where $S_0$ decreases. Susceptibility peaks at intermediate values of cloud thickness. The results are consistent with previous studies of warm cumulus clouds (e.g., Sorooshian et al., 2009; Jiang et al., 2010; Duong et al., 2011), reporting $S_0$ peaks at intermediate values of cloud LWP ($\sim$600-1300 g m$^{-2}$). On the other hand, removal of $N_d$ due to the rain (wet scavenging) makes susceptibility stronger overall (dashed line shows higher $S_0$ than that of solid line in Fig. 4.9) and peaks at lower cloud thickness ($600$ m $<$ $H$ $<$ $800$ m) than that from lightly precipitating clouds ($1000$ m $<$ $H$ $<$ $1200$ m). Recently, Duong et al. (2011) also pointed out that wet scavenging leads to higher $S_0$.

Although the aerosols affect cloud and precipitation processes and characteristics (Section 4.1), small cumuli alter the aerosol properties of their immediate environment through cloud and precipitation processes (Section 4.2 and Section 4.3). Clouds effects on aerosols in the cloud near-field, associated with entrainment and detrainment process are studied in Section 4.2 and summarized in 4.2.5.
Clouds effects on aerosols, in particular, modification of aerosols in the near cloud field are examined in 4.3, in the cloud shell/halos region with aerosol structures. Cloud droplet number concentrations ($N_d$) decreases as $-w'$ increases in the regimes of $w' < 0$ (Fig. 4.6), and thus, this regime is associated with cloud dissipation. Further, downward motions are observed along the cloud edges (so-called cloud halos or cloud shell regions) from chaff tracer experiment in Section 4.2. Mie cloud and chaff clouds are exampled to show the aerosol structures in the cloud halo region. The Mie cloud has a width of about 2 km, thickness of about 1 km, while the chaff cloud has a width of about 500 m and cloud thickness of about 700 m. For the Mie cloud case, significant CCN depletions are found along the cloud edges (decrease down to 50 cm$^{-3}$); the background CCN is $\sim 200-250$ (#/cc). Further, CCN depletion is more efficient in the regime of precipitation at the flight level (e.g., around 16:17:42; near dashed line in Fig. 2.9). For the chaff cloud case, CCN depletion near cloud field is not as significant as in the Mie cloud; the background CCN is $\sim 50-150$ (#/cc). This is possibly due to the smaller size of the cloud. Moreover, the chaff cloud was purely non-precipitating cloud, while Mie cloud was a precipitating cloud. In general, CCN depletions concur with the presence of cloud and precipitation, indicating the presence of clouds and precipitation deplete CCN efficiently; but more efficiently with precipitation.
Chapter 5: Vertical Structure of Aerosols, Temperature and Moisture Associated with an Intense African Dust Event Observed over the Eastern Caribbean

In previous chapter, we see the aerosol effects on clouds and precipitation, as well as clouds effects on aerosols near the cloud field. This chapter describes the modification of boundary layer aerosol by clouds (and cloud processes), by studying the vertical structure of aerosol, temperature and moisture associated with an intense African dust event during the Barbados Aerosol Cloud Experiment.

5.1 Motivation and background

It is common to observe the westward movement of dust-laden air from Africa over the tropical Atlantic Ocean under the influence of the trade winds (Carlson and Prospero, 1972). The Saharan Air Layer (hereafter, SAL) leaving Africa is formed by intense heating over the land, and thus, characterized by a well-mixed layer that is warm and dry, and extending from the surface to about 500 hPa over Africa, for example, during the summer (Carlson and Prospero, 1972; Prospero and Nees, 1977). As the SAL moves off the coast of Africa-typically following an easterly wave-the base of the SAL is observed at higher levels (e.g., ~850 hPa in summer) as cooler and moister maritime air occupies the lower boundary layer (Carlson and Prospero, 1972; Karyampudi and Carlson, 1988; Westphal et al., 1988; Karyampudi et al., 1999; Dunion and Velden, 2004; Nalli et al., 2005, 2006; Wong and Dessler, 2005; Wong et al., 2006). Individual SAL outbreaks can cover an area of the Atlantic as large as the 48 contiguous United States and can migrate to the eastern coast of Florida (Dunion and Velden, 2004).

The appearance of the multi-layering of dust in the SAL has been reported since early 1970’s (Prospero and Carlson, 1972; Karyampudi et al., 1999; Reid et al., 2002;
Reid et al., 2003). However, different mechanisms are suggested for the formation of the multi-layering of dust; Karyampudi et al. (1999) suggested dust in the boundary layer underneath the SAL was a residual of the initial SAL. On the other hand, Reid et al. (2003) argued that heavy dust in the boundary layer observed during the Puerto Rico Dust Experiment (PRIDE; July, 2000) is transported across the Atlantic Ocean with its vertical character already defined as it leaves Africa. Although the general features of dust outbreaks over the ocean are known, questions remain regarding the vertical transport and processing of dust, and details of the vertical structures of dust layers and the characteristics of each layer.

Satellite based studies are useful to study aerosol indirect effects over a large geographical area with long time periods, but the adiabatic assumption is not valid for warm cumulus cloud (as seen in Fig. 3.16a, for instance). Further, the vertical distribution—a key component of aerosol indirect effect, as shown in Chapter 4—is usually unknown. Lidar observations provide an excellent tool for illustrating the vertical distribution of aerosol. However, both ground-based and space-borne lidar observations are not always able to define the fine vertical structure of dust when clouds are present.

The vertical and horizontal distributions of aerosol observed on the 15 flights with an instrumented research aircraft (Fig. 3.5) show a wide range of aerosol conditions that include the most intense African dust event observed at the Barbados Ragged Point surface site during all of 2010. In this chapter, two days of an intense African dust event (1-2 April), when clouds were significantly suppressed, are focused. I combined the in situ aircraft data, which were obtained in an area just upstream from the Barbados Ragged Point surface site, with soundings at Barbados and western Africa to study the
lower tropospheric thermodynamics (i.e., temperature and moisture) at the source (Africa) and the local (Barbados) sites, and the vertical structure of aerosol and thermodynamics over the Eastern Caribbean. The general information on flight patterns, profiles of thermodynamics and various aerosols are described in Section 5.3 along with observations from a Micro Pulse Lidar (MPL). Transports and processing affecting the observed aerosol structures are described in Section 5.4. The section includes descriptions of mixing diagrams, particle size distributions, and hygroscopicity parameter, kappa. Summaries are given in Section 5.5.

5.2 Between source (Africa) and site (Barbados)

To get a general idea of when the air mass left Africa and, thus, how long the SAL took to cross Atlantic, 315-hr (13-day) back trajectories, arriving at 500 m in Barbados at 17 UTC on 1-2 April were calculated (Fig. 5.1). The trajectories indicate that air mass, arriving in Barbados at 17 UTC on 1 and 2 April 2010, moved off the African coast on 22 March and on 26 March, respectively, indicating a 7-10 day transit cross the Atlantic Ocean.

The origin of this dust event and its westward movement, across the Atlantic, is illustrated by the MODIS satellite images on 22 and 29 March, shown in Fig. 5.2. These images show that the event originates over Africa around 22 March (Fig. 5.2b) and moves to a longitude of ~40 °W (western most, in Fig. 5.2a) 7 days later, then arrives at Barbados at around 1 April (not shown). This progression is roughly consistent with the back trajectories in Fig. 5.1.
**Figure 5.1:** The 13-day back trajectory arriving at 500 m in the middle of the BACEX flight domain for 17 UTC 1 April (magenta) and 2 April (cyan) 2010. Trajectories of each day at 00 UTC are denoted as cross symbols and several dates of back-trajectory are shown accordingly. Locations of sounding stations at Africa are added as cross symbols.

**Figure 5.2:** MODIS satellite images on (a) 29 March, 2010 over the Atlantic and on (b) 22 March, 2010 over western Africa. Images were downloaded from the MODIS website (http://modis-atmos.gsfc.nasa.gov/IMAGES/index.html).

The MODIS satellite images from 19 March to 2 April were compared with the trajectories in Fig. 5.1, and show good qualitative agreement. Thus, based on the satellite images and the back trajectories, I estimate the transit time of the dust event from Africa to the Barbados area to be 7-10 days. The unusually long transit of this event (7-10 day of
transit compared with a typical 5-7 day transit during the summer, e.g., Prospero and Carlson, 1972) may relate to the large scale circulation pattern, but it is beyond the scope of this study. The trajectories in this study agree well with the satellite images, nevertheless, the uncertainties on the trajectories are discussed in the Appendix C.

A massive dust area extending from Africa to the Atlantic Ocean is apparent in the MODIS image shown in Fig. 5.2. The axis of this dust area sloped to the south slightly during the transit, starting from 6-20 °N at the coast of Africa and then centered around 12 °N over the Atlantic Ocean. These dust dominant regimes look milky and turbid. In addition, the clouds are significantly suppressed in the regimes of heavy dust and, the low-level clouds, embedded in the dust, appear as fuzzy entities.

![Figure 5.3: Profiles of (a) potential temperature and (b) mixing ratio on 21 March 09:00 UTC at GOTT, 22 March 12:00 UTC at GOOY, in Africa, and on 1-2 April 12:00 UTC at Barbados. Profiles of wind direction on 21 March 09:00 UTC at GOTT, 22 March 12:00 UTC at GOOY in Africa are shown in Fig. 5.3(c). SAL observed at Barbados is denoted as arrows.](image-url)
Profiles of potential temperature, $\theta$ and mixing ratio, $r$ calculated from the Barbados rawinsondes on 1-2 April 2010 are shown in Fig. 5.3, along with two soundings (GOTT and GOOY) from western Africa (locations shown in Fig. 5.1) on 21 and 22 March corresponding the time period of the MODIS images shown in Fig. 5.2. The air in and above the SAL observed at Barbados probably has not experienced significant moist processes. The satellite images during the 7-10 days of the transit, across the Atlantic, indicate no deep convection in the dust area, although some low-level clouds, embedded in the dust, are visible from satellite. Thus, in the SAL, one would expect that mixing ratio would be conserved.

The SAL, shown as well-mixed layer in $\theta$ and $r$, over Africa extends from 500 m to $\sim$4000 m ($\sim$3.5 km depth), but the top extends only to $\sim$2650 m with a depth of $\sim$800 m at Barbados after traveling over Atlantic, suggesting subsidence of SAL $\sim$1250 m during 10 days ($\sim$125 m day$^{-1}$). The potential temperature (in the SAL) at Africa (Fig. 5.3a) is approximately 311-312 K, but is about $\sim$307 K at Barbados, 10 days later, indicating a 4-5 K cooling over 7-10 days (a cooling rate of about 0.5 °C day$^{-1}$). The moisture profile in the African SAL (Fig. 5.3b) has a mixing ratio of $\sim$4 g kg$^{-1}$, but a 5-6 g kg$^{-1}$ mixing ratio of SAL is observed at Barbados. The differences in mixing ratio between Africa and Barbados may involve uncertainties about the locations and times of the SAL over Africa, and thus, we may see slightly different air mass within the SAL (as will be shown in 5.3.2). Regardless, here, we consider the mixing ratio difference of $\sim$1-2 g kg$^{-1}$ in the SALs between Africa and Barbados to not be significant. The wind direction for the two African soundings (Fig. 5.3c) indicates easterly winds throughout the layers, except for northerly winds close to the surface in the GOOY station that is close to the coast. As the
SAL travels over the Atlantic, the boundary layer is modified by dry and moist convection; the difference between two groups of soundings (reddish from Africa and black/gray from Barbados) below the SAL (below ~1700 m, Fig. 5.3), indicates the modification of boundary layer thermodynamics (and aerosols, as will be shown later) by cloud processes.

We can estimate the fluxes needed to modify the lowest layers of a deep mixed layer that originates over Africa. If we assume that the lower layers of the initial African soundings (0 to about 1700 m) are moistened by moisture fluxes from the surface and convective transports within the boundary layer, we can make a rough estimate of the average surface latent heat, along the trajectory using a simple moisture budget written as

$$\frac{d\bar{r}_l}{dt} = -w \frac{\partial \bar{r}_l}{\partial z} - \frac{\partial}{\partial z} \left( w' r_l \right),$$

(Equation 5.1)

where \( r_l = r_v + r_i \) is the total water mixing ratio, and the over bar indicates a time/area average. On the left hand side of this expression, we assume that the time rate of change of the liquid water mixing ratio can be neglected and the total derivative is the horizontal component of the total derivative, and \( w \) is the vertical air velocity. The total water mixing ratio is used in the flux divergence term since moist convective transports (which include cloud transports of water) can alter the large-scale water vapor budget. If we consider a time average along the 10-day trajectory, the budget equation can be applied in this case, by using the difference in the mixing ratio between the African and the Barbados soundings (Fig. 5.3). Further if we assume that the vertical advection term average for the 10-day period is estimated to be roughly 0.5 g kg\(^{-1}\) per day (\( w \) is assumed to be about 60 m day\(^{-1}\) and the mixing ratio vertical gradient is about 10 g kg\(^{-1}\) km\(^{-1}\)). By assuming that the convective moisture flux at the top of the boundary layer is zero, the
moisture budget equation can be integrated from the top of the boundary layer downwards to give a vertical distribution of the total water flux that gives the surface moisture flux needed to moisten the boundary layer during the 10-day transit from Africa to Barbados. The surface moisture flux (latent heat flux) from this estimate is ~60-70 Wm\(^{-2}\). We conclude that the surface fluxes inferred on average along the 10-day trajectory are reasonable (Wu et al., 2007) and thus could easily account for the observed moistening.

Another possibility is that the air masses below the SAL (surface to ~1700 m) might have different source regions than the SAL (1700 m-2500 m) in Barbados, and thus represent trajectories from very different areas. To investigate this possibility, the air mass extending from the surface to the top of the SAL that flows across the Atlantic was examined with NCECP reanalysis data. The reanalysis wind fields, consistently show easterly at the surface, 900 hPa, 800 hPa and 700 hPa, over the Atlantic, from Africa to Barbados, indicating flows near the surface are similar to those in the SAL (not shown here). This deep easterly flow is also consistent with the wind directions in the SAL shown in Fig. 5.3c. Thus, the difference between two groups of soundings in the moist profile below SAL is not likely explained by this hypothesis. Further, the relative simplicity of the flow may help explain why the long-term trajectories show good agreement with the observed advection of the dust mass across the Atlantic.

5.3 Observations

5.3.1 Flight paths

On each of the two days studied, Twin Otter flights were made about 50-100 km upwind of Ragged Point and included several constant-level legs (about 50 km in length),
flown at heights from near the surface to about 3 km that were tied together by soundings, made typically at the beginning and at the end of a sequence of level legs (Fig. 5.4b, Fig. 5.4d).

**Figure 5.4:** (a, c) Flight paths, (b, d) time series of flight altitudes on (a, b) 1 April and (c, d) 2 April, 2010. Numerical numbers and colors are assigned to the individual level runs accordingly. Dashed lines in Fig. 5.4(a) and Fig. 5.4(c) represent the first (up1) and second (up2) ascent soundings while solid lines indicate data from level runs (leg 1 through leg 7). The starting point of each leg is denoted as cross symbol in Fig. 5.4(a) and Fig. 5.4(c). The Barbados Ragged Point surface site is located at 13.2 °N, 59.5 °W.

On 1 April 2010, no clouds were observed over the course of the 3-hr flight. The aircraft made seven level runs (Figs. 5a-b), starting from the top of the SAL (leg 1) to ~30 m above the surface (leg 7) over the course of ~2 hr. The first sounding (up1) and level runs (leg 2 to leg 7) were made in the same upstream area from Ragged Point (Fig. 5.4a). The second sounding (up2) was made slightly north (~2 km) of the first sounding. On 2 April 2010, two layers of heavy dust (with very heavy cirrus overcast advected from South America) were observed. There was a very occasional cumulus clouds, generally in tiny patches and almost all very optically thin. Flight paths and time series of the altitudes
flown on the day are shown in Fig. 5.4(c-d). The aircraft made seven level runs for the day (Fig. 5.4d) as on 1 April, but flight legs between 16:30 and 17:00 UTC (legs 5-7) were performed on the NE edge (~6 km) of the main study area (Fig. 5.4c). However, the difference (~2 km in Fig. 5.4a and ~6 km in Fig. 5.4c) is negligible when the mean wind speed of 2-8 m s\(^{-1}\) on the days is considered.

### 5.3.2 Thermodynamic and aerosol profiles

Atmospheric thermodynamic (potential temperature and mixing ratio) profiles and aerosol concentrations from PCASP are shown in Fig. 5.5 for the soundings obtained on the two flights. Observations from the level legs and the descents between these legs are also shown.

The vertical profiles of potential temperature, vapor mixing ratio, and aerosol concentrations obtained from PCASP (hereafter, simply PCASP) in Fig. 5.5, indicate three layers with distinct aerosol and thermodynamic characteristics. The definition and illustration of the layers, used in this study, are based on the first sounding (up1) from 1 April shown in Fig 5.5a. A layer above the trade wind boundary layer (at about 1900 m to 2700 m) has well-mixed potential temperature, mixing ratio, and aerosol concentrations in the vertical, and will be referred to as the SAL. A layer extending from the surface to about 500-600 m is well mixed and capped by a stable layer (a narrow layer sits slightly above 500 m, referred to as the transition layer) where mixing ratio decreases with height. This inversion (the transition layer) is sufficiently strong to inhibit moist convection during most of the two flights (the strength of transition layer is about 20 °K km\(^{-1}\) on 1 April and 7 °K km\(^{-1}\) on 2 April). But for the traditional trade wind boundary layer (e.g.,
Albrecht et al., 1979), a layer extending from the surface to about 500-600 m would be the sub-cloud layer (SCL). The third layer extending between the SAL and SCL will be referred to as the intermediate layer (IL) and displays the greatest temporal and spatial thermodynamic and aerosol variability. In terms of the classic trade wind boundary layer structure (e.g., Augstein et al., 1974; Albrecht et al., 1979) this IL would be called the cloud layer and is capped by the trade inversion.

The variability in the thermodynamics and aerosol structure is illustrated by the difference between the first and second soundings. The second sounding was made about 2 hours after the first one. A dry layer is found above the SAL on both soundings on 2 April (above 2300-2400 m in Fig. 5.5e) and in the second sounding on 1 April (above 2600 m in Fig. 5.5b); but the first sounding on 1 April did not extend high enough to sample this layer.

In Figs. 5.5(a-c), the base of the SAL is about 200 m lower in the second soundings and, the largest aerosol variations occur in the IL. In addition, the second sounding on 1 April shows an intermediate layer that is warmer, drier and dustier (this assumes that the aerosol in the intermediate layer are dust) than that observed on the first sounding.

Although, about two hours of flight time separate the two soundings made on 2 April, the atmospheric thermodynamics and aerosols observed in the SAL and the sub-cloud layer show little change. However, alternating cool and warm (moist and dry) layers are observed throughout the intermediate layer; the cool and moist layers are observed in Fig. 5.5(e), between 600 m and 1000 m, which includes leg 3 and leg 4, and between 1450 m and 1700 m (close to base of the SAL). The base of the SAL is about
100 m higher in the second sounding during this flight. Relatively lower concentrations of aerosols are observed in these cool and moist layers. These layers are most likely associated with modification by episodic shallow cumulus convection (through detrainment), sometime in the previous history of the air mass as it advected across the Atlantic. The intermittent nature of these convective events may explain the variability observed in the IL compared with that in the SAL and in the SCL.

The horizontal variability within the IL, is also illustrated by the large variations found on the level legs (e.g., legs 3-5 for 1 April and legs 2-4 for 2 April), flown in this layer. The temperature and moisture in the SAL show little variability between soundings and, little variability along the level legs flown in this layer.

On 2 April (Fig. 5.5e), a relatively dry layer (mixing ratio ~4 g kg⁻¹) is observed between leg 1 and leg 2 during the descent, and the value is consistent with mixing ratio found in the SAL from the African soundings in Fig. 5.3, suggesting that there is basically uncertainty in an air mass sampled at Barbados and that sampled from a fixed site in Africa as discussed earlier in 5.2. The SAL may have a slight moisture variability in the horizontal (for example, ~1-2 g kg⁻¹), depending on the relative location from the center of the SAL. Air masses, sampled between leg 1 and leg 2 show two distinctive thermodynamic characteristics (temperature and moisture); similar to that of either SAL or IL (not shown here).
Figure 5.5: Profiles of (left) potential temperature $\theta$, (middle) water vapor mixing ratio $r$, and (right) PCASP aerosol concentrations from aircraft during ascents. Data (10 Hz) from each level run are denoted by black bold colors and numbered accordingly. Thin black dots represent data obtained during descents between two consecutive legs. For 1 April 2010 (upper panel); Leg 1: top of the haze layer, Leg 2: middle of Saharan air layer (SAL), Leg 3-Leg 5: intermediate layer (IL), Leg 6-7: sub-cloud layer (SCL). For 2 April, 2010 (lower panel); Leg 1: middle of Saharan air layer (SAL), Leg 2-Leg4: intermediate layer (IL), leg 5-7: sub-cloud layer (SCL). The layers of Saharan air, intermediate and sub-cloud are labeled in Fig. 5.5(a) on the first sounding from 1 April to illustrate the layers.
5.3.3 Vertical profiles of various aerosol characterizations

The vertical structure of the aerosols observed on the two flights show some complicated structures (Figs. 5.5c and 5.5f), but general features that are common to both days. To examine these structures in more detail, profiles of CN, CCNs, and PCASP were examined for the soundings made on the 1-2 April 2010 (Fig. 5.6).

The layered structures are evident in the soundings made on both days. (Note that aircraft soundings represent both vertical structure and horizontal variability). In the SAL, the concentrations of aerosols are greater than those of any of the other layers, and very low aerosol concentrations are found above the SAL (Fig. 5.6b-d). The SAL in the second sounding on 2 April (Fig. 5.6d) is thinner with its top at about 2300 m and its base is about 200 m higher (or slopes toward south) than that in the first sounding in Fig. 5.6c.

Aerosol concentrations in the intermediate layer show the largest variability; for the first sounding on 1 April (Fig. 5.6a), aerosol concentrations in the IL are slightly lower than those in the sub-cloud layer. However, the concentrations increased during the course of 2-hr separation in the soundings and show aerosol concentrations that range between those of the SCL and those of the SAL in Fig. 5.6b.

For the first sounding on 2 April (Fig. 5.6c), two layers with heavy aerosol are observed near Leg 1 (middle of the SAL) and near the bottom of the IL. However, the layer of higher aerosol concentrations at the bottom of the IL is not present in the second sounding (Fig. 5.6d); throughout the layer below leg 2 (surface ~1250 m), aerosol concentrations show relatively well-mixed features but with a slight increase with height.
In some parts of the intermediate layer, the aerosol concentrations are lower than that in the sub-cloud layer, providing further evidence for possible cloud processing of the aerosols in the intermediate layer. Aerosol concentrations in the sub-cloud layer remain relatively constant (~200-400 cm$^{-3}$) during the course of flights.

The CCN (0.6 %) and PCASP profiles are in good agreements throughout all the layers. But there are large differences between CCNs (0.3 % and 0.6 %; blue and red lines) as well as CCN (0.3 %) and PCASP (blue and black lines) below the heights of ~2400 m on both days. The largest differences are found near the center of the SAL, and the difference decreases closer to the surface. At the lower super-saturation (CCN 0.3 % compared with CCN 0.6 %), large particles are activated first. At the higher super-saturation (CCN 0.6 % compared with CCN 0.3 %), smaller particles can be activated too. Thus, the large difference between the two CCN profiles indicates the presence of small particles at those levels and/or that those particles are more hydrophobic.

CCN (0.3 %) concentrations increase rapidly with height, above about 2400 m on both soundings on 1 April (Fig. 5.6a-b), although the potential temperature and mixing ratio are well-mixed within the SAL (Fig. 5.5), and approach the CCN (0.6 %) and PCASP values at around 2500 m. The collapses of the two CCN profiles and the PCASP profiles indicate that all the particles (small + big) at this level, are activated and/or particles at this level are more hydrophilic. In addition, in the layer above the SAL on 2 April (Figs. 5.6c and 5.6d), the profiles have CCN concentrations that exceed those from PCASP, indicating the possible presence of small (less than 0.1 μm-the lowest resolvable size from the PCASP) hydrophilic particles in this layer. These points will be further discussed in 5.4.2.
Figure 5.6: Profiles of various aerosols from various probes for (a, c) first (up1) and (b, d) second (up2) sounding on (a-b) 1 April and (c-d) 2 April 2010. CN (green), PCASP (black), CCN (0.3 %; blue) and CCN (0.6 %; red) are shown. Average heights of individual levels (~10 minute) runs are denoted as thick gray horizontal lines (top down: leg 1 to leg 7). CCN (0.3 %) indicates cloud condensation nuclei (CCN) activated at super-saturation of 0.3 %. Same notation for CCN (0.6 %). Information of data acquisition time and location is shown in Fig. 5.4.
5.3.4 MPL Aerosol returns

A further indication of the time (and implied spatial) evolution of the vertical structure of aerosols at the Ragged Point surface site, just downstream from the aircraft observations, is given by the lidar backscatter profiles. The Normalized Relative Backscatter (so-called NRB signals, level 1 data), obtained from MPLNET on 1-2 April 2010, are shown in Fig. 5.7. The MPL backscatter profiles for the two days include observations made during the aircraft flights, although the MPL is sheltered from the sun for a 1.5 hr period around solar local noon (17 UTC).

![Figure 5.7: Time-height cross-section of NRB (Normalized Relative Backscatter) obtained from Micro-Pulse LIDAR (MPL) at Ragged point in Barbados during 1-2 April 2010. Figures were obtained from MPLNET (http://mplnet.gsfc.nasa.gov/). The LIDAR is blocked from the sun for a 2-hour period centered on around local noon and indicated by the dark areas in this figure. Local time=UTC-5.](image)

The backscatters from the SCL, IL, and SAL show some of the same features shown by the vertical profiles from the aircraft observations. On 1 April, the top of the SAL increases from about 2 km to ~2.7 km from 00 to 14 UTC. From 18 UTC to about 12 UTC on 2 April, the top of the SAL fluctuates between 2.5 and 2.8 km (depicted as bluish colors). The top of the layer descends slowly after 12 UTC on 2 April, and a two layered structure also appears after 18 UTC. The separation of this upper layer is at a
height of about 2.4 km. The descending features (of the top of the SAL and a layer above SAL) are consistent with the temporal, and thus spatial, variability in the SAL shown by the soundings in Fig. 5.5(f).

The lidar signal is attenuated backscattering. As a result, the signals decrease in strengths with height even if the actual aerosol concentrations are higher aloft. Nevertheless, the NRB is useful to infer the interrelationship between clouds and aerosols by showing signals from both aerosols and clouds. In Fig. 5.7, a multi-layer structure is noticeable from 12 UTC on 1 April to the end of 2 April, especially during the period of Twin Otter flights (15-18 UTC). The elevated high backscatter at ~500-1000 m (shown as white color) indicates clouds with bases near the bottom of the intermediate layer. These cloud returns are quite numerous, prior to the flight on 1 April, but no clouds were observed during the flight. On 2 April, one strip of strong backscatter, an indicative of cloud, is seen near 1800 UTC. During the time of the flight, the entire layers, below 2 km, are well-mixed without reddish colors in the sub-cloud layers and are consistent with a weaker transition layer and IL on 2 April than 1 April (Fig. 5.5a, Fig. 5.5d). The lidar backscatter returns also show variability on scales that are consistent with the variability observed in the intermediate layer during the two flights.

5.4 Discussion of transports and processes

5.4.1 Mixing diagram

The possible processes affecting the nature of the observed aerosol layers are examined with mixing diagrams using conservative variables. In this case, the water vapor mixing ratio and number concentration of aerosol per mass are assumed to be the
conserved parameters and would mix almost linearly under dry condition. Mixing diagrams, formulated using mixing ratio and CCN (0.6 %) per mass, are shown in Fig. 5.8 to investigate the role of mixing process on aerosol distributions observed on 1-2 April 2010. The water vapor mixing ratio is plotted with decreasing values on the y axis; thus giving a rough sense of height.

![Figure 5.8: Mixing diagram of CCN (0.6 %) per mass (µg⁻¹) and mixing ratio (g kg⁻¹) on (a) 1 April and (b) 2 April, 2010. Black squares are obtained from the ascent environmental soundings and colored dots are obtained from each level run. (a): leg 1 (top of the SAL), leg 2 (SAL), legs 3-5 (IL), legs 6-7 (SCL), (b): leg 1 (SAL), legs 2-4 (IL), legs 5-7 (SCL).](image)

The mixing diagrams for both days show distinct clusters that are indicated by both level leg measurements and those from the profiles. For 1 April, the most significant variations in both moisture (mixing ratio) and aerosol concentration are observed in the upper/middle part of the intermediate layer. For example, leg 3 in Fig. 5.8a shows two distinct populations of aerosol and moisture, suggesting strong horizontal variations at that level with two distinctive air masses--clusters of air masses with mixing ratios either higher or lower than 10 g kg⁻¹. Aerosols obtained at this level lie on a mixing line
extending from the leg 1 (top of the SAL) and leg 2 (SAL) points. This mixing line is interpreted as aerosols at leg 3 sharing properties with aerosols at leg 1 and leg 2. Aerosols in leg 5 (lower IL), shown by magenta dots, connect to aerosols in leg 3 as well as those in leg 1 and leg 2, suggesting, aerosols in leg 5 were mixed with aerosols in those levels.

The sounding extends to the air at the top of the SAL and identifies dry and low aerosol concentrations (e.g., clusters with mixing ratio of ~4 g kg\(^{-1}\) and aerosol concentration of ~300 µg\(^{-1}\)). This air mass connects to mixing lines extending to the SAL (clusters with mixing ratios of about 6 g kg\(^{-1}\) and aerosol concentrations of 500 µg\(^{-1}\), shown as blue dots) and the SAL air connects to the air mass in the intermediate layer as seen above at leg 3 and leg 5.

The SAL shows homogenous thermodynamic conditions, but varying CCN. For example, aerosols at leg 1 (top of the SAL) and leg 2 (SAL) on 1 April (Fig. 5.8a), display similar moisture characteristics (mixing ratio of ~5-6 g kg\(^{-1}\)), while they show large variations in aerosol concentrations (400-600 µg\(^{-1}\) at CCN 0.6 %). In contrast, the aerosols in the sub-cloud layer (for example, leg 7) show similar mixing ratios (~17 g kg\(^{-1}\)) and aerosol concentrations (~250 µg\(^{-1}\)) with a circular distribution rather than horizontally elongated pattern as seen in legs 1 and 2. This pattern is more clearly seen in the mixing diagrams shown later in Fig. 5.9 where the variability within the various clusters is illustrated using CCN and CN observations.

The mixing diagram for 2 April is shown in Fig. 5.8b. The largest variability in the thermodynamic and aerosol characteristics is found on leg 2 in a layer below the SAL, as on the leg 3 for 1 April in Fig. 5.8a. Aerosols sampled on leg 1 (SAL, blue dots) are
associated with lower mixing ratios than those of the sounding, sampled at the same height (e.g., clusters with mixing ratio of ~6 g kg\(^{-1}\) and CCN of ~400-500 µg\(^{-1}\)), indicating significant horizontal variations of aerosol at the same height, and suggesting that aerosols in this region (leg 1, SAL) are associated with a drier air mass than that of sounding as shown in Fig. 5.5e.

![Mixing diagram of aerosol concentrations (µg\(^{-1}\)) for CCNs and CN versus mixing ratio (g kg\(^{-1}\)) for each level run, for (a-e) 1 April, and for (f-i) 2 April, 2010.](image)

**Figure 5.9:** Mixing diagram of aerosol concentrations (µg\(^{-1}\)) for CCNs and CN versus mixing ratio (g kg\(^{-1}\)) for each level run, for (a-e) 1 April, and for (f-i) 2 April, 2010.

Mixing diagrams for each level run for 1 and 2 April, 2010 are shown in Fig. 5.9. Aerosols in the SAL (Figs. 5.9a, 5.9b, 5.9f) show homogenous (well-mixed) thermodynamic conditions but varying CCNs as indicated by horizontally elongated
pattern, while aerosols in the SCL (sub-cloud layer; Figs. 5.9e, 5.9i) show similar thermodynamics and aerosol characteristics shown as a circular pattern. Transports of aerosols from the SAL to the intermediate layer are seen in upper to middle part of the intermediate layer (IL) in Fig. 5.9c; Transports of aerosols within the intermediate layer, from upper to lower parts of the layer, are seen in Fig. 5.9g. Further, a connection of aerosols from the lower part of the intermediate layer to sub-cloud layer is seen in Fig. 5.9d and Fig. 5.9h.

5.4.2 Size and composition

Aerosols Particle Size Distributions (PSDs) from PCASP differ among the layers, and provide some insight into the processes operating on these various layers that affect the concentrations. The PSDs, in terms of $dN/d(\log D)$ and $dV/d(\log D)$, from the top of the SAL to the sub-cloud layer on 1 and 2 April 2010 are shown in Fig. 5.10. The PSDs at all levels have a maximum concentration of aerosols, in the range of $D < 0.5 \, \mu m$, with a maximum at 0.15 $\mu m$ in the fine mode (Fig. 5.10a); and show a secondary maximum at 0.8 $\mu m$ for $D > 0.5 \, \mu m$, representing the coarse mode.

For 1 April, the largest population of small particles ($D < ~0.5 \, \mu m$) is found near the top of the SAL (leg 1), although the larger particles ($D > ~0.5 \, \mu m$) in this layer have smaller concentrations than those in the SAL and intermediate layers (legs 3-5 in Fig. 5.10a). Among the intermediate layer profiles (legs 3-5), the upper part of the intermediate layer (leg 3) has more larger particles (especially larger than 1 $\mu m$) compared with the PSD in leg 5, which is closer to the bottom of the intermediate layer, indicating the presence of larger particles in the upper part of the intermediate layer (Fig.
5.10a). Here, the upper part of the intermediate layer (leg 3) is drier than lower part (leg 5) of IL. Aerosol concentrations in the sub-cloud layer (legs 6-7 in Fig. 5.10a) show consistently the lowest values during the flight, indicating that the main source of aerosols on this day is aloft, and not from the surface.

Figure 5.10: Aerosol particle size distributions for (a) 1 April 2010 from top of the SAL to sub-cloud layer and for (b) 2 April 2010 from SAL to sub-cloud layer.
Since similar results are shown for the second case (2 April 2010), only a couple of distinct features are discussed here. One difference between the 1 and 2 April cases, is the presence of clouds on 2 April from samples taken near the bottom of the intermediate layer (legs 3-4), although the clouds are small and few in number. Since the aerosol probes (e.g., PCASP and CPCs) are known to have poor accuracy inside clouds, the aerosol data obtained from in-clouds were excluded from the analysis. On leg 1, in Fig. 5.10b, the SAL has the most numerous particles over all ranges of sizes. In contrast, the sub-cloud layer has the lowest concentration of particles over all ranges, except at D ~0.2 µm. Aerosols obtained at the upper part of the IL (leg 2 in Fig. 5.10b) show the lowest concentrations of particles at D ~0.2 µm, and the second highest concentrations of large particles (especially D > 0.5 µm) and suggesting the removal of fine particles and augmentation of coarse particles at the level.

To examine the effects of gravitational settling on changing the PSD, fall speeds were calculated. Particles with diameters of 0.15 µm and of 0.8 µm, which correspond to the two main peaks in PSDs, have fall velocities of approximately 5.8 cm day⁻¹ and 1.6 m day⁻¹, respectively. Fall speed of particles in the range of PCASP (0.1 µm - 2.5 µm) varies from 0.26 m to 160 m during 10 days. Therefore, differential gravitational settling is not important for explaining any size distribution variability as a function of height.

Profiles of mean CCN (0.3 % and 0.6 %) at each level with ±1σ are shown in Fig. 5.11. To examine the contribution of small particles, the difference between CN and PCASP that shows the aerosol concentrations sizing between 3 nm and 0.1 µm at each level, is also presented in Fig. 5.11. The difference (CN-PCASP), shown as black solid line, decreases from the SAL to the upper IL, and increases slightly towards the bottom
of the IL, and then increases sharply near the surface. The lowest aerosol concentration of the small particles (3 nm < D < 0.1 µm) is found in the intermediate layer (upper IL on 1 April and cloud layers on 2 April), and indicates some processing in this layer that removes the smaller particles. The larger small particle concentrations near sea surface indicates a possible source there, possibly due to the particles formed at the sea surface by bursting bubbles and consisting mostly of organic material (de Leeuw et al., 2011).

**Figure 5.11:** Mean profiles of CCNs (blue; CCN 0.3 %, red; CCN 0.6 %) at each level leg for (a) 1 April and (b) 2 April, 2010. The difference of number concentrations between CN and PCASP at each level is also shown (black squares). Horizontal bar indicates ±1σ from the mean value at each leg.

The differences between the CCNs decrease closer to surface (as seen in Fig. 5.6), indicating the presence of larger particles and/or the presence of more hydrophilic particles near surface. In this case, presence of more hydrophilic particles is expected.
near the surface by referring to the differences between CN and PCASP. Profiles of potential temperature and ultrafine aerosol concentrations for the second soundings on 1-2 April are shown in Fig. 5.12.

![Figure 5.12](image)

**Figure 5.12:** Profiles of (a, c) potential temperature (black solid line), CCNs (0.3 %, blue; 0.6 %, red), and (b, d) ultrafine aerosol concentrations (red, $3 < D < 15$ nm; blue, $3 < D < 10$ nm; and green solid lines, $3 < D < 15$ nm) for the second soundings on (a) 1 and (b) 2 April, 2012. Average heights of individual level-runs are denoted as thick gray horizontal lines (top down: leg 1 to leg 7) in Fig. 5.12(a) and Fig. 5.12(c).

Ultrafine aerosol concentrations in three size ranges are obtained from the difference between CPCs; $3 \text{ nm} < D < 10 \text{ nm}$, $3 \text{ nm} < D < 15 \text{ nm}$, and $10 \text{ nm} < D < 15 \text{ nm}$. 
These differences were averaged five times over 12-second intervals to smooth out small scale variability. Horizontal variations in the ultrafine were observed on the level-legs in SAL (not shown) and may explain the variability observed in the profiles. Concentrations of aerosol particles sizing between 3-10 nm and 3-15 nm are significantly higher in the SAL (~50 cm⁻³) than those in the other layers (e.g., less than 20 cm⁻³). However, aerosol concentrations between 10-15 nm are close to zero, indicating that the ultrafine particles are confined to 3-10 nm range. In contrast, the separations between the CPCs are evident (e.g., Fig. 5.12d) in the clean layer above the SAL; here 20 cm⁻³ of aerosol concentrations in the range of 10 nm < D < 15 nm; 30 cm⁻³ for the range of 3 nm < D < 10 nm.

In the SAL, newly formed extremely fine particles (3 < D < 10 nm) may promptly transfer to the larger size without leaving particles, ranging between 10-15 nm. But in the layer above the SAL (layer A in Figs. 5.12c-d), where CCN and PCASP concentrations are very low, thus, the accumulation of the newly formed ultrafine particles on larger particles will be inhibited. As a result, particles ranging between 10-15 nm are observed. The existence of extremely fine particles (3 < D < 10 nm) in the SAL, possibly implies a source of aerosol in this layer. A layer shown as B in Fig. 5.12a, CCN (0.3 %) increases rapidly near the top of the SAL and lies on top of CCN (0.6 %) as seen in Fig. 5.6, indicating the presence of abundant small hygroscopic particles. In layer C (middle of the SAL in Fig. 5.12), significant difference between CCNs (0.3 % and 0.6 %) are found, implying two possibilities; one, there are few large particles (i.e., plenty of small particles); or two, the particles are hydrophobic.

To examine the difference of size distributions between layers, PSDs, obtained from above the SAL (A-clean environment), at the very top of the SAL (B-a layer with
similar CCNs) and in the middle of the SAL (C-a layer with huge difference between CCNs) are shown in Fig. 5.13.

**Figure 5.13:** Aerosol particle size distributions calculated from the layers of A, B and C in Fig. 5.12, from the second soundings of (a) 1 April 2010 and for (b) 2 April 2010. The layer A represents a layer above SAL, layer B indicate the very top of the SAL (Fig. 5.13a) and a transit layer from the very top of the SAL to a layer above the SAL (Fig. 5.13b), and a layer C indicates the middle of the SAL. The locations of three layers are shown in Figure 5.12.

In Fig. 5.13a, PSDs obtained from the top of the SAL (B) and those obtained from the middle of the SAL (C) are almost identical in shapes and concentrations, indicating that the entire SAL is well-mixed and thus gives similar size distributions, although layers B and C have substantially different CCN characteristics. Similarly, PSDs obtained from layers B and C on April 2 (Fig. 5.13b) also show similar shapes in PSDs, but a PSD obtained from a layer B shows lower aerosol concentrations over all ranges of the sizes. Considering both particle size distributions and CCN characteristics particles that possibly are coated by more hydrophilic materials are speculated to exist at the very top of the SAL (layer B) where the trade wind inversion located.
The PSD obtained from above the SAL (A, in Fig. 5.13b) shows a clearly different shape of distribution from the other layers. Concentrations monotonically decrease with size and show no peak at either at ~0.2 µm or at ~0.8 µm as shown from the other dust associated layers. It indicates that particles from above the SAL have substantially different characteristics than those of the layers below (e.g., particles in the SAL, IL and SCL).

To infer the composition of aerosols at each level, first, critical diameters ($D_c$) for the CCN (0.3 %) measurement were obtained by counting the particle numbers from the largest size, until the accumulated number equals the number of CCN at the level shown in Fig. 5.11, for instance. Then, the hygroscopicity parameter, kappa ($\kappa$), for each level was obtained following Petters and Kreidenweis (2007) using these critical diameters. Critical diameters ($D_c$) and kappa ($\kappa$) for 1-2 April are summarized in Table 5.1 for the CCN (0.3 %) measurements at each level. Number concentration difference between CN and PCASP (CN-PCASP) are also shown in Table 5.1.

For 1 April, the kappa values derived from 0.3 % measurement are fairly consistent throughout all the layers: on average 0.026 (within a 10 % standard deviation) and indicating the same species of aerosols (in this case, dust; kappa for pure dust = 0.01, Petters and Kreidenweis, 2007) at all levels. The kappa values decrease slightly from the SAL to the upper/middle part of the IL, then increase closer to surface, indicating that the particles are becoming more hydrophilic. For 2 April, in general, kappa values have increased slightly compared with previous day, possibly by aging or cloud processing (the addition of hygroscopic species via aqueous-phase processing). Kappa values are fairly consistent throughout the all layers with a value of 0.02-0.03. However, kappa for
leg 2 (a layer below SAL) has the smallest value among the other, indicating particles in this layer are more hydrophobic (Note that no estimation is available in SAL for the day). Further, kappa values are higher in legs 3 and 4, where clouds formed, possibly indicating cloud processing. The patterns of kappa are consistent with those of CN-PCASP in Fig. 5.11.

Table 5.1: The critical diameters ($D_c$) and kappa ($\kappa$) at each level for 1-2 April, 2010, with values of $k$ from Twomey (1959) and number concentration of CN-PCASP.

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<th>LEG</th>
<th>Height</th>
<th>$D_c$ (µm)</th>
<th>Kappa, $\kappa$</th>
<th>CN-PCASP (# cm$^{-3}$)</th>
<th>$k$ from $N_c=C_s^k$</th>
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<td>Above SAL</td>
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<td>0.025</td>
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<th>CN-PCASP (# cm$^{-3}$)</th>
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<td>Leg3</td>
<td>828 m</td>
<td>IL</td>
<td>0.164</td>
<td>0.033</td>
<td>69</td>
</tr>
<tr>
<td>Leg4</td>
<td>769 m</td>
<td>IL</td>
<td>0.165</td>
<td>0.032</td>
<td>67</td>
</tr>
<tr>
<td>Leg5</td>
<td>514 m</td>
<td>SCL</td>
<td>0.167</td>
<td>0.031</td>
<td>68</td>
</tr>
<tr>
<td>Leg6</td>
<td>191 m</td>
<td>SCL</td>
<td>0.172</td>
<td>0.028</td>
<td>77</td>
</tr>
<tr>
<td>Leg7</td>
<td>18 m</td>
<td>SCL</td>
<td>0.164</td>
<td>0.033</td>
<td>104</td>
</tr>
</tbody>
</table>
The CCN count can be approximated as $N_c = C s^k$ (Twomey, 1959), where $s$ is the percent supersaturation, $N_c$ and $C$ correspond to CCN and CN in this study, respectively. The $k$ values are also shown in Table 1. Overall, $k$ values decrease closer to the surface and vary from 0.78 at the top of the SAL to 0.52 near the surface, indicating more continental characteristics of air mass aloft, when compared with those near the surface (e.g., Twomey, 1977; $k \sim \frac{1}{2}$ typical for maritime environment and $k \sim \frac{2}{3}$ for continental air mass). In short, kappa and $k$ shows more likely continental air masses in the SAL and maritime air masses from surface to the lower IL but with dust composition (kappa close to 0.01) throughout the BL.

5.4.3 Changes in particle size distribution

**Figure 5.14:** Time series of (a) aerosol number concentrations (CCN, 0.6 % per mass) and (b) mixing ratio sampled on leg 3 (intermediate layer, IL) on 1 April for dry (magenta), moist (cyan), and transit (grey dots) intermediate layers (ILs). A mixing diagram of aerosol concentrations ($\mu g^-1$) versus mixing ratio ($g \text{ kg}^-1$) for these three air masses is shown in Fig. 5.14(c).

The two well-defined air mass populations are found in the intermediate layer on 1 April (e.g., Fig. 5.8a, Fig. 5.14): a dry air mass with mixing ratio ($w$) of 9 g kg$^-1$ and
aerosol concentrations of 300 µg\(^{-1}\) (hereafter dry IL); and a moist air mass with mixing ratio of 12 g kg\(^{-1}\) and aerosol concentrations of 200 µg\(^{-1}\) (hereafter moist IL); an air mass found between moist and dry ILs will be referred to as the transit IL. To investigate the possibility of cloud processing on the changes in PSDs, aerosol distributions made on leg 3 on 1 April (see Fig. 5.8a), associated with these three populations, are compared. The PCASP particle size distributions, in terms of \(dN/d(\log D)\) and \(dV/d(\log D)\), obtained from three Intermediate Layers (ILs), Saharan Air Layer(SAL) and Sub-Cloud-Layer (SCL) on 1 April, are shown in Fig. 5.15.

**Figure 5.15:** Aerosol particle size distributions (\(dN/d(\log D)\) in upper, \(dV/d(\log D)\) in lower panels) on 1 April, 2010 obtained from SAL, three intermediate layers (moist IL, dry IL and transit IL), and sub-cloud layer.
PSDs of moist IL regime are similar to those in the sub-cloud layer over all ranges. PSDs of transit IL regime look like a mixture of the SAL and the moist IL air mass; PSDs are close to those in the SAL in the range of \( D > 0.5 \, \mu m \) and PSDs are close to those in the SCL for \( D < 0.5 \, \mu m \). Particle size distributions, obtained from the dry IL, show a similar characteristics to those in the SAL for the smaller particles (\( D < 0.5 \, \mu m \)), but show the greatest concentrations of larger particles (\( D > 0.5 \, \mu m \)). The larger aerosol concentrations in dry IL for the range of \( D > 0.5 \, \mu m \) possibly show the horizontal variations of SAL; i.e., the drier, dustier parts of the SAL air masses are not sampled in SAL-leg (leg 2), but are sampled in IL-leg (leg 3-2 in Fig. 5.15). Overall, the PSDs are shifted to the lower concentrations for moister and cooler air masses. A comparison of PSDs from the moist IL with the PSDs from the dry IL suggest the removal of aerosol particles over all ranges by cloud process, although precipitation processes in the earlier history of the air mass may be involved too. As a layer is moistened (by comparing PSDs in the SAL with those in the transit IL), small sizes of particles (\( D < 0.5 \, \mu m \)) decrease significantly and larger particle concentrations (\( D > 0.5 \, \mu m \)) increase slightly (or little change), suggesting coagulation processes by the cloud.

5.5 Discussion and summary

Throughout the year large amounts of dust from Africa are transported over the Atlantic to the Caribbean Sea with important implications for climate in these regions. The island of Barbados - located on the eastern edge of the Caribbean Sea - typically experiences northeast trade winds and periodic African dust outbreaks. In this chapter, an intense African dust event observed upstream from Barbados on two days is studied
using measurements from the CIRPAS Twin Otter research aircraft to examine the vertical distribution of aerosols, temperature, moisture, and the processes leading to the observed stratification.

During the dust event studied, the surface observations from Ragged Point on Barbados indicate dust concentrations of about 155.1 µg m⁻³ – the highest concentrations observed for all of 2010—with an aerosol optical depth (at 550 nm) of about 0.6. The Saharan Air Layer (SAL) sampled during the flights was shown to have its origin over Africa 7-10 days prior to the observations. It originally had potential temperature of θ ~312 K, mixing ratio ~4 g kg⁻¹ with a depth of ~3500 m in Africa. The SAL subsided about 125 m per day and ended up about ~800 m of depth with θ ~307 K and mixing ratio ~5-6 g kg⁻¹ at Barbados, showing a cooling rate of about 0.5 °C day⁻¹.

The vertical profiles of CCN, PCASP and CN sampled indicate three layers with distinct aerosol and thermodynamic characteristics; SAL (Saharan Air Layer), IL (Intermediate Layer) and SCL (Sub-Cloud Layer). The SAL and SCL are characterized by well-mixed layer, while the greatest thermodynamic and aerosol variations is observed in the IL. The highest aerosol concentrations were observed in the SAL with CN and CCN (s=0.3 %) concentrations of about 700 cm⁻³ and 300 cm⁻³, respectively, and unexplained ultrafine (3-10 nm) particle concentration of about 50 cm⁻³ was observed in the elevated SAL. The sub-cloud layer had CN and CCN concentration of 400 cm⁻³ and 200 cm⁻³. The intermediate layer had CN and CCN of those between SAL and SCL, as a whole.

A comparison of the thermodynamic structure observed in the event over Africa with that at Barbados indicates that layers below the SAL (from surface to 1700 m) was
moistened by surface fluxes (latent heat flux estimated to be about 60-70 Wm\(^{-2}\)) as the air mass moved across the Atlantic over 7-10 days. The cool and moist layers in the IL concurred with the lower concentrations of aerosols. These layers are most likely associated with modification by episodic shallow cumulus convection sometime in the previous history of the air mass, as it advected across the Atlantic. The intermittent nature of these convective events explains the variability observed in the IL compared with that in the SAL and in the SCL. Further, in some parts of the intermediate layer, the aerosol concentrations are lower than those in the sub-cloud layer, providing further evidence for possible cloud processing of the aerosols in the intermediate layer. Time-height observations from the Micro Pulse Lidar (MPL) located at Ragged point also indicate mesoscale variability in the IL aerosols.

Mixing diagrams using aerosol concentrations per mass and conserved thermodynamic parameters provide insight into the vertical transports and mixing processes giving the observed aerosol and thermodynamic variability in each layer. Aerosols in the intermediate layer share their thermodynamic and aerosol characteristics with those in the SAL, suggesting the source of aerosols in IL in this study, is those in SAL. The hygroscopicity parameter, kappa (\(\kappa\), a proxy for chemical composition) ranged from 0.01 to 0.026, which confirmed that dust was present in all layers during the event. Small hygroscopic particles that have the same particle size distribution as those in the SAL were observed at the very top of the SAL and possibly suggesting the existence of small particles that were coated by hydrophilic materials, at the very top of the SAL.

Changes in particle size distributions due to the cloud processing was examined with three distinctive air masses that were sampled in intermediate layers; dry, moist and
transit ILs. This study shows that air mass in the moist IL have been processed by convection, and thus, have characteristics similar to those in the sub-cloud layer, while the dry air masses have characteristics similar to those in the Saharan air layer; and the transit air masses have characteristics similar to air masses between the SAL and the SCL.

This study has documented the vertical structure of aerosols associated with intense dust events and helps put into perspective the observations of dust concentrations and AOD observed from surface measurements and satellite measurements from the top of the atmosphere. The SAL layer is often referred to as an elevated layer but the aerosol characteristics in this layer have characteristic similar to those in the SAL, but with aerosol concentrations somewhat reduced. But the thermodynamic structure below the SAL is substantially modified by dry and moist convective processes in the boundary layer. Further the presence of dust in and above the boundary layer may impact the modification of the boundary layer by modulating convective and radiative processes as air flows westward in the North Atlantic trade. Thus large-scale models should properly include dust effects on the downstream evolution of the boundary layer structure if the boundary layer is to be simulated realistically. The evolution may be particularly important in affecting the environment of tropical cyclones, the evolution of tropical waves, and long-term climate realizations for the Atlantic. Further study is clearly needed to understand the Lagrangian evolution of the SAL and the processes that affect aerosol and thermodynamic properties of the layer.
Chapter 6: Summary, Conclusions and Future Work

6.1 Summary

This dissertation investigates several issues related to the aerosol-cloud-precipitation interactions over trade cumuli regime. The research objectives in Chapter 1 are examined based on the analysis of observational data, collected during the Barbados Aerosol Cloud Experiment (BACEX), which took place off the Caribbean island of Barbados within the northeast trades of the eastern Atlantic, during mid March and April 2010. The principal observing platform for the experiment was the CIRPAS Twin Otter (TO) research aircraft that was equipped with aerosol, cloud, and precipitation probes and standard meteorological instruments for observing the mean and turbulent thermodynamic and wind structures. The FMCW 95-GHz (W-band) Doppler cloud radar was also mounted on the top of the aircraft in an upward-looking mode to provide vertical structures of cloud and precipitation properties.

Operations were centered in a domain located to the east and northeast of Barbados. On each day during the BACEX, TO flights were made about 50-100 km upwind of Ragged Point (13.2 °N, 59.5 °W), eastern shore of Barbados where surface aerosol measurements were made. Each flight had a duration of 3-4 hours and included at least one sounding, excluding take off and landing soundings, and included several constant-level legs, flown at relatively constant heights from near the ocean surface to the trade-wind inversion height; e.g., levels of below the cloud (so-called sub-cloud level leg flight); near the cloud base (cloud-base level leg); near the cloud tops (cloud-top level leg).

Satellite based studies can be used to study aerosol indirect effects over a large geographical area with long time periods, but are known to suffer from retrieval bias and
the vertical distribution—a key component of aerosol indirect effect—is usually unknown. Therefore, in this dissertation I combined the in situ aircraft data with some other supplementary data to explore the boundary layer structures as well as the properties of clouds and aerosols over the Caribbean Sea.

Chapter 2 overviews the BACEX project, data and instruments that used in this work (Section 2.1 and 2.2). The airborne cloud radar used in this study is an important tool to provide vertical structures of cloud and precipitation properties. To make full use of this tool, reflectivity calibration (Section 2.3) and velocity correction (Section 2.4) are made and innovative techniques for making full use of the radar are developed. Techniques of aircraft platform correction and retrieval of vertical air motion is described in Section 2.4.

The vertical components of the Doppler velocity from airborne radar are contaminated by the aircraft platform motions involved with aircraft’s pitch angle, speed of aircraft and the aircraft’s physical up and down motion velocity. Thus, first of all, Doppler velocity observed from airborne cloud radar was corrected by considering the first-order corrections to the vertical component of the Doppler velocity.

At the radar wavelength of 3.2 mm (W-band, frequency of 95-GHz), the raindrop backscatter cross section varies between successive maxima and minima as a function of the raindrop diameter (D) that are well described by Mie theory (i.e., Mie signature). The location of the first Mie minimum (e.g., D = 1.65 mm for 95 GHz radar) in the recorded radar Doppler spectrum has been used successfully in the past, with ground-based radars to retrieve the vertical air motion. The first application of this technique (i.e., Mie technique) to airborne W-band radar Doppler is presented in Section 2.4. Moreover,
separate spectral peaks due to the cloud droplets are also observed in the same radar Doppler spectra that contain Mie signatures. This feature has not been observed from ground-based radars and provides an independent validation of the Mie scattering-based vertical air motion retrieval, using the cloud droplets as a tracer of vertical air motion. This is the first demonstration of such Rayleigh and Mie scattering signatures in the radar Doppler spectra. The application of the Mie technique, using an airborne W-band system, can lead to the new opportunities in cloud and precipitation research.

In Chapter 3, aerosol variations and characteristics of cloud and precipitation sampled over the eastern Caribbean Sea are documented. The temporal variations and vertical distributions of aerosols observed on the 15 flights show a wide range of aerosol conditions that include a characterization of the most intense African dust outbreak observed during all of 2010 (e.g., 1-2 April). The intense African dust events concurred with a massive dry air intrusion into the lower atmosphere and very weak easterlies. In addition, clouds and precipitation were significantly suppressed during the middle of dust outbreak.

The 10-day backward trajectories of air masses arriving at 500 m in the middle of the flight domain show that three distinctive air masses dominate over the Eastern Caribbean (e.g., typical maritime air mass, Saharan Air mass, and Middle latitude dry air, as shown in Dunion (2011)). A variety of aerosol vertical structures is observed (Figs. 3.2c and 3.5b) and categorized by three distinct vertical profiles: 1) aerosol concentrations decrease steadily with height, with its maximum concentrations ( < 250 cm\(^{-3}\)) below the trade-wind inversion, and are associated with a typical maritime air mass from the back trajectory analysis. 2) The second type of profile shows aerosol
concentrations increasing with height, with a maximum concentration above the inversion; this profile is associated with air mass originated from middle latitude. 3) The third type of aerosol profile is associated with African dust events (e.g., 31 March to 5 April). High concentrations of aerosols are observed throughout the entire boundary layer with stratified aerosol structures. These three typical vertical structures are summarized in Fig. 6.1.

![Figure 6.1](image_url)

**Figure 6.1**: Profiles of aerosol number concentrations obtained from PCASP. Left: Typical maritime vertical structure; Middle: vertical structure associated with middle latitude dry air; Right: aerosol vertical structure during African dust events. Color bar indicates the number of research flights (RF #), shown in Table 2.1.

The TO research aircraft sampled many clouds in various phases of growth during the BACEX. Data obtained from the cloud radar, during the cloud-base level flights, provided the basis of the general characteristics of clouds studied in this dissertation. Clouds sampled from 12 among 15 flights are used for the radar analysis. No clouds were observed during the middle of African dust events (1-2 April) and data from 19 March (test flight) are not used.
Two populations of reflectivity and velocity are observed during BACEX: precipitating-dominant clouds and non-precipitating and/or lightly precipitating clouds. All clouds (precipitating, non-precipitating, and lightly precipitating clouds) sampled during the BACEX show the cloud bases and tops at about 400 m and 2700 m, respectively with the depths of about 2300 m. Non-precipitating and/or lightly precipitating clouds show cloud bases and tops of about 700 m and 2000 m, respectively, with depth of about 1300 m. Reflectivity of ~ -35 dBz and Doppler velocity of ± 2 m/s are the most frequently observed for these non-precipitating and/or lightly precipitating clouds between 600 m and 1300 m.

Although, the maximum cloud depth observed is about less than 3 km, more than half of the clouds precipitate somewhere in the clouds. However, the precipitation in and near the cloud is less than 10 mm day\(^{-1}\) as a whole, except on 22 March. Further, no surface precipitation is recorded (at the Barbados surface site), indicating clouds precipitated but evaporated in the atmosphere. This study also shows that clouds in the study area (trade-wind boundary layer) are not valid for the adiabatic assumption; even the cloud cores are far from adiabatic possibly due to the precipitation and entrainment of environmental air into the cloud.

During the BACEX, two types of precipitation features were observed (Fig. 3.13). In one, precipitation shafts are observed to emanate mainly from the cloud base with evaporation in the sub-cloud layer. In the other, precipitation shafts emanate mainly near cloud top on the downshear side of the cloud and evaporate in the cloud layer. This type of precipitating cloud is shallower than the first type of cloud, and also sometimes accompanies precipitation shafts emanating near cloud base. These two types of
precipitation are expected to have effects on the trade-wind boundary layer in different ways, in particular in terms of stability and the moisture budget. For example, the cloud-base precipitation can stabilize the sub-cloud layer by the evaporation rain. As a result, the following cloud can grow only if it overcomes the stability. Thus, vertical velocities of sub-cloud eddies need to be strong enough to penetrate the inversion barrier if clouds are to form. On the other hand, when precipitation shafts emerge from cloud top and evaporate in the cloud layer (assume that precipitation occurs and evaporate in the upper part of the cloud), would destabilize the cloud layer, but stabilize near cloud top. Consequently, small clouds could form relatively easily in the cloud layer. Further, the evaporation of precipitation in the cloud layer provides moistures to the near cloud field that may affect the moisture budget and increase cloud lifetime of future clouds. The schematics are illustrated in Fig. 6.2.

![Figure 6.2: A cloud photo that shows two types of precipitation; Precipitation shafts emanate near the cloud top on the downshear side of the cloud; Precipitation shafts emanating near the cloud base. The photo was taken by Bruce Albrecht from the balcony of his office at Rosenstiel School of Marine and Atmospheric Science (RSMAS), Miami. Photo: Courtesy Bruce A. Albrecht.](image-url)
In Chapter 4, aerosol-cloud-precipitation interactions are investigated. Section 4.1 describes aerosol effects on cloud properties and precipitation. The effect of sub-cloud aerosol on the cloud thickness is first presented. It shows that vertical structures of aerosol as well as the depth of the aerosol layer resided aloft the inversion layer play roles on the cloud formation and development. For example, deeper clouds develop under the typical maritime aerosol profile structures. On the other hand, clouds are significantly suppressed when high concentrations of aerosols prevail throughout the entire boundary layer. Further, when the high aerosol concentration is observed above the inversion layer, and the aerosol layer is deep (e.g., 25-26 March), the clouds are suppressed.

The aerosol-cloud interactions are assessed by examining the relations between aerosol ($N_a$) and cloud droplets number concentrations ($N_d$), and the sizes of effective droplets ($D_e$) with cloud-base and flight-averaged properties. A robust increase in $N_d$ with aerosols is found. In general, the effective diameter of cloud droplets ($D_e$) tends to decreases as aerosol concentrations ($N_a$) increase when data is stratified by cloud depth. However, the inverse relation between $D_e$ and $N_a$ is valid only for shallow clouds (e.g., clouds shallower than 500 m) when sub-cloud $N_a$ and cloud-base $D_e$ are considered (Fig. 4.4f), suggesting possible effects of cloud processes such as entrainment and precipitation on the cloud properties.

The sensitivity of $N_d$ to changes in vertical velocity perturbations $w'$ ($d\ln N_d/d\ln w'$) is introduced in this section and the sensitivity of precipitation to changes in aerosols, so-called precipitation susceptibility, is also presented. Cloud-base $N_d$ is highly correlated to the sub-cloud CCN concentrations, in particular, in the regimes of updrafts and/or positive perturbation of $w$ ($w' > 0$). Cloud-base droplet concentrations increase as $w'$
increases and the sensitivity to $w'$ ($dN_d/dw'$ as well as $d\ln N_d/d\ln w'$), is greater in the regimes of dustier conditions (high aerosol concentrations), suggesting a slight increase in positive $w'$ in polluted conditions can make greater increase of $N_d$. The sensitivity of $N_d$ to changes in vertical velocity perturbations $w'$ ($d\ln N_d/d\ln w'$) for the dusty and pristine conditions shows a slope of 0.2 and 0.1, respectively in Fig. 6.3.

![Figure 6.3](image)

**Figure 6.3:** The sensitivity of $N_d$ to changes in vertical velocity perturbations $w'$ ($d\ln N_d/d\ln w'$) for two dusty (4/5-red and 3/31-magenta), and two pristine (4/10-cyan and 3/29-blue) conditions. The slope shows 0.2 and 0.1 for dusty and pristine conditions, respectively.

The slope ($d\ln N_d/d\ln w'$) is used to evaluate the parameterization of Twomey’s formula (Eq. 4.3). The slope ($d\ln N_d/d\ln w'$) from Twomey’s formula shows 0.3, using the typical value of $k \sim 0.5$, indicating his parameterization overestimates the sensitivity of $N_d$ in changes in vertical velocity perturbations for shallow cumulus observed in the trade-wind boundary layer.
The suppression of precipitation toward higher aerosol concentrations is a general feature in data sampled during the BACEX. The sensitivity of precipitation to changes in aerosol, the so-called precipitation susceptibility \( S_0 \), is examined to quantify this suppression while minimizing the effects of macrophysics. \( S_0 \) exhibits three regimes: (1) low cloud thickness, where clouds do not precipitate, and are not sensitive to \( N_d \), \( S_0 \sim 0 \); (2) intermediate cloud depth, where precipitation begins to decrease due to the increased aerosols (and \( N_d \)), \( S_0 \) increases to 2 at \( H \sim 1100 \) m, where \( S_0 \) peaks; (3) high cloud thickness, where \( S_0 \) decreases. This pattern of \( S_0 \), i.e., susceptibility peaks at intermediate range of cloud thickness, is consistent with previous studies of warm cumulus clouds using satellite data (e.g., Sorooshian et al., 2009; Jiang et al., 2010; Duong et al., 2011).

![Figure 6.4](image)

**Figure 6.4**: Precipitation susceptibility, \( S_0 \) with fixed cloud thickness from BACEX (upper panel), and from models and satellite observations (lower panel).

However, peaks from model and satellite observations are smaller than the value of \( \sim 2 \) found in this study; peaks occur at around 1.1 and 0.5 from the model and satellite
observations, shown in Fig. 6.4, indicating $S_0$ is underestimated by model and satellite observations. The value of $\sim 2$ often assumed (single value, not variations with LWP or H is used) for the auto-conversion of cloud to rain water in the climate models (e.g., Khairoutdinov and Kogan, 2000). In addition, removal of $N_d$ due to the rain (wet scavenging) makes susceptibility stronger overall. It is summarized in Fig. 6.4.

The interactions between clouds and environmental air (aerosols) are presented in Section 4.2 from the views of entrainment process. Further, cloud effects on aerosols near cloud field that connects to the entrainment process and flows around the small cumuli (in Section 4.2) are discussed in Section 4.3. Radar chaff, pre-cut metallic coated fiber, is used to show the entrainment process and flows in and around small cumuli. A shallow marine cumulus cloud that sampled about 100 km east of Barbados, on 29 March, 2010 during BACEX was investigated as an example. Chaff were dispensed near the cloud top and edges of the cloud (LEG 1, LEG 2) and then the aircraft made penetrations of the cloud at lower levels (about 300 m and 450 m lower than the initial level of chaff dispense) to observe the chaffed cloud above, with an airborne cloud radar. It is found that the radar returns from the chaff are easily distinguished from the cloud and precipitation returns since the Doppler spectrum width is substantially narrower than that associated with the cloud and precipitation, but at the same time, has a reflectivity that is elevated relative to the cloud. The results show that environmental air above the cloud top descended along the downshear side of the cloud edges through the downdraft (so-called cloud halos or cloud shell regions) to the temperature equilibrium level; and is subsequently entrained laterally into the cloud with updrafts. The entrained airs tagged
with chaff eventually circulate along the incloud edges with inversed P letter shape, showing complete picture of the in-cloud flow pattern.

In this small cloud, the strong downdrafts are observed along the cloud edges (so-called cloud halos or cloud shell regions). The lower virtual temperatures, $T_v$ (negative buoyancy), presumably due to the evaporative cooling, are noticeable along the cloud edges, especially in the regions of the strongest descending motion are found. These downdrafts extend into the clear air surrounding the cloud. Further, the depletion of aerosols near cloud field (i.e., in the cloud shell regimes) are noticeable, and the reduction of aerosols is more significant in precipitating clouds compared with non- and/or light-precipitating clouds.

In Chapter 5, an intense African dust event observed upstream from Barbados on two days (1-2 April) is studied using measurements from the CIRPAS TO research aircraft to characterize particle size distributions; vertical distributions of aerosols, temperature and moisture. Further, processes leading to the observed stratification in the boundary layer—from surface to the trade inversion—are examined by using the changes in vertical structures of temperatures, moistures, aerosols and particle size distributions.

The vertical profiles of various aerosols were similar on both days and show three layers with distinct aerosol and thermodynamic characteristics: the Saharan Air Layer (SAL; ~2.1 km ± 400 m), a sub-cloud layer (surface to ~500 m), and an intermediate layer extending between them. The SAL and the sub-cloud layer display characteristics of well-mixed aerosols and thermodynamics, although the most significant horizontal and vertical variations in aerosols and thermodynamics occur in the intermediate layer. The highest aerosol concentrations were observed in the SAL with CN and CCN (super-
s at 0.3 %) concentrations of about 700 cm$^{-3}$ and 300 cm$^{-3}$, respectively. An unexplained ultrafine (3-10 nm) particle concentration of about 50 cm$^{-3}$ was observed in the elevated SAL.

The SAL sampled was shown to have its origin over Africa 7-10 days prior to the observations. A comparison of the thermodynamic structure observed in the event over Africa with that at Barbados indicates that the lower part of the SAL (from the surface to 1700 m) was moistened by surface fluxes, as the air mass was advected across the Atlantic. The aerosol variability observed in the intermediate layer is likely associated with modification by shallow cumulus convection occurring sometime in the prior history of the air mass. Two approaches are used to show the modification of boundary layer aerosol by cloud process; first, by comparing the variability in the thermodynamics and aerosol structure and, second, by comparing particle size distributions of aerosol sampled from intermediate layer. The vertical structure of thermodynamics and aerosol shows that warmer and drier intermediate layer is accompanied by dustier conditions, and relatively lower concentrations of aerosols are observed in the cool and moist layers. Further, aerosol concentrations in the intermediate layer are slightly lower than those in the sub-cloud layer, suggesting the removal of aerosol in the layer.

To investigate the possibility of cloud processing on the changes in particle size distributions, aerosol distributions sampled from the intermediate layer (IL), associated with three populations of aerosol (e.g., moist IL, dry IL, and transit IL) are compared. PSDs of moist IL regime are similar to those in the sub-cloud layer over all ranges. PSDs of transit IL regime look like a mixture of the SAL and the moist IL air mass; Particle size distributions, obtained from the dry IL, show a similar characteristics to those in the
SAL, supporting the hypothesis that cloud processes in boundary layer can cause complicated stratification in the aerosols below the SAL (Saharan Air Layer).

Mixing diagrams using aerosol concentrations and conserved thermodynamic parameters provide insight into the vertical transports and mixing processes that may explain the observed aerosol and thermodynamic variability in each layer. The hygroscopicity parameter, kappa (a proxy for chemical composition) confirmed that dust was present in all layers during the event.

6.2 Implication to model parameterization

6.2.1 First-indirect effect

In the climate models, two general approaches have been used to relate changes in cloud droplet number concentrations ($N_d$) and aerosol concentrations ($N_a$). One is to use an empirical expression between cloud droplet concentrations and aerosol concentrations, similar to Eq. (1.1).

$$N_a \sim N_a^a$$  \hspace{1cm} (Equation 6.1)

In Chapter 4, the first aerosol indirect effect is addressed by relating aerosol concentrations ($N_a$) in the sub-cloud layer to cloud droplets number concentrations ($N_d$) in the cloud-base. In this work, linear relations between $N_a$ and $N_d$ are established for the updrafts regimes; still, the problem of this approach is that the empirical relationship derived from a certain region may not appropriate for other regions, and thus, may not be representative globally. Nevertheless, linearity can make parameterization be simpler. In addition, when models assess the uncertainties of each scheme used for assessing the indirect effect (e.g., Penner et al., 2006), the bias (or error) that could be caused by the
embedded non-linear relations between $N_a$ and $N_d$, can be easily counted if the relations are linear.

The second method to relate changes in $N_d$ to changes in $N_a$ is based on a prognostic parameterization of the cloud droplet formation process (e.g., Abdul-Razzak and Ghan, 2000). To treat cloud droplet formation accurately, the aerosol number concentration, its chemical composition and the vertical velocity on the cloud scale need to be known. In this study, cloud radar provides vertical velocity distribution (Chapter 3). Further, the sensitivity of $N_d$ to changes in vertical velocity perturbations ($w'$) is introduced. Overall, cloud-base droplet concentrations ($N_d$) increase as $w'$ increases for both pristine and dusty cases. Sensitivity to $w'$, i.e., $dN_d/dw'$ as well as $dlnN_d/dlnw'$, is greater in the regimes of more polluted (or dustier) environments, suggesting a slight increase in positive $w'$ in polluted conditions can make greater increases in $N_d$ (possibly more clouds). This slope of $dlnN_d/dlnw'$ can be used to evaluate the parameterization of Twomey’s formula (Eq. 4.3): $N_d$ expressed in terms of total aerosol population that can account for formation of cloud and updraft with $k$, a parameter that depends on the air-mass type. The slope from Twomey’s formula shows $\sim 0.3$ by using a typical value of $k \sim 0.5$. On the other hand, this study shows slopes of 0.2 and 0.1, for the dusty and pristine conditions, respectively.

### 6.2.2 Second-indirect effect

The various constructs that are examined in this study are of interest because they provide a link between climate model parameterization and observational studies. For example, the process of collision-coalescence between cloud drops to form precipitation
(auto-conversion), is represented in GCMs using a parameterization typically relating rain rate (R) to the amount of liquid water in the clouds (i.e., LWP) and drop number concentration ($N_d$) in the form of power law, similar to Eq. (4.5).

$$R_{ra} = LWP^\alpha N_d^{-\beta}$$

(Equation 6.2)

Cloud thickness can be used instead of liquid water path (LWP) in Eq. (6.2). For the stratocumulus, adiabatic assumption is valid. However, the assumption works poorly in the warm cumulus clouds studied in this work. Moreover, cloud thickness is directly obtained from radar measurement, while LWP adds additional uncertainty related to the vertical distribution of liquid water contents (i.e., mean LWC is used). The exponent $\beta$ (same as precipitation susceptibility, $S_0$) increases with cloud thickness and peaks at intermediate height as $\beta = 2$. The value of $\sim 2$ often assumed for the auto-conversion of cloud to rain water in the climate models (e.g., Khairoutdinov and Kogan, 2000). The value $\beta$ (Same as $S_0$) gets larger (e.g., $\beta \sim 2.5$ in this study), if heavy precipitation are included to the sampling (sampling bias), which is frequently observed over trade-wind cumuli regime; more than half of clouds precipitate from Chapter 3. $S_0$ in Eq. (4.4), estimated in this study for warm cumulus clouds, provides a validation of model and satellite-based estimates of $S_0$ (e.g., Sorooshian et al., 2010; Duong et al., 2011)

6.3 Future work

In the trades over Barbados, African dust (so-called Saharan Air Layer, SAL) is the dominant aerosol constituent during certain seasons. To better understand the interactions between aerosols, clouds, and precipitation, the overall distribution and
characteristics of cloud and aerosol itself should be known. However, little is known about the distributions of cloud intensity (reflectivity), vertical velocity, cloud-thickness (tops and bases). Further, vertical structures of aerosols (e.g., Reid et al., 2002) are rarely documented. Part of this work was intended to better document the vertical structure of the Saharan Air Layer (SAL) and provide standards for interpreting and comparing satellite data; further, to characterize the clouds properties in this regime. The results presented in this work help to understand the variability of aerosol, cloud and precipitation over the trade-wind cumuli. While the data set does include a wide range of aerosol and cloud conditions, it may still too small to draw definite conclusions from. How representative are the results presented here on the cloud property, aerosol variations, and their interactions for the regimes of trade-wind cumuli? One area of potential future work would be to compare the results shown here to other field campaigns, such as dataset sampled in Key West (May 2012) and collected near Florida (Spring 2008) to explore the representativeness of aerosol-clouds and their interactions in the region of the trade-wind cumuli. The extended dataset, further, can provide opportunities to better understand the aerosol-cloud-precipitation processes and their interactions.

Another area of potential future work would be to explore the mechanisms of the fast dissipating clouds. Some clouds in nature dissipate faster than others. Large-scale synoptic condition (e.g., outflows from adjacent clouds) as well as the effect of aerosol (CCN depletion in the cloud shell regime, shown in Section 4.3) can be examined to get some insights of this issue. The preliminary results showed that precipitating and/or non-precipitating clouds efficiently deplete CCN along the cloud edges (e.g., Figs. 4.18-4.20;
more efficient in the precipitating clouds). This cloud shell regime is also closely connected to the entrainment process of small cumulus clouds (Section 4.2). Previous studies (Jiang et al., 2006; Small et al., 2009) proposed that aerosol effects can affect the cloud lifetime through the evaporation process—the higher the aerosol concentrations would give smaller cloud droplets that will evaporate more quickly at the cloud boundaries, and thus, decrease the cloud lifetime. Circulations at the cloud edges, driven by the evaporation may provide a feedback that enhances the evaporation process (Heus and Jonker, 2008; Small et al., 2009; Wang et al., 2009) and thus contributing to a shortening of the cloud lifetime. However, visual evidence of rapidly decaying clouds in polluted environments is not available. Further, this evaporation process also can moisten the near-field surrounding clouds environment (Radke and Hobbs, 1991; Perry and Hobbs, 1996; Lu et al., 2003; Laird, 2005) and thus buffer the evaporation effects. A few fast-dissipating clouds, observed during the Barbados and Key West field campaigns, can be examined and compared with regular clouds to explore this future work.

In Chapter 5, the transport and process of aerosols in the marine boundary layer is focused, and showed how the aerosols were modified by the clouds (processes). Another area of potential future work related to this topic, would be to explore the mechanisms of aerosol transports from above SAL to SAL, and/or from SAL to upper intermediate layer. The vertical profiles of mixing ratio on one of the most intense African dust events in Barbados (2 April, 2010) are shown in Fig. 6.5, together with mixing diagram of aerosol concentrations and mixing ratios.
**Figure 6.5:** (a) Vertical profile of mixing ratio, and (b-c) mixing diagram of aerosol concentrations (PCASP, µg$^{-1}$) and mixing ratios (g kg$^{-1}$) obtained during the second ascent sounding on 2 April, 2010. Colors correspond to the heights. Aerosols and mixing ratios, sampled in the dry layer, right below the SAL in Fig. 6.5(a), are overlaid in Fig. 6.5(c) as black cross-symbols. The upper and lower boundary of the dry layer is denoted as horizontal dashed lines in Fig. 6.5(a).

Aerosols sampled in the middle of SAL are shown as clusters in the mixing diagrams (PCASP versus mixing ratios in Figs. 6.5b-c). Aerosols sampled in the dry layer (black cross-symbols in Fig. 6.5c) are overlapped with aerosols sampled in SAL, in particular, those in the lower and upper parts of the SAL, indicating aerosols sampled in the dry layer (1500-1800 m) show the same characteristics of aerosols as those sampled within the SAL, in particular, those in the lower and upper parts of the SAL. The effect of radiative cooling on the interfaces across the inversion layer (SAL in this case) associated with entrainment process (across the SAL interfaces) could be examined to address the role of radiative cooling as a possible aerosol transports mechanism across the SAL interfaces. Large Eddy Simulation study will be accompanied with this work.
Appendix A

Table A1: Table of acronyms and symbols used in this study.

<table>
<thead>
<tr>
<th>Acronym</th>
<th>Expression</th>
</tr>
</thead>
<tbody>
<tr>
<td>AERONET</td>
<td>Aerosol robotic network</td>
</tr>
<tr>
<td>AOD</td>
<td>Aerosol Optical Thickness</td>
</tr>
<tr>
<td>ARL</td>
<td>Above the airborne Radar Level</td>
</tr>
<tr>
<td>BACEX</td>
<td>Barbados Aerosol Cloud Experiment</td>
</tr>
<tr>
<td>BL</td>
<td>Boundary Layer</td>
</tr>
<tr>
<td>CAS</td>
<td>Cloud Aerosol Spectrometer</td>
</tr>
<tr>
<td>CCN</td>
<td>Cloud condensation nuclei</td>
</tr>
<tr>
<td>CIRPAS</td>
<td>Collaborative Institute for Remotely-Piloted Aircraft Studies</td>
</tr>
<tr>
<td>CN</td>
<td>Condensation nuclei</td>
</tr>
<tr>
<td>CPC</td>
<td>Condensation Particle Counter</td>
</tr>
<tr>
<td>D_a</td>
<td>Aerosol particle mean diameter</td>
</tr>
<tr>
<td>D_e</td>
<td>Effective diameter of cloud droplet</td>
</tr>
<tr>
<td>DSD</td>
<td>Drop Size Distribution (used for cloud droplets)</td>
</tr>
<tr>
<td>FMCW</td>
<td>Frequency Modulated Continuous Wave</td>
</tr>
<tr>
<td>GATE</td>
<td>(GARP) Atlantic Tropical Experiment (summer of 1974)</td>
</tr>
<tr>
<td>GDAS</td>
<td>Global Data Assimilation System</td>
</tr>
<tr>
<td>HYSPLIT</td>
<td>Hybrid Single Particle Lagrangian Integrated Trajectory</td>
</tr>
<tr>
<td>*IL</td>
<td>Intermediate layer</td>
</tr>
<tr>
<td>LCL</td>
<td>Lifting condensation level</td>
</tr>
<tr>
<td>LWC</td>
<td>Liquid water content</td>
</tr>
<tr>
<td>MPLNET</td>
<td>Micro Pulse Lidar NETwork</td>
</tr>
<tr>
<td>N_a</td>
<td>Aerosol number concentration</td>
</tr>
<tr>
<td>N_d</td>
<td>Cloud droplets number concentration</td>
</tr>
<tr>
<td>NRB</td>
<td>Normalized Relative Backscatter</td>
</tr>
<tr>
<td>PCASP</td>
<td>Passive Cavity Aerosol Spectrometer Probe (0.1-2.5 µm)</td>
</tr>
<tr>
<td>PMS</td>
<td>Particle Measuring System</td>
</tr>
<tr>
<td>PSD</td>
<td>Particle Size Distribution</td>
</tr>
<tr>
<td>RICO</td>
<td>Rain In shallow Cumulus over the Ocean</td>
</tr>
<tr>
<td>SAL</td>
<td>Saharan Air Layer</td>
</tr>
<tr>
<td>SCL</td>
<td>Sub-cloud layer *This layer extends from surface to ~500-600 m</td>
</tr>
</tbody>
</table>

*This layer is the same as a traditional Convective Boundary Layer
Appendix B

The precipitation features that were observed during BACEX are illustrated in Fig. B1. Precipitation can be embedded in updrafts, and thus, two definitions for precipitating clouds are shown (data points with $Z > -20$ dBz and vertical velocity < 0 m s$^{-1}$, black); and data points with $Z > -20$ dBz, red). On 22 March and 30 March, vertical distributions of total clouds (non-precipitating + precipitating; shown as gray) and precipitating clouds from reflectivity threshold only (red) agree well one another, indicating precipitation (defined as strong reflectivity) dominates the total cloud population on these days. On 22 March, approximately 30% of precipitation is embedded in updrafts (950 versus 650 data samples), while about 10% of precipitation is embedded in updrafts on 30 March.

Clouds are most frequently observed between 1400-1800 m (number of samples > 1000) on 22 March, while precipitating clouds are most frequently observed right below 1400 (black). On 30 March, clouds are observed between 500 m and 2000 m with similar frequencies with a maximum near 500 m (grey); but precipitating clouds on this day are observed slightly more frequently near the upper (e.g., 1700 m) and lower (e.g., 700 m) regions of the clouds (black). On the other hand, clouds sampled on 25, 26, 29, 31 March and 11 April are purely non-precipitating clouds, with no frequency from precipitating clouds.

Clouds sampled during the rest of the days (e.g., 3/23, 3/24, 4/5, 4/7, 4/10) show similar patterns among them in a general sense; precipitation (black and red) is observed frequently at slightly higher heights compared with all clouds sampled (gray). For
example, clouds on 7 April are observed most frequently near 800 m with secondary peak near 1200 m, while precipitating clouds (black and red) are observed dominantly near 1200 m. Further, precipitating clouds on 23 March, 5 and 7 April show bimodal populations; the maximum occurrence of precipitation is observed at slightly higher heights than where the majority of the cloud is observed, and a secondary maximum is observed near the cloud base. Approximately, 20 %, 50 %, and 10 % of cloud area precipitates on 5, 7, and 10 April, while more than 90 % of cloud area precipitates on 22 and 30 March.

Figure B1: Number of samples with height for all sampled clouds (gray) and for precipitating clouds (black and red). Precipitating clouds are defined as data points with $Z > -20 \text{ dBz}$ and vertical velocity $< 0 \text{ m s}^{-1}$ (black), and are defined as data points with $Z > -20 \text{ dBz}$ (red).
Appendix C

Uncertainty on 10-day trajectories

Uncertainties on 10-day trajectories in a convective layer, arriving at Barbados at 500 m at 17 UTC on 1 April, are considered by supposing that the subtropical lower troposphere has a large-scale thermodynamic budget that reduces to a balance:

\[
\frac{1}{C_p}[w] \frac{d[s]}{dz} = Q \tag{Equation C1}
\]

where \([w]\) is the large-scale (horizontal-averaged) vertical velocity, \([s]\) is the large-scale (horizontal-averaged) dry static energy \((S=C_pT+gz)\), and \(Q\) is the large-scale heating rate. In this study, \([w] \sim -125 \text{ m day}^{-1}\) with \(Q=0.5 \text{ K day}^{-1}\), lapse rate of 6 K km\(^{-1}\), and the African SAL extends up to \(\sim 4.2 \text{ km}\) (Fig. 5.3). The air mass, arriving at Barbados (500 m) originated from 1.75 km \((1250 \text{ m} + 500 \text{ m})\) within in African SAL by using \([w]\). If a pollutant is transported by environmental air with vertical velocity \(w_e\), rather than the mean vertical velocity \([w]\), a rough estimate of how wrong the origin locations of 10-day back trajectories from a 500 m endpoint, can possibly be made by examining, first, winds over the Atlantic to estimate a wind shear, then, second, considering the effect of a mis-assigning parcel's altitude in this sheared wind profile, due to using \([w]\) instead of \(w_e\) as its vertical velocity. We supposed that \([w]\) in the lower troposphere, is an area-weighted average of "environmental" vertical velocity \(w_e\), plus cloudy updrafts with an effective vertical velocity, \(w_c\) covering fractional area, \(a\); \(a w_c + (1-a) w_e = [w]\). From these considerations, it is useful to constrain the product \(a w_c\) by considering a heat budget like
the equation above but for the environmental air alone, assuming that $Q_{env}$ is solely a clear-air radiative cooling rate:

$$\frac{1}{C_p} w \frac{d[s]}{dz} = w_s \left( \frac{dT}{dz} + \frac{g}{C_p} \right) = Q_{env} \quad \text{(Equation C2)}$$

Clear-air radiative cooling rate of $-1 \text{ K day}^{-1}$ $\sim -2 \text{ K day}^{-1}$ was used for this calculation with varying atmospheric lapse rate ($-dT/dz$). As long as the atmospheric lapse rate is close to and/or smaller than $5-6 \text{ K km}^{-1}$, which is typical in the tropics, the air parcel of 500 m endpoint in Barbados originates from heights within the African SAL. As a result, the uncertainty on back-trajectories will not affect the result, which indicates that the air mass at 500 m in Barbados originates from the air mass within the African SAL (e.g., 3 km subsidized for 10 day when $-dT/dz = 5 \text{ K km}^{-1}$ and $Q_{env} = -1.5 \text{ K day}^{-1}$ are used, thus, the air observed at 500 m at Barbados originated from 3.5 km of the African SAL (3 km + 500 m). Besides the winds within the African SAL as well as winds over the Atlantic, blows westward consistently as shown in Fig. 5.3c. More directly, MODIS satellite images (e.g., Fig. 5.2) clearly show the African origins of air mass during the period.
References


175


