Thirteen times during the RICO (Rain In Cumulus over the Ocean) field campaign, MISR (Multiangle Imaging SpectroRadiometer) onboard the EOS Terra Spacecraft and NCAR’s S-PolKa radar collected coincident high-resolution visible and near-IR satellite data and dual-polarization S-band radar reflectivity data to understand trade wind cumuli cloud distribution and precipitation. Using a Z-R relationship derived from optical array probe data from the NCAR C-130, rainfall characteristics of these clouds were examined by collocating the radar reflectivity and rain-rate with the satellite reflectance at the pixel level.

Collocating these data has offered a unique perspective into the macrophysical characteristics of trade wind clouds and this thesis will focus on the role these clouds play in the water cycle of the trade wind regime in the Western Tropical Atlantic. First, through a latent heat flux analysis, it was found that these clouds are between 17-21% efficient at returning evaporated water to the ocean through precipitation. Also the area averaged rain rate of the collocated data was 1.30 mm day\(^{-1}\). Between 5-10% of the cloudy area observed by MISR was responsible for this rainfall and it was discovered that the mesoscale organization of this shallow convection held a strong relationship with precipitation intensity and distribution.

To demonstrate the representativeness of these thirteen coincident scenes, rainfall statistics from the entire field campaign (62 days) were derived. Daily variations in rainfall ranged from days with no rain to days with over 20 mm of rainfall. Eight of the thirteen coincident scenes were found to be from driest half of the days. Within each day a strong diurnal cycle with a pre-dawn peak in the rainfall was observed. The area averaged rain rate for the entire project was 2.26 mm day\(^{-1}\) and the latent heat flux was 51 W m\(^{-2}\). Therefore, clouds observed during RICO were 26-32% efficient at returning water to the ocean.
TABLE OF CONTENTS

CHAPTER                                      PAGE

1. INTRODUCTION_______________________________4

2. MISR AND SPoLKa OVERVIEW________________________12
   2.1. MISR and Cloud Detection_________________12
   2.2. Description of the SPoLKa Radar and the Derivation of Rain Rate___________14

3. PREPROCESSING OF THE RADAR DATASET_______________23
   3.1. Radar Data Description____________________23
   3.2. Noise Filter, Island Removal and Bad Data Removal_________________24
   3.3. Bird Filtering____________________________27
   3.4. Selecting a Rain-No-Rain Threshold________________________30

4. COLLOCATING THIRTEEN COINCIDENT MISR AND SPoLKa SCENES____34
   4.1. Overview_______________________________34
   4.2. Geolocating Each Radar Pixel____________________36
   4.3. Errors in Geolocation______________________44
   4.4. Adjusting for Time Differences____________________45
   4.5. Re-sampling to the MISR Grid____________________47
   4.6. Resolution Differences and Radar Beam Geometry_______________50

5. CHARACTERISTICS OF THE COLLOCATED DATASET________54
   5.1. MISR Cloud Fraction Observations________________________54
   5.2. Cloud Top Height Analysis________________________55
   5.3. Distribution of Reflectivity for Clear and Cloudy Pixels_______________57
Over warmer ocean waters, (i.e., east of major continents) the vast expanse of tropical ocean is populated with shallow cumulus. These clouds are the archetypical form of moist convection and play an integral part in the maintenance of tropical circulations (Stevens, 2005). An excellent explanation of the role shallow convection plays in tropical circulations was provided in Emanuel (1988). He divides the Hadley cell into four regions corresponding to (1) the deep convection of the Inter-Tropical Convergence Zone (ITCZ), (2) the upper air subsidence over the subtropical highs, (3) the trade wind cumulus layer, and (4) the sub cloud layer (Fig. 1). Shallow convection in this model is confined to the well mixed trade wind boundary layer, which is usually capped by a temperature inversion and/or sharp decrease in dew point temperature at the base of the free atmosphere (Stevens, 2005). Clouds forming in this layer cover between 10-25% of the ocean between 30°N – 30°S. (Siebesma et al., 2003; Stevens, 2005; Zhao, 2006)

Shallow convection has two main roles in the global tropical circulation, (1) it moistens and deepens the cloudy layer for later ingestion by the ITCZ, and (2) it offsets the compressional warming from the large scale subsidence by radiational cooling at cloud top. These two mechanisms prime the air in the boundary layer to fuel the convection of the ITCZ and are key components in maintaining the Hadley cell’s heat and moisture budgets.

Modeling studies of shallow convection have focused primarily on entrainment and detrainment and the role the clouds play as an evaporator and/or reflector of insolation (Betts, 1975; Bechtold, 1998; Neggers et al., 2003; Siebesma et al., 2003; Stevens, 2005). Proper
representation of these boundary layer clouds is key to more accurate prediction of climate; however they remain poorly treated in climate models. For example, Bony and Dufresne (2005) ran fifteen coupled climate models and reported that, “The representation of convective and boundary-layer processes, in addition to the parameterization of cloud properties, is known to be critical for the prediction of the cloud’s response to climate change, and it differs widely among models.” Furthermore, model simulations are often validated using case studies, which Siebesma et al., (2003) recognized may be the wrong approach because simulations of limited observations may not be representative of nominal shallow cumulus fields.

Few attempts have been made to estimate shallow convective cloud’s climatological prevalence on regional and seasonal scales (Petty, 1999), and in the past, precipitation from these clouds was often neglected (Emanuel, 1988; Stevens, 2005). Currently, precipitation from shallow convection is measured/reported with large variance and uncertainty (e.g., Fig. 3 of Adler et al., 2003). This is expected given the many difficulties in measuring precipitation on various time and space scales. Fortunately, this shortcoming is slowly being overcome as the
work of Short and Nakamura (2000) and Petty (1999), as well as the efforts of the GPCP (Global Precipitation Climatology Project), the launch of TRMM (Tropical Rainfall Measurement Mission) and the RICO (Rain In Cumulus over the Ocean) field campaign have all attempted to measure precipitation from shallow convection.

Determining shallow convection’s contribution to the nearly two-thirds of the world’s rainfall that falls in the tropics is important for proper representation of the global precipitation amount and distribution (Rauber et al., 1996). Localized precipitation amounts have come from a handful of field campaigns scattered about the tropical oceans during various seasons (see Short and Nakamura (2000) for a recent review). One of the earliest reports was from Battan and Brahm (1956) who stated that precipitation from cumulus congestus over the ocean was not rare, in fact, over half the clouds whose tops reached 3 km ASL were reported to have precipitation. Also, Soong and Ogura (1976) found that clouds with tops below 2 km in the Western Caribbean Sea had a precipitation rate of 0.29 mm day\(^{-1}\). More recently, in the Hawaiian Rainband Project, Rauber et al., (1996) used ground based radar from the northeastern shore of Hawaii to measure an area averaged precipitation rate of 1.1 mm day\(^{-1}\) during July and August of 1990. Furthermore, Lui et al., (1995) used microwave and infrared measurements from satellites over the ocean to find that 14% of precipitating clouds had warm cloud tops which contributed 4% to the total rainfall. Petty (1999) combined surface synoptic reports with GOES IR (infrared) imagery over the Pacific Ocean east of Australia and reported that 20-40% of non-drizzle precipitation was from warm clouds. Short and Nakamura (2000) claim the constant background of rain from shallow clouds over the tropical oceans accounts for 22% of total rainfall in winter. From a global perspective, the creation of the GPCP has provided global precipitation maps by combining rain gauge data, ground based radar and satellite measurements to produce standard
global precipitation maps (Huffman et al., 1997). For the winter 2004-2005 over the Western Tropical Atlantic, which was the focus region of this study and a place where shallow convection were ubiquitous, the GPCP reported an average rain rate of 1.2 mm day\(^{-1}\). Finally, the latest advancement in precipitation measurement is TRMM, which houses the first active spaceborne precipitation radar. During the winter 2004-2005, TRMM observed an average rain rate of 1.05 mm day\(^{-1}\) over the Western Tropical Atlantic.

Even with these advancements, large uncertainty surrounds the true precipitation rates of shallow convection. Precipitation from these clouds is difficult to measure for several reasons. First, the clouds are small; typical cloud width ranges from a few hundred meters to a few kilometers (Wielicki and Welsh, 1986; Wielicki and Parker, 1992; Zhao, 2006). Second, they are warm and shallow; cloud top temperatures are well above 273K and most cloud-top heights are capped by the inversion near 1.5-2.5 km ASL (Stevens, 2005). Furthermore, these clouds reside over the global tropical oceans; a place where rain gauges are extremely scarce and ground-based radars are rare. Therefore, with these limitations the only platform remaining with the spatial coverage and temporal resolution necessary to measure precipitation over the ocean is satellite.

Satellites use a variety of techniques to measure precipitation and make use of several regions in the electromagnetic spectrum. The first and most rudimentary method is to take a thermal infrared brightness temperature measurement from the cloud-top and associate it with a rain rate. This technique was developed during the GATE field campaign in 1979 by Philip Arkin, who found that thermal infrared GOES data was highly correlated with the fraction of the radar area associated with rain (Kidder and Vonder Haar, 1995). This technique has been named the \textit{GOES Precipitation Index} (GPI) and is still used today. However, this method fails in regions of shallow convection for several reasons. First, this technique uses a minimum threshold
of 235K, which is far too cold and will miss virtually all shallow clouds. Second, the horizontal resolution of the 11 μm GOES channel (thermal IR) is 4 km and, with most shallow convective clouds having a diameter less than a few kilometers, too much of the background ocean is averaged in with the cloud in 11 μm radiance measurement. Third, the indirectness of this method, associating a cloud-top temperature with precipitation, was based on a weak relationship between cloud and rain, and large errors are incurred in scenes with multiple clouds layers or cirrus contamination (Arkin and Arduany, 1989).

A second more direct technique employs microwave frequencies. Passive microwave sensors, at frequencies between 10 – 90 GHz (3 cm – 3 mm wavelength), are sensitive to precipitation-sized hydrometeors. Over ocean, the surface emmisivity is relatively constant and therefore attenuation to the expected microwave signal from the ocean can be associated with rain. For low frequency microwave energy (i.e., 10 – 20 GHz), liquid hydrometeors, in conjunction with some gaseous attenuation, absorb the outgoing microwave radiation. Liquid water drops absorb this energy as the cube of their radius and therefore through an inverse relationship, the microwave brightness temperature is related to the intensity of the precipitation (Arkin and Arduany, 1989). Above about 60 GHz, scattering of the microwave signal by ice particles is dominant which provides a useful method of discriminating warm from cold rain. Despite the directness of this technique, using passive microwave signals to find precipitation in shallow convection is very difficult. The instantaneous field of view of even the newest microwave sensors is still several kilometers by several kilometers. Therefore, the largest control on the microwave brightness temperature is not the attenuation due to rain, but the degree to which the rain showers fill the footprint of the sensor (Spenser et al., 1983). The result is that precipitation signatures are contaminated by clear ocean background.
The third technique is active spaceborne precipitation radar. The TRMM satellite was launched in November 1997 and was the first and only satellite to date to carry a 13.8 GHz (2.17 cm) precipitation radar (PR). In a low-altitude (~350 km), low-inclination orbit the PR detects precipitation fields with 250 m vertical resolution and 4.3 km horizontal resolution at nadir. Despite the large improvement in resolution over the previous techniques, the PR signal suffers from strong attenuation and therefore has rather low sensitivity and can only detect echoes of 17 dBz and higher (Schumacher and Houze, 2000). Regardless, the PR has offered a substantial improvement over other techniques and is detecting large regional differences in precipitation when compared to the other studies (Adler et al., 2000). TRMM also carries the TMI (TRMM Microwave Imager) which measures at the microwave frequencies of 10.7, 19.4, 21.3, 37 and 85.5 GHz. TRMM also houses VIRS (Visible and InfraRed Scanner) which makes measurements using five channels ranging from 0.63 to 12 microns. The greatest advantage of carrying such a wide variety of instruments is that TRMM can use each to help tune the other’s precipitation measurement to arrive at a best estimate for rainfall and to compare where one method works better over another (e.g., Adler et al., 2000; Ikai and Nakamura, 2003; Nesbitt et al., 2004). Furthermore, TRMM has undergone several validation studies, where its precipitation products are compared to ground based radar and rain gauge measurements (e.g., Schumacher and Houze, 2000). However, shallow convection still manages to cause problems for the PR. For example, Schumacher and Houze (2000) found that at the Kwajalein validation site, the PR missed nearly 46% of the rain area. This large fraction was attributed to the weaker echo from shallow convection and shallow stratiform precipitation that has reflectivity below the PR’s noise level. However, the contribution of drizzle and light rain to the total near-surface rainfall was small and the PR therefore only missed 2.3% of total rainfall (Schumacher and Houze, 2000).
The difficulties in measuring precipitation from shallow convection at appropriate time and space scales are numerous, and the global presence of these clouds requires a better understanding. From November 2004 through January 2005, the RICO field campaign set out to, “…characterize and understand the properties of trade wind cumulus at all scales, with particular emphasis on determining the importance of precipitation” (RICO Scientific Overview Document). With this goal, the NCAR SPolKa radar was shipped to the island of Barbuda, on the apex of the Lesser Antilles, to sample precipitation from the undisturbed trade wind regime of the Western Tropical Atlantic. For 62 days SPolKa’s 10 cm (S-Band), dual polarization radar collected data with minimal interruption, providing a sizeable sample of wintertime trade wind precipitation.

Having a radar dedicated to measuring precipitation in this region for two months was ideal for sampling the shallow convective precipitation of the Western Tropical Atlantic, but still did not solve the problem of measuring precipitation over the tropical oceans. As previously discussed, satellites are currently the best platform to resolve the time and space issues involved with global precipitation measurement. In this thesis, characteristics of trade wind clouds and precipitation were investigated using data from RICO. This study uses satellite data to understand the properties of precipitating shallow convective clouds. Instead of using satellite to measure rain, we used it to detect the clouds that were raining and then draw relationships between the cloud field and its associated rain. To accomplish this, high horizontal resolution (275 m) MISR (Multiangle Imaging SpectroRadiometer) data has been collocated with SPolKa radar data to gain insight into the precipitation characteristics of trade wind clouds. This approach allows for characterization of the cloud and precipitation fields observed during RICO by building relationships between the MISR observed cloud field and the SPolKa radar observed
precipitation field. The goal in collocating these data was to apply the relationships to the rest of
the radar data taken during RICO as well as other MISR observed trade wind cloud fields. The
second part of this thesis reports an analysis of the radar data from the entire field campaign.
This work establishes a project-wide area average rain rate, discovers the diurnal cycle in
precipitation and derives a latent heat flux to the atmosphere due to precipitation. These statistics
help put into perspective the coincident radar and satellite data and provide characteristics of
shallow convective precipitation during the winter over the Western Tropical Atlantic.
2.1. MISR and Cloud Detection

The Multiangle Imaging SpectroRadiometer (MISR) is onboard the EOS Terra spacecraft that was launched in December 1999, and is the first instrument to view the Earth at nine separate view angles in the along track direction (Fig. 2.1.1.). Using four forward viewing, one nadir viewing and four aft viewing cameras, MISR makes high resolution (275 m) radiance observations using three visible channels (RGB) and one near-infrared channel (0.86 $\mu$m) that are reported on a Space Oblique Mercator (SOM) map projection. In its sun-synchronous descending node, Terra has an equator crossing time around 10:30 AM local time (LT). From an altitude of 705 km, MISR takes radiance measurements in its 360 km swath using a “push-broom” style scan and records 16-bit data from four CCD (Charged Coupled Device) arrays. MISR is an ideal instrument to study clouds because it samples in the visible and near infrared (NIR) with high spatial resolution, excellent bit-depth, and stereoscopic viewing capabilities.

During RICO, MISR passed within the “RICO Region”, bound between 12 – 20N and 55 – 65W (Fig. 2.1.2.), sixteen times on four different Terra paths. Of these sixteen scenes, thirteen were used in the following analysis. Three of the scenes were removed due to lack of coincident radar data. To distinguish clear sky from cloud, the near-infrared channel (NIR) bi-directional reflectance factor (BRF) was used to make a cloud mask for each image. The NIR channel is best suited for cloud detection over the ocean due to the weak reflection by the ocean surface which causes clouds to stand out against the dark ocean background. The BRF or reflectance is
Fig. 2.1.1. Computer generated image of MISR onboard EOS Terra. (http://www-misr.jpl.nasa.gov/mission/miview1.html) Image courtesy of Shigeru Suzuki and Eric M. De Jong, Solar System Visualization Project. JPL image P-49081

Fig. 2.1.2. The RICO Region and SPolKa Radar domain
calculated for a given channel by dividing the radiance observed at a given sun-view geometry by the top-of-atmosphere irradiance observed at the same sun-view geometry.

Using a static manually derived NIR BRF threshold to classify a given pixel as clear or cloudy is subjective and time consuming, but arguably the best procedure for distinguishing clear and cloudy pixels. Therefore in each of the thirteen scenes, a single NIR BRF threshold was chosen and independently verified by experts (Dr. Guangyu Zhao and Yuekui Yang). Based on these visual inspections, sunglint was minimal over the RICO region and time period and not considered a major contaminant. However, very low BRF thresholds, those considered to be the most cloud conservative, may possibly suffer from contamination due to 3-D radiative effects, such as photons exiting cloud sides and scattering off of aerosols and air molecules toward the MISR instrument. Furthermore, as Wielicki and Welch (1986) point out, even at high resolution, sampling a field of small cumulus is subject to errors in cloud fraction due to optically thin cloud edge pixels as well as partially cloud filled pixels. Moreover, reflectance within a cumulus cloud is highly variable which is attributable to bright turrets and dark shadows and accounts for the lack of a distinguishable cloudy mode in Fig. 2.1.3. Due to these limitations, a range of NIR BRF thresholds are used to represent the possible variations in cloud fraction. The end result is that each pixel in each scene has been classified as clear or cloudy at high spatial resolution and therefore, when collocated with the radar data, precipitation can be identified with cloudy area.

2.2. Description of the SPolKa Radar and the Derivation of Rain Rate

The dual wavelength, dual polarization NCAR SPolKa radar (Fig. 2.2.1.) was located on the island of Barbuda (17.608°N, 61.824°W) for the duration of the RICO field campaign. Using
Fig. 2.1.3. Combined histogram of BRF values from the thirteen coincident MISR scenes. Note the lack of a strong clear or cloudy mode. (cf. Fig. 3 Wielicki and Welch (1986))

Fig. 2.2.1. The NCAR SPolKa radar on Barbuda.
two dual polarization radar systems, one S-band (10.62 cm) and one Ka-band (9 mm), SPolKa collected more than 200,000 constant elevation scans over 62 days. SPolKa used two different scan types; the PPI (Plan Position Indicator) which scans less than 360° at various elevation angles and the SUR (Surveillance) which scans 360° at 0.5° in elevation. Fig. 2.2.2. shows examples for each of these scan types. For the PPI and SUR scan types, the primary scan strategy changed according to the wind direction, so that the start of each scan pointed into the wind and ended pointing with the wind (RICO Scientific Overview Document). Once the hemisphere was chosen, a volume of data was swept out over several elevation angles ranging from 0.5° to 16.8°. Only PPI and SUR scans from the S-band system at an elevation angle of 0.5° were used in this study, of which there were over 26,000 scans.

SPolKa (hereafter “radar”) collected data out to a range of roughly 147 km using a beam width and height of 0.92° and a range gate spacing of 149.89 m. The radar recorded a suite of standard and polarimetric variables (Table 2.2.1.) that are available from the NCAR ATD Data Retrieval website (http://www.eol.ucar.edu/rdp/mss_retrieval/). For this study the reflected power, reflectivity factor, differential reflectivity and raw Doppler velocities were used.

Table 2.2.1. A list of name, description and unit for the nine-variable SPolKa radar dataset (* denotes variables used in the study).

<table>
<thead>
<tr>
<th>Name</th>
<th>Description</th>
<th>Unit</th>
</tr>
</thead>
<tbody>
<tr>
<td>DM*</td>
<td>Reflected Power</td>
<td>dBm</td>
</tr>
<tr>
<td>DZ*</td>
<td>Reflectivity Factor</td>
<td>dBZ</td>
</tr>
<tr>
<td>DX</td>
<td>Cross-pole Reflected Power</td>
<td>dBm</td>
</tr>
<tr>
<td>ZDR*</td>
<td>Differential Reflectivity</td>
<td>dBm</td>
</tr>
<tr>
<td>PHIDP</td>
<td>Differential Phase</td>
<td>Deg</td>
</tr>
<tr>
<td>NCP</td>
<td>Normalized Coherent Power</td>
<td>None</td>
</tr>
<tr>
<td>LDR</td>
<td>Linear Depolarization Ratio</td>
<td>dBm</td>
</tr>
<tr>
<td>VE*</td>
<td>Raw Doppler Velocities</td>
<td>meters/second</td>
</tr>
<tr>
<td>SW</td>
<td>Spectral Width</td>
<td>meters/second</td>
</tr>
</tbody>
</table>
Fig. 2.2.2. Top) Typical PPI (Plan Position Indicator) scan; Bottom) Typical SUR (Surveillance) scan from the SPolKa radar.
Chapter 3 will cover the preprocessing of the radar data. A few basics on how S-band radars remotely sense precipitation is given below. All references in this paper to rain rate, rainfall or rain intensity were those derived from the radar reflectivity (Z).

A plane microwave, incident on a solitary target (i.e., a spherical water drop), can interact with that target through absorption, transmission and/or scattering. The amount of backscattered energy from this interaction is a function of the size, shape, complex dielectric constant and viewing aspect of the target and is characterized by the radar cross section given in Eq. 2.2.1.,

\[ \sigma = \frac{\pi^5 |K|^2 D^6}{\lambda^4} \]  

(2.2.1.)

where, \( K \) is the complex refractive index of water (proportional to the induced electric dipole moment), \( D \) is the spherical diameter and \( \lambda \) is the wavelength of the transmitted pulse. The radar cross section is equivalent the ratio of back scattered energy to incident energy at a given distance from the radar. For a volume containing a distribution of particles, the reflected power to the radar from these particles is a function of the peak power transmitted from the radar, antenna gain, wavelength of transmitted pulse, average radar cross section and range. Reflected power is measured by the antenna as a power flux density and converted into decibels (dBm) (Doviak and Zrnić, 1984). The reflected power field from the SPolKa radar was used to determine the minimum detectable signal the antenna can detect above the instrument noise.

For precipitation sized particles, those ranging from 0.1 mm to 1.0+ mm, Rayleigh scattering by S-band microwave energy dominates the reflected power to the antenna. The reflectivity for a given volume of spherical particles is equivalent to the sum of the sixth power of the particle diameters, \( D \), divided by the contributing volume \( V \), over which they reside and is given as Eq. 2.2.2.
Under the Rayleigh scattering assumption, the reflectivity is a representation of the backscattered energy from the average radar cross section of randomly distributed particles within a contributing volume. $Z$'s dependence on the sixth power of the diameter makes it cumbersome and therefore a logarithmic scale is most commonly used as presented in Eq. 2.2.3.

$$Z(dBZ) = 10\log_{10} \left( \frac{Z}{1(mm^6 m^{-3})} \right)$$

(2.2.3.)

The relationship between $Z$ and $D$ has numerous meteorological implications. Precipitation rate, $R$, is defined as the volume of precipitation passing through a horizontal surface per unit time and area and is equal to the sum of all drop diameters, $D_j$, multiplied by their individual fall speeds, $w_j$, in Eq. 2.2.4. (Doviak and Zrnić, 1984). In Eq. 2.2.4., $V$ is the contributing volume over which these drops are observed and the units are often converted to mm hr$^{-1}$. To develop a relationship between $Z$ and $R$, knowledge of the drop size distribution is needed. Size distribution measurements are obtained through a variety of methods including drop collection from ground sampling stations and optical array probes attached to aircraft. Once the distribution is known, calculations of $Z$ and $R$ can be made and graphed and the equation of the line that describes the relationship between each variable can be applied to a measured $Z$ to obtain an $R$. The $Z$-$R$ relationship is typically written in the form of the power law $Z = aR^b$, where coefficient $a$ and exponent $b$ are determined by the shape of the line. For this study, a $Z$-$R$ relationship was developed using the optical array probe data from the 2DC and 2DP probes flown on the NCAR
C-130. The 2DC probe measured drops over a size range of 30 μm to 1800 μm and the 2DP probe measure drops over a size range of 100 μm to 3200 μm and together they sample nearly all precipitation sized particles. Fig. 2.2.3. shows the drop spectra obtained from this data. Hilary Minor hand edited the drop spectra to remove out of focus drops, splattered drops, split drops, or drops that could not have their radius measured. Matt Freer and Sabine Goeke calculated \( Z \) and \( R \) and Fig. 2.2.4. shows the least squares fit line for the calculated \( Z \) and \( R \). The correlation coefficient for this data series was 0.98, suggesting that the fit was very good and represents a large range of rain rates. Table 2.2.2. shows a series of other \( Z-R \) relationships used for shallow convection.

*Table 2.2.2. The RICO and other warm rain oceanic \( Z-R \) relationships.*

<table>
<thead>
<tr>
<th>Paper</th>
<th>( Z-R )</th>
</tr>
</thead>
<tbody>
<tr>
<td>RICO 2006</td>
<td>( Z=88.7R^{1.52} )</td>
</tr>
<tr>
<td>Iguchi et al., 2000</td>
<td>( Z=145R^{1.55} )</td>
</tr>
<tr>
<td>Houze et al., 2004</td>
<td>( Z=175R^{1.5} )</td>
</tr>
<tr>
<td>Rauber et al., 1996</td>
<td>( Z=260R^{1.7} )</td>
</tr>
<tr>
<td>Stout and Mueller, 1968</td>
<td>( Z=126R^{1.87} )</td>
</tr>
<tr>
<td>TRMM Shallow Convection</td>
<td>( Z=148R^{1.75} )</td>
</tr>
</tbody>
</table>
Fig. 2.2.3. Droplet spectra from the 2DP probe (red) and the 2DC probe (blue) used to develop the RICO Z-R relationship.
Fig. 2.2.4. Z-R relationship calculated from the droplet spectra of the 2DP and 2DC probes (blue). Also graphed are the Z-R relationships from Table 2.2.2.
CHAPTER 3
PREPROCESSING OF THE RADAR DATA

3.1. Radar Data Description

Radar data collected during RICO were stored in the “sweep” file format and converted to NetCDF for processing. One file was made for each PPI or SUR scan and file sizes typically ranged from 3MB to 15MB depending on the area swept out by the radar. The header of each NetCDF file contains the names and attributes of all the radar variables and the body of each NetCDF file contains the variable arrays. Table 3.1.1. describes the data structure of the variables that were used in this study. Several radar variables were constant, like the beam width and range gate spacing, which is the width of the pulse at half power and the pulse length in the radial direction, respectively. Other variables were one dimensional such as the azimuth angle which was only a function of the number of beams. Finally variables such as reflectivity factor and reflected power were two dimensional and were a function of both the number of beams and the number of range gates.

Table 3.1.1. Radar variables used in this study.

<table>
<thead>
<tr>
<th>Name</th>
<th>Type</th>
<th>Dimensions</th>
<th>Units</th>
</tr>
</thead>
<tbody>
<tr>
<td>Time</td>
<td>Int.</td>
<td># of beams</td>
<td>Second from 1970-01-01 00:00</td>
</tr>
<tr>
<td>Azimuth angle</td>
<td>Real</td>
<td># of beams</td>
<td>Degrees</td>
</tr>
<tr>
<td>Elevation angle</td>
<td>Real</td>
<td># of beams</td>
<td>Degrees</td>
</tr>
<tr>
<td>Doppler Velocity (VE)</td>
<td>Int.</td>
<td>Beams*Gates</td>
<td>Meters/second</td>
</tr>
<tr>
<td>Reflected Power (DM)</td>
<td>Int.</td>
<td>Beams*Gates</td>
<td>dBM</td>
</tr>
<tr>
<td>Differential Reflectivity (ZDR)</td>
<td>Int.</td>
<td>Beams*Gates</td>
<td>dBm</td>
</tr>
<tr>
<td>Reflectivity Factor (DZ)</td>
<td>Int.</td>
<td>Beams*Gates</td>
<td>dBZ</td>
</tr>
</tbody>
</table>
3.2. Noise Filter, Island Removal and Bad Data Removal

Since the radar data were used to understand precipitation rates, filtering of unwanted signal was necessary so as to not bias the statistics. Noise in the radar system was created by background electronic interference and represented the level at which a signal can no longer be distinguished. The power (dBm) at which the signal was not recognizable over noise was determined using the “Reflected Power” field (DM).

The goal of using a noise filter was to preserve coherent echo and remove system noise. A coherent echo is one that can be assigned a Doppler velocity (Doviak and Zrnić, 1984). The Doppler velocity is a measure of the contributing volume’s average movement toward or away from the radar in the radial direction. Particle motion in the radial direction produces a frequency shift in the backscattered energy, from which its radial velocity (m s$^{-1}$) can be derived. For incoherent echoes, a random Doppler velocity is reported (Doviak and Zrnić, 1984). These random velocities stand out in the VE field as wildly varying velocities from pixel to pixel. Any reflectivity reported below the noise threshold was generated by system noise which cannot be assigned a coherent Doppler velocity. When selecting a noise threshold, the goal was to preserve all coherent velocities and remove all incoherent velocities. Using this as a guide; a noise threshold of -115.1 dBm was selected. Fig. 3.2.1. is a combined histogram of DM for all 0.5° PPI and SUR radar files from the entire project and it shows a clear break from the Gaussian distribution near -115 dBm. The tail of this distribution larger than -115 dBm was attributable to the coherent signal from clear-air and precipitation echoes, which are discussed in detail in Section 3.4. Also from this figure, it is clear that the vast majority of the radar images were occupied by noise, which was expected given the small cloud fraction and low rain rates reported in Chapter 1.
Fig. 3.2.1. Histogram of Power Return (DM) in dBm for all 0.5° PPI and SUR files. The break in the distribution near -115 dBm signifies the minimum power to distinguish a coherent signal.

Other unwanted returns such as island ground clutter, sun spikes and other contaminates were removed using the radar data editing software, SOLOII. SOLOII is a UNIX based radar data editing software package that allows the user to interact and change radar data in its original format. Using SOLOII, boundaries were drawn around the islands of Barbuda, Antigua, St Kitts and Nevis, Anguilla, Montserrat and Guadeloupe to remove the ground clutter they produced. Ground clutter from these islands occupied less than 2% of the total radar area for the SUR scans and less than 0.5% for the PPI scans in the northernmost hemisphere. Fig. 3.2.2. illustrates the effects of removing the island clutter, a bad beam and noise.
Fig. 3.2.2. Top Left) raw unfiltered Doppler velocities (m s\(^{-1}\)); Top Right) unfiltered Reflectivity Factor (dBZ) with islands and bad beam removed; Bottom Left) raw Doppler velocities (m s\(^{-1}\)) with noise removed; Bottom Right) Reflectivity Factor (dBZ) with noise, islands and a bad beam removed. Notice the large amount of clear air echo close to the radar in the bottom images.
3.3. Bird Filtering

Frigate birds (*Fregata magnificens*) or “man-o-war birds” were a common feature in the first 40 km of the radar data. A colony, 10,000+ strong, resides on the northwestern side of Barbuda at the tip of the lagoon. Frigate birds glide around on rising thermals and convective rolls and appear on the radar image as long lines or clusters of very high reflectivity (Fig. 3.3.1.). The high reflectivity was attributable to a large radar cross section that results from their seven foot wingspan. The main problem these birds presented was in their interpretation in the radar data as precipitation. Precipitation in trade wind clouds was often oriented along lines and clusters that can be easily confused with the birds. Therefore they had to be removed from the data.

Fortunately, because the Frigate birds don’t always fly passively with the wind, their flight directions were typically different from the background flow making them stand out in raw Doppler velocity field (Fig. 3.3.1.). Furthermore, Frigate birds produce a strong signature in the polarmetric variable $Z_{DR}$ (Differential Reflectivity) because of their large wingspan. Using these noticeable signals in the radar data, an algorithm was developed to eliminate most of the echo created by the birds.

Due to the large coverage of coherent signal in clear air near the radar (discussed in detail in Section 3.4.), the background wind had a smoothly transitioning appearance in the Doppler velocity field. Frigate birds stood out from the background wind field with highly variable coherent velocities from pixel to pixel. This distinction was used to flag birds in the data. The algorithm examined a pixel in the array of Doppler velocities and compared its velocity to its closest 120 neighbors. If the velocity of the target pixel was more than 4 m s$^{-1}$ different from the
Fig. 3.3.1. a) Radar Reflectivity Factor (dBZ) centered just north of the island of Barbuda; b) Raw Doppler velocities (m s⁻¹) centered just north of the island of Barbuda. Notice the distinct high reflectivity lines crossing image (a) from NE to SW and their correspondence to the highly variable velocities in image (b).
median velocity of its 120 neighbors then it was flagged as bird. In the smoothly transitioning radial velocity field like that shown in Fig. 3.2.2., a single pixel that had a coherent Doppler velocity that was 4 m s\(^{-1}\) higher or lower than its neighbors was clearly not from hydrometeors. It was crucial that this filter target unwanted signal only and upon visual inspection, precipitation signatures and coherent clear air echo were found to be untouched.

The second bird filter used the differential reflectivity variable, \(Z_{\text{DR}}\), which is the ratio of the horizontally polarized reflectivity factor, \(Z_{\text{HH}}\), to the vertically polarized reflectivity factor, \(Z_{\text{VV}}\), and is given in Eq. 3.3.1.

\[
Z_{\text{DR}} = 10 \log_{10} \left( \frac{Z_{\text{HH}}}{Z_{\text{VV}}} \right)
\]

(3.3.1)

For hydrometeors, \(Z_{\text{DR}}\) is often used to categorize the scatterer’s shape because of its dependence on the axis ratio of the scattering particles in the contributing volume (Straka et al, 2000). For a particle that is perfectly spheroid, the ratio of the horizontal axis to the vertical axis is one and its \(Z_{\text{DR}}\) value is zero. A prolate spheroid has a larger vertical axis, larger \(Z_{\text{VV}}\) and a \(Z_{\text{DR}}\) value less than zero while an oblate spheroid, has a larger horizontal axis, larger \(Z_{\text{HH}}\) and a \(Z_{\text{DR}}\) value greater then zero. \(Z_{\text{DR}}\) is also sensitive to the size of the scattering particles and therefore, \(Z_{\text{DR}}\) values from a contributing volume that contains Frigate birds will most likely be representative of the birds, not other scatterers in the volume. Using this to discriminate birds, two thresholds were chosen to filter out all \(Z_{\text{DR}}\) values that were not possibly made from liquid hydrometeors. According to Straka et al., (2000), liquid hydrometeors typically have a \(Z_{\text{DR}}\) range of -1 to 3. Because most birds had a very large \(Z_{\text{DR}}\) signature upon visual inspection, any \(Z_{\text{DR}}\) values less than -2.5 or greater than 4.0 were flagged as bird.

These two birds filtered were combined to form an automated bird filter and tested on the thirteen manually filtered MISR coincident radar images. Each radar image from the thirteen
coincident scenes was manually filtered in SOLOII using visual inspection of the reflectivity, Doppler velocity, $Z_{DR}$, and associated cloud fields. On average, 5% ($\pm 2.5\%$) of the pixels that were manually classified as birds were missed by the automated bird filter. Of note however, when manually filtering images in SOLOII, several pixels that surround the birds were often also flagged as bird and thus the true effectiveness of the filters was probably greater than 95% for all images. Birds that were missed were evident in the reflectivity field in the collocated data because there were often a few pixels with high reflectivity scattered about the images that were not associated with the collocated cloud field. The radial velocities of these missed birds were found to be nearly the same as the background wind. Furthermore, their $Z_{DR}$ values were the less than three and greater than negative one suggesting that occasionally birds could fly so that their velocity was the same as the wind and the ratio of their width to height was close to one. Fig. 3.3.2. is a typical example of the efficiency of the bird filters from 12 Dec 2004.

3.4. Selecting a Rain No-Rain Threshold

With the unwanted echoes removed from the radar data, the remaining echoes could be interpreted as meteorological in nature. Due to the sensitivity of S-band radar systems, coherent signal was observed from precipitation and clear air during RICO. Clear air echo in the maritime boundary layer is typically the result of Bragg scattering associated with gradients in the index of refraction (Wilson et al., 1994). In the maritime boundary layer, incomplete turbulent mixing due to rising thermals and convective rolls create moisture and temperature gradients through adiabatic vertical motions and evaporation (Knight and Miller, 1993). As microwave energy from the radar propagates through the boundary layer, scattering from clear air can constructively interfere if the spatial variations of the index of refraction occur at half the
Fig. 3.3.2. The above image from 12 December 2004 shows the effects of the bird, noise and island filters. a) Reflectivity (dBZ); b) Reflectivity (dBZ) with island removed; c) Reflectivity (dBZ) with island, noise removed and automated bird filter applied; d) same as c but with an additional hand edit to remove any birds the automated filter missed.

wavelength of the radar (i.e., 5 cm for S-band radar). This constructive interference can produce a strong enough signal to be recognized above the noise level of the radar. During RICO, the 0.5° beam contained both coherent Bragg scattering from clear air and Rayleigh scattering from hydrometeors as seen in Fig. 3.4.1.
Fig. 3.4.1. Reflectivity (dBZ) image showing both the Bragg and Rayleigh scattering found in nearly all RICO radar images. Note: for this image, white is assigned to any pixel that was flagged as noise, island or bird.
Although Bragg scattering can be used in many application, separating it from Rayleigh scattering was required to measure precipitation rates. Knight and Miller (1998) examined the presence of Bragg scattering in early echoes from cumulus clouds with both X-band and S-band radars. They reported that above 5 dBZ, both radars measured the same reflectivity factor but below 5 dBZ the S-band system was much more sensitive to Bragg scattering. They interpreted that their data meant that above 5 dBZ Rayleigh scattering from hydrometeors dominated the signal and 5 dBZ provided a useful distinction for setting a rain-no-rain threshold in reflectivity field. In this study a more conservative reflectivity factor of 7 dBZ was considered the minimum threshold for rain, which corresponds to a rain rate of 0.15 mm hr\(^{-1}\). Justification for the selection of 7 dBZ threshold is discussed further in Chapter 5.
CHAPTER 4
COLLOCATING THIRTEEN COINCIDENT MISR AND SPoLKa SCENES

4.1. Overview

The objective of this chapter is to describe the procedure used to collocate the satellite and radar data. Of the sixteen times MISR took measurement in the RICO radar domain, only thirteen scenes were used in this comparison (Table 4.1.1.). The collocation of these thirteen scenes offer a high resolution glimpse into the characteristics of trade wind clouds observed from space and ground-based radar. The techniques described in this chapter were used to develop unique relationships between clouds and precipitation, and offer a guideline for future processing of similar datasets.

Table 4.1.1. Date, SPoLKa file time, rawinsonde time and location, MISR time, MISR Path and Orbit, LCL (m) and comments for the collocated data.

<table>
<thead>
<tr>
<th>Date</th>
<th>SPoLKa Time</th>
<th>Rawinsonde Time</th>
<th>MISR time</th>
<th>Path/Orbit</th>
<th>LCL(m)</th>
<th>Comments</th>
</tr>
</thead>
<tbody>
<tr>
<td>11-26-04</td>
<td>14:51:48</td>
<td>12UTC (Guadeloupe)</td>
<td>14:54:15</td>
<td>3/26289</td>
<td>780</td>
<td>PPI</td>
</tr>
<tr>
<td>11-28-04</td>
<td>14:40:39</td>
<td>12UTC (Guadeloupe)</td>
<td>14:41:57</td>
<td>1/26318</td>
<td>807</td>
<td>SUR</td>
</tr>
<tr>
<td>12-05-04</td>
<td></td>
<td>14:48:17</td>
<td></td>
<td>2/26420</td>
<td></td>
<td>No radar data</td>
</tr>
<tr>
<td>12-07-04</td>
<td>14:35:47</td>
<td>12UTC (Guadeloupe)</td>
<td>14:35:58</td>
<td>233/26449</td>
<td>742</td>
<td>SUR</td>
</tr>
<tr>
<td>12-12-04</td>
<td>14:54:35</td>
<td>10:48UTC (Barbuda)</td>
<td>14:54:36</td>
<td>3/26522</td>
<td>700</td>
<td>SUR</td>
</tr>
<tr>
<td>12-14-04</td>
<td>14:42:24</td>
<td>10:51UTC (Barbuda)</td>
<td>14:42:16</td>
<td>1/26551</td>
<td>812</td>
<td>SUR</td>
</tr>
<tr>
<td>12-21-04</td>
<td>14:47:33</td>
<td>10:59UTC (Barbuda)</td>
<td>14:48:33</td>
<td>2/26653</td>
<td>1127</td>
<td>PPI</td>
</tr>
<tr>
<td>12-23-04</td>
<td></td>
<td>14:36:13</td>
<td></td>
<td>233/26682</td>
<td></td>
<td>No radar data</td>
</tr>
<tr>
<td>12-28-04</td>
<td>14:52:29</td>
<td>10:57UTC (Barbuda)</td>
<td>14:54:49</td>
<td>3/26755</td>
<td>829</td>
<td>SUR</td>
</tr>
<tr>
<td>12-30-04</td>
<td>14:49:26</td>
<td>10:57UTC (Barbuda)</td>
<td>14:42:29</td>
<td>1/26784</td>
<td>889</td>
<td>SUR</td>
</tr>
<tr>
<td>01-06-05</td>
<td>14:48:11</td>
<td>16:56UTC (Barbuda)</td>
<td>14:48:43</td>
<td>2/26886</td>
<td>641</td>
<td>SUR</td>
</tr>
<tr>
<td>01-08-05</td>
<td></td>
<td>14:36:22</td>
<td></td>
<td>233/26915</td>
<td></td>
<td>No radar data</td>
</tr>
<tr>
<td>01-13-05</td>
<td>14:55:31</td>
<td>14:03UTC (Barbuda)</td>
<td>14:54:56</td>
<td>3/26988</td>
<td>484</td>
<td>SUR</td>
</tr>
<tr>
<td>01-15-05</td>
<td>14:42:37</td>
<td>16:59UTC (Barbuda)</td>
<td>14:42:34</td>
<td>1/27017</td>
<td>813</td>
<td>SUR</td>
</tr>
<tr>
<td>01-22-05</td>
<td>14:48:12</td>
<td>16:45UTC (Barbuda)</td>
<td>14:48:45</td>
<td>2/27119</td>
<td>629</td>
<td>SUR</td>
</tr>
<tr>
<td>01-24-05</td>
<td>14:36:38</td>
<td>16:46UTC (Barbuda)</td>
<td>14:36:23</td>
<td>233/27148</td>
<td>796</td>
<td>SUR</td>
</tr>
</tbody>
</table>
In its descending node, four of MISR’s 233 paths intersected the SPolKa radar domain around 14:45 UTC (10:45 AM LT) (Fig. 4.1.1.). During the overpass times, the radar scan strategy was altered to include an extra SUR scan near the time MISR was overhead. However, predicting the exact time MISR was over the radar was difficult and therefore the closest 0.5° scan was chosen which was occasionally a PPI scan rather than a SUR scan. Table 4.1.1. details the observation times of the radar, satellite and rawinsondes used in the collocation. Also included are MISR path and orbit information, the height of the lifting condensation level (LCL) and the reason why a particular date was not used.

![MISR Paths intersecting SPolKa Radar Domain](image)

*Fig. 4.1.1. MISR Paths that intersect the SPolKa Radar Domain.*
4.2. Geolocating Each Radar Pixel

As a part of MISR’s standard products, latitude and longitude information were provided for each pixel. Radar data, however, were reported on a radial coordinate system and each pixel was referenced to a range gate number, beam azimuth angle and starting elevation angle. To collocate these two datasets at the pixel level, each radar pixel was assigned a latitude and longitude.

To geo-locate a radar pixel, the path of the radar beam through the atmosphere was found to accurately determine the beam’s position along the earth’s surface. As an electromagnetic pulse travels through the atmosphere, its propagation is determined by changes in the index of refraction with height. The index of refraction is dependent on the density of the medium the microwave pulse is traveling through. Microwave pulses passing through a medium of constant density will have a straight path. In the standard atmospheric profile, density decreases with height which causes the propagation speed of the beam to increase with altitude and bend the beam path back toward the Earth (Doviak and Zrnić, 1984). Including the effects of moisture and temperature inversions found in the tropical oceanic boundary layer, the beam may travel much differently than predicted by a standard atmosphere. Therefore, to accurately determine the position of beam, information about the vertical profile of temperature, dew point temperature and ultimately the change in the index of refraction is needed.

To obtain these profiles, rawinsondes were launched from the Spanish Point sounding launch sight on the island of Barbuda, just a few miles east-southeast of the radar site. If a sounding was not available from Barbuda, the 12 UTC Guadeloupe sounding was used. (See Table 4.1.1. for a list of the sounding file names and cloud base for a surface parcel). Each sounding was chosen for its closeness in time with the MISR overpass and then logarithmically
interpolated to one meter vertical resolution. Next the saturation vapor pressure, dew point
temperature, water vapor pressure and partial pressure of dry air were derived from the
measurements of temperature and relative humidity and given as Eqs. 4.2.1., 4.2.2., 4.2.3. and
4.2.4., respectively.

\[ e_s(mb) = EXP \left( 5422.99K \left( \frac{1}{273.15K} - \frac{1}{T(K)} \right) \right) \times 6.11mb \] (4.2.1.)

\[ T_d(\degree C) = \left( \frac{1}{273.15K} - \ln \left( \frac{e_s \times RH}{6.11mb} \right) \right) - 273.15K \] (4.2.2.)

\[ e(mb) = a + T_d(b + T_d(c + T_d(d + T_d(f + T_d(g + T_d)))))) \] (4.2.3.)

\[ p_d = P - e \] (4.2.4.)

In the above equations, \( e_s \) is saturation vapor pressure, \( T(K) \) is the temperature in Kelvin, \( T_d \) is
the dew point temperature in degrees Celsius, \( RH \) is relative humidity, \( e \) is vapor pressure, \( p_d \) is
partial pressure of dry air and \( P \) is pressure. The constants in Eq. 4.2.3. are given as,

\[ a = 6.107799961 \]
\[ b = 4.436518521 \times 10^{-1} \]
\[ c = 1.428945805 \times 10^{-2} \]
\[ d = 2.650648471 \times 10^{-4} \]
\[ f = 3.031240396 \times 10^{-6} \]
\[ g = 2.0340880948 \times 10^{-8} \]
\[ h = 6.136820929 \times 10^{-11} \]

Using these four equations, the index of refraction can be calculated with Eq. 4.2.4.,
where $n$ is the index of refraction, $P_d$ is the partial pressure of dry air, $e$ is the vapor pressure and $T$ is the temperature in Kelvin (Battan, 1959). Using Eq. 4.2.4., $n$ can be calculated at each one meter vertical increment in the sounding and the path of the beam through the atmosphere can be properly represented.

To calculate the beam path, the equations from the Effective Earth’s Radius Model are used as described in the second chapter of Doviak and Zrnić (1984). The equations in this model are only valid under the assumption that the ray is nearly parallel to the earth’s surface, propagating through the lowest 15 km of the atmosphere and the height of the beam $h$, must be much smaller than the radius of the earth $a; (h << a)$. The exact differential equation to solve for a ray propagating though a spherically stratified equation was derived by Hartree et al., (1946). This solution to this equation provides a relationship for the path of the beam ($r$) through the atmosphere and its projected arc distance along the Earth’s surface ($s$) by accounting for the changes in the beam path due to the vertical gradient of the index of refraction. Using the assumptions under the Effective Earth’s Radius Model this equation was reduced to Eq. 4.2.5.

$$\frac{d^2h}{ds^2} = \frac{1}{a} + \frac{dn}{dh}$$  \hspace{5cm} (4.2.5.)

In Eq. 4.2.5., $h$ is the height above the Earth’s surface, $s$ is the arc distance along the Earth’s surface and $n$ is the index of refraction. To solve Eq. 4.2.5., set $\frac{dh}{ds} = \tan \phi$, where $\phi$ is the elevation angle of the beam. Next set $C_1 = \frac{1}{a} + \frac{dn}{dh}$, where $a$ is the radius of the Earth, so that the equation becomes Eq. 4.2.6.
\[
\frac{d}{ds} \left[ \tan \varphi \right] = C_1 
\]  
(4.2.6.)

After integrating both sides from \( \varphi_0 \) to \( \varphi \) and \( s_0 \) to \( s \) the resulting equation becomes Eq. 4.2.7.

\[
\tan \varphi = \tan \varphi_0 + C_1 (s - s_0) 
\]
(4.2.7.)

Next, substitute in \( \frac{dh}{ds} = \tan \varphi \) so that the equation can be written as Eq. 4.2.8. and then integrate from \( h_0 \) to \( h \) and \( s_0 \) to \( s \) to arrive at Eq. 4.2.9. Then solve for \( s \) and after expanding and regrouping this equation, it can be re-written as the quadratic equation of Eq. 4.2.10.

\[
\frac{dh}{ds} = \tan \varphi_0 + C_1 (s - s_0) 
\]
(4.2.8.)

\[
h = \tan \varphi_0 (s - s_0) + \frac{C_1 (s - s_0)^2}{2} + h_0 
\]
(4.2.9.)

\[
0 = \frac{C_1 s^2}{2} + s (\tan \varphi_0 - C_1 s_0) - s_0 \tan \varphi_0 + \frac{C_1 s_0^2}{2} + h_0 - h 
\]
(4.2.10.)

Solving the quadratic for \( s \) gives Eq. 4.2.11.

\[
s = \frac{-(\tan \varphi_0 - s_0 C_1) \pm \sqrt{\tan \varphi_0^2 + 2C_1 (h - h_0)}}{C_1} 
\]
(4.2.11.)

Now \( s \) can be solved iteratively for each one meter height interval in the sounding to find the projection of the beam path along the Earth’s surface. (Note: The variables of \( \varphi_0 \) and \( s_0 \) are the \( \varphi \) and \( s \) from the previous iterative step in height.)

The above procedure gives an arc distance along the Earth’s surface as the beam passes through each one meter vertical level. However, radar data were reported every 149.89 m along the beam in cells called range gates. Therefore, the arc distance along the Earth’s surface needs to be related to the ~150 m range gate spacing along the beam. To find this, the Eq. 4.2.12.
\[ s = k_\epsilon a \sin^{-1}\left(\frac{r \cos \theta}{k_\epsilon a + h}\right) \tag{4.2.12.} \]

given in Doviak and Zrnic (1984) can be solved in 150 m intervals for the beam path \( r \), where \( a \) is the radius of the earth and \( k_\epsilon \), is the radius of curvature given as \( k_\epsilon \equiv \frac{1}{1 + a \left[ \frac{dn}{dh} \right]} \). Eq. 4.2.12. describes the relationship between \( r \) (distance along the beam) and \( s \) (arc distance along the surface), and after \( r \) is interpolated to the correct range gate spacing a corresponding \( s \) can be assigned to each range gate. The end result is that each range gate along the beam has an associated arc distance along the ground and the data are now in the two dimensional coordinate system of range and azimuth referenced at the surface.

A way to use the range and azimuth to assign a latitude and longitude to each radar pixel is to mathematically rotate the globe under the radar so that it sits on the North Pole. If positioned on the North Pole, each successive range gate along a beam would be equivalent to a change in latitude and each successive beam would be a change in longitude (Fig. 4.2.2.). Two mathematical rotations of the reference frame are use to position the radar on the North Pole. In its initial reference frame, the radar resides at the Cartesian position of \((x, y, z)\) defined in Eq. 4.2.13., with latitude, \( \theta \), measured from the z-axis and longitude, \( \phi \), measured from the x-axis as shown in Fig. 4.2.3a. The first rotation is around the z axis (connecting the poles) so that the radar is now positioned along the 0° longitude line (Prime Meridian). This rotation defines a new coordinate system \((x_1, y_1, z_1)\) as shown in Fig. 4.2.3b. Next, the \( y_1 \)-axis is rotated so that the \( z_1 \)-axis points through the center of the radar. The new coordinate system is now \((x_2, y_2, z_2)\) as shown in Fig. 4.2.3c, and in this frame, a change in range and azimuth can be expressed as a change in latitude and longitude. Using Eqs. 4.2.13. - 4.2.16. a radar located at latitude \( \theta_R \) and
Fig. 4.2.2. Display of a radar beam if the radar was located on the North Pole

longitude $\varphi_R$ in the original $(x, y, z)$ reference plane can be translated to the final $(x_2, y_2, z_2)$ plane. Using the relationships in Eqs. 4.2.13. – 4.2.16., a radar pixel’s latitude and longitude can be found in this new reference system and translated back to its true position on the globe. The only additional information needed is the arc distance range $(s)$ along the Earth’s surface and azimuth angle for each pixel. The range and azimuth information is then incorporated into Eq.
4.2.16. through the original (x, y, z) definitions, but instead of using the radar latitude ($\theta_R$) and longitude ($\phi_R$) the pixels range and azimuth are used in their place as in Eq. 4.2.17. To account for the arc distance ($s$) of each pixel a simple conversion factor of a $1^\circ$ change in latitude is equal to

Fig. 4.2.3. a) Original (x, y, z) reference frame of the radar; b) New frame (xL, yL, zL) after z-axis rotation; c) Final frame (xL, y2, z2) after yL-axis rotation. The result of these rotations is that the radar is re-located to the North Pole so that a change in range and azimuth are related to a change in latitude and longitude.
\[
\begin{bmatrix}
  x \\
  y \\
  z
\end{bmatrix} =
\begin{bmatrix}
  \text{radius} \cdot \cos(\varphi_R) \cdot \sin(\theta_R) \\
  \text{radius} \cdot \sin(\varphi_R) \cdot \sin(\theta_R) \\
  \text{radius} \cdot \cos(\theta_R)
\end{bmatrix}
\]  
(4.2.13.)

\[
\begin{bmatrix}
  x_1 \\
  y_1 \\
  z_1
\end{bmatrix} =
\begin{bmatrix}
  \cos(\varphi_R) & \sin(\varphi_R) & 0 \\
  -\sin(\varphi_R) & \cos(\varphi_R) & 0 \\
  0 & 0 & 1
\end{bmatrix}
\begin{bmatrix}
  x \\
  y \\
  z
\end{bmatrix}
\]  
(4.2.14.)

\[
\begin{bmatrix}
  x_2 \\
  y_2 \\
  z_2
\end{bmatrix} =
\begin{bmatrix}
  \cos(\theta_R) & 0 & -\sin(\theta_R) \\
  0 & 1 & 0 \\
  \sin(\theta_R) & 0 & \cos(\theta_R)
\end{bmatrix}
\begin{bmatrix}
  x_1 \\
  y_1 \\
  z_1
\end{bmatrix}
\]  
(4.2.15.)

\[
\begin{bmatrix}
  x_2 \\
  y_2 \\
  z_2
\end{bmatrix} =
\begin{bmatrix}
  x \cos(\theta_R) + z \sin(\theta_R) \cdot \cos(\varphi_R) - y \sin(\varphi_R) \\
  x \cos(\theta_R) + z \sin(\theta_R) \cdot \sin(\varphi_R) + y \cos(\varphi_R) \\
  -x \sin(\theta_R) + z \cos(\theta_R)
\end{bmatrix}
\]  
(4.2.16.)

111.20 km is applied. Then the latitude of each pixel is found by subtracting Eq. 4.2.18. from 90° and the longitude is found by taking the opposite of Eq. 4.2.19.

\[
\begin{bmatrix}
  x \\
  y \\
  z
\end{bmatrix} =
\begin{bmatrix}
  \text{radius} \cdot \cos(\text{azimuth}) \cdot \sin\left(\frac{s}{111.20\text{km}}\right) \\
  \text{radius} \cdot \sin(\text{azimuth}) \cdot \sin\left(\frac{s}{111.20\text{km}}\right) \\
  \text{radius} \cdot \cos\left(\frac{s}{111.20\text{km}}\right)
\end{bmatrix}
\]  
(4.2.17.)

\[
\cos^{-1}\left(\frac{z}{\text{radius}}\right)
\]  
(4.2.18.)

\[
\cos^{-1}\left(\frac{x}{\text{radius} \cdot \sin(\text{latitude})}\right)
\]  
(4.2.19.)
4.3. Errors in Geolocation

To collocate the MISR and radar data, accurate geolocation information from both instruments was a necessity. In ideal conditions, errors are negligible or occur on scales less than the resolution of the instrument. First order errors in the geolocation of a radar pixel stem from two sources; 1) the projection of the beam path to the Earth’s surface; and 2) precision of radar location. First, the formulas for projecting the exact position of the beam path on the Earth’s surface could be biased due to the inherent assumptions in the equations. Fortunately, by exclusively using the 0.5° data the assumption that the beam was nearly parallel to the Earth’s surface was best satisfied and incurs the least amount of error possible of all elevation angles. Also, the value of the Earth’s radius used in the calculations was corrected for the radar’s location near 17N and 61W, and therefore data taken from this location were on the same sphere the equations assume the data were taken on. If the radius of the earth was increased or decreased by 10 km, changes in geolocation for each pixel are about 0.000000006° (lat/long) and therefore errors due to the radius were essentially non-existent.

Errors in the exact latitude and longitude of the radar’s location can cause a systematic bias in the prediction of each pixels geolocation. For example, an error of 0.0009° in latitude will bias a pixel location by about 100 m. Fortunately, radar location was known with high precision (0.000000000001°) keeping geolocation biases below one meter.

Another geolocation error source could be the result of using a sounding that was unrepresentative of the environment the radar beam was traveling through. In this case, the change of index of refraction with height would be incorrect resulting in the calculated beam path being different from the actual one. To test this, the temperature and dew point temperature
for each sounding was perturbed by ± 2°C. As a result of this perturbation, the geolocation of each pixel was altered on average by 0.000001° which would bias the geolocation by about 10m.

MISR geolocation information, which was reported for each 275 m pixel, has a mean geolocation error for all nine cameras of 60 m with a standard deviation of 100 m for the nadir viewing camera (Jovanovic, 2002). Therefore, MISR geolocation errors were less than its resolution and not considered a major source of bias.

4.4. Adjusting for Time Differences

Simultaneous observations from satellite and radar are ideal for collocation, however as Table 4.1.1. shows, there were significant time differences between the MISR and radar images. To account for this difference, pixel locations in the radar data were adjusted according to how far a cloud element would have approximately moved during the time it was observed by MISR and when it was observed by radar. In general, the boundary layer trade winds blow ~10 m s\(^{-1}\) and most boundary layer clouds travel passively with the background wind. Therefore, an adjustment in the latitude and longitude can be applied to each radar pixel based on a time difference and background wind measurement for each scene.

Background flow was found using the Velocity Azimuth Display (VAD) technique (Doviak and Zrnić, 1984). The large amount of clear air echo from Bragg scattering acted as a tracer for the low level background wind. Plotting the Doppler velocity at a given range over all azimuth angles reveals the wind speed and direction (Fig. 4.4.1.). Each point on Fig. 4.4.1. represents the Doppler velocity taken from each beam’s 30 km pixel. By plotting this velocity as a function of azimuth, a background wind speed and direction were found for each of the thirteen
Fig. 4.4.1. Top) Raw Doppler Velocity display with a white circle denoting the 30 km range ring. Bottom) Velocity Azimuth Display (VAD) with a 3rd order polynomial fit (red line) used to determine wind speed and direction. On this day, (6 Jan 2005) background trade winds were from the NE (60°) at 12 m s⁻¹.
coincident scenes. Doppler velocities are highest when the beam points into the wind because the wind blows the drops in the radial direction which produces the largest Doppler shift. Therefore, in Fig. 4.4.1., the wind speed and direction were found by locating the maximum wind speed and its associated direction. In Fig. 4.4.1., the wind was from the NE (60°) at 12 m s⁻¹.

Time differences were determined by taking the difference between the MISR block center time for the 76th block and the time stamp assigned to each radar pixel (top Fig. 4.4.2.). By choosing the 76th MISR block center time as the reference time, an additional time difference of approximately 20 s was possible, which could increase geolocation error up to ~200 m during the time adjustment process. Regardless, the time difference was then used to adjust the latitude and longitude originally assigned to each radar pixel as seen in the bottom of Fig. 4.4.2. The new latitude and longitude of the adjusted radar pixel was the pair used in the collocation processes described in Section 4.5.

The assumption that all clouds move passively with the background flow was another source of error in this procedure. Certain mesoscale features such as clouds forming along cold pools and large clusters of clouds typically have an additional component in their movement that was not considered in this adjustment. Fortunately, time differences were small and almost all pixels were shifted less 500 m.

4.5. Re-sampling to the MISR Grid

After assigning a time adjusted latitude and longitude to each radar pixel the radar data were re-sampled to the MISR grid. As discussed in Chapter 2, MISR NIR BRF data were reported on the Space Oblique Mercator (SOM) projection at 275 m ground resolution. Using a nearest neighbor re-sampling routine, each radar pixel’s reflectivity was assigned to the closest
Fig. 4.4.2. Top) Diagram of MISR data blocks and the radar beams showing where times are reported for each and how time differences are calculated; Bottom) diagram showing how a radar pixels latitude and longitude are adjusted according to the background wind.
Fig. 4.5.1. Outlined in red are idealized radar pixels and outlined in black are idealized MISR pixels. As an example of the nearest-neighbor re-sampling algorithm, the radar pixel outlined in yellow would have its reflectivity assigned to the 4 black filled MISR pixels because they are the closest in absolute distance to that radar pixel.

MISR pixel based on the minimum absolute difference in geolocation between the two datasets (Fig. 4.5.1.). The nearest neighbor algorithm was chosen because it does not smooth the reflectivity field during the re-sampling. It was extremely important that the original data values were preserved during the collocation so as to not wash out the already weak radar signal from drizzle and Bragg scattering. Furthermore, as discussed in Chapter 2, a rain-no-rain threshold of
7 dBZ was chosen for this project and any dilution of the original data may reduce the number of pixels that have reflectivity above 7 dBZ and bias the statistics. Figs. 4.5.2. – 4.5.14. show the thirteen collocation scenes. *(Figs. 4.5.2 – 4.5.14 are in a separate file called chapter_4_figs.pdf)*

4.6. Resolution Differences and Radar Beam Geometry

Although the top down view like in Figs. 4.5.2. – 4.5.14. is typically how radar and satellite data are viewed, the way they remotely sense precipitation and cloud is quite different. Aside from the fact that the radar actively senses precipitation at microwave wavelengths and satellite passively views cloud at visible and NIR wavelengths, much attention must be given to the difference in view geometry between the instruments. MISR’s view of trade wind cloud is from the top down. At visible wavelengths, clouds, especially those with high liquid water paths, are very reflective. Therefore, MISR’s view of clouds is from the appearance of cloud-top. Only on the edges of the MISR swath do cloud sides become visible. Regardless view angle biases in cloud cover at the edge of the MISR swath are very small (Minnis, 1989).

A radiance measurement from satellite has contributions from the atmosphere between the cloud and the sensor. At some NIR wavelengths, water vapor has strong absorptive properties which could decrease the brightness of cumulus clouds. Fortunately, MISR’s NIR channel’s spectral response function samples in an atmospheric window, which decreases the affects of gaseous attenuation.

The view-geometry of a radar beam is a function of the range from the antenna and height above the Earth’s surface. The area of a microwave pulse volume transmitted from a radar obeys a square law as it propagates. As distance increases, the three-dimensional volume expands as the square of its range from the antenna. Also, the elevation angle of the antenna
determines the beam’s initial direction of propagation and then the atmosphere’s exponentially decreasing density acts to bend the ray back toward the surface. The bottom left of Fig. 4.6.1. shows the average beam path of the 0.5° beam for the thirteen coincident scenes and illustrates the range dependent beam geometry. The top of Fig. 4.6.1. illustrates how the pulse volume changes in height and width as a function of range in the presence of shallow convective precipitation.

The affect of the radar beam geometry is that close to the radar, the contributing volume is small, the beam height is low and sensitivity is very high. As the beam propagates, the contributing volume expands, the beam height increases and the noise level climbs. Beam geometry presents an interesting challenge when sampling low trade wind clouds. Close to the radar the beam is below cloud base, the contributing volume is small and the radar is very sensitive to drizzle and clear air echo from Bragg scattering. At about 60 km the center of the radar beam enters cloud base (~800 m) while the top of the beam is at an altitude of ~1200 m and the bottom is ~500 m below cloud base. At 100 km in range, the top of the beam is now above the typical trade wind inversion (2.5 km), the center of the beam is sampling in the middle of the cloud while the bottom of the beam is still below cloud base. Also, the beam is now 1.5 km wide and given that the typical trade wind cumulus cloud diameter is only 0.5-1.0 km, only a portion of the beam is filled with cloud. It is clear that a power return from 100 km could include many more targets in the contributing volume than a return from 10 km. Therefore, caution must be used when interpreting precipitation rates derived from radar reflectivity because the majority of the sampling area of radar is not below cloud base, which makes the assumption that the precipitation is hitting the ground not always valid.
Fig. 4.6.1. Top) A cross-section view along the radar beam of the radar pixel height and width with respect to a typical precipitating trade wind cumulus cloud. Bottom Left) 0.5° vertical beam geometry as a function of arc distance along the Earth’s surface. Bottom Right) A side cross-section of the range gate width and height with respect to a typical precipitating trade wind cumulus cloud. Note the large change in the radar’s pixel size and height as range increases. The dashed square box on the top image and darker yellow shading in the lower left image highlight the ranges of 25-90 km. This is the range over which the results of Chapter 6 are processed over.

The range dependency goes beyond physical resolution as attenuation at microwave frequencies also controls how radar measures precipitation. Although S-band radar systems suffer very little attenuation (Ryzhkov and Zrnic 1995), the noise level increases as a function of range. Fig. 4.6.2. is a plot of the average reflectivity (converted to dBZ) for power returns below the noise threshold (-115.1 dBm) as a function of range for all SUR files. Close to the radar, the noise level was very low and the radar was very sensitive to clear air echo and weak Rayleigh echo from clouds. However, as range increased so did the noise level and beyond 115 km, echo from light drizzle and Bragg scattering may become indistinguishable from the noise.
Fig. 4.6.2. The mean reflectivity (dBZ), as a function of range, is plotted for all pixels with a power return of less than -115.1 dBm.

The net affect is that radar data are highly dependent on range. Considering the sub-millimeter daily precipitation rates typically measured over this region, precipitation estimates with radar need to be made with some reservation. To account for this range dependency, results in Chapters 5 and 6 will be reported as a function of range.
CHAPTER 5
CHARACTERISTICS OF THE COLLOCATED DATASET

5.1. MISR Cloud Fraction Observations

As discussed in Chapter 4 and as Figs. 4.5.2. – 4.5.14. show, each satellite overpass covers a different portion of the SPolKa radar domain. Table 5.1.1 lists the date, MISR path number, total overlapping area and two cloud fraction (CF) calculations for the thirteen coincident scenes. It is clear from this table (as well as in Fig. 4.1.1., and Figs. 4.5.2. – 4.5.14.) that paths 1 and 2 have the largest overlapping area with the radar data, while path 3 covers less than half and path 233 covers only a small sliver on the eastern side. Sampling biases in the statistics of this chapter due to the various sizes of the overlapping regions were unavoidable. It is important to note, however, that the data were not manually sub-sectioned or chosen to focus on a particular cloud type or feature. All available coincident data were used in this analysis.

The MISR observed cloud amounts were considered representative of typical trade wind cloud amounts found in the winter over the Western Tropical Atlantic, when compared to the International Satellite Cloud Climatology Project (ISCCP) (Fig. 5.1.1.) and ship observations (http://www.atmos.washington.edu/%7Eignatius/CloudMap/WebO/index.html). The average CF of the thirteen coincident scenes is 49% which matches well with the ISCCP CF for December ’04 and January ’05 over this region of 40% and 47%, respectively. Furthermore, cloud cover observations from ship reports show that the 44 yr average CF over the RICO region is 48%. Therefore, the relationships developed from cloud observed by MISR over the RICO region can be considered representative for the Western Tropical Atlantic clouds. However, as Zhao (2006)
Table 5.1.1. List of days, MISR paths, overlapping area (km$^2$), cloud fraction in overlapping area and cloud fraction from the surrounding five MISR blocks (see Fig. 4.1.1.).

<table>
<thead>
<tr>
<th>Date</th>
<th>Path</th>
<th>Total overlapping area</th>
<th>Cloud Fraction in overlapping area</th>
<th>5 Block Cloud Fraction</th>
</tr>
</thead>
<tbody>
<tr>
<td>11-26-04</td>
<td>3</td>
<td>18303.14 km$^2$</td>
<td>51.1%</td>
<td>32.2%</td>
</tr>
<tr>
<td>11-28-04</td>
<td>1</td>
<td>54014.10 km$^2$</td>
<td>21.6%</td>
<td>32.5%</td>
</tr>
<tr>
<td>12-07-04</td>
<td>233</td>
<td>8996.12 km$^2$</td>
<td>36.7%</td>
<td>51.0%</td>
</tr>
<tr>
<td>12-12-04</td>
<td>3</td>
<td>27286.03 km$^2$</td>
<td>41.4%</td>
<td>21.7%</td>
</tr>
<tr>
<td>12-14-04</td>
<td>1</td>
<td>54584.69 km$^2$</td>
<td>58.1%</td>
<td>64.8%</td>
</tr>
<tr>
<td>12-21-04</td>
<td>2</td>
<td>34497.55 km$^2$</td>
<td>19.9%</td>
<td>34.9%</td>
</tr>
<tr>
<td>12-28-04</td>
<td>3</td>
<td>27464.28 km$^2$</td>
<td>77.8%</td>
<td>39.6%</td>
</tr>
<tr>
<td>12-30-04</td>
<td>1</td>
<td>55659.47 km$^2$</td>
<td>44.8%</td>
<td>46.6%</td>
</tr>
<tr>
<td>01-06-05</td>
<td>2</td>
<td>70282.85 km$^2$</td>
<td>53.9%</td>
<td>68.2%</td>
</tr>
<tr>
<td>01-13-05</td>
<td>3</td>
<td>27434.86 km$^2$</td>
<td>84.3%</td>
<td>47.6%</td>
</tr>
<tr>
<td>01-15-05</td>
<td>1</td>
<td>58674.41 km$^2$</td>
<td>49.9%</td>
<td>44.9%</td>
</tr>
<tr>
<td>01-22-05</td>
<td>2</td>
<td>66717.59 km$^2$</td>
<td>48.9%</td>
<td>39.9%</td>
</tr>
<tr>
<td>01-24-05</td>
<td>233</td>
<td>10729.67 km$^2$</td>
<td>49.1%</td>
<td>36.6%</td>
</tr>
</tbody>
</table>

demonstrated when only looking at cumulus cloud with 15 m resolution data, cloud fraction was 10%. Furthermore, of the 440 total scenes used in Zhao (2006), only 124 were covered exclusively by shallow cumulus. Therefore, a large portion of the cloud fraction reported from the coincident data came from stratiform clouds.

5.2. Cloud Top Height Analysis

MISR’s nine view angles and stereoscopic capabilities allow it to retrieve cloud-motion and cloud-top height. Stereoscopic retrieval of cloud-top heights requires cloud feature matching between cameras which are discriminated easily with MISR’s 16-bit depth. A cloud-top height is retrieved by measuring the parallax created by projecting the position of a cloud feature onto the spheroid from multiple cameras. Cloud-motion is determined by the amount a particular cloud element moves during the seven minutes it takes for all nine cameras to pass over. The cloud-top
Fig. 5.1.1. Top) ISCCP-D2 Monthly mean total cloud amount for December 2004. Bottom) ISCCP-D2 Monthly mean total cloud amount for January 2005.
height is then adjusted according for cloud-motion and reported on the SOM projection at 1.1 km with an accuracy of ± 562 m (Moroney, 2002).

The top of Fig. 5.2.2. is a distribution of the “Best Winds” cloud-top heights measured in the area that overlapped with the radar data. The MISR “Best Winds” stereo height data were used because the stereoscopic heights were corrected by the cloud-track winds. The bottom of Fig. 5.2.2. is the distribution of the “Best Winds” heights of the surrounding five blocks. One important distinction in this project is that all cloud types were considered, not just shallow convection (cf. Zhao 2006). Fig. 5.2.2. clearly shows a second mode near 6 km in height, suggesting that some mid level cloud occupied a portion of the collocated data. Upon visual inspection of Figs. 4.5.2. – 4.5.14. one can easily identify days of deeper convection (e.g., 13 January 2005) and days with considerable mid level stratus (e.g., 6 January 2005). All clouds observed during RICO were considered when determining the cloud cover amount reported in Table 5.1.1. This was done so that the relationships found between cloud and precipitation were not particular to scenes that only contain shallow cloud.

5.3. Distribution of Reflectivity for Clear and Cloudy Pixels

Chapter 3 discussed the selection of a rain-no-rain threshold based on the dual wavelength radar observations made by Knight and Miller (1998). According to their findings the transition when Raleigh scattering dominates the return signal occurs in the vicinity of 5 dBZ. The collocated data offer another way to determine the significance of the 5 dBZ threshold by comparing the reflectivity factor observed for cloudy and clear pixels. The blue curve in Fig. 5.3.1. represents the distribution of reflectivity factor for pixels flagged as clear by MISR. The shape of this distribution below 0 dBZ was attributable to clear air echo from Bragg scattering.
Fig. 5.2.2. Top) MISR “Best Winds” Stereo Height distribution for clouds within the SPoLKa radar domain binned every 500 m starting at 250 m; Bottom) Same as Top but for the five surrounding MISR blocks. Although the low cloud mode dominates the height retrievals, a distinct mid-level cloud mode is included in this study.
Fig. 5.3.1. Distribution of reflectivity factor (dBZ) for clear (blue) and cloudy (red) pixels.

Above 5 dBZ, however, a small tail extends out to 45 dBZ. This portion of the spectrum was populated perhaps by missed birds, collocation errors, beam filling at large ranges or residual rain shafts below optically thin and dying cloud.

The red curve in Fig. 5.3.1. represents the distribution of reflectivity for cloudy pixels. This curve clearly shows that a large number of cloudy pixels had reflectivity below the rain-no-rain threshold of 7 dBZ, which suggests that only a small fraction of the cloudy area was associated with rain. The pronounced tail in the red curve strengthens the findings of Knight and
Miller (1998) because around 7 dBZ the distribution changes character. This transition marked the point at which Rayleigh scattering from hydrometeors dominated the return signal. Therefore, in this study, a pixel with a reflectivity factor greater that 7 dBZ (0.15 mm h\(^{-1}\)) was considered raining at a detectable rate.

In Fig. 5.3.1., below -13 dBZ, the two distributions have roughly the same number of points for a given reflectivity factor and at these low values the radar is insensitive to the presence of clouds. However, for reflectivity factor between -11 dBZ and 7 dBZ the distribution of reflectivity factor for cloudy pixels is greater than the clear, suggesting that when the beam is passing in or under cloud, echo due to very light drizzle or echo from cloud droplets contributes to the return signal.

5.4. Characteristics of the Collocated Scenes as a Function of Range

Section 4.6. introduced the range dependency associated with measuring shallow convective precipitation with radar. Therefore, in the analysis that follows, data were stratified using five-kilometer wide doughnut-like range rings. Processing the radar data using these rings attempts to identify range dependencies and reveal how best to use the radar data to measure shallow convective precipitation.

Any pixel flagged as island, bird, noise or bad beam, as prescribed in Chapter 3, is hereafter also referred to as a “rejected” pixel, while coherent echo from clear air and echo from hydrometeors is hereafter also referred to as an “accepted” pixel. Fig. 5.4.1. shows the count of the total number of collocated, cloudy, accepted and rejected pixels as a function of the 5 km range rings for all collocated data combined. Recall that the data presented in this chapter was analyzed on the MISR projection which has a grid spacing of 275 m. As expected, as the range
increased, the total number of pixels also increased (linearly due to fixed grid spacing). Close to the radar, almost all radar pixels were accepted and few were rejected. As range increased between 15 km and 45 km, the ratio of accepted to rejected pixels remained high, however, beyond 50 km the number of accepted pixels begins to drop. This decrease was attributable to both beam geometry and signal sensitivity as the radar beam climbed in altitude and became less sensitive to Bragg scattering. Therefore, at ranges beyond 60 km only taller precipitating clouds were producing echo and clear air echo was rare.

The cloud field was unchanging as a function of range. The steadily increasing number of cloudy pixels with range indicated that when the thirteen scenes were combined, the background

Fig. 5.4.1. Pixel counts as a function of the 5 km range rings. Total pixel count (Blue); accepted pixel count (green); rejected pixel count (green); cloudy pixel count (red); rainy pixel count (gray).
cloud field was well distributed. Given this, one would expect that if the clouds were well
distributed, the rain from them should behave likewise. However, if the radar beam geometry
caused a range dependent bias in the sampling of precipitation from shallow convection, the
number of raining ($\geq 0.15 \text{ mm hr}^{-1}$ or $\geq 7 \text{ dBZ}$) pixels in each ring would not grow steadily with
range but rather behave much like the line for the number of accepted pixels. Fig. 5.4.1. shows
different behavior, as the raining pixel count (lowest grey line) grew steadily as the range rings
increased in size. Furthermore, the slope of the raining pixel count increased at a rate of about
5% of the slope of the cloudy pixel count line. This suggested that for these collocated scenes,
5% of the cloudy area was occupied with measurable precipitation ($\geq 0.15\text{mmhr}^{-1}$ or $\geq 7 \text{ dBZ}$).
Although, in these thirteen coincident scenes there appeared to be no strong range bias in the
radar’s ability to measure precipitation, a thorough investigation of the range dependency is
reported in Section 6.2.

5.5. Percent of Cloudy Area Associated with Rain

The fraction of a scene covered with cloud and rain provides a useful relationship
between trade wind clouds and their precipitation. The top of Fig. 5.5.1. shows the cloud fraction
and rain fraction variation for the thirteen coincident scenes in order of cloudiest to clearest day.
It was difficult to draw a relationship from this graph between cloud fraction and rain fraction for
a given day. For example, 15 January 2005 and 22 January 2005 had nearly identical cloud
fraction but much different rain fraction. To better develop this relationship, the thirteen scenes
were combined into a cumulative frequency distribution that shows the percent of cloudy area
associated with a rain rate greater than a given rain rate (bottom Fig. 5.5.1.). By calculating this
distribution over a series of cloud detection thresholds ranging from very cloud conservative to
Fig. 5.5.1. Top) Cloud fraction (blue) and rain fraction (red) sorted from cloudiest to least cloudiest day; Bottom) Cumulative frequency distribution of the percent of cloudy area associated with a rainfall rate greater than a given rain rate. This yellow envelop represents how the distribution changed when calculated over a range cloud detection thresholds. The bottom black line is made from the most cloud conservative threshold and the top black line is made from a 200% increase in the cloud conservative threshold which would represent a clear conservative threshold.
very clear conservative, an envelop of possible values were reported. The range of cloud detection thresholds was defined by a 200% increase in most cloud conservative threshold. It was found that between 5.5% and 10.5% of the cloudy area had rain rates greater than 0.1 mm hr\(^{-1}\) and between 1.5% and 3.5% of the cloudy area had rain rates greater than 1.0 mm hr\(^{-1}\). This also revealed that 90-95% of the cloudy area was not producing detectable precipitation. However, the clouds that are raining are doing so at light (0.1 mm hr\(^{-1}\)) to moderate (1 mm hr\(^{-1}\)) and even occasionally heavy (10 mm hr\(^{-1}\)) amounts.

5.6. Mesoscale Organization Variability

Upon visual inspection of the collocated data (Figs. 4.5.2. - 4.5.14.) cloud organization at mesoscale levels held a strong correlation with precipitation intensity and distribution. Tompkins (2001) reported that the role of propagating cold pools was determined important to the spatial organization of tropical deep convection. Similar conclusions were found in this data set as the collocation reveals that nearly all of the pixels with reflectivity factor \(\geq 7\) dBZ were in deeper clusters of cumulus aligned along arcs, apparently associated with propagating cold pools. These cold pools had a typical length dimension of 30-200 km and were characterized by a large arcing ring of deeper convection (up to 5 km in depth) ahead of clear region. When compared to wind parallel cloud streets and small cumulus cluster (Fig. 5.6.1.), it is clear that these clouds were responsible for nearly all of the measurable precipitation of the collocated data.

5.7. Area Averaged Rain Rate and Latent Heat Flux

To summarize the precipitation characteristics of the collocated data, an area-averaged rain rate was calculated using Eq. 5.7.1.,
Fig. 5.6.1. Typical cloud mesoscale organization. Top left) wind parallel cloud streets; Top Right) small cumulus clusters; Bottom) deep cumulus clusters along propagating cold pools. The top two cloud types were rarely associated with reflectivity greater than 7 dBZ suggesting the deeper clusters forming along cold pools are the main precipitation producers.

\[
\overline{R} = \frac{\sum R_{i,j} A_{i,j}}{A_{\text{total}}}
\]  

(5.7.1.)

where \( R_{i,j} \) is the rain rate observed over pixel \( i,j \) with area \( A_{i,j} \). These products were summed and then divided by \( A_{\text{total}} \), which was the total area in the radar domain less the area occupied by
birds, islands and bad beams. The area averaged rain rate was only calculated for pixels with reflectivity factor $\geq 7$ dBZ and for the collocated data, it was 1.30 mm day$^{-1}$.

As discussed in Chapter 1, shallow convection’s role in global circulation is an evaporator that modifies marine boundary layer air for later ingestion into the ITCZ. However, precipitation does fall from shallow convection and determining the efficiency of these clouds at returning water to the ocean surface is crucial to properly representing them in the water budget of the tropics. Therefore, using Equation 5.7.2., adapted from Rauber et al. (1996), the latent heat flux to the atmosphere through precipitation was calculated.

$$LHF = \frac{L_v \rho_w \sum R_{i,j} A_{i,j}}{A_{total}}$$ (5.7.2.)

The latent heat flux is calculated by multiplying the latent heat of vaporization, $L_v$, by the density of liquid water $\rho_w$ and the sum of the product of the rain rate $R_{i,j}$ and the area over which it fell, $A_{i,j}$, for all pixels with reflectivity factor $\geq 7$ dBZ. This quantity was then divide by $A_{total}$, which was the total area in the radar domain less the area occupied by birds, islands and bad beams. Eq. 5.7.2., describes the heat energy that was released due to condensation that forms precipitation. For the collocated scenes, the latent heat flux was 33.85 W m$^{-2}$.

To gauge the efficiency of shallow convection’s ability to return water to the ocean, the latent heat flux was compared to satellite observations of ocean-surface latent heat fluxes described by Chou et al., (1995). During the winter months over the RICO domain, the satellite estimated ocean surface latent heat flux ranges between 160-200 W m$^{-2}$. Given these numbers, the precipitation efficiency of the collocated data was between 17-21%.
CHAPTER 6
PROCESSING OF THE ENTIRE RICO RADAR DATASET

6.1. Motivation

The collocated data presented in the previous chapters represent thirteen snapshots at 10:45 AM (LT) into the life of trade wind clouds observed in the RICO region. To discover the representativeness of the collocated data, the remainder of the radar data observed during RICO were processed. This chapter investigates how the area averaged rain rate and latent heat flux vary as a function of day to discover the magnitude of daily variations in rainfall. Also, this chapter will reveal the diurnal cycle in precipitation to help put the 10:45 AM (LT) overpass data into the perspective of the hourly variations in rainfall. First, however, a detailed examination of the radar range effect in the presence of shallow convection is carried out to determine an acceptable range within which to process the radar data.

To fulfill the above objectives, all (26058) PPI and SUR radar files at the 0.5° elevation angle were examined. Over the 62 days that the radar operated, a variety of scan strategies were employed, but before the data was used to understand the project wide precipitation characteristics, the differences in these strategies were considered to eliminate biases due to preferential sampling. For example, for the first two days of the project, the radar continuously made 0.5° SUR scans. However, throughout the majority of the project, the volume scan strategy included only one 0.5° scan per radar volume. Thus the beginning of the project contained many more 0.5° scans. Furthermore, the radar typically scanned in only one hemisphere (i.e., 0-180° or 90-270° etc.) based on the wind direction, which therefore changed everyday (RICO Scientific
Overview Document). Also, numerous 0.5° SUR scans in the beginning of the project were made at a slower scan rate, which effectively over-sampled the radar domain by double. Finally, there were several periods of time when the radar was not operational.

To eliminate the bias in the statistics due to the irregular time and area sampling, the results reported in this chapter were normalized so not to allow any one time period to bias the results. To equally represent the data as a function of time, the files were grouped into half hour time periods for each day. Then, all files within a half hour time period per day were combined so that only one value (i.e., area averaged rain rate) was reported for each time period and that value represented the data collected during that time. Then a matrix was constructed from the normalized data with the dimensions of 62 X 48, where 62 is the number of days and 48 is the number of half hour periods per day. Processing the data using the matrix eliminates the influence of the irregular sampling by allowing each half hour period from each day to be represented by all the data taken within that time frame. That way, if the 40th day had 15 SUR files taken from 6-6:30 PM, it would carry the same weight as the 6-6:30 PM slot on 42nd day where only 3 PPI files were taken. However, the radar did not operate continuously throughout the project and there were time periods on certain days that were not represented.

6.2. Resolving the Range Effect

To determine the influence of the range issues discussed in Section 4.6. on the observed precipitation field, data was stratified within five kilometer range rings. The goal of this section was to determine the range over which the radar measured precipitation was relatively unchanging. Several factors, including beam geometry, bird contamination and the presence of the islands contribute to the statistics of the precipitation derived from the reflectivity field.
Isolating the impact of each of these factors as a function of range was necessary to expose the precipitation signature in the radar data.

Fig. 6.2.1. shows the area averaged rain rate and rain fraction for reflectivity factor ≥ 7 dBZ as a function of the 5 km range rings. Close to the radar the area averaged rain rate was very high, then it dropped into a valley near 25 km, rose briefly and dipped again at 55 km and then after another short increase it fell again. Without any consideration, one might see these curves and suspect that the islands greatly influenced the precipitation, but upon closer inspection, one would discover that beam geometry and bird contamination also played an interesting role in the shapes of these curves.

![Fig. 6.2.1. The area averaged rain rate (blue) and rain fraction (red) for reflectivity factor ≥ 7 dBZ as a function of the 5 km range rings.](image)
To decouple these effects, the impact of the islands was broken down by range. Fig. 6.2.2. plots the total area (blue), total area less the area of the islands and birds (red) and the fractional coverage of the islands and birds (black) as a function of the 5 km range rings. Fig. 6.2.3. plots the individual fractional coverage of the birds (red), the islands (black), their combined effect (blue) and the rain fraction (brown) between 20-140 km. Within the first 15 km, the 20-90% of the total area within a range ring was occupied by the island of Barbuda. Between 55-70 km, Antigua occupied up to 4% of the total area and between 90-110 km, the islands of St Kitts and Nevis, Anguilla and Montserrat also took up about 4% of the total area.

Fig. 6.2.2. Total area (blue), total area less the area of the islands and birds (red) and the fractional coverage of the islands and birds (black) as a function of the 5 km range rings.
The impact of the islands on the area averaged rain rate was that rain over the islands was not included in the statistics. However, minor variations in the rain fraction line in Fig. 6.2.3. were possibly the result of clouds and precipitation that form due to the extra dynamic forcing from the islands. Furthermore, the islands may have also left a clear wake on there lee shores which was possibly the cause for the slight dip in the rain fraction line near 55 km, 95 km, and 120 km. Further support for this hypothesis was found in the variations in cloud fraction seen close to the islands in Fig. 5.4.1. and confirms this as a possible explanation for the behavior of the area averaged rain rate near the islands.

Fig. 6.2.3. Fractional coverage of the birds (red), the islands (black), their combined effect (blue) and the rain fraction (green) as a function of 5 km range rings between 20 km and 140 km.
A second possible explanation could be due to bird contamination. The fraction of the total area occupied by pixels flagged as bird decreased exponentially as a function of range. As the red curve in Fig. 6.2.3. shows, between 20-40 km the Frigate birds occupied 3-8% of the total area. As range increased the influence of the birds waned and beyond 60 km, less than 2% of the total area was flagged as bird. However because such a large fraction of the total area in the first 40 km was occupied by bird and because the filter was only 95% effective at removing them, bird contamination in the precipitation field will bias the area averaged rain rates high. The combined affect of the bird and island contamination suggests that measurements of precipitation within the first 25 km should not be considered in this analysis and the affect of this contamination is seen by the exponentially decreasing rain rates in the first 25 km.

The effect of beam geometry was that with increasing range, the contributing volume grew in width and height. Beam geometry can impact the precipitation field in two ways; 1) the rain area and, 2) the rain intensity. If the precipitation field was insensitive to beam geometry changes, the rain area should steadily grow as a function of range and match the growth rate of the radar pixels size. The rain intensity should however remain constant.

Rain area is defined by the area (in radar space) of all pixels in a given range ring with a reflectivity factor \( \geq 7 \) dBZ. The top of Fig. 6.2.4. plots the rain area (blue) and rain fraction (red) as a function of the 5 km range rings. In the first 90 km, rain area grew steadily with range, but beyond 90 km the curve flattened. This suggests that beyond 90 km the radar’s ability to detect precipitation decreased. From a beam geometry perspective, this was likely due to height of beam with respect to cloud size. Fig. 4.6.1. shows that past 85 km the contributing volume occupied nearly all of a typical trade wind cumulus cloud and the center of the beam was nearly half way between cloud base and cloud top. In addition, beyond this range, the precipitation
Fig. 6.2.4. Top) Rain area in m$^2$ (blue) and rain fraction in % (red) as a function of the 5 km range rings; Bottom) average reflectivity for pixels with reflectivity factor $\geq$ 7 dBZ.
particles that were falling from cloud base were not sampled which contributed to the reduction of the rain area and rain fraction beyond 90 km.

To test the range dependency of the intensity of the rain, the bottom of Fig. 6.2.4. plots the average reflectivity factor for all pixels with reflectivity factor $\geq 7$ dBZ. Within the first 25 km, the average reflectivity was high because the missed birds were bright in the reflectivity field which biased the average high. Between 25 km and 90 km, the average reflectivity increased steadily. This tendency was attributable to beam geometry effects. As the beam grew in width, and height, it preferentially sampled shallow convection. The beam geometry was such that between these ranges, the contributing volume grew so that it sampled large drops that were present in the updrafts as well as large drops that fell from cloud base. Close to the radar the beam sampled only rain that fell from cloud base, but as range increased both rain within cloud and rain falling from cloud were sampled thus increasing the average intensity with range. The peak at 95 km was due to a small portion of ground clutter from the island of St. Kitts and Nevis missed in the island removal process.

The conclusion of this range analysis is that between 25 and 90 km the sampling of the radar observed precipitation field was adequate. Outside of these ranges, the sampling had sufficient problems that made it inadequate for this study. Therefore, in the analysis that follows, all data processing was limited to these ranges.

6.3. **Area Averaged Rain Rate and Latent Heat Flux by Day**

Findings in Chapter 5 were based on thirteen separate days during the 62 day long project. To understand how representative these days were with respect to the rest of the project,
the area averaged rain rate and latent heat flux were averaged for each day (Fig. 6.3.1.). Highlighted in yellow on the x-axis of Fig. 6.3.1. are the days where MISR observations were made. It is clear from Fig. 6.3.1., that daily rainfall varied considerably during the field campaign. For example, there was a notable dry period from 27 November 2004 through 5 December 2004 when almost no rain was measured within the radar domain. Similarly, there were several days where the area averaged rain rate exceeded 5 mm day\(^{-1}\). When these days were sorted from rainiest to least rainy (Fig. 6.3.2.), the distribution appears somewhat exponential. Fig. 6.3.2. shows that 41% of the total rain that fell during RICO occurred during the three rainiest days. MISR did not sample these days. Also, 56% of the rain fell in the six rainiest days. Finally, only 9% of the total rainfall fell in the driest half of this distribution, where eight of the thirteen coincident scenes were observed. (Figs. 6.3.1 and 6.3.2 are in a separate file called chapter_6_figs.pdf)

6.4. The Diurnal Cycle in Area Averaged Rain Rate and Latent Heat Flux

Diurnal cycles in oceanic precipitation vary by region, type and proximity to land (see Nesbitt and Zipser, 2003 for a recent review). Over the RICO region, where shallow convection trapped within the boundary layer over the open ocean was observed, there was a strong nocturnal peak in the area averaged rain rate. In Fig. 6.4.1., the area averaged rain rate and latent heat flux from precipitation are plotted against time binned in half hour increments. Between the hours of 8 PM and 6 AM (LT) a steady increase in rain fall was observed. After sunrise, rainfall quickly dropped over 50% from its pre-dawn peak before rebounding in the middle afternoon. Although there are many explanations for the cause of the diurnal variations, the appearance of this curve suggests a strong tie with radiative forcing. During the nighttime, cooling at cloud top
Fig. 6.4.1. Area averaged rain rate $\text{mm day}^{-1}$ (blue) and latent heat flux to the atmosphere through precipitation $\text{W m}^{-2}$ (red) reported in half hour time bins in UTC and LT.

steepened the lapse rates within the boundary layer and promoted convective overturning. After the sun rises, the temperature profile responded by stabilizing until the sun heated the lower levels which reduced the stability and gave way to the mid afternoon resurgence in rainfall.

MISR observations occurred within the radar domain around 10:45 AM (LT) which corresponded to driest part of the day in this distribution. Therefore, the representativeness of the relationships found in the collocated data may not hold for the remainder of the day. To
determine if these relationships hold for the remainder of the day, satellite data from other instruments will have to be analyzed.

6.5. Project Wide Area Averaged Rain Rate and Latent Heat Flux

To summarize the precipitation characteristics from the analysis of all 0.5° PPI and SUR scans, the area average rain rate and latent heat flux were averaged for all files over the ranges of 25-90 km using the matrix described in Section 6.1. The area averaged rain rate for the project was 2.26 mm day\(^{-1}\) and the project latent heat flux was 50.94 W m\(^{-2}\). Since the main role of these clouds in the general circulation is to moisten and deepen the boundary layer, quantifying the amount of water removed from the boundary layer through precipitation processes allows for a better representation of the energy and water budget of the circulation. Therefore to calculate the local return of water to the ocean surface through precipitation, the latent heat flux to the atmosphere through precipitation was compared to the ocean-surface latent heat flux as measured by satellite and detailed in Chou et al., (1995). For the precipitation observed during the 62 days of radar operation, the average precipitation efficiency of the trade wind clouds was 25-35%.

However, as Fig. 6.3.2. suggests, a few very rainy events may have elevated the project precipitation statistics and therefore, to determine their influence, the six rainiest and six least rainy days were removed to find their contribution. Once removed, the area averaged rain rate dropped to 1.21 mm day\(^{-1}\) and the latent heat flux dropped to 31.51 W m\(^{-2}\). It was clear that the occasional heavy rain events provided nearly half of the rainfall observed during RICO.

The impact of the removal of these days on the diurnal cycle did not change the pattern of the pre-dawn peak in the rain fall, but it did reduce the magnitude over which the rainfall varied throughout the day. Fig. 6.5.1. shows the diurnal cycle with six rainiest and six least rainy days
Fig. 6.5.1. Area averaged rain rate mm day\(^{-1}\) (blue) and latent heat flux to the atmosphere though precipitation W m\(^{-2}\) (red) reported in half hour time bins in UTC and LT with the six rainiest and six least rainy days removed.

removed. Although the trend in this curve remained the same, the major difference was that the afternoon peak in rainfall was proportionally higher than before.

6.6. Conclusions

When compared to previous observations of trade wind clouds reported in Chapter 1, the rain rates observed during RICO were significantly higher. As detailed in Chapter 1 most
measurements of precipitation in this region during the winter months are found to be below 1.0 mm day$^{-1}$. When compared to the most recent observations by TRMM of precipitation over this region, rainfall in this region was nearly double. For example, Fig. 6.6.1. shows the December 2004 and January 2005 rain totals from the TRMM’s 3B43 “Best Estimate” precipitation product. TRMM’s 3B43 “Best Estimate” product uses an optimal contribution from its microwave, infrared and precipitation radar instruments to estimate rain rate on a 0.25° grid (http://daac.gsfc.nasa.gov/precipitation/TRMM_README/TRMM_3B42_readme.shtml). For December 2004 and January 2005 that average precipitation rate in a 3° X 3° box centered on the RICO region was 1.07 mm day$^{-1}$ and 1.03 mm day$^{-1}$ respectively. However it is difficult to compare these numbers directly with the RICO observations. First, the PR on TRMM used a different $Z-R$ relationship ($Z=148R^{1.75}$) to derive rain rate (TRMM Instructional Manual). Second, the PR was only sensitive to reflectivity factor above 17 dBZ and will therefore miss a significant amount of rain. Third, if TRMM missed any of the heavy rain days as seen in Fig. 6.3.1. due to its non-sun synchronous proceessional orbit, then its average rain rates would be expected to be lower given the large contribution of these days to the total rainfall.

Finally, when the RICO rain rates are compared to those measured by the GPCP, the rates were again found to be significantly higher. Fig. 6.6.2. shows that for a 5° X 5° box centered on the RICO region, the GPCP average rainfall for December 2004 was 1.36 mm day$^{-1}$ with an absolute error of 0.465 mm day$^{-1}$. For January 2005, the GPCP reported 1.14 mm day$^{-1}$ with an absolute error of 0.485 mm day$^{-1}$. Given the scarcity of gauge observations in this region and the use passive microwave techniques and IR brightness temperatures used to make these calculations, which are known to be inadequate for shallow convection, the lower rain rates and large absolute error were expected.
Fig. 6.6.1. Top) TRMM 3B43 monthly rain rates (mm hr\(^{-1}\)) for December 2004; Bottom) TRMM 3B43 monthly rain rates (mm hr\(^{-1}\)) for January 2005. The small white box near the white arrow denotes the 3° X 3° box near the RICO region.
Fig. 6.6.2. Top) GPCP Version 2 monthly precipitation rates (mm day\(^{-1}\)) for December 2004; Bottom) GPCP Version 2 monthly precipitation rates (mm day\(^{-1}\)) for January 2005. The white arrows point to the RICO region.
CHAPTER 7
SUMMARY AND FUTURE WORK

7.1. Thesis Summary

Shallow cumulus clouds are the archetypical form of moist convection and understanding their precipitation characteristics is important for properly representing these clouds in global tropical circulations. This study examined the relationships between cloud coverage derived from MISR and rainfall measured by the NCAR SPoK radar for thirteen coincident scenes. By collocating the coincident observations from these instruments, the characterization of the cloud and precipitation fields was compared at the pixel level. To put these scenes in perspective with the rest of the field campaign, the entire set of 0.5° scans from the full RICO radar dataset was analyzed.

Between 5.5% and 10.5% of the trade wind cloud area observed in the coincident scenes was associated with rain rates $\geq 0.1 \text{ mm hr}^{-1}$. Furthermore, between 1.5% and 3.5% of the cloudy area was producing rainfall at a rain rate $\geq 1.0 \text{ mm hr}^{-1}$. Rain falling from these clouds also exhibited a strong relationship to the mesoscale cloud structure. Based upon visual inspection, it was found that clouds forming along propagating cold pools were responsible for nearly all of the rainfall. The area averaged rain rate for the collocated data was 1.30 mm day$^{-1}$, which translated into a latent heat flux to the atmosphere through precipitation of 30.85 W m$^{-2}$. When compared to the latent heat flux from the ocean-surface, clouds from the coincident observations were 17-21% efficient at locally returning water to the ocean through precipitation.

To determine the representativeness of the collocated data, the remainder of the radar dataset was processed. The project wide area averaged rain rate was 2.26 mm day$^{-1}$, which was
nearly double that estimated from past observations. The latent heat flux to the atmosphere from this rain was 50.94 W m\(^{-2}\). The clouds were 26-32% efficient at locally returning water to the ocean though precipitation.

Daily rainfall totals varied substantially during the field campaign from completely dry days to days where over 20 mm day\(^{-1}\) of rain were measured by the radar. Eight of the thirteen coincident MISR observations were made in the drier half of the 62 day long field campaign. Within each day, a strong diurnal cycle was observed. Radiative cooling at cloud top apparently reduced stability during the nighttime hours and produced a strong pre-dawn peak in the precipitation. Solar heating also apparently reduced the stability of the boundary layer in the daytime and produced a small afternoon resurgence in rainfall. The thirteen coincident scenes were observed during the driest part of diurnal cycle around 10:45 AM (LT). Therefore, the combination of sampling during the driest part of the day and on the drier days of the field campaign caused the 1 mm day\(^{-1}\) difference in the collocated data when compared to the project wide statistics.

Finally, removing the six rainiest and six least rainy days from the project totals revealed that more half the project wide rainfall fell in the six rainiest days. This suggested that periodic heavy rain events contribute significantly to the precipitation statistics. Furthermore, when these days were removed from the diurnal cycle calculation, the afternoon rainfall peak grew considerably. However, the pre-dawn peak remained the dominate feature.

7.2. Future Work: Characterizing Evaporation Below Cloud Base

One of the fundamental assumptions in the study was that precipitation measured by the 0.5° beam was falling to the ocean surface. However, the beam geometry was such that beyond
60 km, the center of the beam was above cloud base. Therefore, reflectivity measured beyond that range was not necessarily from hydrometeors that would eventually reach the ocean surface. Furthermore, relative humidity below cloud base was less than 100% and evaporation in the rain shafts below cloud base could impact the precipitation statistics. To discover these effects, the impact of evaporation on the drop size distribution needs to be calculated over a range of relative humidity. Then the precipitation amounts can be scaled according the evaporation potential below cloud base.

7.3. Future Work: Height Analysis

MISR’s “Best Winds” stereo cloud-top height product was used to report a distribution of heights observed during RICO in Fig. 5.2.2. When the radar data were re-sampled to the cloud-top height data grid, some preliminary results were found as shown in Fig. 7.3.1. This normalized distribution was made using the rain-no-rain threshold of 7 dBZ, where pixels below the threshold were used to make the red curve and pixels above the threshold were used to make the blue curve. From this analysis, precipitating clouds exhibited a dominant modal height at 2750 m. Non-precipitating clouds had a bimodal distribution with a low peak at 1750 m and a higher peak at 6250 m. Building upon these relationships to understand why at a certain heights some clouds precipitate and others do not should be a focus of future work.

7.4. Future Work: Mesoscale Organization

The correlations drawn between the mesoscale organizations of the cloud fields in the coincident data with precipitation were a surprising outcome of this thesis. To solidify the relationships drawn in the collocated data, an analysis of the radar data to explore the area,
Fig. 7.3.1. Normalized distribution of the MISR “Best Winds” stereo height field for raining (blue) and non-raining (red) pixels.

The perimeter and shape of the rain field should be the topic of future work. If the relationships are robust, the implications to the modeling community may be far reaching. This might be the missing link modelers need to better match predicted precipitation amounts with observations.
REFERENCES


