A Global Survey of Clouds by CloudSat

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UNIVERSITY OF MIAMI

A GLOBAL SURVEY OF CLOUDS BY CLOUDSAT

By

Emily Marie Riley

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A GLOBAL SURVEY OF CLOUDS BY CLOUDSAT

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With the launch of CloudSat, direct observations of cloud vertical structure became possible on the global scale. This thesis utilizes over two years of CloudSat data to study large-scale variations of clouds. We compose a global data set of contiguous clouds (echo objects, EOs) and the individual pixels comprising each EO. For each EO many attributes are recorded. EOs are categorized according to cloud type, time of day, season, surface type, and region. From the categorization we first look at gross global climatology of clouds. Maps of cloud cover are subdivided by EO (cloud) type, and results compare well with previous CloudSat work. The seasonality of cloud cover is also examined.

Focus topics studied in this thesis include: (1) mid-level clouds, (2) stratocumulus clouds, and (3) clouds across the Madden-Julian Oscillation (MJO). The mid-level cloud work found an unexpected frequency peak in EO top heights between 7-8 km in the tropics, further shown to correspond to a global peak in EO top temperature between -15°C – -20°C. Hypotheses are discussed regarding cause of this feature.

Stratocumulus clouds are defined as low-level (tops < 4.5 km), wide (width > 11 km) EOs. Stratocumulus cloud cover agrees (with understandable differences) with other estimates (ISCCP and CALIPSO). The seasonal cycle of stratocumulus over the main stratocumulus decks is examined. The Peruvian and Namibian decks have increased
cloud cover in austral spring in 2007 vs. 2006, corresponding sensibly to sea surface
temperature differences and changes in lower static stability. Looking at rain and drizzle
statistics, wider EOs are found to drizzle more.

Clouds across the MJO are defined relative to temporally filtered OLR data. Cloud cover (volume) doubles (triples) from suppressed to active MJO phases, with some shifts of the relative contributions of different EO types from the front to back of the MJO. Pixel statistics in dBZ-height space correspond to these cloud-type shifts. High anvils and low clouds in front lead deep convection followed by relatively lower anvils in the back.
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TABLE OF CONTENTS

LIST OF TABLES........................................................................................................ vii
LIST OF FIGURES........................................................................................................ viii

Chapter

1 Introduction ................................................................................................................. 1
  1.1 Historical Background on Cloud Observations ................................................. 1
  1.2 Previous Work on Cloud Climatology and Vertical structure ....................... 4
    1.2.1 Satellites........................................................................................................ 4
    1.2.2 Radiosondes .................................................................................................. 6
    1.2.3 Radars .......................................................................................................... 6
  1.3 CloudSat Background ......................................................................................... 7

2 Data and Methods ...................................................................................................... 9
  2.1 CloudSat Cloud Profiling Radar ........................................................................ 9
  2.2 Data Products Utilized ....................................................................................... 10
    2.2.1 CloudSat 2B-GEOPROF Product ................................................................. 10
    2.2.2 ECMWF Auxiliary Product ........................................................................ 15
    2.2.3 TRMM 3B42 Product ................................................................................... 15
    2.2.4 TRMM Microwave Imager ........................................................................... 16
    2.2.5 NCEP Reanalysis ......................................................................................... 16
    2.2.6 Filtered OLR .................................................................................................. 16
  2.3 Subdividing CloudSat Climatology ................................................................... 17
  2.4 Data Versions ..................................................................................................... 21
3 Global CloudSat Climatology

3.1 Defining Echo Object Types
3.1.1 Tropics
3.1.2 Sub-tropics
3.1.3 Mid-latitudes
3.1.4 High latitudes

3.2 Geographical Maps of Echo Cover
3.2.1 Daytime and Nighttime Cover for 60°S – 60°N
3.2.2 Differences in Daytime and Nighttime Echo Cover for 60°S – 60°N
3.2.3 Annual Differences in Echo Cover for 60°S – 60°N
3.2.4 Daytime and Nighttime Cover for High Latitudes
3.2.4.1 Arctic Echo Cover
3.2.4.2 Antarctic Echo Cover

3.3 Seasonal Cycle of Echo Objects

3.4 Echo Object Width Characteristics

3.5 Tropical Examples of Pixel Statistics in dbZ – Height Space
3.5.1 NCFAD Enhancements for Tropical EO Types
3.5.2 Tropical Daytime and Nighttime CFAD Differences
3.5.3 NCFAD Enhancements of Deep Precipitation for Geographical Regions

4 Unexpected Peak Near -15°C in CloudSat Climatology

4.1 Background
4.2 Tropical Mid-level Echo Object Climatology ........................................ 65
4.3 Extratropical Echo Object Climatology .................................................. 69
4.4 Discussion of Possible Interpretations and Mechanisms .......................... 72

5 A Closer Look at Stratocumulus Echo Objects ........................................ 78

5.1 Background .......................................................................................... 78
5.2 Stratocumulus Coverage and Top Height ............................................... 80
5.3 Stratocumulus Seasonal Cycle ................................................................. 87
5.4 Stratocumulus Precipitation Characteristics ......................................... 97
5.5 Summary .............................................................................................. 99

6 Cloud Types Across the Madden-Julian Oscillation .................................. 101

6.1 Background .......................................................................................... 101
6.2 Defining MJO Phases ........................................................................ 103
6.3 Cloud Types by MJO Phase ................................................................ 106
   6.3.1 Echo Object Statistics ................................................................. 106
   6.3.2 Pixel Statistics in dBZ-Height Space ............................................ 110
   6.3.3 Tropical Echo Cover ..................................................................... 113
6.4 Summary and Discussion ....................................................................... 116

7 Conclusions ............................................................................................. 119

References .................................................................................................... 125
# LIST OF TABLES

<table>
<thead>
<tr>
<th>Table</th>
<th>Page</th>
</tr>
</thead>
<tbody>
<tr>
<td>2.1 All EO attributes recorded during pair processing</td>
<td>11</td>
</tr>
<tr>
<td>2.2 Geographic regions. Yellow are mid-latitude, green are sub-tropical regions, blue are tropical regions, and pink are high latitudes</td>
<td>18</td>
</tr>
<tr>
<td>2.3 Seasons included in data set</td>
<td>19</td>
</tr>
<tr>
<td>2.4 2D histograms computed by two IDL codes. Each histogram is made three times: weighted by EO count, cloud cover (area), and volume. * is Julian days since 1 January 2006</td>
<td>19</td>
</tr>
<tr>
<td>6.1 Top values are actual area contribution per phase per cloud type, bottom values are percent area contribution per phase per cloud type</td>
<td>109</td>
</tr>
<tr>
<td>Figure</td>
<td>Page</td>
</tr>
<tr>
<td>--------</td>
<td>------</td>
</tr>
<tr>
<td>1.1 Conceptual diagram of A-train constellation. Adapted from L’Ecuyer et al. (2004)</td>
<td>8</td>
</tr>
<tr>
<td>2.1 Example of CloudSat web. Red line indicates CloudSat path. Black arrows indicate direction of ascending day and descending night time passes</td>
<td>9</td>
</tr>
<tr>
<td>2.2 Histogram of the mean latitude of each EO for all the EOs (red line), EOs excluding double counted EOs (blue line) and EOs excluding double counted EOs and “bad” EOs (black line). “Bad” is defined in text. Bin size is 1°</td>
<td>13</td>
</tr>
<tr>
<td>2.3 Histograms of profile (x) location outliers. (a) and (b) are for the maximum position of an EO, while (c) and (d) are for the minimum position of an EO. Red circles in a-d indicate the outliers. (e) and (f) are the width and number of pixels in the x outlier EOs.</td>
<td>14</td>
</tr>
<tr>
<td>2.4 Geographical regions of the world. Box name descriptions in table 2.2</td>
<td>17</td>
</tr>
<tr>
<td>2.5 Flow chart explaining processing and analyzing steps. Yellow boxes are made during loop step (see text for details), while the green box is done in a separate step.</td>
<td>21</td>
</tr>
<tr>
<td>3.1 Horizontal cloud cover in the tropics (20°S – 20°N) accounted for by clouds with tops and bases in the indicated bins for 16 June 2006 – 31 May 2008. (a) distributes cloudiness by EO bases and tops, while (b) distributes it according to local (echo profile) bases and tops. Contour values (labeled) are in units of 10^3 horizontal echo pixels per bin. Lines and letters delineate EO types, roughly associated to cloud types: (A) deep precipitation (dp), (B) anvil (an, thick cirrus), (C) cirrus (ci), (D) cumulus congestus (cg), (E) alto cumulus/alto stratus (ac), (F) cumulus (cu) and stratocumulus (sc). The 2D distribution of panel b is integrated at right to show the 1D distribution by local echo top height. Bin size is 240 m by 240 m.</td>
<td>26</td>
</tr>
<tr>
<td>3.2 As in figure 3.1 except for the sub-tropics (20° - 35°S/N)</td>
<td>27</td>
</tr>
<tr>
<td>3.3 As in figure 3.1 except for the mid-latitudes (35° - 60°S/N)</td>
<td>28</td>
</tr>
</tbody>
</table>
3.4 As in figure 3.1 except for the high latitudes (latitude > 60°N and latitude < 60°S). The anvil cloud type has been removed.

3.5 Echo cover for (a) all EOs from 16 June 2006 – 31 May 2008 from 60°S – 60°N, (b-h) for each indicated cloud type defined using figures 3.1 – 3.3. Bin size is 3.0° longitude by 3.6° latitude. Echo cover values over 100% can occur due to the fact that an entire EO’s area is placed in the mean latitude and longitude bin of the EO, even if the EO spans several bin. Also because there can be more than one cloud above a given location on Earth.

3.6 As in figure 3.5 except for daytime (1330 LT) CloudSat passes only

3.7 As in figure 3.5 except for nighttime (0130 LT) CloudSat passes only


3.9 Difference from year 2 (1 June 2007 – 31 May 2008) and year 1 (16 June 2006 – 31 May 2007) in echo cover for (a) all EOs and (b-h) each indicated cloud type defined in figures 3.1 – 3.3. Land is masked out for better interpretation. Zero difference line is contoured in black. Light smoothing has been applied for ease of viewing.

3.10 Zonal mean differences from year 2 (1 June 2007 – 31 May 2008) and year 1 (16 June 2006 – 31 May 2007) in echo cover for (a) all EOs and (b-h) each indicated cloud type defined in figures 3.1 – 3.3. Horizontal line represents 0 difference. Light smoothing has been applied for ease of viewing.

3.11 Zonal cross section from 10°S – 10°N of echo cover difference from year 2 (1 June 2007 – 31 May 2008) and year 1 (16 June 2006 – 31 May 2007). Horizontal line represents zero difference. Light smoothing has been applied for ease of viewing.

3.12 Echo cover for (a) all EOs from 16 June 2006 – 31 May 2008 with latitude > 60°N, (b-h) for each indicated cloud type defined using figure 3.4. Bin size is 3.0° longitude by 3.6° latitude.

3.13 Same as figure 3.9 except for latitudes < 60°S

3.14 EO cover distribution by time (Julian days) vs. mean latitude of each EO in our complete data set, 16 June 2006 – 31 October 2008. EO count is weighted by the horizontal span of each EO. Units are 10^4 horizontal pixels per bin. Bin size is 10 days by 3.6° latitude. Black bars indicate missing 2B-GEOPROF data.
3.15 EO cover distribution by mean latitude vs. top height for each EO for each season (June, July, August (JJA); September, October, November (SON); December, January, February (DJF); March, April, May (MAM)) and year (2006, 06; 2007, 07) indicated. EO count is weighted by the horizontal span of each EO. Units are $10^3$ horizontal pixels per bin. Bin size is $3.6^\circ$ latitude by 240 m............................ 49

3.16 Width vs. top height of each EO for 16 June 2006 – 31 May 2008. Distribution is weighted by total number of pixels (volume) per EO to emphasize larger EOs. Bin size is 10 pixels (~ km) by 240 m. Width is the horizontal span of the EO........... 50

3.17 (a) Normalized tropical (20°S – 20°N) CFAD for 16 June 2006 – 31 May 2008. Units are $10^2$ percent per bin. Bin size is 1 dBZ by 240 m. (b) Integral over (a) over all dBZ values (rescaled to total number of pixels per height bin) to give echo coverage. Below white line is heavy rainfall attenuation ........................................ 51

3.18 Same as 3.15 except for EO occurring only over ocean................................. 52

3.19 Same as 3.15 except for EO occurring only over land................................. 52

3.20 Tropical echo cover associated with each EO type (labeled in panels a-g) over land (solid line) only and ocean (dashed line) only for 16 June 2006 – 31 May 2008 ........................................................................... 54

3.21 Shading is CFAD positive enhancements associated with each indicated cloud type relative to the normalized tropical CFAD (Fig. 3.14a). Black contours represent the normalized tropical CFAD. Note that color scale changes between panels ....... 56

3.22 CFAD differences for day (1330) and night (0130) over the tropical (a) ocean only and (b) land only. Values are normalized percentages of the maximum difference. Positive (negative) differences are solid (dashed) contours. (c) CloudSat tropical day and night cloud cover over ocean. (d) same as (c) except over land............. 57

3.23 Same as figure 3.17a, except just for the deep precipitation EO type, box a figure 3.1 only over (a) ocean, (b) land........................................................................................................ 60

3.24 Shading is CFAD positive enhancements associated with deep precipitating EO types in each ocean basin to the normalized tropical deep precipitation ocean CFAD (Fig. 3.20). Black contours represent the normalized tropical deep precipitation ocean CFAD. Color scale is a relative scale, established by the range of values present within each individual panel. West Pacific includes ocean Dp EOs in boxes 16, 17, 24, and 25; East Pacific includes ocean Dp EOs boxes 10, 11, 18, 19; Atlantic includes ocean Dp EOs in boxes 12, 13, 20, 21; Indian Ocean Dp EOs includes ocean in boxes 14, 15, 22, 23 in figure 2.4............................................................................. 62
3.25 Shading is CFAD positive enhancements associated with deep precipitating EO types in each continental region to the normalized tropical deep precipitation land CFAD (Fig. 3.21). Black contours represent the normalized tropical deep precipitation land CFAD. Color scale is a relative scale, established by the range of values present within each individual panel. Amazon is Dp EOs over land in boxes 11, 12, 19, 20; Congo is Dp EOs over land in boxes 13, 14, 21, 22; Maritime Continent is Dp EOs over land in boxes 15, 16, 23, 24.

4.1 Horizontal cloud cover in the tropics distributed by EO top height. (a) all tropical EOs, and midlevel subdivision by layer vs. tower type EOs (types D and E in Fig. 1a). (b-d) cloudiness in layer-type EOs (heavy in all panels) and indicated subdivisions. The Indian, and west and east Pacific are separated at 110°E and 180°W.

4.2 Time (day) of EO vs. top height of EO for one year, July 2006 – June 2007 for EOs in box E of figure 3.1a. Bin size is 10 days by 240 m. Units are 10$^2$ profiles per bin. Vertical lines indicate missing CloudSat days.

4.3 Time (day) of EO vs. mean latitude of EO for EOs in box E of Fig. 3.1a for one year, July 2006 – June 2007. Bin size is 10 days and 3.6° latitude. Units are 10$^2$ profiles per bin. Vertical lines indicate missing CloudSat days.

4.4 Echo cover for (a) upper peak, 6 km – 10 km, (b) lower peak, 4.5 km – 6 km, and (c) the difference panel a – panel b. Bin size is 3° longitude by 3.6° latitude. This figure was made with EOv3 data set.

4.5 1D histogram of EO top temperature for the upper (Fig. 3.1a box E 6 km – 10 km) and lower (Fig. 3.1 box E 4.5 km – 6 km) sub-populations in the tropics (20°S – 20°N) for one year, July 2006 – June 2007. The peak near -30°C in the upper population is the start of the cirrus EOs “leaking” into box E of figure 3.1a. This figure made using EOv3 data set.

4.6 EO base height vs. EO top temperature for all EOs observed in one year, July 2006 June 2007 for (20°N/S – 60°N/S). Binsize is 240 m by 2°C. Contour units are 10$^3$ echo-containing profiles per bin.

4.7 Horizontal cloud cover in the extratropics (20°-60°S/N) distributed by EO top temperature. Subdivisions include (a) day and night, (b) ocean and low land (topography < 1 km), (c) winter (JJA (DJF) S. (N.) Hemisphere and summer (DJF (JJA) S. (N.) Hemisphere), and (d) Theoretical growth curve (Rogers and Yau, 1989, Figure 9.4) overlaid with extratropics echo top temperatures between 0° and -40°C. Bin size is 2°C.

4.8 As in figure 3.1b but redefined to only include profiles where the minimum reflectivity exceeds -20 dBZ. Contouring to the right of the thick 1 to 1 line on the left side is contouring error.
4.9 TRMM 3B42 rainfall composites around daytime and nighttime EOs within the (a) upper peak (Fig. 3.1a, box E, 4.5 km – 6 km) and (b) lower peak (Fig. 3.1a, box E, 6 km – 10 km). This figure was made with EOV3 data set .......................... 75

4.10 (a) Average relative humidity profile for upper peak (solid) and lower peak (dashed) box E EOs. (b) Average lapse rate profile for the tropics (solid), upper peak (dashed) and lower peak (thin solid). The tropics wide lapse rate profile only includes EOs occurring over the ocean. The first km in each figure was excluded due to topography. This figure made with EOV3 data set .......................... 76

5.1 Cartoon of well-mixed, nonprecipitating, strato-cumulus layer, overlaid with data from research flight 1 of DYCOMS-II. Adapted from Stevens (1995) .................... 79

5.2 Same as Figure 3.1 except for 35°S – 35°N for July 2006 – June 2007 ............... 81

5.3 Stratocumulus echo cover for (a) day and night, (b) night, and (c) day from July 2006 – June 2007. Bin size is 3° longitude by 3.6° latitude. Echo cover is defined the same as in chapter 3 ............................................................... 82

5.4 Same as Fig. 5.3 except for both stratocumulus and cumulus EOs .................... 83

5.5 ISCCP low (tops < 680 mb) cloud cover average from 1983 - 2007. Courtesy of David Painemal ................................................................. 83

5.6 2D histogram of EO top height versus longitude for July 2006 – June 2007. Units are $10^{-1}$% horizontal echo pixels per longitude (3°), height (240 m) bin. A, B, C, and D, denote Californian, Peruvian, Namibian, and Australian decks, respectively. Figure made with EOV3 ........................................................................ 84

5.7 Average stratocumulus EO top height per bin. Figure made with EOV3 .......... 86


5.9 Seasonal cycle for (a) Australian [10°S-35°S, 50°E-110°E], (b) Peruvian [5°N -35°S, 70°W-130°W], (c) Namibian [0°-35°S, 40°W-20°E], and (d) Californian [5°N-35°N, 180°W-110°W] stratocumulus echo area coverage. Units are $10^{-1}$ percent contribution for each time (10 days), latitude (3.6°) bin. Black vertical bars denote missing CloudSat days. Figure made with EOV3 dataset ........................................ 89

5.10 Californian stratocumulus echo area coverage. (a) Echo objects with overlapping echo at other altitudes and (b) these without overlapping echo. Figure made with EOV3................................................................. 90
5.11 Same as Fig. 5.3 except just for stratocumulus EOs in September, October, November (SON) (a) 2006 and (b) 2007. (c) 2007 minus 2006 SON stratocumulus cover.......................................................... 90

5.12 TRMM-TMI sea surface temperature for the average of August, September, and October of (a) 2006 and (b) 2007 and (c) the difference between 2007 and 2006 .................................................................................. 92

5.13 NCEP reanalysis 1000 mb temperature for August, September, and October of (a) 2006 and (b) 2007 and (c) the difference between 2007 and 2006 from NCEP/NCAR reanalysis. In (c) Solid curves are positive, dashed are negative, and thick solid is zero .............................................................. 94

5.14 700 mb theta computed from NCEP reanalysis 700 mb temperatures for August, September, and October of (a) 2006 and (b) 2007 and (c) the difference of 2007 and 2006. In (c) Solid curves are positive, dashed are negative, and thick solid is zero .......................................................................................... 95

5.15 Static stability for August, September, and October of (a) 2006 and (b) 2007 and (c) the difference of 2007 and 2007 from NCEP/NCAR reanalysis. In (c) Solid curves are positive, dashed are negative, and thick solid is zero .................................................................................. 96

5.16 ASO 500 mb omega for (a) 2006 (b) 2007 and (c) the difference of 2007 and 2006 from NCEP/NCAR reanalysis. Solid curves are positive, dashed are negative, and thick solid is zero. Contour interval is 0.01 Pa/s........................................................ 97

5.17 Cumulative distribution of the total (black diamonds), drizzling (red triangles), and raining (blue pixels) number of pixels per sc EO......................................................................................... 98

5.18 Histogram of (a) raining-to-total pixel ratio for raining stratocumulus EOs and (b) drizzling-to-total pixel ratio for drizzling stratocumulus EOs. Stcu EOs are divided into quartiles according to volume. Area under each curve equals one. Bin size is 0.1 ........................................................................................................ 99

6.1 Schematics of clouds through MJO phases from (a) Lin and Johnson (1996), (b) Benedict and Randall (2007), and (c) Morita et al. (2006). (d) is cloud development through a mesoscale convective system from Zipser (1981)................................. 103

6.2 Raw OLR from 15°S – 15°N for July 2006 – February 2008 from NOAA PSD website overlaid with filtered MJO OLR. Contour interval is 10 Wm⁻², with negative values dashed................................................................. 104

6.3 Phase definition for MJO. Standardized filtered MJO OLR (solid contours in Fig. 6.2) is scatter plotted against its standardized local time derivative. Numbers represent the phase of the MJO, for times and places when amplitude (distance from origin) is greater than two standard deviations .................................................. 105
6.4 Horizontal cloud cover weighted histograms (like Fig. 3.1b) in each indicated phase of the MJO. Cloudiness is distributed by local (echo profile) bases and tops. Contour values (labeled) are units of 10 (echo-containing profiles) per 240 m altitude bin. Lines and letters delineate EO types, same as figure 3.1b. The 2D distribution of each phase is integrated at the right of each figure to show the 1D distribution of each MJO phase cloud cover by local cloud top height (in units of % per 240 m) .............. 107

6.5 (a) Percent areal coverage for each cloud type (color) during each phase of the MJO (different bars). (b) As in (a) but total areal contribution per cloud type. (c) As in (a) but for volume. (d) As in (b) but for volume ............................................. 108

6.6 Normalized CFAD from all phases of the MJO. Units are $10^{-2}$ percent contribution for each dBZ, height bin. Bin size is 1 dBZ by 240 m................................................. 111

6.7 Shading is CFAD positive and negative enhancements associated with each MJO phase relative to the normalized MJO CFAD for all phases (Fig. 6.5). Black contours represent the normalized MJO CFAD for all phases....................................................... 113

6.8 Percent echo cover per indicated phase of the MJO................................................. 115

6.9 (a) Percent cloud cover through each phase of the MJO, contour units are percent. (b) Anomalous cloud cover through each phase of the MJO relative to the mean cloud cover over all phases. Contour units are percent difference, solid (dashed) contours are positive (negative) enhancements .................................................. 116

6.10 Cartoon of the stretched building block hypothesis. (a) Unfiltered view. MCSs have varying amounts of shallow (left), deep (middle), and stratiform (right) clouds according to their position within a large-scale wave event. (b) Low-pass filtered view. When the mean is removed and smoothing broadens features, the resulting large-scale pattern resembles the MCS life cycle. Adapted from Mapes et al. 2006 .......................................................... 118
Chapter 1: Introduction

The goal of this thesis is to study large-scale variations of clouds observed by CloudSat. To accomplish this goal, this thesis examines two similar yet oppositely posed questions: (1) Given a certain cloud type, where are those clouds occurring? and (2) For a certain geographical or space-time region what are the dominant cloud types and what are the differences in clouds from one region to another. Looking at cloud types is important because different clouds are governed by different dynamical processes and have different microphysical properties and radiative impacts (Hartmann et al. 1992). This thesis begins by giving a brief history on cloud observations, climatology, and vertical structure.

1.1 Historical Background on Cloud Observation

Clouds represent the visual manifestation of the atmospheric component of the Earth’s hydrological cycle. Clouds are born out of various dynamic, thermodynamic, and microphysical processes. Through these process clouds are a fundamental component in the redistribution of mass, momentum, and energy throughout the atmosphere making them an important part of the atmosphere’s general circulation. Clouds are complex in that they exist on a variety of time and space scales and are forced by and feedback on various dynamical scales in the atmosphere. Such complexity could lead one to describe clouds as both evanescent and permanent features in the sky, affecting both day-to-day weather and climatology. Evanescent in the sense that clouds undergo constant motion and transformation through a life cycle of formation, growth, and decay, yet permanent in
that the life cycle of cloud formation as a whole is ongoing. On day-to-day time scales clouds affect local precipitation and surface temperature. On longer time scales clouds play a vital role in the Earth’s radiation budget, acting to both shade and blanket the earth by reflecting shortwave radiation and/or trapping longwave radiation (Wielicki et al. 1995).

Because of clouds visual appearance in the sky, they have been of interest to humans for many centuries, at least as far back as the ancient Chinese in the 2nd century (Hobbs 1989). In the 19th century Luke Howard (1803) developed the first cloud classification scheme based on a cloud’s physical appearance. He established three major cloud types: cirrus, cumulus, and stratus with an additional descriptor nimbus to denote falling precipitation (Stephens 2003 and Hobbs 1989). Today the cloud classification scheme has evolved to include 10 different cloud types set forth by the World Meteorological Society in 1956. The scheme is based on cloud base height, physical appearance, composition, and presence of precipitation. Using these cloud types Hahn et al. (1982, 1984) and Warren et al. (1986, 1988) compiled atlases of the global distribution of cloud types and cloud fraction from surface observations. While surface observation provided the first benchmark of cloud fraction and distribution of cloud types and can be useful for tracking cloud climatology of a specific location, they are sparsely located across the globe (especially oceanic regions) and can have difficulty seeing multi-layer cloud systems.

Great strides in understanding and observing cloud processes and climatology have been possible through the advancement of instrumentation over the past several decades. After World War II radars started being used for meteorological studies,
providing insight to the internal structure of clouds and precipitation (Rinehart 2004). Soon after, scientists began to view clouds on a global scale with the launch of satellites in the 1970s (Hobbs 1991). The ability to view large-scale cloud organization was now possible. With these satellites great efforts have been made to determine cloud climatology and their radiative impacts through international research projects such as the International Satellite and Cloud Climatology Project (ISCCP) (e.g. Schiffer and Rossow 1983, Rossow and Schiffer 1999, Jakob and Tselioudis 2003) and the Global Energy and Water Cycle Experiment (GEWEX) Cloud Systems Study (GCSS) (GCSS Science Team 1993). Likewise, radars have served an important part in advancing the knowledge of precipitation processes, cloud composition, cloud life cycles, and cloud microphysics. Studies from land based radars are numerous but place specific. Studies of oceanic clouds and convection have been accomplished through large field campaigns (e.g. TOGA COARE) or tropical island based radars (e.g. Manus ARM site).

Radio sondes have also been useful in assessing cloud climatology and characteristics (e.g. Poore et al. 1995, Wang and Rossow 1995, and Wang et al. 2000). Most recent cloud observing technologies have the synergy of radars and satellites, with the placement of radars in space. The Tropical Rainfall Measuring Mission (TRMM) launched in 1997 established the first space-based meteorological radar to study tropical precipitating clouds (Simpson et al. 1988). CloudSat followed in 2006, establishing the first spaced based cloud radar to study global clouds (Stephens et al. 2002).

Each of the above cloud observing techniques (human observations, satellites, radiosondes, and radars) comes with a unique set of advantages and disadvantages. Human observations benefit from longevity and consistency: Cloud reports are available
for many decades with no change in official observing techniques. Surface observers can see low clouds sometimes obstructed from satellites due to the presence of high clouds, as well as clouds smaller than a satellite’s pixel size. Cloud classification is based on meteorological conditions and not inferred from satellite derived radiative properties. The disadvantage of surface observations is they are location specific (Warren and Hahn 2003).

Satellites overcome the location specific problem by being able to view the entire globe via geostationary satellites that continuously monitor one section of the globe or polar orbiters that pass over all portions of the globe. Infrared (IR) emissions from cloud top allow cloud top height to be measured based on temperature soundings of the atmosphere provided by radiosondes or satellite sounders. Satellite measurements in the visible (VIS) spectrum provide information on the cloud optical depth (Petty 2006).

1.2 Previous Work on Cloud Climatology and Vertical Structure

1.2.1 Satellites

By combining optical depth properties and cloud top pressure, ISCCP defines 9 different cloud types, delineating high, middle, and low-based clouds by top pressure, and specific types via optical thickness (Rossow and Schiffer 1999). Another way of determining cloud types via satellite is through pattern and texture recognition (e.g. Ebert 1987, Garand 1988, Bankert 1994, and Uddstrom and Gray 1996). Pattern recognition algorithms are trained by experts (human interpretation of satellite images) then used to
analyze various IR and VIS satellite scenes. With the ISCCP classifications it is possible to discern the global radiative impacts of specific cloud types (Zhang et al. 1995).

Because ISCCP satellites are limited to 2 channels, one IR and one VIS, cloud properties such as cloud base and layer thickness, as well as particle size, shape, and phase are undetermined. Additionally, with the 2 channel limit ISCCP algorithms must assume single cloud layer scenes (Rossow and Schiffer 1999). This assumption is poor since surface observations (Hahn et al. 1982) and aircraft observations (Tian and Curry 1989) report cirrus co-occurring with stratus 53% and 61% of the time, respectively, when a cirrus cloud is present (statistics for approximately 30°N – 60°N). Warren et al. (1985) also reports surface observation of cloud occurrence and co-occurrence, but is broken up by season and time of day. ISCCP will report cirrus over stratus co-occurrence as a single mid-level cloud (Chang and Li 2005).

Several methods have been developed to detect multiple cloud layer scenes and cloud particle phase via satellites. Baum et al. (1992, 1994, 1995) developed a multispectra multiresolution (MSMR) approach. In the MSMR method, HIRS (high-resolution infrared radiometer sounder) and AVHRR (advanced very high resolution radiometer) data are combined to analyze overlapping cloud layers. Other methods use a combination of IR and VIS satellite data with microwave measurements from satellites (Liu et al. 1995, Shue et al. 1997, Lin et al. 1998) or ground radiometers (ARM SGP site, Huang et al. 2005) to discern cloud types, properties and layers.

A common problem is the IR/VIS and microwave data are taken from different instruments in the above studies (Baum et al. 1992, Liu et al. 1995, Shue et al. 1997, and Huang et al. 2005). This requires interpolating one instruments field of view (FOV) to
another. With the launch of MODIS (the Moderate Resolution Imaging Spectrometer) (Platnick et al. 2002) 36 channels became available on one instrument. Chang and Li (2005a, 2005b) used MODIS to establish a near global climatology of clouds.

1.2.2 Radiosondes

Radiosondes benefit from their ability to penetrate cloud layers, making it possible to discern cloud layer thickness. Cloud layers are determined by dew point depression (or relative humidity) measurements. Radiosondes studies can be limited to certain areas (i.e. northern hemisphere, Poore et al. 1995, Wang and Rossow 1995) and times of day (radiosondes are usually launched at 0000 and 1200 UTC), although Wang et al. (2000) used about 1200 radiosonde stations from around the globe to determine global cloud vertical structure.

1.2.3 Radars

Radars provide direct measurements of cloud vertical structure and composition. Radars measure the returned power of emitted electromagnetic radiation after hitting a volume of distributed point targets. Returned power is referred to as reflectivity and has units of $\text{mm}^6/\text{m}^3$, or more conveniently in log scale as dBZ ($10 \log_{10} (Z)$ where $Z$ is in $\text{mm}^6/\text{m}^3$) (Rinehart 2004). Radars come in a variety of shapes and sizes (from S-band at cm wavelength to W-band at mm wavelength), each designed for specific tasks. Vertically pointing W-band radars are capable of detecting hydrometers for most non-precipitating clouds through the entire depth of the troposphere independent of the number of cloud layers. Under moderate to heavy precipitation the radar beam can
become fully attenuated. Additionally, thin cloud (such as cirrus) or shallow water clouds not containing large ice particles or drizzle droplets may go undetected at millimeter wavelengths. Evaluation of millimeter wavelength radars and techniques for determining cloudy echo are available in Clothiaux et al. (1995) and Clothiaux et al. (2000). Hogan and Illingworth (2000) used millimeter radar data collected over southern England to evaluate the overlap of cloud layers observed there. Mace and Bensen-Troth (2002) utilized millimeter radar data from the ARM sites to provide a more extensive and longer record of cloud type and overlap. With the launch of CloudSat in April of 2006 (Stephens et al. 2002) millimeter radar data is now available on the global scale.

1.3 CloudSat Background

CloudSat was launched 28 April 2006 and went to work 2 June 2006, after maneuvering into place. CloudSat flies as part of the Sun-synchronous A-train constellation of satellites at a mean equatorial altitude of 705 km, situated behind Aqua and about 15 seconds in front of CALIPSO (Fig. 1.1). Being a part of the A-train, one of the main objectives of CloudSat is to evaluate cloud properties along with other A-train satellites (i.e. Aqua). Other principle objectives include: evaluating the representation of clouds and cloud processes in global atmosphere circulation models, evaluating the relationship between the vertical profiles of cloud liquid water and ice and the radiative heating of the atmosphere and surface, and contribute to improving the understanding of the indirect effect of aerosols on clouds (Stephens et al. 2008).

Chapter 2 discusses the data and methods used. Chapter 3 looks at gross global CloudSat climatology. Then chapters 4 – 6 examine specific topics that were spurred by
the creation of our CloudSat cloud dataset. Chapter 4 looks at a novel observation seen in tropical climatology: A peak near -15°C in CloudSat climatology. Chapter 5 examines the well-known sub-tropical stratocumulus decks. Chapter 6 takes the idea of looking at clouds in various geographical regions and applies it to various wave phases of the Madden Julian Oscillation (MJO). Chapter 7 concludes the thesis.

Figure 1.1 – Conceptual diagram of A-train constellation. Adapted from L’Ecuyer et al. (2004)
Chapter 2: Data and Methods

2.1 CloudSat Cloud Profiling Radar

On board CloudSat is a 94 GHz (3 mm) nadir pointing cloud profiling radar (CPR) (Im et al. 2005). The CPR has a 1.4 km across track by 1.8 km along track nominal footprint, but oversampling results in profiles 1.1 km apart. Throughout the thesis “profile” will refer to an entire single vertical column of CloudSat bins. Vertical resolution (range gate width) is 480 m, but backscatter is oversampled to provide a 240 m data resolution. There are 125 vertical bins per profile. The minimal detectable signal (MDS) designed to be -28 dBZ, was found to be -30 to -31 dBZ in early CloudSat results from Haynes and Stephens (2007). CloudSat makes approximately 14 orbits per day with an equator passing time of 0130 and 1330 local time (LT). The orbit path repeats itself about every 17 days, giving a CloudSat “web” of information (Fig. 2.1). Unlike traditional satellites which have a large field of view and see most points on the globe, CloudSat always has diamond sized gaps approximately 1.5° longitude by 3.6° latitude, which gives exactly 233 diamonds in longitude. Below is a brief description of each product/data set and how they are implemented.

Figure 2.1 – Example of CloudSat web. Red line indicates CloudSat path. Black arrows indicate direction of ascending day and descending night time passes.
2.2 Data Products Utilized

2.2.1 CloudSat 2B-GEOPROF Product

This thesis centers on the data product known as 2B-GEOPROF, version 4 (Mace 2004). The 2B-GEOPROF product includes 2D (alongtrack x vertical) arrays of the radar reflectivity factor and a gaseous attenuation correction (which we used), as well as of a “cloud mask” value between 0 and 40. Zero indicates no hydrometer detected, while higher values indicate an increased likelihood of hydrometer detection. Following product documentation advice, we take pixels with mask values \( \geq 20 \) as a hydrometeor echo detection. The percent of false detection for an echo with a mask value \( \geq 20 \) is estimated to be less than 5% from product quality control statements. More information on the 2B-GEOPROF product can be found online at the CloudSat data processing center (http://cloudsat.cira.coloradostate.edu).

To simplify and enrich our access to the CloudSat dataset, we devised a hierarchical method of analyzing and storing the cloudy echo portion of the data, with two tiers: (1) as echo objects (EOs) and (2) as the pixels comprising EOs. An EO is defined as a contiguous region (in the radar’s vertical sampling plane) of cloud mask \( \geq 20 \), consisting of at least three pixels with their edges (not merely corners) touching. Throughout the thesis EOs are often described/plotted in terms of their area or volume. EO area refers to the horizontal span (or width) of the EO, while EO volume is the total number of pixels comprising an EO. One and two pixel objects were initially retained in our first processing, and constituted around 63% of the EOs, but less than 5% of horizontal echo coverage, and less than 1% of echo volume. They were concentrated at
low levels near topography, suggesting clutter contamination. Eliminating them seemed no great loss, while halving dataset memory requirements.

For each EO a plethora of attributes is recorded (Table 2.1), including geometric and geographic information as well as substructure information. Geographic attributes include minimum, mean, and maximum latitude and longitude, as well as underlying surface altitude and land mask data allowing discrimination of land (by altitude), sea, and coastal EOs. Geometric data include top, mean, and bottom heights, width, and total number of pixels, which allow computation of geometrical and mean thickness. We also save statistics about each EO’s internal structure and on vertically overlapping cloudiness. Contiguous regions of reflectivity greater than 0 and -17 dBZ are identified using IDL’s LABEL_REGION. We refer to these internal contiguous regions as cells.

<table>
<thead>
<tr>
<th></th>
<th>Echo Object Attributes</th>
</tr>
</thead>
<tbody>
<tr>
<td>1</td>
<td>Date of EO</td>
</tr>
<tr>
<td>2</td>
<td>Time of Day</td>
</tr>
<tr>
<td>3</td>
<td>Minimum Latitude</td>
</tr>
<tr>
<td>4</td>
<td>Mean Latitude</td>
</tr>
<tr>
<td>5</td>
<td>Max Latitude</td>
</tr>
<tr>
<td>6</td>
<td>Minimum Longitude</td>
</tr>
<tr>
<td>7</td>
<td>Mean Longitude</td>
</tr>
<tr>
<td>8</td>
<td>Max Longitude</td>
</tr>
<tr>
<td>9</td>
<td>Top Height</td>
</tr>
<tr>
<td>10</td>
<td>Mean Height</td>
</tr>
<tr>
<td>11</td>
<td>Bottom Height</td>
</tr>
<tr>
<td>12</td>
<td>Width</td>
</tr>
<tr>
<td>13</td>
<td>Total number of pixels per EO</td>
</tr>
<tr>
<td>14</td>
<td>Total number of pixels above EO</td>
</tr>
<tr>
<td>15</td>
<td>Number of cells &gt; 0 dBZ per EO</td>
</tr>
<tr>
<td>16</td>
<td>Biggest cell &gt; 0 dBZ per EO</td>
</tr>
<tr>
<td>17</td>
<td>Number of pixels &gt; 0 dBZ per EO</td>
</tr>
<tr>
<td>18</td>
<td>Number of pixels &gt; -17 dBZ per EO</td>
</tr>
<tr>
<td>19</td>
<td>Underling surface height</td>
</tr>
</tbody>
</table>

Table 2.1 – All EO attributes recorded during pair processing.
To locate and tally EOs, CloudSat orbit files (which begin and end at the equator) must be processed as pairs. An orbit begins at the first profile south of the equator on the descending pass of the satellite. To understand pair processing, consider three CloudSat orbits, A, B and C. An orbit pair is created by stitching the first 2000 CPR profiles of orbit B to the end of orbit A. The number 2000 is arbitrary, but longer than the biggest EOs we have found near the equator. Stitching together orbits maintains contiguity of EOs that may span the end of one orbit and the beginning of the next. Contiguous EOs in the (A+B) stitched array are identified by Interactive Data Language (IDL) routine LABEL_REGION. If an entire EO is contained in profiles 2 - 2000 of orbit B, that object is discarded, since it will be counted when processing the next orbit pair (B+C). Any object beginning in the first profile of orbit A is also discarded, as this would have been accounted for in the previous pair processing. Because the CloudSat data processing center treats each orbit independently, the last three and beginning three profiles of each orbit are masked as zero, because the mask routine uses a seven pixel wide spatial filtering box to screen false detections (Marchand et al. 2008). Near orbit edges, we had to create our own cloud mask. For the last five and first five profiles of each orbit, pixels between the surface and 18 km that have a radar reflectivity factor between -25 dBZ and 30 dBZ were counted as an echo detection and assigned a cloud mask value of 30. These threshold values were chosen based on study of tropical EOs not affected by the mask problem. Only EOs very near the equator are affected by this procedure. After processing we realized EOs spanning the equator were double counted, due to the inherent nature of the LABEL_REGION IDL procedure. Additionally, our near equator cloud mask caused an unreasonable spike in the latitudinal distribution of
EOs (i.e. some EOs where counted as echo that should not have been). Figure 2.2 is a histogram of the mean latitude for each EO. The red curve includes all EOs, the blue excludes double counted EOs and the black excludes double counted as well as “bad” EOs.

**Figure 2.2** – Histogram of the mean latitude of each EO for all the EOs (red line), EOs excluding double counted EOs (blue line) and EOs excluding double counted EOs and “bad” EOs (black line). “Bad” is defined in text. Bin size is 1°.

“Bad” equator EOs caused by our own near equator cloud mask mainly occurred in the first and last several profiles of a stitched together orbit pair (Fig. 2.3 a-d). To eliminate outliers in the EO distribution of profile location we excluded all EOs in those locations with a width of 10 pixels and a size of 3 pixels (Fig. 2.3 e-f). Additionally, all EOs with a top height less than or equal to 500 m were eliminated, as surface clutter contaminates the first CloudSat vertical bin. There are 6,650,125 EOs in our data set, 16 June 2006 to 31 October 2008.
Figure 2.3 – Histograms of profile (x) location outliers. (a) and (b) are for the maximum position of an EO, while (c) and (d) are for the minimum position of an EO. Red circles in a-d indicate the outliers. (e) and (f) are the width and number of pixels in the x outlier EOs.

For each EO we save the horizontal and vertical positions and reflectivity values of all the individual cloud-masked echo pixels in a separate file. With these data, we can
reconstruct every EO observed by CloudSat, providing an uncompromised visual display or statistical description of all the pixels in all the EOs satisfying any given set of criteria.

We convert cloudy pixel counts to cloud cover units by dividing by the total number of atmospheric columns sampled by CloudSat in the region and period being considered. This sampling rate was computed during processing, and is approximately 8.05 pixels per square degree per day. It is a weak function of latitude over most of the earth, but rises steeply near the satellite’s near-polar turnaround latitudes.

2.2.2 ECMWF Auxiliary Product

Weather analyses from ECMWF (temperature, pressure, and specific humidity) are available concurrent with CloudSat data, interpolated to the CloudSat grid on a pixel-by-pixel basis, in the ECMWF-AUX product (ref). An atmospheric “sounding” for each EO was computed after the EO identification, in a separate process stage, by averaging the ECMWF data over each height level in the horizontal region spanned by the EO. Given the uncertainties and coarser native resolution of the analyses relative to radar pixels, this averaging approach seems appropriate for all but the largest EOs (which can span several grid cells in the weather model). Excluding these large EOs does not change the main results or conclusions of this thesis.

2.2.3 TRMM 3B42 Product

The TRMM 3B42 product (Huffman et al. 2007) provides 3-hour, 0.25° X 0.25° rainfall rates, which we have rebinned to 1° X 1°. TRMM 3B42 combines rainfall estimates from radar, infrared and microwave instruments and rain gauges. For each EO
(used in chapter 4) observed by CloudSat, the closest TRMM 3B42 rainfall rate grid cell is identified, allowing composites of the surrounding weather conditions (rainfall pattern) in space and time to be constructed.

2.2.4 TRMM Microwave Imager

The TRMM Microwave Imager (TMI) provides 3-day mean maps of sea surface temperature (SST) on a 0.25° resolution. The microwave imager is advantageous over traditional infrared observations for SST measurements because at 10.7 GHz it is able to see thru clouds and is not affected by aerosols and is insensitive to water vapor (Wentz 1997).

2.2.5 NCEP Reanalysis

Many atmosphere and ocean variables are available from 1948 to present from the National Center for Environmental Prediction (NCEP) reanalysis project. In chapter 5 we use daily mean values of 1000 mb and 700 mb temperatures on a 2.5° grid.

2.2.6 Filtered OLR

The filtered MJO olr signal was obtained courtesy of George Kiladis. Space-time filtering OLR was done following Wheeler and Kiladis (1999) to isolate eastward wavenumbers 0 – 9 with periods between 30 – 96 days. Unlike Wheeler and Kiladis (1999) the filtering is done on raw OLR with no symmetric and antisymmetric decomposition. More details given in chapter 6.
2.3 Subdividing CloudSat Climatology

We classified EOs by a variety of categories: (1) cloud type, (2) geographical region, (3) time of day, (4) underlying topography, and (5) season. Cloud types are defined according to EO base and EO top height criteria (discussed more in chapter 3). Geographical regions are defined by latitude, longitude boxes (Fig. 2.4). Table 2.2 provides the name description of each box. Boxes were designed subjectively, keeping in mind dynamical regimes (e.g. tropics separated from mid-latitudes), known regional differences (e.g. East Pacific separated from West Pacific), seasonal oscillations (e.g. monsoons), and convenience for later analysis. Time of day (i.e. sampling orientation, Fig. 2.1) separates the nighttime (~ 0130 LT) and daytime (~ 1330 LT) CloudSat overpasses. Maximum and minimum ground altitude under an EO distinguishes EOs occurring over land (z maximum > 0) and ocean (z maximum < 0). Coastal EOs were supposed to be distinguished from EOs occurring entirely over land or ocean, but a code mistake (z maximum > 0 vs. z minimum > 0) put all clouds spanning land and ocean into the land classification. There are nine 3-month seasons in our data set (Table 2.3).

Figure 2.4 – Geographical regions of the world. Box name descriptions in table 2.2
<table>
<thead>
<tr>
<th>Number</th>
<th>Geographical Regions</th>
</tr>
</thead>
<tbody>
<tr>
<td>1</td>
<td>North east Pacific &amp; western U.S.</td>
</tr>
<tr>
<td>2</td>
<td>North west Atlantic &amp; eastern U.S.</td>
</tr>
<tr>
<td>3</td>
<td>North east Atlantic &amp; Europe</td>
</tr>
<tr>
<td>4</td>
<td>Russia &amp; north west Pacific</td>
</tr>
<tr>
<td>5</td>
<td>Northern Hemisphere (NH) sub-tropical east Pacific</td>
</tr>
<tr>
<td>6</td>
<td>Sub-tropical Atlantic/Gulf of Mexico</td>
</tr>
<tr>
<td>7</td>
<td>East north Africa</td>
</tr>
<tr>
<td>8</td>
<td>West north Africa</td>
</tr>
<tr>
<td>9</td>
<td>Sub-tropical west Pacific/Tibetan Plateau</td>
</tr>
<tr>
<td>10</td>
<td>NH central east Pacific</td>
</tr>
<tr>
<td>11</td>
<td>NH tropical east Pacific</td>
</tr>
<tr>
<td>12</td>
<td>NH tropical Atlantic</td>
</tr>
<tr>
<td>13</td>
<td>Africa monsoon area</td>
</tr>
<tr>
<td>14</td>
<td>East Africa</td>
</tr>
<tr>
<td>15</td>
<td>North Indian Ocean/Indian monsoon</td>
</tr>
<tr>
<td>16</td>
<td>West Pacific warm pool</td>
</tr>
<tr>
<td>17</td>
<td>NH central west Pacific</td>
</tr>
<tr>
<td>18</td>
<td>Southern Hemisphere (SH) central east Pacific</td>
</tr>
<tr>
<td>19</td>
<td>(SH) tropical east Pacific</td>
</tr>
<tr>
<td>20</td>
<td>East South America</td>
</tr>
<tr>
<td>21</td>
<td>Namibian Stratocumulus Region</td>
</tr>
<tr>
<td>22</td>
<td>East south Africa</td>
</tr>
<tr>
<td>23</td>
<td>SH tropical Indian Ocean</td>
</tr>
<tr>
<td>24</td>
<td>Papua New Guinea</td>
</tr>
<tr>
<td>25</td>
<td>SH central west Pacific</td>
</tr>
<tr>
<td>26</td>
<td>SH Sub-tropical east Pacific</td>
</tr>
<tr>
<td>27</td>
<td>SH Sub-tropical Atlantic</td>
</tr>
<tr>
<td>28</td>
<td>Sub-tropical Indian Ocean</td>
</tr>
<tr>
<td>29</td>
<td>Australia</td>
</tr>
<tr>
<td>30</td>
<td>South East Pacific</td>
</tr>
<tr>
<td>31</td>
<td>South Atlantic</td>
</tr>
<tr>
<td>32</td>
<td>South Indian Ocean</td>
</tr>
<tr>
<td>33</td>
<td>South West Pacific</td>
</tr>
<tr>
<td>34</td>
<td>Antarctic</td>
</tr>
<tr>
<td>35</td>
<td>Arctic</td>
</tr>
</tbody>
</table>

**Table 2.2** – Geographic regions. Yellow are mid-latitude, green are sub-tropical regions, blue are tropical regions, and pink are high latitudes.
<table>
<thead>
<tr>
<th>Season</th>
<th>2006</th>
<th>2007</th>
<th>2008</th>
</tr>
</thead>
<tbody>
<tr>
<td>June, July, August (JJA)</td>
<td>X</td>
<td>X</td>
<td>X</td>
</tr>
<tr>
<td>(starting 16 June)</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>September, October, November (SON)</td>
<td>X</td>
<td>X</td>
<td>X</td>
</tr>
<tr>
<td>(excluding Nov.)</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>December, January, February (DJF)</td>
<td>X</td>
<td>X</td>
<td></td>
</tr>
<tr>
<td>(2006/2007)</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>March, April, May (MAM)</td>
<td>X</td>
<td>X</td>
<td></td>
</tr>
</tbody>
</table>

**Table 2.3** – Seasons included in data set

<table>
<thead>
<tr>
<th>2D Histogram Type</th>
<th>Bin Size</th>
<th>Min Values</th>
<th>Max Values</th>
</tr>
</thead>
<tbody>
<tr>
<td>1 Longitude x Latitude</td>
<td>3° x 3.6°</td>
<td>-180°, -90</td>
<td>180°, 90°</td>
</tr>
<tr>
<td>2 Latitude x Top Height</td>
<td>3.6° x 250 m</td>
<td>-90°, 0 m</td>
<td>90°, 20 km</td>
</tr>
<tr>
<td>3 Latitude x day of EO</td>
<td>3.6° x 10 days</td>
<td>-90°, 166*</td>
<td>90, 1035*</td>
</tr>
<tr>
<td>4 Longitude x Top Height</td>
<td>3° x 250 m</td>
<td>-180°, 0 m</td>
<td>180°, 20 km</td>
</tr>
<tr>
<td>5 Day of EO x Height</td>
<td>10 days x 250 m</td>
<td>166, 0</td>
<td>1035*, 20 km</td>
</tr>
<tr>
<td>6 Base x Top Height</td>
<td>250 m x 250 m</td>
<td>0 m, 0 m</td>
<td>20 km, 20 km</td>
</tr>
<tr>
<td>7 Width x Top Height</td>
<td>10 km x 250 m</td>
<td>1 m, 0 m</td>
<td>3 km, 20 km</td>
</tr>
<tr>
<td>8 Pixel dBZ x Height (“CFADS”)</td>
<td>1 dBZ x 250 m</td>
<td>-50 dBZ, 0 m</td>
<td>25 dBZ, 20 km</td>
</tr>
<tr>
<td>9 Profile Base x Profile Top</td>
<td>250 m x 250 m</td>
<td>0 m, 0 m</td>
<td>20 km, 20 km</td>
</tr>
</tbody>
</table>

**Table 2.4** – 2D histograms computed by two IDL codes. Each histogram is made three times: weighted by EO count, cloud cover (area), and volume. * is Julian days since 1 January 2006.
To analyze the various categories, code was written to loop through all 35 regions, 7 cloud types (6 in the high latitudes), 2 times of day, 2 underlying topographies, and 9 three-month seasons to make 9 different 2D histograms listed in table 2.4. Histogram types 1-7 are created from EOs, while 8 and 9 are created in a separate code at pixel level. Postage stamps (pixel renderings of EO morphology) are created in a separate code. A specified number of EOs are chosen at random from a desired set of EOs. The probability of selection is area weighted, so potentially numerous small EOs do not dominate the postage stamps. All together 78,732 figures were created.

Each was saved with a unique filename to make recalling them at a later time easy. For example, Afmonsoon.cb.day.land.JJA07.lonlathis.sav would be the longitude vs. latitude histogram (i.e. a map) for cumulonimbus EOs, over land, during the day, for June, July, and August of 2007 in the African monsoon region (box 13, Fig 2.4). With so many files created it becomes necessary to combine histograms of the same type (e.g. longitude vs. latitude) but made from different specifications (e.g. tropics and mid-latitudes) to make more meaningful plots, as a histogram of latitude vs. longitude for one region is quite dull. Because histogram bin sizes and ranges remain constant for each file created (Table 2.4), histogram arrays are simply additive or subtractive. For example, if we desired a global distribution of daytime cumulonimbus clouds we would add all regions, seasons, and topography types of cumulonimbus clouds occurring during the day. We could then do the same thing but for nighttime cumulonimbus and then compare (or even quantitatively difference) the daytime and nighttime distributions. Examples are given in chapter 3.
A summary of the steps taken to build the CloudSat climatology is shown in figure 2.5. The EOs are created, then sub-divided (separated into categories) and counted to populate 2D histograms in attribute space. Area and volume weighting are applied to each histogram; most results in this thesis are area (horizontal coverage) weighted. These elementary histograms can then be combined later to make meaningful 2D histograms. 1D plots can also be created by pooling data across unused 2D dimensions to increase sampling and achieve smoothing and statistical significance.

**Figure 2.5** – Flow chart explaining processing and analyzing steps. Yellow boxes are made during loop step (see text for details), while the green box is done in a separate step.

### 2.4 Data Versions

The process of making EOs went through several iterations for various reasons including: fixing code errors, adding EO attributes, and including more CloudSat days. In all, we have 5 EO versions (EOV) saved. Different versions were used to make
various plots throughout this thesis, for historical reasons that never impact conclusions. Below is a brief description of each EOV.

EOV0 was completed January 2007, so naturally contained only the first half year of CloudSat data. There were several code errors. EOV1 sought to improve on EOV0 code errors and was completed August 2007. At this point we had a years worth of data and some additional EO attributes were saved including daytime and nighttime overpass distinctions of each EO, an account of cells greater than 0 dBZ per EO, and total number of cloudy and non-cloudy CloudSat profiles. A couple more code errors were fixed in EOV2, completed in January 2008, and we now saved pixels greater than -17 dBZ per EO along with ground height under each EO. EOV2 contains one and half years of CloudSat data. These three versions (EOV 0-2) are not used in the thesis, but mentioned here for completeness.

After completing EOV2, we discovered that EOs spanning the equator were being counted twice. So, EOV3 sought to resolve this issue while improving a couple other code errors. CloudSat data in EOV3 is from June 2006 to June 2008. EOV4 in August 2008 became necessary when we discovered that fixes made in EOV3 were not successful because the CloudSat team handles each orbit individually (section 2.2.1). Therefore, for EOV4 we created our own cloud mask near the equator. Unfortunately, some EOs were still double counted (section 2.2.1), so they were removed in a post-processing step. EOV4 contains our most complete CloudSat data set, June 2006 to October 2008. EOV3 and EOV4 only differ in that EOV3 has some double counted clouds near the equator, so for figures made with no interest in the equator region EOV3 is sometimes used.
Chapter 3: Global CloudSat Climatology

Here we survey the global dataset, to confirm validity of our techniques (comparing to published results) and to explore facets of global cloudiness fields for which our EO-based treatment is uniquely well suited.

3.1 Defining Echo Object Types

We define EO types according to top and base height attributes (histogram type 6, Table 2.4). We use different definitions with latitude to maintain consistency with the lowering of tropopause heights from the tropics to higher latitudes.

The CloudSat science team offers a cloud classification product, which identifies different cloud types. K. Sassen and Z. Wang (available at http://cloudsat.cira.colostate.edu/data1CDlist.php?go=listpath=/2B-CLDCLASS) describe the CloudSat cloud type algorithm in detail. Briefly, their algorithm has two main steps: (1) to perform a cloud clustering analysis to identify extended cloud layers and (2) to classify a cloud cluster into a cloud type. A cloud cluster is a horizontally continuous cloud layer (i.e. bins with a cloud mask ≥ 20 that are not separated by more than 500 m). Cloud types are determined based on cloud cluster height, maximum reflectivity, temperature at the height of the maximum reflectivity, as well as the presence of precipitation apparently reaching the surface.

We use our own EO-based definition of cloud types to utilize all the unique attributes we keep track of for each EO (table 2.1). Our EO-based definition of cloud
types is compared with Sassen and Wang (2008) who used the CloudSat cloud classification algorithm in section 3.2.

### 3.1.1 Tropics

Figure 1a shows a joint histogram of EO base height versus EO top height for the tropical belt (latitude < 20°). The EO count histogram is weighted by the horizontal span of each EO, to form a true cloud-cover density. Low-based EOs occur along the vertical axis, while thin layer EO types lie along the diagonal. A minimum of cloudiness with ~ 3 km EO bases (c.f. Zuidema 1998) separates these distinct tower (or precipitating) vs. thin layer type EO populations.

EO types are defined by partitioning figure 1a into boxes A-F. The horizontal lines at 4.5 km (approximately the tropical 0°C level) and 10 km (approximately the height of -40°C) separate warm (liquid water), potentially mixed-phase, and fully glaciated EO tops. Along with temperature relevance, these lines also tend to delineate natural types by running through minima in the joint histogram. For middle and high topped EOs, the vertical line (base = 3 km) separates geometrically thick and thin EOs. High-topped EOs are further divided into thick and thin by a base = 7 km line. We may roughly associate each box of Fig. 1a to a familiar cloud type name as follows: (A) deep precipitating, (B) detached anvil (or thick cirrus) (C) cirrus, (D) cumulus congestus, and (E) altostratus and altocumulus. Low EOs (F) are divided by width into stratocumulus (width > 10 pixels) vs. cumulus (width < 10 pixels). As a reminder a pixel is approximately 1.1 km across.
Low EOs (type F) represent the most echo cover. High-topped thick EOs (box A) also have high coverage due to their large areal coverage when present (an entire echo mass is considered low-based if it precipitates to the surface anywhere.) Thin cirrus are also common.

By retrieving pixels from each EO, we can also construct the corresponding distribution of the base and top height of each CloudSat profile (Fig. 1b). Panel b is smoother, as there are more columns than EOs, but Figs. 1a and 1b have similar features: maximum echo-cover in the lowest EOs, a minimum in echo-cover at 10 km top height and 3 km base height, and a bimodal distribution in box E. Thin cloud layers are more meaningfully separated from tower-like clouds, and better resolved (density is closer to the diagonal) in panel b, since in panel a the entire echo area of an EO is assigned to the EO’s uppermost height bin.

Summing the joint histogram over all base heights in figure 1b gives the line plot of echo-top density on the right side of Fig. 1b. Its units are %, and its interpretation (for the peak value at 2 km) is that about 1.5% of the tropical belt is covered by cloudy echo whose top lies within a 240 m interval of 2 km. The distribution has two dominant peaks of about equal magnitude at low and high levels, and two smaller mid-level peaks, rendering a curious quad-modal structure. This quad-modal structure is studied in detail in chapter 4.
Figure 3.1 – Horizontal cloud cover in the tropics (20°S – 20°N) accounted for by clouds with tops and bases in the indicated bins for 16 June 2006 – 31 May 2008. (a) distributes cloudiness by EO bases and tops, while (b) distributes it according to local (echo profile) bases and tops. Contour values (labeled) are in units of $10^3$ horizontal echo pixels per bin. Lines and letters delineate EO types, roughly associated to cloud types: (A) deep precipitation (dp), (B) anvil (an, thick cirrus), (C) cirrus (ci), (D) cumulus congestus (cg), (E) alto cumulus/alto stratus (ac), (F) cumulus (cu) and stratocumulus (sc). The 2D distribution of panel b is integrated at right to show the 1D distribution by local echo top height. Bin size is 240 m by 240 m.

3.1.2 Sub-tropics

Figure 3.2 is the same as figure 3.1 except for EOs occurring in the sub-tropics (20° - 35°N/S). As in the tropics low EOs (box F) represent the most echo coverage and there is a minimum in echo coverage at base heights between 2 km – 4 km, in both panels a and b. Panel b shows more echo cover in boxes C and E and less in boxes A and D compared to panel a. As in figure 3.1 differences are due to the fact that all of an EOs area is assigned to the top most bin of the EO in panel a.

Unlike the tropics the line plot of echo-top density (right side panel b) is bimodal. Peaks occur at 2 km and 11 km with a minimum around 3.5 km. In the sub-
tropics the lower peak dominates over the upper peak, consistent with more suppressed conditions from the descending branch of the Hadley circulation.

Figure 3.2 – As in figure 3.1 except for the sub-tropics (20° - 35°S/N).

3.1.3 Mid-latitudes

Figure 3.3 is the same as the previous two figures except with slight definition adjustments for the mid-latitudes (35° - 60° N/S). The lines separating low, middle, and high-topped EOs are now at 3 km and 7 km, respectively. The line separating thick and thin middle and high-topped EOs is at 2 km with the additional thickness line for high-topped EOs now at 5 km. Thickness adjustments (vertical lines at base = 2, 5 km) are done to maintain approximate separation between maxima and minima in the joint distribution.

Panel a is dominated equally by low EOs (box F) and high, thick EOs (box A), with thinner cirrus EOs also being common. The line plot of profile top density on the right side of panel b is bi-modal. The dominant peak occurs around 1.5 km with a broad upper-level peak between 9 km – 10 km.
3.1.4 High Latitudes

Figure 3.4 is similar to the previous three figures except now for the high latitudes (latitude > 60°). Again, adjustments are made to height criteria separating low, middle, and high-topped EOs to be in accord with the lowering of the tropopause. Low, middle, and high-topped EOs are separated at 3 km and 6 km, respectively. The vertical line, base height = 2 km, is the same as in the mid-latitudes, but is now the only line delineating geometrically thick (deep precipitating) and thin (cirrus) EOs for the high topped EOs. Two boxes describing high-topped EOs in the high latitudes seemed sufficient, as there are no sharp changes in the distribution going from top heights above 6 km and base heights greater than 2 km like in the other latitudinal belts.

Both the EO and profile base and top height distribution have a broad maximum in echo cover for low-based EOs over all top heights. This broad maximum makes the difference between the magnitude of the lower and upper peak in the line plot of profile top density (right side Fig. 3.4b) small.
Figure 3.4 – As in figure 3.1 except for the high latitudes (latitude > 60°). The anvil cloud type has been removed.

3.2 Geographical Maps of Echo Cover

Having established different cloud types for each latitude belt this subsection answers the first question raised by this thesis: (1) Given a certain cloud type, where are those clouds occurring. Echo cover is computed by dividing the number of cloudy samples per bin by the total number (both cloudy and clear) of CloudSat samples per bin.

3.2.1 Daytime and Nighttime Cover for 60°S – 60°N

Using the EO type definitions of each latitude belt from figures 3.1 – 3.3, geographical maps of each EO type for both day and nighttime CloudSat overpass times are shown in figure 3.5, for two years of CloudSat data (16 June 2006 – 31 May 2008) from 60°S – 60°N. It should be noted that in panels d (cirrus cover) and f (altocumulus cover) there are sharp discontinuities in echo cover at 35°N/S caused by definition adjustments going from the subtropics to the mid-latitudes.
Figure 3.5a shows total echo cover for the two CloudSat years. Maximum values in echo cover occur in a nearly unbroken belt in the southern hemisphere mid-latitude storm tracks. Maximum values also occur over ocean poleward of 40°N. An ocean dominant southern hemisphere provides a constant source of moisture for cloud formation, therefore explaining the high echo cover values in the southern hemisphere storm tracks compared to the northern hemisphere storm tracks. Secondary maximum values (~70%) occur over the west Pacific warm pool and inter-tropical convergence zone (ITCZ). Echo cover values over northern mid-latitude land are around 50% with little differences between North America, Europe, and Asia. Tropical land echo cover is largest over the South American rainforests and in central Africa. Minima in echo cover occur over the global deserts and sub-tropical regions. A couple noteworthy features in the echo cover distribution are the relative minimum in echo cover over the central United States near north Texas and Oklahoma and the high echo cover values in the tropics over Papua New Guinea compared to the rest of the maritime continent.

Deep precipitating (Dp) EOs are primarily found along the ITCZ, over the west Pacific warmpool, and in the mid-latitude storm tracks (Fig. 3.5b). In the east Pacific and Atlantic Ocean the ITCZ is climatologically north of the equator in a narrow band as a consequence of the shape of the central and Southern American coastline (Philander et al. 1996). Dp are also found over the Amazon rainforests in South America and the Congo basin in Africa. Coverage values can be high due to the large horizontal span of some Dp. Our Dp coverage values are about three times greater than cloud fraction values of deep clouds defined using the CloudSat cloud classification algorithm (Sassen and Wang 2008), since in their definition, EO’s are categorized into several cloud types. For
example, a well developed deep precipitating cloud with anvil outflow and attached cirrus, will be classified by them as consisting of different cloud types, but all as Dp by the present method. Nevertheless, both definitions have deep precipitating echo in the same locations.

Anvil (or thick cirrus) echo cover in figure 3.5c is rather uniform throughout the mid-latitudes and ITCZ region with values between 5% - 10%. Maximum values occur over the Tibetan Plateau and North American Rockies as a result of middle-based clouds, relative to these high elevations, being classified as high-level clouds by our echo type definitions.

The pattern of cirrus (ci) echo cover is strikingly similar to that of deep precipitating EOs. Cirrus cover is broader along the ITCZ than deep precipitation (cf 3.5d, 3.5b). The high spatial correlation between cirrus and deep precipitation is due to the fact that convective detrainment and synoptic scale ascent associated with mid-latitude frontal systems are sources of cirrus formation (Mace et al. 2006, Starr and Wylie 1990, Sassen 2002). Luo and Rossow (2004) estimated that 44% of tropical cirrus are directly detrained from convection, while 56% form in-situ.

Cumulus congestus (cg) cover (Fig. 3.5e) is prominent along the mid-latitude storm tracks and the ITCZ. Cg tends to be more prevalent in the east Pacific ITCZ region compared to the Indian Ocean and west Pacific warm pool. Cg cover is greater over northern South America than central Africa. Papua New Guinea has more cg cover than the surrounding warmpool and other maritime continent islands. This local Papua New Guinea maximum was first noted by Mace et al. (2007).
Figure 3.5f shows the echo cover for mid-level thin clouds, collectively called altocumulus (ac). Similar to figure 3.5c, maximum coverage is evident over the Tibetan Plateau, the North American Rockies, and the South American Andes. All low clouds, relative to these high elevations, are being classified as mid-level clouds by our echo type definitions. Besides these topography related hot spots, ac cover is generally small (cover < 10%) and uniform across both land and ocean for the entire globe, expect for the oceanic tropical belt between 0° and 15°S.

Stratocumulus (sc) coverage is prominent in the sub-tropics over the cool upwelling regions of the eastern ocean basins (Fig. 3.5g). Values peak between 35% - 50% for sc found west of California, Peru, Namibia, and Australia. Other areas of sc are off the west coast of north Africa around the Azores, and over the north west Pacific ocean. More details on sc are given in chapter 5.

Cumulus (cu) cover is present over most of the tropics and mid-latitudes with the exception of deserts. Oceanic cover is around 5% with smaller land coverage. Small cover values are due to the small size of cu. The ubiquity of oceanic cumulus stems from the persistent oceanic trade inversion, which limit shallow convection to the lowest several kilometers.
Figure 3.5 – Echo cover for (a) all EOs from 16 June 2006 – 31 May 2008 from 60°S – 60°N, (b-h) for each indicated cloud type defined using figures 3.1 – 3.3. Bin size is 3.0° longitude by 3.6° latitude. Echo cover values over 100% can occur due to the fact that an entire EO's area is placed in the mean latitude and longitude bin of the EO, even if the EO spans several bin. Also because there can be more than one cloud above a given location on Earth.

3.2.2 Differences in Daytime and Nighttime Echo Cover for 60°S – 60°N

Figure 3.6 shows echo cover for daytime CloudSat overpasses, while figure 3.7 is for nighttime. For most EO types day and night differences are subtle.

For total echo cover (Figs. 3.6a and 3.7a) the mid-latitudes show little difference between day and night. In the tropics, though, nighttime cover is increased over the eastern Pacific Ocean and Atlantic Ocean near the South American and African coast, respectively. There is also a hint of increased nighttime cover over South America and
central Africa. Subsequent panels in figures 3.6 and 3.7 for individual EO types reveal
that increased se echo cover at night (cf. 3.6g and 3.7g) accounts for the total echo cover
increases at night in the eastern ocean basins. Increased South American and central
Africa nighttime cover appears to result from increases in ci nighttime cover (cf. 3.6d and
3.7d). Sassen et al. (2008) noted a similar increase in nighttime cirrus cover. They
postulated that deep convection may be more prevalent at night resulting in more cirrus
(which is consistent with a slight increase in dp cover over South America and Africa,
Figs. 3.6b and 3.7b) or that daytime anvil created from afternoon convection thins,
becoming nighttime cirrus. We see no increase in daytime anvil relative to nighttime
(Fig. 3.6c and 3.7c). However, CloudSat overpass times miss the peak amplitudes in the
diurnal cycle of tropical convection. Deep precipitation coverage and rainfall over land
peak in mid-afternoon around 1500 LT (Nesbitt and Zipser 2003), so maximum cover in
anvil from convection would be expected afterwards. So, it remains likely that increased
nighttime cirrus is a result of afternoon convection.

Overall differences in cg, ac, and cu cover from day to night are slight. A few
noticeable exceptions in cg are the increase in daytime cover over Papua New Guinea and
Malaysia, an eastward extension of cg during the day along the northern border of South
America, and an increase in nighttime cover near Lake Victoria in Africa (Figs. 3.6e and
3.7e). Ac cover differences are only apparent over the North American Rockies and
Tibetan Plateau, with day having more cover than night (Figs. 3.6f and 3.7f). Cu cover
increases slightly during nighttime overpasses in a narrow band from 10°N – 15°N in the
west and central Pacific Ocean. Contrary to these nighttime increases, daytime increases
extend westward off the southern tip of South America (Figs. 3.6h and 3.7h).
Figure 3.6 – As in figure 3.5 except for daytime (1330 LT) CloudSat passes only.

Figure 3.7 – As in figure 3.5 except for nighttime (0130 LT) CloudSat passes only.
3.2.3 Annual Differences in Echo Cover for 60°S – 60°N

The time period CloudSat has been in orbit offers a unique opportunity to view cloud climatology during El Nino and La Nina like conditions. From June 2006 – February 2007 conditions were El Nino like, becoming neutral afterwards until about July 2007 when conditions became La Nina like (Fig. 3.8, courtesy of Rob Burgman). These different periods match perfectly with our first year (16 June 2006 – 31 May 2007) and second year (1 June 2007 – 31 May 2008) CloudSat dataset.

![NINO3.4 SST](image)

**Figure 3.8** – Nino 3.4 sea surface temperature (SST) anomalies from Smith and Reynolds extended reconstructed SST data for 1990 – 2008. Courtesy of Rob Burgman.

Figure 3.9 shows the percent echo cover differences from the second year minus the first year of CloudSat for all EO types (panel a) and for each EO type (panels b – h). (Land changes have been masked out because there were no obvious patterns of echo cover changes from the first to second year and it makes ocean differences more comprehensible.) The most striking difference is the decrease in echo cover from about 10°S – 10°N over the central Pacific with a horseshoe shape increase in echo cover around it (i.e. increase in echo cover in the Indian and west Pacific and from about 10° - 30°N/S in the central Pacific). There are also slight increases in echo cover over the eastern ocean basins during the second year.
The second year minus the first year of CloudSat in dp and ci cover (Figs. 3.9b and 3.9d) over the central and west Pacific mimic the overall pattern of changes in echo cover. Together these two echo types account for the percent echo cover changes in figure 3.9a. Changes in dp and ci are consistent with expectations that during La Nina (El Nino) conditions convection is enhanced in the west (central) Pacific relative to neutral conditions. Increases in sc cover during the second year account for the increases in echo cover over the eastern ocean basins. There is especially a noticeable increase in sc cover in a narrow band just above the equator in the eastern and central Pacific. More details on changes in sc cover are given in chapter 5.

Over the global oceans cu cover tends to show small increases during the second year (Fig. 3.9h). Changes in an, cg, and ac (Figs. 3.9c, e, and f) tend to have random patterns of increases and decreases in echo cover.

Figure 3.10 shows the zonal mean changes in echo cover during the second year of CloudSat data (1 June 2007 – 31 May 2008) compared to the first year of CloudSat data (16 June 2006 – 31 May 2007) for total echo cover (panel a) and each EO type (panels b-h). Light smoothing has been applied to the plots. Plots in figure 3.10 include echo occurring over both land and ocean. Panel a shows that over all latitudes except near the equator and between 25°N and 30°N there is an increase in echo cover during the second CloudSat year. The largest increases (4%) in total echo cover occurs in the northern hemisphere tropics between 10°N and 15°N. The pattern of deep precipitation changes (panel b) is similar to the overall echo cover pattern: decreases (-1%) occur near the equator, with the largest increases (2%) between 10°N and 15°N during the second CloudSat year. Deep precipitation echo cover also shows decreases (-1%) near 55°N.
Zonal mean changes in anvil (panel c) cover are small (< 0.5%) compared to overall changes in panel (a) during the second CloudSat year and are generally positive through the tropics and subtropics (latitude < 35°) with decreases in echo cover from 35°N/S – 55°N/S and back to increased echo cover from 55°N/S – 60°N/S. Zonal mean changes in cirrus (panel d) cover during the second CloudSat year are also small (< 0.5%). Cirrus echo cover changes have a wave-like pattern: increases poleward of ~50°, decreases between ~30° - 50°, increases between ~15° - 30°, and decreases equatorward of 15°. Zonal mean changes in cumulus congestus (panel e) cover are less than 1% from the first CloudSat year to the second. The largest increases (decrease) in echo cover in the second CloudSat year are between 5°N – 15°N (35°N – 45°N). Zonal mean changes in altocumulus (panel f) and cumulus (panel h) cover are positive over the entire globe (except at 40°S – 45°S for altocumulus) during the second CloudSat year. Lastly, stratocumulus (panel g) shows increases in echo cover during the second CloudSat year everywhere except 15°N – 35°N. Increases in stratocumulus cover are greatest in the southern hemisphere mid-latitudes (35°S – 60°S). All the above changes are fairly consistent with what the 2D maps of echo cover changes show in figure 3.9.

Figure 3.11 shows a zonal cross section along the equator (10°S – 10°N) of changes in echo cover during the second CloudSat year compared to the first CloudSat year. Again, light smoothing has been used to alleviate noise. Echo cover increases are greatest in the Indian Ocean and west Pacific (~70°E – 140°E), while the largest decreases occur in the central and east Pacific (~150°E – 180°E). These changes are consistent with warm ocean waters shifting westward into the west Pacific and a stronger, more westward extended cold tongue during La Nina-like conditions.
Figure 3.9 – Difference from year 2 (1 June 2007 – 31 May 2008) and year 1 (16 June 2006 – 31 May 2007) in echo cover for (a) all EOs and (b-h) each indicated cloud type defined in figures 3.1 – 3.3. Land is masked out for better interpretation. Zero difference line is contoured in black. Light smoothing has been applied for ease of viewing.
Figure 3.10 – Zonal mean differences from year 2 (1 June 2007 – 31 May 2008) and year 1 (16 June 2006 – 31 May 2007) in echo cover for (a) all EOs and (b-h) each indicated cloud type defined in figures 3.1 – 3.3. Horizontal line represents zero difference. Light smoothing has been applied for ease of viewing.

Figure 3.11 – Zonal cross section from 10°S – 10°N of echo cover difference from year 2 (1 June 2007 – 31 May 2008) and year 1 (16 June 2006 – 31 May 2007). Horizontal line represents 0 difference. Light smoothing has been applied for ease of viewing.
3.2.4 Daytime and Nighttime Cover for High Latitudes

3.2.4.1 Arctic Echo Cover

Figure 3.12 show total echo cover (panel a) and echo cover for each EO type (panels b-g) in the Arctic region (latitude > 60°N). Total echo cover (panel a) is highest (near 100%) in the North Atlantic. Minimum (~30%) cover occurs over the interior of Greenland. Values elsewhere vary between 50% and 75%. These results are consistent with CloudSat result from Sassen and Wang (2008).

Dp EO cover (panel b) is minimum (near 0%) over the interior of Greenland. Small coverage values are also found in the Arctic Ocean north of North America. Coverage over the North Atlantic, northern Europe and Russia, and the Arctic Ocean north of Europe and Russia varies mostly between 30% and 50%. Small isolated maximas in cover are found along the southeastern coast of Greenland and over Russia around 90°E and 65°N. These small maximas are also seen in Sassen and Wang (2008) for their nimbostratus cloud classification cover. Our dp echo type is likely including their nimbostratus and deep cloud types, since both are classifications for precipitating clouds.

Ci EO cover (panel c) is greater over high latitude land vs. ocean, with an exception over the North Atlantic where values are similar to land. Maximum values (~40%) in ci cover occur over the interior of Greenland.

Mid-level thick (cg, panel d) and thin (ac, panel e) EOs have opposite echo cover patterns. Ac cover peaks (~ 7% to ~ 12%) over Greenland, with small values (< 3%) elsewhere, while cg has a minimum over Greenland (0%) with larger values (between ~
7% and 20%) elsewhere. There is more cg cover over the North Atlantic than over the Arctic Ocean north of North America. The ac peak over interior Greenland is likely surface-associated EOs relative to the high elevation of interior Greenland. It is difficult to compare our mid-level results with those of Sassen and Wang (2008) because they do not have a distinct classification for thick middle-topped clouds.

Low-topped echo cover (sc, panel f and cu, panel g) overall is greater over high latitude ocean vs. land, consistent with Sassen and Wang (2008). The exception is relative maxima in cu cover over Alaska and northwestern Canada and Russia around 135°E and 60°N – 65°N. Sc maximum cover (between 15% - 30%) occurs over the North Atlantic, consistent with Sassen and Wang (2008). Cu maximum cover (near 10%) occurs along the west and east coast of Greenland and the southeastern tip of Baffin Island.
Figure 3.12 – Echo cover for (a) all EOs from 16 June 2006 – 31 May 2008 with latitude > 60°N, (b-h) for each indicated cloud type defined using figure 3.4. Bin size is 3.0° longitude by 3.6° latitude.
3.2.4.2 Antarctic Echo Cover

Figure 3.13 shows total echo cover (panel a) and cover per EO type (panels b-g) over Antarctica (latitude < 60°S). Total echo cover (panel a) is highest (values > 75%) over the Southern Ocean and lowest (values < 25%) over land in east Antarctica, consistent with Sassen and Wang (2008).

High-topped thick (dp, panel b) and thin (ci, panel c) EOs have opposite echo cover patterns. Dp has more cover over the southern ocean and coast of Antarctica, while ci has more over the interior of Antarctica. Dp values are fairly constant around the Southern Ocean (between 40% and 60%). Maximum ci values (values > ~ 45%) occur over land in west Antarctica, while minimum values (values < ~ 15%) occur over the Antarctic Peninsula. Over the Southern Ocean ci values are generally less than ~ 10%.

Like the high-topped EOs, thick and thin middle-topped EO types (cg, panel d and ac, panel e) have opposite echo cover patterns. Cg (ac) is more prominent over ocean (land). Cg values generally range from ~ 10% to 20% over the Southern Ocean. Ac cover is greater over west Antarctica than east.

Low-topped EO cover (sc, panel f and cu, panel g) occurs over the Southern Ocean and along the Antarctica coast. High elevation in the interior of Antarctica keeps EO tops above the 3 km low EO threshold. Sc cover is highest (between 15% and 30%) over the Southern Ocean between the Antarctic Peninsula and the Ross Sea. Cu cover is fairly uniform over the Southern Ocean with values ~ 3%. The maximum cu cover occurs along the Transantarctic Mountains. There is a relative cu cover maximum over the Antarctic Peninsula.
Figure 3.13 – Same as figure 3.9 except for latitudes < 60°S.
3.3 Seasonal Cycle of Echo Objects

Figure 3.14 shows the echo cover distribution by time (Julian days) vs. mean latitude of each EO in our complete data set, 16 June 2006 – 31 October 2008. Notable features include the sharp difference in the seasonal movement of northern and southern hemisphere mid-latitude echo cover, the movement of the ITCZ with time, and the persistent minimum in echo cover in the sub-tropics. The southern hemisphere is cloudier than the northern hemisphere during all times, except for the highest latitudes (latitude > 75°). Echo cover in the northern hemisphere extratropics (latitude > 30°N) moves in a wave-like pattern over time: echo cover extends equatorward during the boreal fall and winter months, then retreats poleward in the spring and summer in agreement with the fluctuations in strength and movement of the polar and sub-tropical jet streams. In contrast to the northern hemisphere, the southern hemisphere extratropics (latitude < 30°S) echo cover meanders little through the seasons, perhaps a slight equatorward (poleward) extension during the austral fall (spring) and winter (summer). While the southern hemisphere polar and sub-tropical jet vary seasonally, the mostly maritime southern hemisphere keeps cloudiness persistent year round.

Changes in the ITCZ are consistent with seasonal shifts in the Hadley circulation. Echo cover in the ITZC is greatest between ~ 5°N and 10°N and extends northward during boreal summer months. During boreal winter and spring months the ITCZ echo cover lessens and shifts southward. There is a hint of a secondary ITCZ during the end of February beginning of March around 10°S (Zhang 2001).
Figure 3.14 – EO cover distribution by time (Julian days) vs. mean latitude of each EO in our complete data set, 16 June 2006 – 31 October 2008. EO count is weighted by the horizontal span of each EO. Units are $10^4$ horizontal pixels per bin. Bin size is 10 days by $3.6\degree$ latitude. Black bars indicate missing 2B-GEOPROF data.

Figure 3.15 perhaps shows a clearer picture of the seasonal changes in echo cover associated with changes in the Hadley cell. Here, mean latitude vs. top height of all EOs is shown for each three-month season in our data set (Table 2.3). Seasons are June, July, August (JJA); September, October, November (SON); December, January, February (DJF); and March, April, May (MAM) for specified years.

The tropics have a persistent column of echo cover associated with the ascending branch of the Hadley circulation. The top of the column (tops $> 10$ km) shifts to the summertime hemisphere, while the bottom portion (tops $< 10$ km) remains north of the equator at about $5\degree$N through all seasons. Mid-level (between 5 km and 8 km) echo cover extends from the column into the summertime hemisphere sub-tropics (compare panels a, c, e, g). During spring and fall months (panels, b, d, f, and h) the mid-level echo cover in the sub-tropics disappears. Besides these summertime extensions of mid-level
echo cover into the sub-tropics, the sub-tropics have the least echo cover, associated with the descending branch of the Hadley circulation. Echo cover values are smallest for the winter hemisphere sub-tropics extending from ~ 3 km to the tropopause.

In the northern hemisphere mid and high-latitudes mid-level echo cover is greater during wintertime (compare panels a, c, e, and g). This wintertime enhancement is especially great between summer 2006 (panel a) and winter 2006/2007 (panel c) where even low-level echo cover is enhanced during winter (compare panels a and c). Like in figure 3.14 enhanced wintertime echo cover in the northern hemisphere is associated with a stronger jet and frontal systems. Seasonal changes in southern hemisphere mid and high-latitude echo cover are not as noticeable at the northern hemisphere.

Low-level (tops < 3 km) echo cover is greater in the southern hemisphere vs. northern hemisphere for all seasons. The low-topped EO hemispheric contrast is especially great during JJA 2006 (panel a). JJA 2007 (panel e) has more low-level echo cover than JJA 2006 (panel a) for both hemispheres. Low-level echo cover extends to the equator in the southern hemisphere during SON (panels b and f). The extension is greater in SON 2007 vs. 2006. Changes in low-level echo cover are largely associated with the sc echo type and are discussed further in chapter 5.

Besides low-level echo cover changes in the southern hemisphere, the fall and spring months show little variation (compare panels b, d, f, and h). The tropical echo cover column is situated over the equator. An interesting feature is the low-level maximum over the Arctic during the spring and fall months, which vanishes during the summer and winter months (compare right and left panels). Reduced boreal summertime cloudiness in 2007 in the Arctic is related to anomalously high sea-level pressure over the
western Arctic. The reduced summertime 2007 cloudiness is not unprecedented, but is anomalous for the recent past (62 year record) (Kay et al. 2008).

Figure 3.15 – EO cover distribution by mean latitude vs. top height for each EO for each season (June, July, August (JJA); September, October, November (SON); December, January, February (DJF); March, April, May (MAM)) and year (2006, 06; 2007, 07) indicated. EO count is weighted by the horizontal span of each EO. Units are $10^3$ horizontal pixels per bin. Bin size is $3.6^\circ$ latitude by 240 m.
3.4 Echo Object Width Characteristics

Figure 3.16 shows the distribution of EO width vs. top height for 16 June 2006 – 31 May 2008. The distribution is weighted by the total number of pixels (volume) per EO to emphasize larger EOs. Most EOs are low-topped (tops < 4 km) with widths less than ~50 km, though, some low-topped EOs can get as wide as 700 km. EOs with tops near the mid-latitude tropopause are the most horizontally extensive, some wider than 2000 km. Such wide clouds are likely associated with deep precipitation associated with frontal systems. EOs with tops greater than 13 km have widths as great as 1500 km. These tall and wide EOs are likely associated with deep precipitation in the tropical latitudes.

Figure 3.16 – Width vs. top height of each EO for 16 June 2006 – 31 May 2008. Distribution is weighted by total number of pixels (volume) per EO to emphasize larger EOs. Bin size is 10 pixels (~km) by 240 m. Width is the horizontal span of the EO.

3.5 Tropical Examples of Pixel Statistics in dBZ-Height Space

2D histograms of pixel dBZ vs. height are called contoured frequency by altitude diagrams (CFADs, Yuter and Houze 1995). Figure 3.17a is the normalized CFAD (NCFAD) for the tropics (20°S – 20°N) for 16 June 2006 – 31 May 2008. The most pixels occur at -25 dBZ between 12 km and 13 km. Pixels are also common at low-levels
over all reflectivities and above the rainfall attenuation line (white line in Fig. 3.17a). A broad minimum in the NCFAD occurs at reflectivities below 0 dBZ in the mid-levels (between 4 km and 8 km). Rescaling the NCFAD to total number of pixels per bin, then integrating over all reflectivity values gives the vertical profile of tropical echo cover (Fig. 3.15b). Echo cover is the ratio of cloudy echo pixels sampled by the CPR to the total number of pixels (both clear and cloudy) sampled by the CPR. There are two main peaks in echo cover, one at ~ 1.5 km the other at ~ 11.5 km and a small mid-level peak around 5 km with a hint of a second mid-level peak around 8 km. The interpretation for the upper-level peak is that a little over 14% of the tropics is covered by echo at ~ 11.5 km. The low-level peak is slightly greater, though narrower, than the upper-level peak. The 1.5 km, 5 km, and 11.5 km peaks are associated with known levels of increased static stability in the tropics: the trade inversion, the 0° melting isotherm, and the tropopause (Johnson et al. 1999).

**Figure 3.17** – (a) Normalized tropical (20°S – 20°N) CFAD for 16 June 2006 – 31 May 2008. Units are are 10^{-2} percent per bin. Bin size is 1 dBZ by 240 m. (b) Integral over (a) over all dBZ values (rescaled to total number of pixels per height bin ) to give echo coverage. Below white line is heavy rainfall attenuation.
Figures 3.18 and 3.19 are the same as 3.17 except for the pixels comprising only ocean EOs and land EOs, respectively. The ocean (land) CFAD is normalized relative to the total number of pixels occurring only over ocean (land). The NCFAD (Fig. 3.18a) has two peaks at -25 dBZ, one below 3 km, the other between 11 km and 12 km. The minimum in the NCFAD occurs at reflectivities between -5 dBZ and -20 dBZ in the mid-levels (between 4 km and 8 km). The line plot of total echo cover over the tropical ocean again shows an upper-level and low-level peak with a small mid-level peak at 5 km. The lower level peak dominates over the upper level peak, with over 17% echo cover vs. about 11% echo cover, respectively.

**Figure 3.18** – Same as 3.15 except for EO occurring only over ocean.

**Figure 3.19** – Same as 3.15 except for EO occurring only over land.
The land only NCFAD (Fig. 3.19a) peaks at -25 dBZ between 11 km and 12 km, like the ocean only and total tropical NCFADs. However, the low-level distribution of pixels with reflectivity less than 0 dBZ is much less than the ocean and total tropical NCFADs and really just becomes an extension of the mid-level minimum. The distribution is greater over land than ocean for reflectivities greater than ~ 5 dBZ below 5 km. This suggests there is perhaps heavy rainfall and therefore more attenuation over land than ocean. Overall, land is cloudier than ocean. Over land echo cover is highest at upper-levels (between 11 km and 12 km), followed by the peak around 5 km, and finally low-level echo cover.

Why are there such stark differences in echo cover over land and ocean? Differences in echo cover for each EO type may underlie the answer. Figure 3.20 shows tropical echo cover for each EO type for land (solid line) and ocean (dashed line). High-topped EOs (dp, an, and ci, panels a, b, and c, respectively) have move echo cover over land than ocean. Dp echo cover is especially more over land than over ocean at high levels ($z \geq 8$ km). Middle-topped EOs (cg and ac, panels d and e, respectively) have more echo cover over land than ocean expect for the lowest levels ($z < 1.5$ km) of cg cover. Low-topped EOs (sc and cu, panels f and g, respectively) have more echo cover over ocean than land. So, decreased echo cover at low levels over land ($z < 3$ km) is largely a result of differences in sc cover over land and ocean, with small contributions to changes in cg and cu over land and ocean. Figure 3.20 results are consistent with figure 3.5, which shows that low-topped echo types are more prominent over ocean. Again in figure 3.5 we see little sc and cu (panels g and h) echo cover over land in the tropics, yet large areal cover of dp and ci (panels b and d) over land.
Figure 3.20 – Tropical echo cover associated with each EO type (labeled in panels a-g) over land (solid line) only and ocean (dashed line) only for 16 June 2006 – 31 May 2008.

3.5.1 NCFAD Enhancements for Tropical EO Types

Figure 3.21 shows the positive NCFAD enhancements associated with each EO type relative to the tropical NCFAD. The background contours are the tropical NCFAD, while the color shading represents the positive enhancements, defined as positive areas of individual EO type NCFADs minus mean tropical NCFAD. By looking at positive enhancements we can emphasize the characteristic contributions of each EO type to the total tropical NCFAD.

Deep precipitation EO type (panel a) enhancements are greatest at high reflectivities between 3 km and 5 km, suggesting heavy rainfall. Deep precipitation
enhancements extend rearward (i.e. from high to low reflectivities) and upward from 4 km and 15 dBZ to 15 km and -30 dBZ, suggesting anvil and cirrus attached to the convective core.

Anvil EO type enhancements (panel b) are wedge shaped between 7 km and 13 km from -30 dBZ to +5 dBZ. The greatest enhancements occur between -15 dBZ and -5 dBZ.

Cirrus EO type enhancements (panel c) have a similar wedge shaped enhancements to anvil, except the wedge has moved upward and rearward, fitting almost perfectly into the tropical NCFAD contours at high altitudes and low reflectivities. Enhancements are greatest at -25 dBZ and 13 km.

Cumulus congestus EO type enhancements are shown in panel d. By definition top height for cumulus congestus EO type may go up to 10 km, however enhancements only reach around 7 km. Enhancements occur over all reflectivity values, but are greatest at -25 dBZ between 3 km and 5 km. Enhancements at high reflectivities (dBZ > 0) to the surface suggest that precipitation reaching the surface is prominent in this EO type.

Alto cumulus EO type enhancements are shown in panel e. Enhancements occur at reflectivities less than 0 dBZ, between 5 km and 8 km. Low reflectivity enhancements suggest alto cumulus is typically a non-precipitating EO type.

Stratocumulus EO type enhancements (panel f) occur at low-levels (height < 3 km, by definition of this EO type) and span all reflectivities. Maximum enhancements are at -25 dBZ. Enhancements at higher reflectivities (dBZ > -17) suggest that some the stratocumulus EOs are capable of drizzle and/or rain. We consider reflectivities greater than -17 dBZ as drizzle (Frisch et al. 1995).
Cumulus EO type enhancements (panel g) are confined to the lowest-levels (height < 2 km) and low reflectivities (dBZ < 0). Low reflectivity enhancements suggest cumulus is generally a non-precipitating EO type. It is interesting the difference in the stratocumulus and cumulus EO type enhancements when the only criterion separating the two types is width (width ≤ 10 pixels is cumulus, width > 10 pixels is stratocumulus). The fact that the stratocumulus type has some rain and/or drizzle, suggests that some level of organization (width) is necessary to activate precipitation.

Figure 3.21 – Shading is CFAD positive enhancements associated with each indicated cloud type relative to the normalized tropical CFAD (Fig. 3.14a). Black contours represent the normalized tropical CFAD. Note that color scale changes between panels.
3.5.2 Tropical Daytime and Nighttime CFAD Differences

Daytime (1330 LT) and nighttime (0130 LT) differences over the tropical ocean for 16 June 2006 – 31 May 2008 are examined in 3.22a/b. Panel a shows greater daytime echo starting at 3 km and moving rearward and upward over all reflectivities. Low-level echo (height < 3 km) is greater at nighttime at reflectivities less than ~ 10 dBZ. There is also more nighttime echo in a small wedge between 12 km and 8 km and reflectivities between -15 dBZ and -25 dBZ. A figure similar to panel a has already been published in Zuidema and Mapes (2008, their Fig. 5), but only included the first year of CloudSat (July 2006 – June 2007).

![Figure 3.22](image)

**Figure 3.22** – CFAD differences for day (1330) and night (0130) over the tropical (a) ocean only and (b) land only. Values are normalized percentages of the maximum difference. Positive (negative) differences are solid (dashed) contours. (c) CloudSat tropical day and night cloud cover over ocean. (d) same as (c) except over land.

Profiles of day and nighttime echo cover over the tropical ocean are shown in panel b. Echo cover is higher during the day above 9 km by about 500 m and also
slightly greater between 11 km and 12 km. Below 9 km nighttime echo cover is greater than daytime. Similar results to panel b have been published in Liu et al. (2008). Both Zuidema and Mapes (2008) and Liu et al. (2008) postulate that higher daytime echo may result from residual cirrus from the previous nights convection.

Figure 3.22c/d are the same as 3.22a/b except for EOs occurring only over land. In panel c, daytime echo is greater below 5 km for reflectivities less than 5 dBZ. These low-level daytime enhancements are similar enhancements of cumulus congestus (Fig. 3.21d), suggesting that while CloudSat is missing the mature stage of deep convection over land (Nessbit and Zipser 2003, Liu and Zipser 2008), it may be observing the peak in cumulus congestus development. Daytime echo is also greater than nighttime at high altitudes and reflectivities below 5 dBZ. These upper-level daytime enhancements are similar to enhancements for deep precipitation at reflectivities below 0 dBZ (Fig. 3.21a). Nighttime echo is greater at high reflectivities (dBZ > 0) and in a wedge from 7 km to 14 km at reflectivities below 0 dBZ. The nighttime wedge is similar to anvil enhancements (Fig. 3.21b). Nighttime anvil is likely from deep convection formed during the afternoon and early evening. Greater nighttime echo at high reflectivities suggests that rainfall is greater at 0130 LT, than 1330 LT over land. At 0130 perhaps convection triggered by earlier convection in the afternoon and evening is at a maximum.

Profiles of day and nighttime echo cover over the tropical land are shown in panel d. There is slightly more echo cover during the day below 3 km. From 3 km to 13 km echo cover is greater at night. Between 13 km and 14 km echo cover during day and night is similar. Above 14 km, day cover is slightly higher than at night. Again more upper-level echo cover at night may be associated with anvil from afternoon and/or early
evening convection. The increase during the day of the highest-altitude echo cover is perhaps due to the fact that deep precipitation occurring at 1330 is more intense and therefore reaching higher altitudes than precipitation occurring at 0130 (Liu and Zipser 2005).

3.5.3 NCFAD Enhancements of Deep Precipitation for Geographical Regions

This sub-section examines reflectivity differences for the deep precipitation EO type over different tropical ocean basins and land areas, as an example of answering the second question posed by this thesis. Figure 3.23 shows the NCFADs for deep precipitation over ocean (panel a) and land (panel b). The two panels are similar. As in 3.21a we see that heavy rainfall occurs with this EO type as indicated by the low-level high reflectivity values. Associated with the heavy precipitation is anvil and cirrus echo, indicated by the NCFAD extending rearward and upward from 15 dBZ above 5 km. There are a couple differences. A more defined bump at high reflectivities above 5 km over land suggests that deep precipitation over land has stronger updrafts, since more high-reflectivity (dBZ > 10dBZ) are being lofted above 5 km. Also the peak in anvil and cirrus type echo associated with deep precipitation is higher over land than ocean (compare the 6.93 value black contour in panel a and b).
Figure 3.23 – Same as figure 3.17a, except just for the deep precipitation EO type, box a figure 3.1 only over (a) ocean, (b) land.

Having established the NCFAD over the entire tropical ocean for deep precipitation, figure 3.24 shows the positive NCFAD enhancements in deep precipitation over the West Pacific (panel a), East Pacific (panel b), Atlantic Ocean (panel c), and Indian Ocean (panel d). Anvil and cirrus heights are highest over the West Pacific, followed by the Indian, Atlantic, and East Pacific. Anvil reflectivity is highest over the East Pacific. The East Pacific and Atlantic have more intense precipitation towers from stronger updrafts lofting high reflectivities (dBZ > 10) above the melting level (~ 5 km). The Indian Ocean has more high-reflectivity rain shafts (attenuating dBZ columns) below the melting level than the other basins, suggesting precipitation over the Indian Ocean may have fewer strong deep updrafts that lead to attenuation than the other basins. The East Pacific has enhancements for weak rainfall (high reflectivities below 3 km). The brightband (reflectivity enhancements around the melting level) is lower in the East Pacific than the Atlantic and Indian Ocean.

Figure 3.25 shows positive NCFAD enhancements over the total land tropical NCFAD for the Amazon basin (panel a), Congo basin (panel b), and Maritime Continent.
These names are general. For example, the Maritime Continent panel also includes land in northern Australia and southern Asia (figure caption provides specifics). Nevertheless, the Congo and Amazon basins have more intense updrafts (high reflectivities above 5 km) than the Maritime Continent. Congo updrafts are slightly more intense than over the Amazon, as high reflectivity enhancements above 5 km are shifted slightly toward higher reflectivities and upward. The Amazon has more low level cloudiness (height < 3 km) and weak rainfall (dBZ > 0 below 3 km) than the other land areas. The Maritime Continent has more high-reflectivity rain shafts below 5 km, perhaps indicating less attenuation by deep overlaying precipitation than the Congo and Amazon. Anvil and cirrus are highest over the Maritime Continent, followed by the Congo and Amazon basin. Brightband enhancements are greater over the Congo than Amazon.
Figure 3.24 – Shading is CFAD positive enhancements associated with deep precipitating EO types in each ocean basin to the normalized tropical deep precipitation ocean CFAD (Fig. 3.20). Black contours represent the normalized tropical deep precipitation ocean CFAD. Color scale is a relative scale, established by the range of values present within each individual panel. West Pacific includes ocean Dp EOs in boxes 16, 17, 24, and 25; East Pacific includes ocean Dp EOs boxes 10, 11, 18, 19; Atlantic includes ocean Dp EOs in boxes 12, 13, 20, 21; Indian Ocean Dp EOs includes ocean in boxes 14, 15, 22, 23 in figure 2.4.
Figure 3.25 – Shading is CFAD positive enhancements associated with deep precipitating EO types in each continental region to the normalized tropical deep precipitation land CFAD (Fig. 3.21). Black contours represent the normalized tropical deep precipitation land CFAD. Color scale is a relative scale, established by the range of values present within each individual panel. Amazon is Dp EOs over land in boxes 11, 12, 19, 20; Congo is Dp EOs over land in boxes 13, 14, 21, 22; Maritime Continent is Dp EOs over land in boxes 15, 16, 23, 24.
Chapter 4: Unexpected peak near -15°C in CloudSat climatology

4.1 Background

Among the three cloud étages (high, middle, low), middle clouds are least frequent and arguably the most mysterious. They are often obscured from surface and space optical observation by low and high clouds, respectively, so cloud radar offers the best chance at unbiased documentation of their occurrence. Upward looking cloud radars (e.g. ARM sites, Clothiaux et al. 1999) are useful but limited in coverage. IR based satellite may misclassify thin cirrus overlaying low cloud as a middle-topped cloud, causing an over-representation of middle clouds in data sets such as ISCCP (Wang et al. 1999). Global circulation models (GCMs) badly underpredict mid-level cloud amounts when compared to these (possibly imperfect) observations (Web et al. 2001, Zhang et al. 2005, Williams and Tselioudis 2007). Under future climate change projections from GCMs mid-level mid-latitude clouds have a consistent decrease in fractional coverage (Fig. 10.10, IPCC 2007), indicating that a better understanding of mid-level clouds and their climate feedbacks may be important to a changing climate.

Mid-level clouds are interesting from both a microphysical and dynamical perspective. Their microphysics are complicated by the possibility of both ice and supercooled liquid water (Wallace and Hobbs 2006). Dynamically, mid-level clouds may owe their existence to preferred detrainment levels in cumulus convection (Mapes and Zuidema 1996, Mapes 2001). Melting may also play a key role in mid-level cloud formation (Mapes and Houze 1995, Yasunaga et al. 2006).
This chapter revisits figure (3.1a), this time focusing on mid-level cloud structure. While variables such as ice (liquid) water content (IWC, LWC) may be useful in this chapter, the current CloudSat product providing IWC and LWC is too crude for this current study, as they currently assume a simple linear transition from water to ice between 0°C and -20°C (Wood 2008). Therefore, we only utilize the CloudSat 2B-GEOPROF product. Section 4.2 presents tropical characteristics of mid-level clouds, while 4.3 examines mid-latitude mid-level clouds. Section 4.4 discusses some possible interpretations and implications of the findings in section 4.2 and 4.3, especially of the surprising distribution peak near -15°C.

4.2 Tropical Mid-level Echo Object Climatology

Revisiting figure 3.1a we note there are two peaks in mid-level EOs (box E). Again, summing over all base heights in figure 3.1b gives the line plot of cloud-top density on the right side. As noted in section 3.1, the distribution has two dominant peaks at low and high levels, and two smaller mid-level peaks, rendering a curious quad-modal structure. The 2 km, 5–6 km, and 13 km peaks in tropical cloudiness are commonly explained by the three well know stability levels in the tropics: the trade inversion, the 0°C melting level isotherm, and the tropopause (Johnson et al. 1999). The second mid-level peak around 8 km (or -15°C in the tropics) was first noted in early CloudSat results from Haynes and Stephens (2007) and can be seen (once it is known to exist) in ARM site vertically-pointing cloud radar data (Hollars et al. 2003).

The corresponding profile of echo-cover distribution by EO top (rather than by local echo profile top) is shown as a solid curve in Fig. 4.1a, with similar values (in the
same units) as Fig. 3.1b. The EO types contributing to the bimodal mid-level structure are shown by thin and dashed lines in figure 4.1a. The dashed line includes only midlevel EOs in box E of Fig. 3.1a (thin layer EOs), while the dashed-dot line indicates the contribution by box D (tower like or precipitating EOs). The bimodal structure at mid-levels is clearest for the layer type EOs, so the remaining panels in Fig. 4.1 focus on those. Figs. 4.1b-d show that the mid-level bimodal structure in layer type EO tops is a robust feature occurring during day and night (4.1b), over land and ocean (4.1c), and in each tropical basin (Atlantic Ocean, Indian Ocean, and West and East Pacific) (4.1d). A few slight differences are seen too. Land coverage is greater than ocean coverage. The West Pacific has more midlevel cloudiness than the other basins, and an upward shift to the altitudes of the two peaks. Night has more midlevel clouds of this type than day. The main point of Fig. 4.1 remains: the bimodal structure is ubiquitous and robustly sampled.

Figure 4.1 – Horizontal cloud cover in the tropics distributed by EO top height. (a) all tropical EOs, and midlevel subdivision by layer vs. tower type EOs (types D and E in Fig. 1a). (b-d) cloudiness in layer-type EOs (heavy in all panels) and indicated subdivisions. The Indian, and west and east Pacific are separated at 110°E and 180°W.
We split box E of figure 3.1a in two at 6 km to look for similarities and differences in the two peaks. The upper (lower) peak refers to EOs in box E greater (less) than 6 km. The two peaks occur year round (Fig. 4.2). The upper peak appears more prevalent during the boreal summer months, while the lower peak remains constant throughout the year (Fig. 4.2). The two sub-populations have a seasonal progression similar to the ITCZ (Fig. 4.3), enhanced occurrence located north of the equator during boreal summers that progresses southward through the boreal fall and winter month. Geographically, the two peaks occur in similar areas in the tropics (Fig. 4.4a/b), with some differences. Because the upper peak contains more EOs (38,130 upper peak EOs vs. 18,163 lower peak EOs) the difference plot in Fig. 4.4c is favored toward the upper peak. They are both prominent around the ITCZ regions (i.e. north of the equator in the central and west Pacific and Atlantic Ocean) and in the tropical west Pacific warm pool. The lower peak has a slight equatorward shift in the central and west Pacific and Atlantic Ocean relative to the upper peak. The upper peak is enhanced relative to the lower one over land; geographic distributions indicate that this enhancement is prominent over North Africa and adjacent deserts (Fig. 4.4c).

**Figure 4.2** – Time (day) of EO vs. top height of EO for one year, July 2006 – June 2007 for EOs in box E of figure 3.1a. Bin size is 10 days by 240 m. Units are $10^5$ profiles per bin. Vertical lines indicate missing CloudSat days.
Figure 4.3 – Time (day) of EO vs. mean latitude of EO for EOs in box E of Fig. 3.1a for one year, July 2006 – June 2007. Bin size is 10 days and 3.6° latitude. Units are 10² profiles per bin. Vertical lines indicate missing CloudSat days.

Figure 4.4 – Echo cover for (a) upper peak, 6 km – 10 km, (b) lower peak, 4.5 km – 6 km, and (c) the difference panel a – panel b. Bin size is 3° longitude by 3.6° latitude. This figure was made with EOV3 data set.
Distribution in EO top temperature for the lower and upper populations is shown in figure 4.5. The lower population EO top temperature is more narrow than the upper population, with temperatures between 6°C and -16°C for the lower population and -7°C and -46°C for the upper population. The lower population distribution maximum is between -3°C and -4°C, while the first peak in the upper population is between -14°C and -18°C. We just mention the first peak in the upper population as the second one is the start of the cirrus EO population that has “leaked” into box E of figure 3.1. These temperature differences are useful for discussion of microphysical hypotheses in section 4.4.

**Figure 4.5** – 1D histogram of EO top temperature for the upper (Fig. 3.1a box E 6 km – 10 km) and lower (Fig. 3.1 box E 4.5 km – 6 km) sub-populations in the tropics (20°S – 20°N) for one year, July 2006 – June 2007. The peak near -30°C in the upper population is the start of the cirrus EOs “leaking” into box E of figure 3.1a. This figure made using EOV3 data set.

### 4.3 Extratropical Echo Object Climatology

We sought evidence of a similar bimodal fine structure in the extra-tropics (20°S/N – 60°S/N). None was apparent in height coordinates, but when we use ECMWF-AUX temperature to bin the EO tops, Fig. 4.6 is the result. Again, we have restricted the plot to include EOs over ocean and low land (topography < 1 km). As in
the tropics, low-based (towerlike or precipitating) EOs are prominent (density along the vertical axis) and produce the largest maxima in Fig. 4.6. Cirrus type EOs, with bases well above the surface and top temperatures below -60°C, also account for a lot of cloudiness. A ridge of density in Fig. 4.6 for EO tops along the -15°C to -20°C band can be discerned.

**Figure 4.6** – EO base height vs. EO top temperature for all EOs observed in one year, July 2006 – June 2007 for (20°N/S – 60°N/S). Binsize is 240 m by 2°C. Contour units are $10^3$ echo-containing profiles per bin.

The integral of this 2D histogram over base heights shows that this gentle ridge of density yields a distinctive if small peak in the 1D distribution of cloudiness by EO top temperature (Fig. 4.7), between the dominant low and high cloudiness peaks. Some enhancement near the 0°C level is also seen, but it is mixed with the broad low cloudiness more than in the tropics, since the Earth’s surface (and so the associated low cloud) tends to be cooler in the extratropics. Because the binning is different from Fig. 4.1, absolute values have different units and thus meaning: The absolute maxima in the profiles of Fig. 4.7 show that about 3% of the extratropical regions are covered by cloudiness whose EO-top temperature is within 2°C of -60°C.
The all-extratropics distribution (thick solid lines in figure 4.7) can be subdivided into day-night, land-ocean, and winter-summer. Daytime and nighttime overpass differences are almost unnoticeable for the cold and midlevel peak (Fig. 4.7a). Only the warm (low cloud) peak fluctuates much between day and night, with more low cloud during the night than day (Fig. 4.6a). Over both extra-tropical land and ocean the cold and midlevel peaks persist, but the low-cloud peak disappears over land (Fig. 4.7b). Low clouds in the winter hemisphere are somewhat cooler than in summer (Fig. 4.7c), but the small peak near -15°C is distinct and robustly present in both winter and summer, as in all the other cross-cutting categories of Fig. 4.7.

**Figure 4.7** – Horizontal cloud cover in the extratropics (20°-60°S/N) distributed by EO top temperature. Subdivisions include (a) day and night, (b) ocean and low land (topography < 1 km), (c) winter (JJA (DJF) S. (N.) Hemisphere and summer (DJF (JJA) S. (N.) Hemisphere), and (d) Theoretical growth curve (Rogers and Yau, 1989, Figure 9.4) overlaid with extratropics echo top temperatures between 0° and -40°C. Bin size is 2°C.
4.4 Discussion of Possible Interpretations and Mechanisms

The peak in CloudSat observed cloudy-echo top frequency in the -15°C to -20°C range raises several questions: Are we certain it isn’t a data processing artifact? If it is a real feature of radar reflectivity climatology, does this indicate an actual enhanced population of clouds, or merely a layer where the particles are more reflective (i.e., a radar ‘bright band’ as seen at this particular wavelength)? In either case, what might be the physical cause?

A data processing artifact seems extremely unlikely. Haynes and Stephens (2007) found the tropical peak near 8-9 km in an independent study of CloudSat data. No aspect of the cloud mask algorithm (Marchand et al. 2008) could produce a spurious sharp peak, especially not in temperature, as the reflectivity data are handled in geometric height. Furthermore, the peak is discernable in the climatology of cloud radar data from the Manus ARM site (Hollars et al., 2003), with very different sampling and sensitivity constraints.

Is this a true cloud climatology feature, or merely a feature of radar reflectivity detection at the wavelength and sensitivity limits of CloudSat? One line of evidence suggesting a true cloudiness enhancement near -15°C is the result of Hanna et al. (2008), showing a similar sharp peak in infrared (IR) derived cloud tops overlying stations experiencing rain or snow in winter. The study of precipitating locations presumably selects for optically dense clouds, minimizing the smearing effects of optically thin clouds on gross satellite IR temperature distributions. It would be fairly easy to seek such sharp structure in larger samples of IR satellite data, screened in other ways, although finding the tropical peak near 8-9 km would be difficult since it is most common in
convectively disturbed conditions (Fig. 4.9), where visibility is often obscured by higher clouds.

The counter-hypothesis – that the peak is merely from detection enhancement – also has support. Modeling of altocumulus by Sassen and Khvorostyanov (2007) showed that at potentially mixed-phase temperatures (0°C – -40°C), large ice crystals dominate cloud radar reflectivity. The liquid component (numerous but small drops) was found to be difficult to detect at a CloudSat-like sensitivity limit of -28 dBZ. In answer, we can at least confirm that the four modes in Fig. 3.1b persist when the absolute detection threshold is changed, i.e. when echo profile top is redefined as the highest pixel exceeding another threshold like -20 dBZ (Fig. 4.8). This is weak evidence however, as a microphysical boost in reflectivity at -15°C could simply enhance the frequency of exceedance of all thresholds.

![Figure 4.8](image)

**Figure 4.8** – As in figure 3.1b but redefined to only include profiles where the minimum reflectivity exceeds -20 dBZ. Contouring to the right of the thick 1 to 1 line on the left side is contouring error.

What is special about temperatures near -15°C? One possibility is that this is where dendritic ice crystal growth is most rapid. Cloud tunnel laboratory results from
Ryan et al. (1976) and Fukuta and Takahashi (1999) show a sharp peak in the mass and length growth rates of ice crystals at -15°C. Theoretically derived growth equations show a broad peak near -15°C. Figure 4.7d shows the growth equation curve (Fig. 9.4 of Rogers and Yau 1989) overlaid with the extra-tropical echo top temperatures between 0°C - -40°C. The broadness of the growth equation curve cannot explain our sharp spike in observation, even though the peak of the curve is at about the right temperature.

Hogen et al. (2006) show that radar reflectivity (Z) goes as mass squared (their equation 2), in the Rayleigh scattering regime where the small particles near cloud tops presumably reside.

Other microphysical factors such as fall velocity, ice crystal density, and ice nucleation may be important to understanding the -15°C to -20°C peak. Fukuta and Takahashi (1999) show a minimum in ice crystal fall velocity and apparent ice crystal density at -15°C. How these temperature dependent density and fall speeds affect are results is beyond our expertise. Ice nuclei become activated between -6°C and -16°C (Rogers and Yau 1989). The dependence of ice nucleation on temperature supports the hypothesis that CloudSat may be only sensing the ice particles, while missing the warmer liquid particles.

Dynamical cloud formation offers another class of hypotheses for the -15°C peak. Enhanced detrainment from deep convection could produce shelf clouds and/or moist layers that could be lifted later to saturation to produce layer clouds. In the tropics at least, the layer-type EO populations with tops near 0°C and -15°C are both associated with deep convection, occurring near and just after maxima in satellite rainfall data composites around the locations of EO occurrence (Fig 4.9). Separate day and night plots
of rainfall composites around the two peaks are very similar to the total (Fig. 4.9 c,d,e, and f). Jakob et al. (2005) also showed a relationship between tropical convective activity and mid-level clouds at the near-equatorial Manus Atmospheric Radiation Measurement (ARM) site. Furthermore, deep (convective) EO tops, not just layer cloud tops, show a second peak near -15°C (thin line in Fig. 4.1a), and deep echo profiles (Fig. 3.1b) also indicate a weak second mode near -15°C in box D, not just box E. Of course, a microphysical explanation could still underlie both tower and layer clouds, so this finding is not definitive indication of enhanced detrainment. Meanwhile, convection can detrain mass at levels below its top, so a detrainment hypothesis for the layer enhancement need not hinge on finding an enhancement of cumulus tops.

Figure 4.9 – TRMM 3B42 rainfall composites around daytime and nighttime EOs within the (a) upper peak (Fig. 3.1a, box E, 4.5 km – 6 km) and (b) lower peak (Fig. 3.1a, box E, 6 km – 10 km). This figure was made with EOV3 data set.

The main environmental control on detrainment is thought to be static stability (Bretherton and Smolarkiewicz 1989). We sought stability anomalies in composite ECMWF-AUX temperature profiles in association with the two sub-populations of layer-like EOs topping near the 0°C level and the -15°C level. A slight broad stability enhancement (lapse rate decrease) does emerge around the altitude of whichever EO type
is used as the composite basis, but the results are subtle and inconclusive (Fig 4.10). In any case, a stable-layer association with cloud enhancement would not constitute an explanation, since a layer of cloud creates a stable layer near its top by radiative cooling, and a peak in climatological stable layer occurrence near -15°C would be harder to explain than the cloudiness peak itself. Clear-air radiation and dynamics have no sharp temperature dependences, so some condensed-water effects must ultimately underlie these sharp fine structures tied to temperature.

![Figure 4.10](image)

**Figure 4.10** – (a) Average relative humidity profile for upper peak (solid) and lower peak (dashed) box E EOs. (b) Average lapse rate profile for the tropics (solid), upper peak (dashed) and lower peak (thin solid). The tropics wide lapse rate profile only includes EOs occurring over the ocean. The first km in each figure was excluded due to topography. This figure made with EOV3 data set.

The most obvious sharp-in-temperature process in the atmosphere is melting of precipitation near 0°C. Is it possible that the -15°C peak is somehow related to that? One dynamical hypothesis would be “melting layer reverberations,” discussed in section 6c of Mapes and Houze (1995). In that paper, vertically sharp features in midlevel wind divergence measured by airborne Doppler radar were the mysterious observation.

Idealized dynamical calculations and entrainment-detrainment arguments were used to illustrate how the vertical wavenumber dependent dispersive nature of slow (near-hydrostatic) gravity wave motions could create vertically wavy inflows and outflows
from convective cells in an environment with a sharp cooling process at the 0°C level.
Details of geometry and scale were necessarily vague, so that material does not predict
what would be expected in a global climatology. The vertical wavelength of excited
gravity waves depends on the depth of the melting layer (layer of negative latent heating),
as well as on how far above that the positive latent heat of fusion was released into the air
during the formation of those same frozen hydrometeors. Neither scale is very
fundamentally constrained, so all of this does not “explain” the ~2km or 15°C separation
of the two peaks in midlevel clouds that are our subject here. Experiments with melting-
like sharp forcings could be done in cloud resolving models with high vertical resolution
grids, to see if the ensemble effect is to create cloud enhancements 2-3 km above.

As a way forward observationally, phase discrimination of clouds near -15°C
might offer more clues. Lidar data from CALIPSO (Winker et al. 2003) could detect
small liquid droplets missed by the CloudSat CPR. Lidar depolarization from CALIPSO
can help discriminate particle phase (Cho et al. 2008). As mentioned in the introduction,
standard ice water content (IWC) and liquid water content (LWC) CloudSat products are
much too crude to advance the case, as they currently assume a simple linear transition
from water to ice between 0°C and -20°C (Wood 2008). There is clearly room for
improvement in our understanding of mixed-phase and ice microphysics, and a robust
global signature like that documented here may offer a good target for theory,
observation, and models.
Chapter 5: A Closer Look at Stratocumulus Echo Objects

5.1 Background

The importance of marine boundary layer (BL) clouds on the Earth’s radiation budget is widely known. Extensive stratocumulus cloud decks occur at the top of the marine BL. They have high albedo values leading to a negative shortwave cloud forcing around -100 Wm\(^{-2}\) locally, with a global net cloud forcing around -17 Wm\(^{-2}\), during the boreal summer (Klein and Hartmann 1993, Albrecht et al. 1995). Possible climate feedbacks from changes to this number are an important driver of research on these clouds (e.g. Bony and Dufresne 2005).

Stratocumulus clouds are prevalent over the upwelling regions of the sub-tropical oceans. Here, there is a large thermal and moisture contrast between cool, moist boundary layer air and warm, dry large-scale subsiding free-tropospheric air from the descending branch of the Hadley cell and climatological surface anticyclones. Large-scale subsidence keeps stratocumulus development in the BL. Convective turbulence driven by longwave radiative cooling at cloud top generally maintains a well mixed BL. Consequently, thermodynamic variables, \(\theta_l\) and \(q_t\), where \(\theta_l\) is liquid potential temperature and \(q_t\) is total water specific humidity, are constant through the depth of the BL and change rapidly in the free troposphere (Fig. 5.1, Stevens 2005). Competing with radiative cooling is cloud top entrainment of warm, dry air, which acts to deepen the BL. During daytime hours solar absorption and entrainment may act to warm and stabilize the sub-cloud layer leading to a transition layer between the well mixed cooler, moister surface and the well mixed warmer, drier cloud layer. This process is known as
decoupling of the BL and may lead to thinning and eventual demise of the cloud layer, as the cloud is cut off from the surface moisture supply. However, the layer may become re-coupled if cumulus convection ensues above the level of decoupling, providing a moisture source or link to the stratocumulus deck (Nicholls 1984, Albrecht et al. 1995, Cieselski 2001).

Figure 5.1 – Cartoon of well-mixed, nonprecipitating, stratocumulus layer, overlaid with data from research flight 1 of DYCOMS-II. Adapted from Stevens (1995).

Stratocumulus clouds were once thought of as non-precipitating; however, recent studies (e.g. Bretherton 1995, Stevens et al. 2005, van Zanten et al. 2005, Comstock et al. 2007) show drizzle not only exists in stratocumulus clouds but can be prevalent. Precipitation can lead to cloud layer decoupling and break up of the stratocumulus deck by depleting the cloud layer of liquid water, in turn cooling and moistening the sub-cloud layer through evaporation.

Direct observational stratocumulus statistics such as the proclivity for drizzle, cloud top and base height, and diurnal variations have thus far been limited to regional field campaigns (i.e. FIRE off the coast of California (Albrecht et al. 1988), ASTEX in the Azores (Albrecht et al. 1995), EPIC in the east Pacific (Bretherton et al. 2004)). While MODIS is good for assessing cloud top heights, it is not well suited for
stratocumulus clouds (Garay et al. 2008, Holz et al. 2008). Zuidema et al. (2008) use temperature from radiosonde launches to determine the relationship between cloud top temperature and cloud top height. Klein and Hartmann (1993) did look at global stratocumulus statistics using cloud fraction data from Waren et al.’s (1986, 1988) cloud atlases and COADS (Comprehensive Ocean-Atmosphere Data Source, Woodruff et al. 1987), as well as SST, sea level pressure, and air temperature from COADS and upper-air data from ECMWF. Their study focused on the seasonal variations of stratocumulus clouds, with emphasis on the differences from region to region, along with the radiative impacts of the clouds.

CloudSat provides a new tool to globally view stratocumulus clouds. With a MDS around -30 dBZ (Stephens et al. 2008), CloudSat is ideal for detecting precipitating stratocumulus. At the time most of the figures in this chapter were made, we had only processed 1.5 years of CloudSat data. In this chapter we aim to answer the first question posed by this thesis (i.e. Given a certain cloud type, where are those clouds occurring?) by specifically examining the stratocumulus EO type. Section 5.2 looks at basic stratocumulus statistics such as cloud coverage and cloud top height for the sub-tropical stratocumulus regions. Section 5.3 discusses seasonal variations of the stratocumulus clouds. Section 5.4 looks at precipitation characteristics and section 5.5 provides a summary and discussion.

5.2 Stratocumulus Coverage and Top Height

Figure 5.2 is the same as Fig. 3.1a, except for one year of CloudSat (July 2006 – June 2007) for 35°S – 35°N. Like in section 3.1, stratocumulus clouds occur in box F of
figure 5.2 and are defined as EOs which have a top height less than 4.5 km and a width greater than 10 pixels (about 10 km). Admittedly there is likely stratocumulus less than 10 km wide, but width was the threshold chosen to discriminate cumulus form stratocumulus. The stratocumulus EOs constitute about 20% of the subtropical and tropical (35°S – 35°N) EOs; specifically 231,671 out of 1,133,093 EOs for year one of CloudSat (July 2006 – June 2007).

![Figure 5.2](image)

**Figure 5.2** – Same as Figure 3.1 except for 35°S – 35°N for July 2006 – June 2007.

Using the stratocumulus EO definition from above, CloudSat identifies the major stratus deck regions of the world: the Californian, Peruvian, Namibian, and Australian decks. To make comparisons with other data sets, which do not attempt to discriminate stratocumulus from cumulus, we plot echo cover for all low (stratocumulus and cumulus) EOs in Fig. 5.4. Nighttime cloud cover is greater, in agreement with past studies. Daytime coverage peaks around 35% while nighttime values exceed 70%. Low EO cover is roughly 10% - 20% less than ISCCP low (cloud tops < 680 mb) cloud coverage (cf. Fig. 5.4a and 5.5). ISCCP has the maximum low cloud coverage occurring in the Namibian deck while the CloudSat maximum is in the Peruvian deck. However, ISCCP is an imperfect comparison, as the data set is known to misclassify low clouds. Perhaps a
better comparison is CALIPSO, which flies directly behind CloudSat with lidar capabilities making it more sensitive to low level cloud heights and free from surface clutter (Winker et al. 2003). Figure 3 of Leon et al. (2008) shows daytime low (cloud top height < 4 km) cloud coverage amounts from CALIPSO can be as high as 80% in some regions that go to almost 100% at night. Unlike CloudSat or ISCCP, CALIPSO has similar maximum cloud cover values in the Californian, Peruvian, and Namibian decks. With about 30% - 40% more cloud cover being detected by CALIPSO, CloudSat is missing some of the lowest clouds. Further evidence of CloudSat limitations on low cloud detection is the maximum in low cloud cover of CloudSat is shifted farther off the coast than that of CALIPSO (cf. Fig. 3 Leon et al. with Fig. 5.4a and b).

![Figure 5.3 - Stratocumulus echo cover for (a) day and night, (b) night, and (c) day from July 2006 – June 2007. Bin size is 3° longitude by 3.6° latitude. Echo cover is defined the same as in chapter 3.](image-url)
Figure 5.4 – Same as Fig. 5.3 except for both stratocumulus and cumulus EOs.

Figure 5.5 – ISCCP low (tops < 680 mb) cloud cover average from 1983 - 2007. Courtesy of David Painemal.
Stratocumulus EO top heights versus longitudes show a coastal shoaling for each sub-tropical deck (Fig. 5.6). While coastal shoaling has been noted in previous studies (e.g. Zuidema et al. 2008, submitted), CloudSat gives the opportunity to view and compare each deck over the same time span. In figure 5.6 the Namibian deck (label C) exhibits the greatest tilt in coastal shoaling, followed by the Peruvian (label B) then Californian (label A) decks and finally the Australian deck (label D), which shows almost no shoaling and higher EO top heights in general.

Figure 5.6 – 2D histogram of EO top height versus longitude for July 2006 – June 2007. Units are $10^{-1}\%$ horizontal echo pixels per longitude (3°), height (240 m) bin. A, B, C, and D, denote Californian, Peruvian, Namibian, and Australian decks, respectively. Figure made with EOV3.

Coastal shoaling differences may be influenced by topography differences on the continental regions east of the decks. Richter and Mechoso (2004, 2006) looked at the influence topography has on the Namibian and Peruvian stratocumulus regions. They concluded the mechanisms influencing lower static stability, and therefore the maintenance of the stratocumulus decks, is different in the southeast Atlantic and southeast Pacific. At the surface westerlies in both the southeast Atlantic and Pacific are deflected equatorward by topography. However, flow above the BL is poleward in the southeast Atlantic, while it is equatorward in the southeast Pacific. Lower African
topography allows the poleward branch of the south African anti-cyclone to extend over the Namibian stratocumulus region. The higher Andes mountains cause blocking, effectively deflecting air east of the mountains poleward and air west of the mountains equatorward. So, in the case of the Namibian stratocumulus region static stability is influenced by warm, continental air, while isentropic sinking above the BL associated with equatorward moving air influences static stability over the Peruvian region. However, the results of Richter and Mechoso (2004, 2006) do not really explain why coastal shoaling is greater in the Namibian deck vs. Peruvian deck, merely that static stability is influenced by different mechanisms in these regions.

Another hypothesis for cloud top height differences from region to region is the shape of nearby coastlines is influencing cloud top height. Figure 5.7 shows the average EO top height per bin for all stratocumulus EOs in the sub-tropics (panel a) and zoomed in on the Peruvian and Namibian decks (panel b). In panel (a) cloud top heights become lower toward the coast for the Californian, Peruvian, and Namibian region, while the Australian deck remains constant (consistent with Fig. 5.6). CloudSat EO top height values are slightly higher than Aqua MODIS derived cloud top heights (cf. 5.7b and 5.8). The comparison is not exactly one to one, since the CloudSat figure (5.7b) includes all stratocumulus EOs for one year, while the MODIS figure (5.8) is the average for October 2005, 2006, and 2007. Both CloudSat and MODIS show the lowest cloud top heights along the South American coastline around Peru with higher heights near the Arica Bight. At the Arica Bight the South American coastline juts westward, causing convergence of surface winds that flow parallel to the Chilean coastline, thus elevating top heights (Zuidema et al. 2008, submitted). Along the south African coastline, both CloudSat and
MODIS find the lowest top heights around -15°S, where the coastline bends westward. Here, low top heights extend northwestward. No explanation is available for the curious nature of the Namibian cloud top height variations (Zuidema et al. 2008, submitted).

Why the Australian deck exhibits no coastal shoaling is intriguing. Looking at Fig. 5.12a,b we see that zonal gradients in SST are less over the Australian stratocumulus region than the Peruvian or Namibian region. Albrecht et al (1995) discuss how increasing SSTs away from the coast will lead to a deeper BL, with increasing cloud top heights. In the absence of strong SSTs changes westward of the Australian coast, the BL height may remain fairly constant, thus making stratocumulus top heights constant.

**Figure 5.7** – Average stratocumulus EO top height per bin. Figure made with EOV3.
5.3 Stratocumulus Seasonal Cycle

In this section we use 1.5 years of CloudSat data (July 2006 – December 2007) to examine the seasonal cycle of the stratocumulus decks and compare EO occurrence in austral spring 2006 to 2007. Figure 5.9 shows the seasonal cycle of each stratocumulus region. The southern hemisphere decks (i.e. Australian, Peruvian, and Namibian, panels a, b, and c, respectively) show maximum EO coverage during August, September, and October (ASO) and minimum coverage in March, April, and May (MAM). The Californian deck (panel d) is fairly persistent throughout the year with a maximum in December and January.

As mentioned in section 5.1, seasonal differences in the extent of the stratocumulus decks have been documented by Klein and Hartmann (1993) using Warren et al.’s (1986, 1988) and COADS surface observation. As in our CloudSat figure 5.9 their figure 4 shows maximum (minimum) cloud amount in ASO (MAM) for the Peruvian and Namibian decks. Unlike our seasonal distribution of the Australian and Californian deck, they show little seasonal progression of cloud amount in the Australian
deck and a maximum amount in June, July, and August (JJA) for the Californian deck, quite opposite to our December, January maximum. We thought that the boreal wintertime maximum in the Californian region in our CloudSat data might be a result of wintertime storms, with low clouds obscured from traditional satellite observations by higher clouds. To test this idea, we reproduced figure 5.9d this time screening for stratocumulus that had overlapping EOs and those that did not (Fig. 5.9). In the overlap situations (Fig. 5.10a) the stratocumulus maximum again occurs in wintertime. However, the no overlap situations (Fig. 5.10b) show two peaks, one still in the boreal wintertime, with the other during boreal summertime (July and August). The summertime maximum is consistent with Klein and Hartmann (1993), but the remaining wintertime maximum even with overlapping clouds removed is puzzling.

Comparing austral spring 2007 to 2006 we find the Peruvian deck (Fig. 5.9b) has increased EO coverage in ASO 2007 in both the main deck region (5°S – 35°S) and the subsidiary cold-tongue region (5°N – 0°). Similarly, the Namibian deck EO coverage is enhanced during ASO 2007 compared to 2006. Contrary to the Peruvian and Namibian decks, the Australian deck has greater EO coverage in ASO 2006. Figure 5.11 show the geographical EO cover for ASO 2006 (panel a), ASO 2007, and their differences (panel c). Increases in echo cover occur in the majority of the Peruvian deck. Increases as high as 40% occur along the cold tongue and near 18°S and 90°W. Echo cover changes in the Namibian deck are not as profound as the Peruvian. The northern extent of the Namibian deck extends further west in 2007 (easiest to see comparing panel a and b). Changes close to the African coast (east of 0°) are mostly small (values < 10%).
Figure 5.9 – Seasonal cycle for (a) Australian [10°S-35°S, 50°E-110°E], (b) Peruvian [5°N-35°S, 70°W-130°W], (c) Namibian [0°-35°S, 40°W-20°E], and (d) Californian [5°N-35°N, 180°W-110°W] stratocumulus echo area coverage. Units are 10^{-1} percent contribution for each time (10 days), latitude (3.6°) bin. Black vertical bars denote missing CloudSat days. Figure made with EOV3 dataset.
Figure 5.10 – Californian stratocumulus echo area coverage. (a) Echo objects with overlapping echo at other altitudes and (b) these without overlapping echo. Figure made with EOV3.

Figure 5.11 - Same as Fig. 5.3 except just for stratocumulus EOs in September, October, November (SON) (a) 2006 and (b) 2007. (c) 2007 minus 2006 SON stratocumulus cover.
We hypothesize that enhancements in the Peruvian and Namibian decks in 2007 may be caused by cooler SSTs and/or enhanced lower-troposphere static stability (or inversion strength). (Of course, there are “chicken-vs.-egg” concerns, since enhanced cloudiness may cause cooler SSTs and enhance lower-tropospheric static stability by radiative cooling effects.) To address the hypothesis, Fig. 5.12 presents SSTs for ASO of 2006 (Fig. 5.12a) and 2007 (Fig. 5.12b). As anticipated, 2007 had cooler SSTs under the enhanced stratocumulus decks. As already mentioned in chapter 3.2.3, the Nino 3.4 index indicates a shift from moderate El Nino conditions in 2006 to major La Nina conditions in 2007 (Fig. 3.8). In the time span shown only the 1999/2000 La Nina was stronger. Cooler waters are found along the South American west coast, along with a more extensive cold tongue near the equator and a retreat of the warmest water temperatures westward to the west Pacific in 2007 compared to 2006. Little difference is seen in SSTs along the west African coastline (Fig. 5.10c). Figure 5.12 supports the hypothesis that cooler SSTs in 2007 caused the Peruvian deck’s increased EO cover, just as Burgman et al. (2008) showed for La Nina like conditions support enhancing sub-tropical cloudiness in the west Pacific on decadal time scales.
Following Klein and Hartman (1993) we define lower-troposphere static stability as the difference in potential temperature at 700 mb ($\theta_{700}$) and the surface (their equation 1). Figure 5.13 shows the 1000 mb temperatures for 2006 (panel a), 2007 (panel b), and of 2007 minus 2006 (panel c). 1000 mb temperatures are used as surface $\theta$. Figure 5.13 has a similar pattern to 5.12, with a more expansive cold tongue corresponding to La Nina like conditions in 2007. 700 mb $\theta$ is shown in figure 5.14 for 2006 (panel a), 2007
Noteworthy differences are a slight increase in $\theta_{700}$ along the west coast of South Africa and a decrease in $\theta_{700}$ along the west coast of South America. The increase along west Africa has a similar pattern as the Namibian stratocumulus deck, i.e. it extends north and westward from the coast of Africa. Figure 5.15 shows the computed static stability for 2006 (panel a), 2007 (panel b), and 2007 minus 2006 (panel c). Consistent with little change in 1000 mb temperatures yet increases in $\theta_{700}$, the static stability along the west African coast increases (between ~25°S – 5°S). Such a change suggests that increases in the inversion strength driven by changes in the $\theta_{700}$ may explain the increased cloudiness in the Namibian deck in 2007. Increases in static stability are also seen along the west South American coast, with maximum increases in the cold tongue region. Here, both the 1000 mb temperature and $\theta_{700}$ decreased, so an increase in inversion strength means 1000 mb temperatures must have decreased more than to $\theta_{700}$.

Another metric by which to explain 2006 and 2007 cloudiness differences in the Peruvian and Namibian region is change in 500 mb omega ($\omega_{500}$, subsidence). Figure 5.15 shows $\omega_{500}$ for 2006 (panel a), 2007 (panel b), and the difference of 2007 and 2006 (panel c). Little change is seen over the Namibian stratocumulus region, while there was a decrease (reduced subsidence) over the Peruvian region. This is contrary to expectations as an increase in subsidence would be more consistent with an increase in stratocumulus coverage. A slight increase in $\omega_{500}$ in the Namibian stratocumulus region is more consistent with expectations.
Figure 5.13 – NCEP reanalysis 1000 mb temperature for August, September, and October of (a) 2006 and (b) 2007 and (c) the difference between 2007 and 2006 from NCEP/NCAR reanalysis. In (c) Solid curves are positive, dashed are negative, and thick solid is zero.
Figure 5.14 – 700 mb theta computed from NCEP reanalysis 700 mb temperatures for August, September, and October of (a) 2006 and (b) 2007 and (c) the difference of 2007 and 2006. In (c) Solid curves are positive, dashed are negative, and thick solid is zero.
Figure 5.15 – Static stability for August, September, and October of (a) 2006 and (b) 2007 and (c) the difference of 2007 and 2006 from NCEP/NCAR reanalysis. In (c) Solid curves are positive, dashed are negative, and thick solid is zero.
Figure 5.16 – ASO 500 mb omega for (a) 2006 (b) 2007 and (c) the difference of 2007 and 2006 from NCEP/NCAR reanalysis. Solid curves are positive, dashed are negative, and thick solid is zero. Contour interval is 0.01 Pa/s.

5.4 Stratocumulus Precipitation Characteristics

To look at precipitation characteristics of stratocumulus EOs we utilize attributes 17 and 18 in table 2.1 (number of pixels greater than 0 dBZ and -17 dBZ) in this section. As mentioned in chapter 3.5.1 a pixel is considered drizzle if reflectivity is greater than -
17 dBZ (Frisch et al. 1995). We use 0 dBZ as our rain threshold. Figure 5.17 provides a breakdown of how many pixels per stratocumulus EO qualify as drizzle or rain. Black diamonds are total number of pixels per stratocumulus EO, by definition this value is always greater than 10. Red triangles and blue asterisks indicate how many pixels per stratocumulus EO qualify as drizzle and rain, respectively. We see that 90% (55%) of all stratocumulus EOs contain at least one drizzle (rain) pixel.

**Figure 5.17** – Cumulative distribution of the total (black diamonds), drizzling (red triangles), and raining (blue pixels) number of pixels per sc EO.

To get a better idea of which sc EOs are producing the drizzle and/or rain, we broke the stratocumulus EOs into size quartiles according to volume and looked at the pixel ratios of raining or drizzling pixels to total number of pixels. Figure 5.18 shows the distribution of rain and drizzle pixel ratios for the smallest size quartile, middle sized half, and largest size quartile. Overall, the distributions are similar for raining stratocumulus EOs in different quartiles (Fig. 5.18a). Less than 1% of the raining stratocumulus EOs in each quartile have pixel ratios exceeding 60%. In figure 5.18b each quartile has a distribution peak between 60% - 70%. The lower quartile (smallest drizzling stratocumulus EOs) has the largest fraction of pixel ratios below 50%, while the upper quartile (largest drizzling stratocumulus EOs) has the largest fraction of pixel ratios
above 60%. The results in figure 5.18b are somewhat counterintuitive as precipitation is usually associated with broken stratocumulus or pockets of open cells (POC; Stevens et al. 2005), so we might have expected the smaller sized stratocumulus EOs to be the more drizzly EOs. Perhaps the larger EOs have better organization leading to an increased likelihood of precipitation. However, these figures say nothing about the larger-scale environment in which the stratocumulus EO types are occurring, so it is possible that the large (upper quartile) raining and drizzling EOs are occurring within a larger region of stratus, thus qualifying them as POCs. Figures like 5.18 for each region were very similar to each other (not shown), indicating statistical robustness of these findings.

![Histograms](image)

**Figure 5.18** – Histogram of (a) raining-to-total pixel ratio for raining stratocumulus EOs and (b) drizzling-to-total pixel ratio for drizzling stratocumulus EOs. Stcu EOs are divided into quartiles according to volume. Area under each curve equals one. Bin size is 0.1.

5.5 Summary

Our CloudSat defined stratocumulus EO type identifies the main sub-tropical stratocumulus decks (i.e. Californian, Peruvian, Namibian, and Australian) (Fig. 5.3). This result is a good check on our cloud type definitions and means we do not have to use prior knowledge on the location of stratocumulus regions to identify them.
Low EO cover (Fig. 5.4) is similar to ISCCP values (~ 10 – 20% less) and quite lower than CALIPSO values (~ 30 - 40% less). Day and night coverage differences are consistent with previous studies that show increased cloudiness overnight due to a better well-mixed BL.

Average EO top heights for each latitude, longitude bin are slightly higher than MODIS derived cloud top heights (cf. 5.6 and 5.7). The Californian, Peruvian, and Namibian decks all show a coastal shoaling of top heights, with the Namibian deck displaying the greatest tilt. The Australian deck shows little if any shoaling (Fig 5.6).

The Peruvian and Namibian deck share a common seasonal cycle (at least for the 1.5 years shown, Fig 5.9), with peak coverage occurring in ASO, consistent with findings from surface observations used by Klein and Hartmann (1993). The Californian deck is persistent year round, with a boreal wintertime maximum, which is contrary to Klein and Hartmann (1993). The Australian deck peaks in JAS. Year-to-year differences in the Peruvian and Namibian deck appear to mainly reflect changes in SSTs (La Nina) and static stability (Figs. 5.12 – 5.15).

Drizzle and rain characteristics of stratocumulus EOs were considered. Most stratocumulus EOs contain at least one drizzling pixel (90%), while a little more than half (55%) contain a raining pixel (Fig. 5.17). Larger EOs are much more likely to have more than 50% of their pixels drizzling, suggesting an importance of mesoscale organization to rain production, rather than a break up of cloud masses by drizzle processes as might have been anticipated (e.g. Stevens et al. 2005)
Chapter 6: Cloud Types Across the Madden-Julian Oscillation

6.1 Background

The Madden-Julian Oscillation (MJO) is a low frequency (30 – 90 day) variation on large scales, in the tropical belt, especially the Indo-Pacific warm pool. A broad envelope of convection is initiated in the Indian Ocean, propagates eastward at an average speed of 5 m s\(^{-1}\) over the Maritime continent, then into the central Pacific Ocean where convection wanes (reviewed in Madden and Julian 2004, Zhang 2005).

The MJO affects tropical weather, including rainfall over the Indian-Pacific sector, monsoon rains over the nearby continents of Asia and Australia, and tropical cyclone around the world. The MJO also has mid-latitude teleconnection impacts, and may be important to the evolution of the El Nino – Southern Oscillation (ENSO). Despite the MJO’s importance, complete understanding of the phenomenon remains elusive as indicated by the failure of global circulation models (GCMs) to accurately simulate the MJO (Zhang 2005, and references therein).

Observational studies characterizing the thermodynamics, dynamics, and precipitation characteristics throughout the MJO provide critical information to understanding the MJO and serve as solid benchmarks for GCM simulations. Previous observational studies have utilized NCEP and ECMWF reanalysis variables (e.g. Lin et al. 2004, Kiladis et al. 2005, Benedict and Randall 2007), radiosondes (e.g. Lin and Johnson 1996, Lin et al. 2004, Kiladis et al. 2005), satellite sounders (e.g. Myers and Waliser 2003, Tian et al. 2006), and precipitation retrievals (Benedict and Randall 2007, Morita et al. 2006). Together these studies provide a picture of the large-scale structure
of the MJO. During the suppressed (dry) phases of the MJO climatological easterly winds predominate, with anomalous dry humidity and descending motion. Immediately ahead of the active (wet) phase surface winds become westerly, low-level moistening occurs and ascending motions begin to predominate. During peak convective activity low-level temperature anomalies are generally cool with warm anomalies through the majority of the upper troposphere with maximum moisture in the middle troposphere. Following the wet phase low level drying ensues and builds vertically and descending motions and easterlies return (see Fig. 3 Benedict and Randall, Fig. 7 Zhang 2005).

With the above knowledge several schematics have been created to visually express convective activity and clouds through the MJO (Fig. 6.1a). All three MJO schematics (Fig. 6.1a,b,c) generally show a progression of cloud types from shallow, to middle topped, to deep clouds, followed by high topped somewhat thick clouds, with an eventual return to shallow clouds. Interestingly, this progression of cloud types is similar to cloud evolution observed over the life cycle of mesoscale convective systems (MSCs) (Fig. 6.1d). Implications for this similarity over various space and time scales will be discussed in section 6.4.

In this chapter we use direct observations from CloudSat to characterize cloud types across various phases of the MJO. Section 6.2 explains our methodology for defining MJO phases. Section 6.3 discusses cloud type variability across the MJO. Section 6.4 provides a summary and discussion.
Figure 6.1 – Schematics of clouds through MJO phases from (a) Lin and Johnson (1996), (b) Benedict and Randall (2007), and (c) Morita et al. (2006). (d) is cloud development through a mesoscale convective system from Zipser (1981).

6.2 Defining MJO Phases

We use a filtered view of twice daily satellite measured outgoing longwave radiation (OLR) on a 2.5° by 2.5° grid to identify MJO events. OLR is used because it is a good proxy for tropical deep convection. The MJO signal was obtained by space-time filtering of OLR to isolate eastward wavenumbers 0 – 9 with periods between 30 – 96 days (filtered data courtesy of George Kiladis). Unlike Wheeler and Kiladis (1999), the filtering is done on latitude belts with no symmetric and antisymmetric decomposition.
We averaged their filtered OLR over 15°S – 15°N to create a time-longitude section (black contours on Fig. 6.2), which highlights MJOs in the raw OLR section (also 15°S – 15°N).

![Daily OLR](image)

**Figure 6.2** – Raw OLR from 15°S – 15°N for July 2006 – February 2008 from NOAA PSD website overlaid with filtered MJO OLR. Contour interval is 10 Wm⁻², with negative values dashed.

Based on the above MJO filtering, 10 discrete phases are defined as indicated in figure 6.3. At each grid point, filtered MJO OLR has first been standardized then plotted against its standardized local time derivative. A pure sine wave would be a unit circle on such a diagram, but because many frequencies are present the points take the spiral forms. To isolate strong MJO events, we only consider phases when the amplitude is greater than two standard deviations of the filtered MJO OLR (colored numbers in Fig. 6.3). Phase numbers are assigned to figure 6.3 by simply dividing the spiral into 10
discrete parts. Interpretation is that phase number 0 (black colored numbers) represents the most suppressed (dry) phase of the MJO, with the highest filtered OLR values and a time derivative near zero. Opposite phase 0, phase 5 (dark green colored numbers) represents the most active (wet) phase of the MJO because it has the lowest filtered OLR values. Likewise phases 1 – 4 (6 – 9) are the before (after) phases of the MJO’s convective activity. The tooth-shaped points extending out of the spiral are times and locations when OLR is at a maxima or minima (e.g. ~100°E and January 1, 2008, Fig. 6.2) or when there is a sharp transition (gradient) to or from median values of OLR to or from maxima or minima (e.g. ~110°E and December 25, 2007, Fig. 6.2).

To identify cloud characteristics through the various phases, each EO in 15°S – 15°N is assigned an MJO phase 0 – 9 according to Fig. 6.3 based on the EO’s time and mean longitude. There were 111,795 EOs in 15°N – 15°S that qualified as occurring in an MJO phase (standardized filtered anomaly > 2 on Fig. 6.3).

Figure 6.3 – Phase definition for MJO. Standardized filtered MJO OLR (solid contours in Fig. 6.2) is scatter plotted against its standardized local time derivative. Numbers represent the phase of the MJO, for times and places when amplitude (distance from origin) is greater than two standard deviations.
6.3 Cloud Types by MJO Phase

6.3.1 Echo Object Statistics

A similar EO base vs. top height histogram as figure 3.1a was initially made for each phase of the MJO, but these were noisy due to a small number of EOs occurring in each phase (~10,000 per phase, compared to more than 600,000 over the entire tropics). To refine the image, Fig. 6.4 shows local profile base vs. profile top height histograms with adjoining profile top density plots. Broadly, the figures show an increase in upper level EO types (boxes A, B, and C) from the suppressed phase (0) to the active phases (4-6), which then dwindle afterwards, phases 7-9. Mid-level EO types (boxes D and E) also become more prevalent during active phases. These broad features are also discernable in the profile top density line plots at the right of each panel. Profile top density reaches a maximum in the 12 km – 14 km range of over 2% per 240 m altitude bin during the most active phase (5), from less than 1.5% during the most suppressed phase (0) (Figs. 6.4a vs. 6.4f).

A clearer display of cloud type evolution through MJO phases is in figure 6.5. The top (bottom) panels show areal coverage (echo volume) coverage for each EO type by MJO phase. The left panels are normalized to show the percent contribution of each cloud type to the total area or volume coverage, while the right panels indicate the actual values of area and volume by EO type in each phase. Each color in the phase bars represents an EO type indicated on the figure’s key. Again EO types are defined as in figure 3.1a, but with a separation of the deep precipitation (dp) type into narrow (width < 200 km) and wide (width > 200 km).
Figure 6.4 – Horizontal cloud cover weighted histograms (like Fig. 3.1b) in each indicated phase of the MJO. Cloudiness is distributed by local (echo profile) bases and tops. Contour values (labeled) are units of 10 (echo-containing profiles) per 240 m altitude bin. Lines and letters delineate EO types, same as figure 3.1b. The 2D distribution of each phase is integrated at the right of each figure to show the 1D distribution of each MJO phase cloud cover by local cloud top height (in units of % per 240 m).
The most noticeable feature in figure 6.5 is a total increase in area and volume during the most active phases (4-6). Total area covered echo roughly doubles from suppressed to active phases (Fig 6.5a), while volume triples (Fig. 6.5d). It is not one EO type contributing to the increase in area and volume during active phases, rather each EO type’s area and volume are more abundant, although wide-dp EOs are modulated most (orange slab in panels a, c). A curious feature in both the area and volume coverage is that the two most suppressed phases (0, 9) actually have more total area and volume than their adjacent less suppressed phases (1, 8). Whether this is true of all MJO events, or simply for our sample and definition, is unclear at this point. Table 6.1 provides exact values for actual area contribution as well as percent area contribution per EO type per phase.
<table>
<thead>
<tr>
<th>Phase</th>
<th>Cloud Types</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>ncb</td>
</tr>
<tr>
<td>0</td>
<td>2.6 \times 10^4</td>
</tr>
<tr>
<td></td>
<td>9.9</td>
</tr>
<tr>
<td>1</td>
<td>2.8 \times 10^4</td>
</tr>
<tr>
<td></td>
<td>12.0</td>
</tr>
<tr>
<td>2</td>
<td>3.3 \times 10^4</td>
</tr>
<tr>
<td></td>
<td>12.3</td>
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<tr>
<td>3</td>
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<td></td>
<td>10.0</td>
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<td>4</td>
<td>3.9 \times 10^4</td>
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<tr>
<td>9</td>
<td>2.2 \times 10^4</td>
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<tr>
<td></td>
<td>7.9</td>
</tr>
</tbody>
</table>

**Table 6.1** – Top values are *actual* area contribution per phase per cloud type, bottom values are *percent* area contribution per phase per cloud type.
A few other points can be seen in Fig. 6.5a and c, the normalized panels. Starting at the top of the rainbow, narrow dp (red) have their largest area and volume percent contribution to all the dp cloud types (red and orange) in the suppressed phase and early phases (0, 1, and 2). Wide dp (orange) has a greater contribution in the most active phases. Together these dp EO types account for more than 30% area and more than 60% of volume of echo during all phases. Cirrus values (green) are the next most prominent for areal contribution, accounting for over 27% of echo cover during phases 0 and 9. Anvil (yellow) EOs are most prominent after the most active phases, during phases 6, 7, and 8. Mid-level EO types, cumulus congestus (light blue) and alto cumulus (dark blue), are almost the same in all phases. Stratocumulus (purple) and cumulus (black) have the greatest fractional contribution during the suppressed and early phases (0, 1, and 2).

6.3.2 Pixel Statistics in dBZ-Height Space

As in chapter 3, pixel statistics are shown as CFADs. Figure 6.6 shows the normalized CFAD (NCFAD) for all phases of the MJO. The basic structure of Fig. 6.6 is broadly similar to the all tropics NCFAD (Fig. 3.15a) discussed in section 3.5. In order to emphasize MJO phase differences we subtract the NCFAD over all MJO phases (Fig. 6.6) from each MJO phase’s NCFAD, to yield the panels of Fig. 6.7. Black contours in each panel are the same as in Fig. 6.6, to show how each phase’s positive and negative enhancements are related to the overall MJO NCFAD. By comparing each phase’s positive enhancements to the NCFADs of each EO type in the tropics (Fig. 3.18), we can discern which EO types are dominant in each phase. Results of Fig. 6.7 are consistent with Fig. 6.5, but offer another way to look at them.
Inactive (phase 0 and 9) positive enhancements show up as enhancements of low dBZ at low and moderately high altitudes (10 km – 13 km) (cumulus and cirrus in Fig. 6.5). But high reflectivity enhancements are also seen, indicative perhaps of the intense cells (Fig. 6.1c) and showers with little overlaying attenuation causing precipitation. At low levels, the cumulus related enhancements in phases 1 and 2 are similar to phase 0, but are about a km thicker. At upper levels, 10 km – 13 km, cirrus enhancement during phase 9 – 0 give way during phase 1 – 2 to higher weak echoes (14 km and above). High reflectivity values at upper levels (indicating strong deep convective updrafts) remain enhanced during phase 1 and 2. In phases 3, 4, and 5 the rain-related areas (high reflectivities near and below 5 km level) become enhanced, while high-altitude, low reflectivity cloudiness remains enhanced.

Mid-level enhancements increase steadily from phase 3 to phase 5 to lower reflectivity, reaching a maximum during phase 5 between 0 dBZ – 10 dBZ and 5 km – 8 km. These results during active phases are consistent with Fig. 4 of Masunaga et al. (2008)’s CloudSat CFAD differences from the wet and dry phases of the December 2006
January 2007 MJO event (where they defined wet and dry by rainfall measured by the TRMM PR).

Conditions are more anvilish through phases 6 – 8, as indicated by the wedge shaped enhancements in 8 km – 13 km that is very similar to the anvil EO type NCFAD enhancements in the tropics in figure 3.18b. In phase 6, very high altitude, low reflectivity enhancements appear along the upper most levels of Fig. 6.7g, suggesting high cirrus is also prominent during this phase. During phase 8 there is a return to cumulus-like low level enhancements at low reflectivities, indicating a return to suppressed conditions of phase 9 – 0. Phase 9 has a striking quadrapole of enhancements at: (1) low levels and low reflectivities (cumulus-like clouds), (2) the upper level wedge at low reflectivities (anvil-like clouds), (3) low levels and high reflectivities (shallow showers without attenuation by precipitation), and (4) the strong updraft region (high dBZ above the 0°C level).
Figure 6.7 – Shading is CFAD positive and negative enhancements associated with each MJO phase relative to the normalized MJO CFAD for all phases (Fig. 6.5). Black contours represent the normalized MJO CFAD for all phases.

6.3.3 Total Echo Cover

Next we look at total cloud cover by MJO phase. Cloud cover is computed, as in chapter 3, by summing un-normalized MJO phase CFAD over all reflectivity bins at each altitude, then dividing by the total number of times CloudSat sampled that phase of the MJO (Fig. 6.8). Echo cover at all heights approximately doubles during active phases of
the MJO compared to suppressed conditions before and after the active phases, consistent with the results of figure 6.5b. Each phase has two prominent echo cover peaks, around 1.5 km and 12 km. During phases 0 – 2 echo cover for the lower and upper peaks are comparable. Subsequent phases show the upper peak coverage growing relatively more than the lower peak. The maximum difference in lower and upper level cloud cover occurs during phase 6, when upper cloud cover is about 9% greater than lower cloud coverage. After phase 6, upper level cloud coverage begins to retreat, though still dominates over lower cloud coverage through phases 7 – 9. A bump in mid-level echo cover around 5 km occurs during phases 3 – 6, with a slight secondary bump at about 8 km in phases 5 – 6, arising from mid-level NCFAD enhancements in Fig. 6.7.

Figure 6.9a shows the distribution of echo cover by MJO phase as in Fig. 6.8 but now contoured together. Figure 6.9b shows anomalous echo cover by MJO phase, after subtracting the mean across all 10 phases. Positive enhancements begin during phase 3. Notable features during positive enhancements are the concave structure of enhancements between 5 km – 15 km during phase 3 and 4 with the opposite convex structure during phase 6 between 5 km -15 km. Also low levels (z < 3 km) enhancements are greater immediately prior to maximum enhancements during phase 5 compared to immediately after phase 5. Starting in phase 7 negative enhancements return. Unlike negative enhancements in phase 0 – 3 maxima in negative enhancements are spread quite evenly through the depth of the troposphere.
Figure 6.8 – Percent echo cover per indicated phase of the MJO.
Figure 6.9 – (a) Percent cloud cover through each phase of the MJO, contour units are percent. (b) Anomalous cloud cover through each phase of the MJO relative to the mean cloud cover over all phases. Contour units are percent difference, solid (dashed) contours are positive (negative) enhancements.

6.4 Summary and Discussion

Collectively, the figures in this chapter provide concrete evidence to the validity of the schematics in figure 6.1a,b,c. In those schematics, cloud structure and evolution were inferred from thermodynamic, dynamic, and precipitation data. Here, actual cloud types and reflectivity structure were examined with CloudSat to discern cloud evolution through various MJO phases. Figures 6.5a,c and 6.7a,b,c confirm that shallow cloud types (i.e. stratocumulus and cumulus) occur more frequently during phases prior to peak cloud cover. Narrow deep convection occurs more prior to peak echo cover (red in Fig. 6.5a,c). High reflectivities above 0°C during phases 1 – 3 imply strong updrafts capable
of lofting high reflectivity particles (Fig. 6.7b,c,d). Though updrafts appear to be stronger prior to peak echo cover, the rainiest conditions occur during the most active phases as indicated by high reflectivities below 0°C (suggesting attenuating rain shafts) in Fig. 6.7e,f. Wide deep precipitation is also more common during these active phases (Fig. 6.5a,c). Following the active phases deep precipitation enhancements decrease, while the anvil coverage (yellow in Fig. 6.5a,c) and NCFAD enhancements are maximized (wedge in Fig. 6.7g,h,i). Deep precipitation differences prior to and during active phases are consistent with results of Morita et al. (2006), who showed that rainfall during suppressed phases occurs in deeper more intense clouds, while the mature stage of the MJO has more widespread rainfall and cloud cover. Anvil enhancements are consistent with results of Lin and Johnson (1996) (Fig. 6.1a) and Morita et al. (2006) (Fig. 6.1c).

In summary, there is an evolution of cloud type predominance of shallow clouds and cirrus mixed with deep, intense, but narrow convective systems during the suppressed phases, to widespread cloud coverage and rainfall during the active phases, that become more anvil dominated, and finally return to suppressed conditions.

As mentioned in the introduction the evolution of predominant cloud types over the life cycle of the MJO is similar to the life cycle of individual MSCs (cf. 6.1a,b,c and 6.1d). This resemblance of the MSC life cycle onto the MJO is perhaps explained by considering that the various parts of MCSs across large time and spatial scale events, such as the MJO, are successively being enhanced in the various phases of wave-envelopes, as in Fig 6.10, from Mapes et al. (2006). The top panel, an unfiltered view, depicts MCSs in various phases of the large-scale life cycle. Different proportions of
shallow, deep, and stratiform clouds are seen in the left, middle, and right MCSs. The bottom, filtered (smoothed or averaged) view, gives a blurred picture where the enhanced shallow convection in the leading (left) MCS appears to evolve to the enhanced stratiform cloud in the last (right) MCS. This apparent evolution arises because the underlying MCS sequence of evolution (composed of its “building blocks”) is “stretched” onto the larger scale. This view represents the stretched building block hypothesis, which says that the MCS life cycle is not literally mimicked in larger scale events, such that only middle-topped clouds would appear in the front of the event, deep clouds in the middle, and stratiform clouds in the rear. Rather, each phase (front, middle, back) of the large scale wave somehow enhances or favors a certain cloud type within the quasi-universal MCS life cycle (Mapes et al. 2006). The results of this chapter give credence to the stretched building block hypothesis.

**Figure 6.10** – Cartoon of the stretched building block hypothesis. (a) Unfiltered view. MCSs have varying amounts of shallow (left), deep (middle), and stratiform (right) clouds according to their position within a large-scale wave event. (b) Low-pass filtered view. When the mean is removed and smoothing broadens features, the resulting large-scale pattern resembles the MCS life cycle. Adapted from Mapes et al. 2006.
Chapter 7: Conclusions

CloudSat’s cloud profiling radar provides direct measurements of cloud vertical structure on the global scale (Stephens et al. 2002). This thesis examined global cloud climatology observed by CloudSat. Specific topics, spurred by the global climatology, where also examined and included: (1) mid-level clouds (i.e. Chapter 4), (2) stratocumulus clouds (i.e. Chapter 5), and (3) clouds associated with the MJO (i.e. Chapter 6). To establish the global climatology, cloudy echo portions of the CloudSat data were analyzed on two levels: (1) as EOs and (2) as the pixels comprising EOs. There are 6,650,125 EOs in our complete data set, 16 June 2006 to 31 October 2008. For each EO a plethora of attributes were recorded. Statistical analysis of EO’s attributes and the pixels comprising EOs were mainly viewed through the use of a variety of 2D and 1D histograms. The analysis of EOs and their pixels provided a fresh perspective on already heavily studied topics (e.g. cloud climatology, stratocumulus clouds, and the MJO), as well as, a novel observation of mid-level cloudiness.

Chapter 3 showed results of our global cloud climatology. EO types were defined by base and top height criteria for each latitudinal belt (i.e. tropics, sub-tropics, mid-latitudes, and high-latitudes). Cloud cover for all EOs and each EO type were examined, as well as, the seasonal cycle of global cloud cover. Pixel statistics by dBZ-height space were also studied for tropical (20°S – 20°N) EOs. The main results of chapter 3 are as follows:
• The Southern Hemisphere mid-latitudes (35° - 60°S) and the Northern Hemisphere storm tracks are the cloudiest locations on Earth, with secondary maxima along the ITCZ and over the west Pacific warm pool.

• Deep precipitation, cirrus, and cumulus congestus EO cloud cover are greatest in the mid-latitudes (35°- 60°N/S) and along the ITCZ and over the west Pacific warm pool.

• Anvil, altocumulus, and cumulus EO cloud cover are generally ubiquitous. Anvil and altocumulus cloud cover is between 5-10% over most of the globe (ignoring topography related maximums), except for over cool ocean waters where cloud cover is absent. Cumulus cloud cover is greater over ocean (~5%) than land (~2%) and is small over the world deserts.

• Stratocumulus cloud cover is between 35-45% over eastern ocean basins.

• El Nino and La Nina affects cloud cover. The first year of CloudSat (June 2006 – May 2007) was El Nino like, while the second year (June 2007 – May 2008) was La Nina like. There is an overall increase in cloud cover over the western Pacific, as well as, eastern Pacific (although, less so) and decrease in cloud cover the central Pacific during La Nina like conditions.

• Cloud cover in the Northern Hemisphere shows higher amplitude seasonal changes than the Southern Hemisphere. Cloud cover increases and extends equatorward during the Northern Hemisphere winters. Southern Hemisphere cloud cover remains fairly constant through the seasons, increasing slightly during the austral winter months and shifting slightly equatorward.
• Over the ITCZ, high cloud cover (tops > 10 km) shifts into the summertime hemisphere, while middle- and low-topped cloud cover (tops < 8 km) stays generally north of the equator.

• Tropical ocean cloud cover at middle- and low-levels (z < 8 km) is greater during nighttime CloudSat overpasses (0130 LT) and less at high-levels (z < 10 km). Tropical land cloud cover at middle- and high-levels (z > 3 km) is greater during nighttime CloudSat overpasses (0130 LT) and less at low-levels (z < 3 km). Caution should be taken in interpreting these results since CloudSat misses the peak in the diurnal cycle of deep convection.

Chapter 4 focused on mid-level topped clouds. Mid-level clouds are often obscured from IR and VIS satellites or surface observation by low and high clouds, respectively, so CloudSat offers the best chance at an unbiased documentation of their occurrence. The vertical distribution of EO top height in the tropics (20°S – 20°N) and EO top temperature in the extra-tropics (20 – 60°S/N) were examined. Main results from chapter 4 are as follows:

• There are two mid-level peaks in EO top height in the tropics, one at 5-6 km (0°C), the other at 7-8 km (-15°C – -20°C). While the 5-6 km peak is generally explained by increased stability at the melting level, the 7-8 km peak is unexpected and lacks an explanation.

• There is a small (yet distinct) peak between -15°C – -20°C in extratropical EO top temperatures, which corresponds to the 7-8 km peak in the tropics.
Both the tropical and extratropical peak at -15°C – -20°C is robustly sampled, occurring day and night, over land and ocean, over each ocean basin, and during both the summer and winter seasons.

With no prior explanation of the -15°C – -20°C peak, we offered several hypotheses ranging from an artifact of data processing to a feature of radar reflectivity detection to truly enhanced cloudiness at that level. No aspect of the cloud mask algorithm (Marchand et al. 2008) could produce a spurious sharp peak, making a data processing artifact unlikely. At -15°C the Bergeron-Findeisen process is very effective leading to rapid dendritic ice crystal growth. The radar may be sensing the large dendrites, while smaller liquid drops go undetected. Dynamical hypotheses include increased stability at 7-8 km leading to detrainment of cloud mass at that level, or gravity waves excited by melting at the 0°C isotherm in the tropics leading (i.e. “melting layer reverberations”, Mapes and Houze 1995). Future work may utilize CALIPSO data or the merged CloudSat, CALIPSO product to understand the existence of the mysterious -15°C – -20°C peak.

Chapter 5 looked at stratocumulus clouds. Stratocumulus EOs are defined as EOs with top height below 4.5 km with a width greater than 10 km. In hindsight, a 10 pixel width threshold is rather crude to distinguish stratocumulus from cumulus clouds. Studying low-clouds, collectively, is perhaps more reasonable. Nevertheless, results from chapter 5 are interesting and are as follows:

- Our stratocumulus definition identifies the main stratocumulus decks (i.e. Californian, Peruvian, Namibian, and Australian).
• Low EO cover is similar to ISCCP values (~ 10% - 20% less) and quite lower than CALIPSO values (~ 30% - 40% less).

• Average stratocumulus EO top heights are slightly higher than MODIS derived cloud top heights.

• The Peruvian and Namibian have more echo cover in SON 2007 vs. 2006. The year-to-year differences mainly reflect changes in SSTs (La Nina) and static stability.

• Most stratocumulus EOs contain at least one drizzling pixel (dBZ > -17) (90%), while a little more than half (55%) contain a raining pixel (dBZ > 0)

  Future work on stratocumulus could explore NCEP reanalysis variables in more detail to get a better understanding of the year-to-year differences in echo cover. One specific idea is to look at the strength of the sub-tropical highs in SON 2006 vs. 2007.

Chapter 6 characterizes cloud types across various phases of the MJO. Phases were defined using filtered MJO OLR from 15°S – 15°N. The main results of chapter 6 are consistent with previous studies and are as follows:

• There is a doubling (tripling) of cloud area (volume) from suppressed to active MJO phases.

• There is an evolution of cloud type *predominance* of shallow clouds and cirrus mixed with deep, intense, but narrow convective systems during suppressed phases, to widespread cloud coverage and rainfall during active phases that become more anvil dominated, and finally return to suppressed conditions.

  Future work will examine cloud evolution across other tropical waves (e.g. Kelvin, mixed Rossby-gravity) using the same methods as developed in chapter 6.
In summary, this thesis provided a broad brush of several different topics in meteorology from the perspective of the first space based cloud radar, CloudSat. Perhaps the best attribute of this thesis is that it provided a jumping off point for several interesting “finds” that were not possible before the launch of CloudSat or creation of our EO-based data set. “Finds” such as the mysterious -15°C peak in the tropics and extratropics and the difference in stratocumulus cloud cover in SON 2006 vs. 2007 should be looked at in greater detail in the future. As CloudSat continues to orbit, we (as a scientific community) will continue to process data, and strive for a deeper understanding of meteorology.
References


