DUAL-POLARIZATION RADAR CHARACTERISTICS OF CONVECTION IN HAWAI'I DURING HERO

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ABSTRACT

In the fall of 2013 a Doppler on Wheels (DOW) mobile radar was deployed to O‘ahu as part of the Hawaiian Educational Radar Opportunity (HERO). The project was one of the first dual-polarization field experiments to date performed in Hawai‘i. Dual-polarization radars send and receive pulses with both horizontal and vertical polarization, allowing them to retrieve hydrometeor characteristics in two dimensions. With this technology, it is possible to gather information about the size, shape and type of hydrometeors. Though the primary purpose of HERO was educational, it provided a unique opportunity to observe the convective environment of O‘ahu at very high spatial and temporal resolution. It is found that the high-resolution Doppler radar data enables a better characterization of the mesoscale characteristics of convection over O‘ahu than was available previously. Of the many storms and weather types observed during HERO, two convective cells were chosen to represent some of the most common weather patterns found on the island: a trade wind case on 24 October 2013, and a sea-breeze case on 27 October 2013. For the trade wind shower case, it was found that although both the maximum radar reflectivity ($Z_H$) and overall size grew larger as the shower approached land, the maximum differential reflectivity ($Z_{DR}$) did not increase much until encountering the mountains. As the storm passed directly over the radar site, the high-resolution vertical velocity structure of the convective core was observed. It is shown that the trade wind shower exhibited a bimodal positive velocity structure as it came onshore. As it passed to the west of the radar, a 7 m s$^{-1}$ updraft was observed at the center. For the sea-breeze storm, the convective lifecycle from initiation through dissipation was studied for a single cell. From convective initiation, it took 15-19 minutes to reach the convective peak, and another 6 minutes for heavy precipitation to fall at the surface. It is shown from CFADs of $Z_{DR}$ that a large mass of small drops are always located just above the high-reflectivity convective core. Larger drops are located within the core, and fall out quickly following the convective peak of the updraft. Joint PDFs in $Z_H$-$Z_{DR}$ space show that as the updraft reaches its convective peak elevation, drops grow large enough to be affected by breakup. $Z_H$ and $Z_{DR}$ were infrequently observed above 52 dBZ or 3 dB.
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CHAPTER 1
INTRODUCTION

The eight primary Hawaiian Islands are the southern part of an archipelago that stretches some 1,500 miles across the Central Pacific trade wind belt in a region of overall mean synoptic-scale descent. Despite an observed minimum in rainfall over the open ocean in trade wind regions, convective storms are a common feature of the Hawaiian environment. Extensive mesoscale interactions involving complex topographic features and diurnal circulations result in a wide variability in rainfall across the islands (Schroeder, 1993). Improvements to our understanding of precipitating storms in tropical regions, including their initiation and lifecycle through dissipation, would greatly benefit future weather forecasts for the Islands and the Pacific region.

Doppler radar has emerged as one of the few meteorological tools that can examine the complex structure of convection at adequate spatial and temporal resolutions. O‘ahu, the most populated island in the state, has coverage from two national Weather Surveillance Doppler Radars (WSR-88Ds), however their distance from the island limits their use in mesoscale studies. Few research radars have visited Hawai‘i in the past, due in part to the cost-prohibitive distance from the mainland United States. The most recent of these was the Hawaiian Rainband Project (HaRP) in 1990, which was conducted on Hawai‘i Island to the south (Chen and Nash, 1994).

A National Science Foundation (NSF) Educational Deployment of a Doppler-on-Wheels (DOW) mobile radar from 21 October 2013 to 13 November 2013 was the first major radar field project ever to occur on O‘ahu (Bell et al., 2015). The Hawaiian Educational Radar Opportunity (HERO) was envisioned as an opportunity for students from the University of Hawai‘i at Mānoa to gain hands-on field experience with a modern weather radar. Though primarily an educational mission, HERO provided an opportunity to observe the convective environment of O‘ahu at very high spatial and temporal resolution, and additionally, was one of the first ever dual-polarization field experiments conducted in Hawai‘i. Dual-polarization radars expand upon the ability of traditional radars by transmitting and receiving both horizontally- and vertically-polarized electromagnetic waves. Comparison of the return signals yields additional microphysical information including the size, shape, orientation and composition of precipitation particles (Kumjian, 2013).

A number of convective systems were observed during the HERO campaign including trade wind showers, sea-breeze storms, orographically-enhanced deep convection, and a heavy rain event associated with a frontal system. Together, these represent some of the most common weather features of the region. The intensity of convection in Hawai‘i is dependent on a number of factors, but is largely controlled by the variable presence of the trade wind inversion, as well as its intensity and depth. The height of the inversion is typically 2 km from the surface over Hawai‘i, although this height is known to vary from day to day. The prevailing northeasterly trade winds and their associated temperature inversion are the dominant feature of Hawaiian weather, blowing from
the east or northeast about 90% of the time during the summer. The axis of a semi-permanent subtropical high pressure ridge is often well to the north of Hawaii during this season (Fig. 1.1a). Trade wind weather in Hawai‘i is normally characterized by pleasant temperatures and humidities as well as a consistent diurnal variation in clouds and rainfall. Depending on the synoptic situation, the subtropical ridge axis can shift further south (Fig. 1.1b), cutting off the trade winds and weakening the inversion, or causing it to disappear entirely. The term Kona weather is often used to describe the light, variable winds sometimes present during this period. During a Kona weather-type situation, the diurnal land- and sea-breeze circulation, as well as the local topography become more important in initiating convection, which can often occur in the early to mid-afternoon when the sea-breeze is at its greatest (Schroeder, 1993).

Precipitation over O‘ahu, and in the Tropics in general, typically originates from clouds that are entirely warmer than freezing (Beard and Ochs, 1993). This is primarily due to the presence of the trade wind inversion, which caps convection well below the freezing level. Some of the earliest studies regarding the warm rain mechanism were conducted in Hawai‘i. Blanchard (1953a) collected drop size distributions (DSDs) using filter paper techniques from non-freezing orographic clouds on the windward side of Mauna Kea on Hawai‘i Island. He found that the maximum drop size in these clouds rarely exceeded 2 mm, and the DSD was generally characterized by extremely large number concentrations of smaller drops. The first ever radar study in Hawai‘i was conducted during the Warm Rain Project in summer 1965. Using the Cornell Aeronautical Laboratory (CAL) 3-cm Doppler radar, Rogers (1967) found a maximum updraft speed of 7 m s\(^{-1}\) in warm rain convective clouds.

Woodcock (1952) first hypothesized that sea salt aerosols were the major initiators of cloud droplet formation in marine environments due to their hygroscopic properties. Cloud droplets formed from aerosols grow first by condensation, and then by rapid accretion once they reach a certain size threshold. Using aircraft observations over the open ocean near Hilo, Takahashi (1981) found that this threshold was approximately 30 \(\mu\)m near Hawai‘i, and that the largest growth

![Figure 1.1: Climatological mean July (left) and January (right) wind direction and contours of sea level pressure for the Eastern Pacific. Pressure is in units of hPa above 1000 hPa. From Schroeder (1993).](image)
Figure 1.2: Schematic of the lifecycle of a typical convective-type storm as viewed by a radar performing an RHI scan. The lifecycle is idealized as a succession of times from initiation at $t_0$, to peak ascendance of the updraft at $t_2$, finally to precipitation fallout at $t_5$. The storm continues to dissipate for an undetermined length of time at $t_{n-1}$, until upper-level clouds are all that remain by $t_n$. Should ice processes occur within a storm, a bright band may be observed at the freezing level at $t_{n-1}$. From Houze (1981).

occurred at the cloud top. Drop sizes formed from this process were initially observed to reach between 1.5-2.5 mm at the cloud base of warm rain cumulus on Hawai‘i Island (Takahashi, 1977). Using aircraft observations conducted during the Joint Hawaiian Warm Rain Project in 1985, Beard et al. (1986) later found drop sizes as large as 4-5 mm in similar clouds. It was previously thought that such large drops could not commonly exist, because they would be destroyed by collisions with their neighbors. Johnson (1982) proposed that giant raindrops could form from ultragiant sea salt particles which grow from the accretion of smaller drops in updrafts. Alternatively, Rauber et al. (1991) proposed that recirculation of drops in eddies generated along updraft/downdraft shear zones near the cloud top could also form giant raindrops in warm rain convective clouds.

Convective systems exhibit distinct vertical cores of high radar reflectivity that rise and fall as the convection overturns. The vertical evolution of high reflectivity cores describes the lifecycle of convective storms. Houze (1981) ascribed six to seven unique stages to the convective lifecycle, from initiation through dissipation, as shown in Figure 1.2. From initiation at $t_0$ in the figure, precipitation particles are carried upward by strong updrafts and grow large due to collision-coalescence. Reflectivity is dependent on both drop concentration, and the sixth power of diameter ($D^6$), so high reflectivity cores are often present at the head of updrafts. At $t_2$, the largest particles overcome the updraft, and begin to fall back through the cloud as rain, until at $t_5$, when heavy precipitation reaches the ground. If the updraft extends beyond the freezing level, and ice-cloud processes take over, then the vertical profile may display a notable bright band at $t_6$, where melting ice aggregates temporarily have large diameters. Saunders (1965) used a 3 cm radar to measure the convective overturning time of tropical precipitation for single-cell storms from Barbados in the Caribbean. He found that intense convective cores grew from fractocumulus in 25-35 minutes, before descending.
2500 m in another 6-10 minutes.

The model of Houze (1981) is an idealization of the convective lifecycle, however real convective systems display variabilities that depend heavily on environmental conditions. In Hawai`i, convective systems are heavily modulated by a number of environmental factors. As a result, each island in the state contains an extensive diversity of unique microclimates, despite their relatively small sizes. Some of these can exist only a few miles apart. The unique ways in which the prevailing northeasterly trade winds interact with the island landscape are primarily responsible for the pattern and variety of climate regimes present in the Hawaiian Islands.

O`ahu consists of two mountain ranges, the Ko`olau and the Wai`anae, each the collapsed remnants of once immense shield volcanoes that stood before (Fig. 1.3). Since those prehistoric times, the remaining massifs have been eroded and shaped by the forces of nature into two ranges of short peaks connected by a single ridgeline, with the major axis roughly perpendicular to the mean east-northeasterly trade wind flow. The Ko`olau range stretches some 60 km along the windward coast, with an average height of around 500-600 m, while the Wai`anae range stretches along the leeward coast for about 35 km, with an average height of around 800-900 m. Orographic lifting of the moist trades produce distinct windward to leeward gradients in rainfall because fallout of orographic precipitation reduces the amount of moisture available for convection downstream (Fig. 1.4). Thermal contrasts on mountain slopes and along coasts driven by diurnal oscillations in solar heating and radiative cooling can enhance and alter this pattern.

Leopold (1949) was one of the first to study the relationship between the trade winds, topog-

![Figure 1.3: Map of O`ahu showing its two parallel mountain ranges, the Wai`anae (left), and the Ko`olau (right).](image-url)
raphy, and the diurnal thermal circulation in Hawai‘i. He found that the interactions and their resulting convection patterns depended primarily on the overall island land area available for diurnal heating and cooling, the aspect of coastline relative to trade wind flow direction, the shape and size of topographic barrier, and the intensity and height of the trade wind inversion. For very large topographic obstacles, where the altitude of the barrier is much greater than the height of the trade wind inversion, he found that lateral flow splitting forces the trade winds to go around the barrier rather than over it.

Island-wide air flow and precipitation patterns have been studied extensively for Hawai‘i Island, which includes the two largest mountains in the Hawaiian chain, Mauna Kea and Mauna Loa, each over 4000 m. Several researchers have noted the strong surface thermal circulation present on the island (Lavoie, 1967; Mendonca, 1969; Garrett, 1980), and numerous projects have been conducted there: Project Shower (1954), Warm Rain Project (1965), Joint Hawaiian Warm Rain Project (1985), and Hawaiian Rainband Project (HaRP, 1990). An observed nocturnal maximum in rainfall is often found near the windward coast just before sunrise (Leopold, 1949; Takahashi, 1977; Schroeder et al., 1977). Leopold (1949) proposed that the interaction between the nocturnal katabatic wind and trade wind flow resulted in a convergence zone at some distance from the associated topography. This hypothesis was supported by subsequent researchers (Lavoie, 1967; Garrett, 1980). Chen and Nash (1994) examined island rainfall during HaRP and found a steady

![Mean Annual Rainfall Island of O‘ahu](image)

**Figure 1.4:** Contoured mean rainfall over O‘ahu from 1978-2007, showing a maximum along the Ko‘olau ridgeline. From the 2011 Rainfall Atlas of Hawaii (Giambelluca et al., 2013).
eastward progression from late afternoon over the windward mountain slopes to the coast later at night, with a maximum just prior to morning. As the thermal circulation later shifted to a seabreeze-upslope regime after sunrise, the convective maximum moved inland, reaching a maximum in both rainfall and inland progression later in the afternoon.

In contrast to Hawai‘i Island, relatively few studies have concentrated on O‘ahu. Studying rainfall over Honolulu, Loveridge (1924) suggested that the observed nighttime maximums in rainfall over the city were due to nocturnal cloud-top radiative cooling. Leopold (1948) analyzed patterns of rainfall from a number of stations around O‘ahu and concluded that rainfall near the Ko‘olau range was mainly from trade wind orographic showers. He also noted that when the trade wind speeds were low, afternoon sea-breezes were more prevalent at southern and leeward stations. Lavoie (1974) employed a simple mesoscale model to simulate rainfall over O‘ahu for a number of varying trade wind regimes and found that under weak trades, rainfall was enhanced over the Wai‘anae range but diminished over the Ko‘olau. The opposite was true under strong trades, where rainfall was enhanced over the Ko‘olau but diminished over the Wai‘anae.

Hartley and Chen (2010) made use of the island’s extensive array of 68 Hydronet rain gauges and 13 wind measuring stations to study airflow and rainfall distributions over O‘ahu during undisturbed summer trade wind days. Their study found that wind speed, trade wind inversion height, and moisture content of the inflow were all important factors when considering summer trade wind rainfall. Generally speaking, rainfall over the Ko‘olau range was highest during strong trade wind days due to more extensive orographic uplift, and rainfall totals were greatest at the ridgetops. Orographic lifting however was not enough to explain the high rainfall totals found there because not all strong trade wind days produced significant rainfall. As was first noted by Blanchard (1953b), for ridgetops such as the Ko‘olau, whose heights are roughly similar to the mean height of the lifted condensation level (LCL), there is not enough time during ascent for collision-coalescence alone to initiate precipitation. Rather, the presence of trade wind cumuli or pre-existing trade wind showers are first required to kick-start the process (Woodcock, 1975; Hartley and Chen, 2010). Higher rainfall frequencies were also noted on days with elevated trade wind inversions, and were attributed to a deeper moist layer, which allowed trade wind cumuli to achieve greater heights. Similar to strong wind days however, not all days with elevated trade wind versions produced more orographic rainfall.

Along the windward coast, flow deceleration of the inland trade wind was apparent during the early morning hours in response to the island katabatic and land-breeze thermal circulation (Hartley and Chen, 2010). An early morning maximum in rainfall over windward coastal stations and over the Ko‘olau range was also detected. They hypothesized that the early morning maximum in rainfall was due to the combined effects of nocturnal cloud-top radiative cooling and flow deceleration, which would allow more time for collision-coalescence. In the central valley between the Ko‘olau and Wai‘anae ranges, trade wind rainfall was rather infrequent, having already been mostly lost to
fallout on the other side of the Koʻolau range. They did however note a relatively small maximum in hourly rainfall during the afternoon, attributed to the development of afternoon sea-breezes.

Nguyen et al. (2010) used the MM5/LSM model to simulate island-wide airflow and weather patterns during summer trade wind conditions. They found that there were considerable spatial variations in temperature, moisture and winds across Oʻahu, and that these variations were related to changes in topography, ground cover, and the distributions of clouds and rain. Over central Oʻahu they found a significant diurnal contrast in temperature, due in part to daytime heating, and also katabatic flow from the Waiʻanae mountains at night. Weaker surface winds were also found in the island interior as compared to the open ocean, particularly at night. Along the windward coast, temperature and moisture variations were comparable to open ocean values. Flow deceleration was the most significant in the early morning, when the land surface was the coolest. They attributed an early morning rainfall maximum on the windward side to anomalous rising motion associated with the flow deceleration.

Although much research has been committed to the study of trade wind showers in the Tropics, there is much that could still be learned from a high-resolution radar study. Sea-breeze storms in island interiors have received comparably less attention. Together, these represent two of the more dominant convective types on Oʻahu. The ability of Doppler radar to probe the 3-dimensional structure of convection at high spatial and temporal resolution could lead to a better overall understanding of the mechanics of these storms. Figure 1.5 shows the same storm over Oʻahu observed at nearly same time from the DOW (left) and from PHMO, the nearest WSR-88D radar on neighboring Molokaʻi Island (right). The storm is a typical convective single-cell, with a core of high-reflectivity

![Figure 1.5: Reflectivity returns from the same storm observed over Oahu on 27 October 2013 from the DOW (left) and from PHMO (right) approximately two minutes earlier. Range gate spacing at this time is 60 m for the DOW and 250 m for PHMO.](image)
at the center. At this time, the storm is approximately 12 km from the DOW and over 100 km from PHMO. The azimuthal spacing between 1° beams at these distances is 209 m for the DOW and 1745 m for PHMO. The convective core is better resolved with DOW, due largely to the shorter range to the cell.

Dual-polarization radar returns provide an additional microphysical context to the high-resolution radar observations provided by the DOW. Differential reflectivity (Z_{DR}) is defined as the difference between radar reflectivity at horizontal (Z_H) and vertical (Z_V) polarization in logarithmic units. When observing spherical scatterers, positive values indicate that the horizontal axis of the mean drop size is larger than the vertical, an indication of large, oblate drops. Although Z_H by itself does give some information about the DSD due to the D^6 relationship of reflectivity on drop diameter, an exact characterization is not possible from traditional radar moments alone. When combined with Z_{DR}, a more explicit description of the DSD is possible. Brandes et al. (2004) calculated several empirical relations for a number of DSD parameters from the definitions of Z_H and Z_{DR} and from radar observations. For the median drop diameter (mm), he found that:

$$D_o = (0.171)Z_{DR}^3(0.725)Z_{DR}^2 + (1.479)Z_{DR} + 0.717$$ (1.1)

When changes in Z_H and Z_{DR} with time are plotted together, microphysical processes may be inferred. Kumjian and Prat (2014) identified four distinct regions of ΔZ_H-ΔZ_{DR} space relating to microphysical processes (Fig. 1.6). A positive change in both Z_H and Z_{DR} is indicative growth of drops and therefore collision-coalescence. Negative changes in both Z_H and Z_{DR} are indicative of precipitation fallout, where the largest drops rain out first. The DSD in this case is characterized by fewer and smaller drops. Negative changes in Z_{DR} in general could also be a sign of collisional breakup of large drops. A positive change in Z_{DR}, combined with a negative change in Z_H represents evaporation, due to the removal of smaller drops from the DSD. This could also indicate size sorting of larger drops if the area sampled is immediately below the downdraft. Finally, a negative change in Z_{DR} with a positive change in Z_H is thought to represent drop breakup.

In this study two convective storms on O‘ahu are analyzed using a modern, high-resolution, dual-polarization radar — a trade wind shower on the windward side of the island, and a sea-breeze storm in the central valley. For both cases, the lifecycle of convection is analyzed in horizontal and vertical cross-sections of Z_H and Z_{DR} which detail the fine-scale evolution of the storm to maturity. For the trade wind shower, the storm passed immediately over the radar during a convectively active phase, allowing for high-resolution vertical scans of the updraft. For the sea-breeze case, a single cell within the multi-cell storm is analyzed in detail. In addition to the direct analysis of Z_H and Z_{DR}, the detailed Doppler velocity structure of the storm from 1-4 km is also analyzed. Contoured frequency-altitude diagrams (CFADs) are generated for Z_H and Z_{DR}. Because Z_{DR} is related to the median drop size distribution, the CFADs help to reveal the vertical microphysical evolution of different drop sizes in the storm. Finally, joint probability density functions (PDFs)
are generated in $Z_H-Z_{DR}$ space. A detailed analysis of the PDFs is performed in which the various microphysical processes involved in the maturation of the cell are inferred.

In Chapter 2, the process of quality control and gridding is detailed extensively. The procedure is separated into multiple sections including an azimuth correction, calibration, attenuation correction, radar variable quality control, and finally, the procedure used to convert whole scanning volumes to 3-dimensional grids. In Chapter 3, results and analysis performed on the collections of gridded volumes from either case are presented. Finally, the project and important results are summarized in Chapter 4.
CHAPTER 2
DATA AND METHODS

The radar data used in this study was collected during HERO on the island of O‘ahu in Hawai‘i from 22 October 2013 to 13 November 2013. The project included the use of the DOW7 – a mobile, X-band dual-polarization radar operated by the Center for Severe Weather Research (CSWR). An expanded list of the DOW7 radar characteristics is provided in Table 2.1. Radar observations from PHMO, a WSR-88D radar located on neighboring Moloka‘i Island were also used. A total of 14 intensive observing periods (IOPs) were conducted throughout the project at various locations around the island, however only the first and second of these are used in this study. Daily forecast meetings gave guidance to the deployment strategy of future IOPs. Eight separate deployment locations around the island were scouted prior to the HERO campaign. IOPs were targeted for early morning or early afternoon, and would last for six hours on station. Radar teams were composed of a mix of graduate and undergraduate meteorology students from the University of Hawai‘i at Mānoa, CSWR staff, National Weather Service (NWS) personnel, and the primary investigator Dr. Michael M. Bell. Decisions regarding scan strategy were made ad hoc by radar team personnel, including choices of scanning angles. Both plan-position indicator (PPI) and range-height indicator (RHI) scans were advantageously utilized. During a PPI scan, the elevation angle of the radar dish is held constant as it is rotated in a full circle. During an RHI scan, the azimuth of the radar dish

<table>
<thead>
<tr>
<th>Table 2.1: DOW 7 radar specifications.</th>
</tr>
</thead>
<tbody>
<tr>
<td><strong>Transmitter</strong></td>
</tr>
<tr>
<td>Frequency 1</td>
</tr>
<tr>
<td>Frequency 2</td>
</tr>
<tr>
<td>Transmitter Type</td>
</tr>
<tr>
<td>Peak Power</td>
</tr>
<tr>
<td>Pulse Length</td>
</tr>
<tr>
<td>PRF</td>
</tr>
<tr>
<td>Nyquist Interval</td>
</tr>
<tr>
<td><strong>Antenna</strong></td>
</tr>
<tr>
<td>Diameter</td>
</tr>
<tr>
<td>Beam Width</td>
</tr>
<tr>
<td>Peak Scanning Rate</td>
</tr>
<tr>
<td>Polarization</td>
</tr>
<tr>
<td>Controller Software</td>
</tr>
</tbody>
</table>
is held constant, but the elevation angle is adjusted between 0° and 90°. A number of radiosondes were launched during the HERO campaign, although some of these failed during flight (Bell et al., 2015).

Once the HERO campaign was complete a comprehensive post-processing was necessary before the data could be analyzed. The post-processing began with a correction to the azimuth angles in the original sweep files to orient them to true north, based on the truck heading. In one case, additional corrections beyond the orientation to north had to be made due to errors in the ray azimuth data. Following the azimuth correction, the next step was to correct a persistent error that had resulted in erroneously negative values of differential reflectivity. This involved the application of a calibration correction to the raw reflectivity and differential reflectivity sweep files, based upon calibration errors calculated from recalibrated IOPs. Next, an attenuation correction was applied. Following the attenuation correction, a comprehensive quality control was applied to eliminate ground clutter and other non-weather radar echoes, and lastly, the data was cropped and interpolated to a 3-dimensional grid for final analysis.

2.1 Azimuth Correction

The DOW7 utilizes the National Centers for Atmospheric Research (NCAR)-developed TITAN and Hawkeye signal processing suites to provide real-time data archiving and display for the radar operator. Following the processing of the raw signal into radar moments, archived radar data is stored in the Doppler Radar Data Exchange (DORADE) format. Following the completion of the HERO campaign, the beginning phases of the post-processing utilized NCAR solo3 software, which is capable of viewing and editing of DORADE sweep files.

Before any quality control or attenuation corrections could be applied to raw data, it was first necessary to re-orient the radar data to true north to make them easier to analyze in the final product. In the raw sweep files, 0° was taken as the forward heading of the radar truck. The truck heading was available in the HERO log files so for the most part, this was a simple matter of applying a constant azimuth correction to the raw data. The truck heading for 24 October 2013 was 347°, so a -13° azimuth correction was applied. For 27 October 2013, the truck heading was 336°, so a -24° azimuth correction was applied. This would have been adequate, had it not been for a number of unexplained azimuth deviations in many of the early sweep files on this day. It was apparent that the errors existed because persistent ground clutter echoes from the nearby mountain ranges appeared to change location from volume to volume\(^1\). There was no indication that there were any deviations within the volumes themselves however, even amongst sweeps whose high elevation angles forbade them from seeing ground clutter. The vertical profile of related meteorological echoes between scans was consistent with the observed wind shear pattern of the environment. Whatever the reason for the errors, they seem to have been corrected sometime around 2132 UTC, after

\(^1\) A volume is a complete set of consecutive radar scans of a single type. A volume number is the index number assigned to that set of scans.
Table 2.2: A list of the azimuth corrections applied to selected scans on 27 October 2013, listed by volume number, scan type, angle, and time.

<table>
<thead>
<tr>
<th>Vol. Number</th>
<th>Scan Type</th>
<th>Scan Angles</th>
<th>Start Time (UTC)</th>
<th>Stop Time (UTC)</th>
<th>Az. Corr.</th>
</tr>
</thead>
<tbody>
<tr>
<td>165</td>
<td>PPI</td>
<td>0.5—36.5</td>
<td>20:44:22</td>
<td>20:47:56</td>
<td>+2.0</td>
</tr>
<tr>
<td>166</td>
<td>PPI</td>
<td>0.5—38.5</td>
<td>20:48:38</td>
<td>20:52:23</td>
<td>+2.0</td>
</tr>
<tr>
<td>168</td>
<td>PPI</td>
<td>0.5—38.5</td>
<td>20:54:52</td>
<td>20:58:38</td>
<td>-3.0</td>
</tr>
<tr>
<td>170</td>
<td>PPI</td>
<td>0.5—14.5</td>
<td>21:01:08</td>
<td>21:02:30</td>
<td>-7.4</td>
</tr>
<tr>
<td>170</td>
<td>PPI</td>
<td>16.5—38.5</td>
<td>21:02:42</td>
<td>21:04:52</td>
<td>-21.7</td>
</tr>
<tr>
<td>172</td>
<td>PPI</td>
<td>0.5—38.5</td>
<td>21:07:22</td>
<td>21:11:07</td>
<td>-30.4</td>
</tr>
<tr>
<td>174</td>
<td>PPI</td>
<td>0.5—38.5</td>
<td>21:13:37</td>
<td>21:17:22</td>
<td>-30.0</td>
</tr>
<tr>
<td>176</td>
<td>PPI</td>
<td>0.5—38.5</td>
<td>21:19:30</td>
<td>21:23:15</td>
<td>-30.0</td>
</tr>
<tr>
<td>178</td>
<td>PPI</td>
<td>0.5—38.5</td>
<td>21:25:22</td>
<td>21:29:07</td>
<td>-30.3</td>
</tr>
<tr>
<td>180</td>
<td>PPI</td>
<td>0.5—8.5</td>
<td>21:31:17</td>
<td>21:32:28</td>
<td>-19.9</td>
</tr>
<tr>
<td>180</td>
<td>PPI</td>
<td>16.5—20.5</td>
<td>21:33:40</td>
<td>21:34:15</td>
<td>+20.0</td>
</tr>
</tbody>
</table>

which there is no observable deviation in azimuth between volumes. Following the correction for truck heading, a comparison of ground clutter echoes of the 0.5° elevation scans after 2132 UTC to actual topography revealed that these volumes were correctly oriented. The strategy for correcting the earlier volumes was therefore to make adjustments to the azimuth angles within each volume, so that the location of ground clutter echoes of the lowest elevation angle scans matched their locations in the later volumes. The final set of azimuth corrections for each of the relevant volumes from 27 October 2013, not including the corrections for truck heading, are given in Table 2.2.

2.2 Calibration

Following the azimuth correction, the next step in the post-processing was to correct a persistent error that resulted in anomalously negative values of differential reflectivity in all scans throughout the HERO campaign. The unusual values were first noticed during the campaign, and were apparent because not only were they ubiquitous throughout the dataset, they did not correspond to any known meteorological phenomenon. Negative values of differential reflectivity indicate vertically-oriented particles, which are not very common in convective storms except in certain regions above the freezing level, where wind-blown ice crystals may have an overall vertical orientation. By contrast, liquid hydrometeor particles are normally horizontally-oriented, owing to the distorting effects of wind resistance in updrafts and downdrafts. Negative $Z_{DR}$ have also been observed in hail shafts, where a few slightly oblate hail stones may occasionally become vertically oriented (Bringi and Chandrasekar, 2001). In neither case were these meteorological conditions consistent.
with what was observed in the HERO dataset. Figure 2.1 shows RHI scans of the thunderstorm observed against the Wai‘anae Range during a period of extreme convective activity. Figure 2.1a shows the unaltered, horizontally-polarized reflectivity while Figure 2.1b shows unaltered differential reflectivity. The most striking feature behind the $Z_{DR}$ scan in Figure 2.1b, is the long shaft of strongly negative values (blues and greens) extending through a 5 km depth near the center of the storm. If this was indeed a hail shaft, it should correspond well with high values of reflectivity in Figure 2.1a, which it does not. Additionally, the upper levels of the storm seem to actually indicate anomalously positive $Z_{DR}$ values (yellows and reds), which is not typically the case above the freezing level.

Initially, it was thought that extreme attenuation in the horizontal channel might be responsible for the negative values. In a heavily convective environment where large oblate drops are common, one would expect a more attenuated horizontal channel, resulting in some negative $Z_{DR}$ bias. This was not able to explain however, why in some cases, $Z_{DR}$ values as low as -2 dB were often seen near the radar, or why overall positive values of $Z_{DR}$ were present aloft. Additionally, an attenuation correction performed on the data was unable to resolve the error.

Eventually, it was determined that the anomalously negative differential reflectivities were due to an apparent accidental switching of the horizontal and vertical channels that had occurred at some point before signal processing. In the case of the DOW, the return power for each polarization
channel is calculated from filters which separate a 45° polarized return signal into horizontal and vertical components, and then from the known sensitivity of each receiver. After the return power is obtained, reflectivity is calculated by adding a radar constant and also adding twenty times the log of the range. The radar constant of a radar is a combination of radar system parameters and physical constants that determines the proportionality factor between radar reflectivity of a target at a given range and the power measured at the receiving antenna. Some of the parameters evaluated in the equation determining the radar constant include the peak power, the gain of the antenna, beamwidth, pulse duration, pulse width, wavelength of the radar transmission, etc. (Rinehart, 2010). Since the radar constant is different for each polarization, switching the channels resulted in the addition of incorrect constants to each channel’s return power, and ultimately, reflectivities that were not only incorrect due to the switching, but were also miscalculated by several dBZ. In the case of differential reflectivity, the quantity was both inverted and miscalculated.

Ultimately, the only way to truly correct the erroneous reflectivities with fine precision is to recalculate the reflectivity and differential reflectivity from each channel from the raw In-phase and Quadrature (I & Q) data that was sometimes recorded in radar time series on site. As recording I & Q data uses a demanding quantity of computer resources and storage, this option was not commonly chosen. Of the two IOPs examined here, only on 27 October 2013 is the time series data available, and for the cell in question, only in part. It is therefore not possible to exactly recalculate the reflectivity and differential reflectivity for the cells in question, however it is possible to come to a close enough approximation (within plus-or-minus 1 dBZ for reflectivity, within plus-or-minus 0.1 dB for differential reflectivity) by means of calibration. Essentially accurate horizontal and vertical reflectivities could be calculated by first switching the channels back to their correct polarization, and then adding a constant calibration correction to each channel.

The calibration corrections themselves would be constants, because the radar constants that were misapplied were also constants. The calibration constants for the horizontal and vertical channels, as well as the differential reflectivity, were calculated by comparing raw radar returns to recalibrated time series data. The time series chosen for the recalibration were recorded during IOPs on 27 October 2013 and 7 November 2013. On 27 October 2013, the collection of data encompassed 69 volumes for 2.5 hours of continuous collection; on 7 November 2013, 52 volumes for 3 hours. Reflectivity and differential reflectivity was recalculated from the time series data (with horizontal and vertical switching corrected) for these dates using a series of scripts that approximated the calculation that would have been applied by the DOW during the radar operation.

Following the recalculation, a quality control was applied to both the recalculated and the raw radar returns to eliminate ground clutter. The quality control performed on the data is discussed later in section 2.4. Following the quality control, the raw returns were subtracted from the recalculated returns. Next, all of the data was converted from log form to linear form and averaged, and then converted back. The averages were taken to be a measure of the calibration
constant for that particular scan.

For the 7 November 2013 dataset, many of the scans had to be removed, owing to the scarcity of meteorological echoes on that day. Although the calibration correction should theoretically be the same for each scan, some residual noise not removed by the quality control contributed to some degree of variation. Despite the variations, the mean values for each day agreed well with each other once the majority of the noise was removed by the quality control, and standard deviations were within 1-2 dBZ. Histograms of the data shown in Figure 2.2 reveal a narrow, single peak frequency distribution, as would be expected from a single, uniform calibration constant. The final calibration constants were taken to be -8.84 dBZ for the horizontal channel, -7.3 dBZ for the vertical channel and 0.28 dB for the differential reflectivity.

Lastly, the calibration was ready to be applied to the azimuth-corrected returns in the post-processing. The horizontal and vertical channels were first switched back to their correct nomenclature, and the differential reflectivity was multiplied by negative one. Next, the calibration constants

![Calibration Bias Frequency Distributions](image)

**Figure 2.2:** Histograms of bulk radar data collected on 27 October 2013 (top) and 7 November 2013 (bottom) showing the recalculated from time series $Z_H$ minus raw $Z_V$ (left), recalculated $Z_V$ minus raw $Z_H$ (middle), and recalculated $Z_{DR}$ minus raw $Z_{DR}$ (right). From left to right, the histograms represent the calibration corrections for $Z_H$, $Z_V$ and $Z_{DR}$.
were added to the data. An example of the results of this procedure are shown in Figures 2.1c-d. In Figure 2.1c, the unaltered, horizontally-polarized raw returns in 2.1a are replaced with the unaltered vertical returns, and the -8.84 dBZ calibration constant was added. In Figure 2.1d, the unaltered differential reflectivity is multiplied by negative one, and the 0.28 dB calibration constant is added. What was before an unexplained, anomalously negative $Z_{DR}$ shaft is now revealed as a narrow column of high $Z_{DR}$ extending upwards to the freezing level. This column is in the flank of a very high reflectivity, low $Z_{DR}$ region, suspected to be a shaft of graupel or small hail embedded in the primary updraft.

### 2.3 Attenuation Correction

Following the calibration correction to the raw radar returns, an attenuation correction was applied to the reflectivity and differential reflectivity fields. Attenuation is a common problem in radar meteorology, and is a consequence of the fact that a propagating radar beam will experience some loss of power due to scattering and absorption by hydrometeors and aerosols. The amount of power loss is dependent on the number and size of the particles that the beam passed through before returning to the radar dish, and also the wavelength of the radar. X-band radars, such as the DOW, suffer far more from attenuation than do S-band radars, such as the WSR-88D, due to forward scattering by hydrometeors at the 3-cm wavelength. Although inherently a fundamental part of radar physics, attenuation is not accounted for in the radar equation because it depends on environmental parameters. Instead, it must be handled by specialized algorithms in post-processing.

The algorithm used in this study to correct attenuation is based on the simple approach described in Gorgucci and Chandrasekar (2005). The attenuation correction procedure described therein assumes a linear relationship between specific attenuation and specific differential phase ($K_{DP}$). The differential phase shift of a radar beam describes the change in phase resulting from the deceleration of electromagnetic energy as it passes through mediums populated in varying degrees by hydrometeors. This deceleration effect is indirectly linked to attenuation, since both power loss and phase shift are modulated by the same environmental parameters. The effect is cumulative, so there is a close relationship between the differential phase and the total attenuation. The range derivative of the differential phase shift, called the specific differential phase, is a measure of the instantaneous phase shift at any one point, and likewise is analogous to specific attenuation. The specific attenuation is given as

$$\alpha_h = a_h K_{DP}$$  \hspace{1cm} (2.1)

where $\alpha_h$ is the specific attenuation, and $a_h$ is a bulk constant determined by a number of factors, including temperature, drop size distribution (DSD) and the mean shape of raindrops. While there is a slight variation in the bulk constant with DSD, it is common to use a single constant for a wide range of DSDs.
The attenuation correction was applied in the post-processing via RadxPartRain. The Radx C++ software package, which includes RadxPartRain, was developed at NCAR as part of the Lidar Radar Open Software Environment (LROSE). An array of programs included in the package are equipped to perform a variety of tasks, including radar and lidar analysis, interpolation to a common grid, and translation between various radar data formats. RadxPartRain is primarily an analysis tool, capable of attenuation correction, hydrometeor classification, and rain rate estimation. An example of the results of the attenuation correction performed by RadxPartRain is shown in Figures 2.1e and 2.1f, where the effect is an overall slight increase in both reflectivity and differential reflectivity.

2.4 Quality Control

The next step in the post-processing after the attenuation correction, was an extensive quality control to remove non-meteorological echoes. Before dual-polarization became a common component of modern radars, attempts to remove ground clutter from radar returns relied on filters which used a variety of specialized techniques, such as regression (Torres and Zrnic, 1999). In most cases, ground clutter returns were expected to have a very narrow spectrum width and zero Doppler velocity. While these techniques often produced very reliable clutter removals, the advent of dual-pol ever-expanded the arsenal of tools available to remove ground clutter. Correlation coefficient for example, which measures the degree of correlation between the horizontal and vertical pulses, is typically low when ground clutter is detected. Meteorological echoes will have a high correlation coefficient ($\rho_{HV}$) because their scatter patterns are highly correlated, indicative of spherical particles.

The quality control used to eliminate ground clutter echoes in this case made use of both dual-pol and traditional Doppler radar moments. At the first stage, radar echoes with a correlation coefficient lower than 0.85 were assumed to be clutter and removed, so as not to preclude the possibility that graupel and small hail might exist somewhere in the radar returns. Liquid hydrometeors, whose shapes are more spherical with decreasing size, have very high correlation coefficients, typically greater than 0.95. Graupel and small hail on the other hand, typically have correlation coefficients greater than 0.85, increasing with decreasing wetness. Only larger hailstones, by virtue of their distorted shapes, produce $\rho_{HV}$ values lower than this threshold, and those were not expected in the shallow convective environments characteristic of either IOP.

In the second stage of the quality control, radar echoes with a normalized coherent power (NCP) less than 0.3 were assumed to be clutter and eliminated. NCP is defined as the ratio of the return power in the first moment (called the coherent power) to the total returned power (Bell et al, 2013). Essentially, NCP is a measure of the correlation between the returned powers in individual samples, so clutter echoes will return a much lower correlation. The third stage of the quality control took advantage of the zero Doppler velocity characteristic of ground clutter.
Doppler velocity explicitly describes the radial velocity of a radar target. Ground clutter echoes, being unmoving in any direction, have a zero Doppler velocity, but so do meteorological targets moving with a vector perpendicular to the radar beam. To prevent the quality control from also eliminating these targets, near-zero Doppler velocity signatures were made to correlate with very low $K_{DP}$. Following this step in the quality control, each scan was despeckled to remove any remaining orphaned radar echoes.

2.5 Gridding

The final step of the post-processing was the interpolation of whole radar volumes to Mercator grids. This was accomplished with Radx2Grid, another component of the LROSE package mentioned earlier. A 3-dimensional Cartesian grid with 101 vertical levels and 100 m horizontal and vertical resolution was chosen for each of the target grids. Multiple radar scans were needed to fill the 3-dimensional volume, so the collection of radar scans chosen for each grid were demarcated by shifts in scanning type. A single grid for instance might contain a group of simultaneous PPI scans, where lower elevations on the grid at distance from the radar were more representative of earlier scans, and higher elevations on the grid both near and far from the radar were representative of later scans. On average, the temporal range of a single PPI volume was about 3-5 minutes, so the gridded volumes are thought to represent the evolution of the storm during this time period, and care should be taken to make inferences based on instantaneous interpretations. Gridded RHI volumes were also created in situations where RHIs passed through the region of interest, however these proved of little use except in situations where the storm was very near the radar. This was attributed to the fact that RHIs performed on site typically had $10^\circ$ azimuthal spacing, so even if the radar scanned a storm in RHI mode, if it was more than a couple kilometers from the radar, very little material was available to recreate a 3-dimensional representation.
CHAPTER 3
ANALYSIS

3.1 Trade Wind Showers on 24 October 2013

3.1.1 Synoptic Conditions

The deployment on 24 October 2013 was the first official IOP of the HERO campaign, not including training and educational outreach deployments in Mānoa Valley Park on 22 and 23 October 2013. The overall forecast for O‘ahu called for a weak trade wind regime with lower than average boundary layer moisture. The 1800 UTC Global Forecast System (GFS) 1000 hPa analysis on 24 October 2013 showed 3-5 m s$^{-1}$ easterly winds over Hawai‘i, supported by an area of weak high pressure to the northeast (Fig. 3.1). A weak ridge associated with the area of high pressure was present just to the northwest of the Hawaiian Islands. To the northwest of the state and the ridge, the analysis showed a deep trough building across the central Pacific, associated with a low pressure system over the Aleutian Islands. At 600 hPa, 0-3 m s$^{-1}$ northwesterly winds were present over the Hawaiian Islands (Fig. 3.2). At 250 hPa, winds were greater than 20 m s$^{-1}$ and from the west (Fig. 3.3). A trough embedded in the subtropical jet was just to the west of Hawai‘i. The 1200 UTC radiosonde

![Figure 3.1: 1800 UTC GFS 1000 hPa temperature (°C), wind vector, and contours of geopotential height (m) for the North Pacific region on 24 October 2013.](image)
Figure 3.2: 1800 UTC GFS 600 hPa temperature (°C), wind vector, and contours of geopotential height (m) for the North Pacific region on 24 October 2013.

Figure 3.3: 1800 UTC GFS 250 hPa temperature (°C), wind vector, and contours of geopotential height (m) for the North Pacific region on 24 October 2013.
launched from Lihue on neighboring Kaua’i Island showed a modest amount of convective available potential energy (CAPE) (787 J/kg) through the depth of the troposphere (Fig. 3.4). Much of this CAPE however was not available to surface-based parcels due to a warm inversion layer from 3-5 km. The lifting condensation level (LCL) was found to be at 514 m.

3.1.2 Deployment Strategy

The best chances for rain and showers were expected in the early morning on the windward side of O’ahu, and would likely be supported by enhanced thermodynamic instability, due to nocturnal cloud-top radiational cooling (Loveridge 1924, Hartley and Chen 2010). The land-breeze circulation is also known to enhance early morning convective development on the windward side (Hartley and Chen 2010). The decision was made to deploy the DOW early in the morning to Kahalu’u Regional Park, which is located east of the Ko’olau Mountain Range (Fig. 3.5).

The site was initially chosen for limited ground clutter, freedom of use, and sweeping views of the mountain range and ocean. Kahalu’u Regional Park lies just off of Hawai’i Route 83, just north of Kāne’ohe. The park is within a few meters of Kāne’ohe Bay and the Pacific Ocean to the east. To the west, the Ko’olau Mountain Range stretches north-northwest to south-southeast, roughly parallel to the highway and the coastline. At its closest point to Kahalu’u Regional Park,
the mountain ridgeline is roughly 3.3 km away. On the windward side of the range, sheer cliffs dominate, providing an orographic barrier for the trade wind flow. Orographic enhancement is often the catalyst for substantial windward convection in Hawai‘i (Schroeder, 1993). Trade wind showers on the windward side are a frequent occurrence.

3.1.3 Local Weather Conditions

The HERO team arrived at the site and set up early, and the DOW began collecting data at 1550 UTC (5:50 a.m. HST), just before sunrise (1631 UTC). Mostly overcast skies were present at the radar site, and shallow cumulus was observed further offshore. Locally heavy rain fell soon after the start of the mission, and reemerged intermittently through the morning. A large number of showers were observed on radar in the first three hours after deployment. Following 1830 UTC, they became increasingly infrequent. A radiosonde launch was attempted, but failed due to problems with the launch system computer.

The DOW7 mast was lifted soon after arrival, and began to sample the environment at 10 m height in one second intervals. Figure 3.6 shows a time series plot of temperature, dewpoint temperature and relative humidity for the first four hours after deployment. Dewpoint temperature was not recorded by the mast, but was later calculated from temperature and relative humidity. Following the initial deployment, relative humidity remained fairly constant near 90% through about 1730 UTC, before falling thereafter. After 1800 UTC, relative humidity begins to fall more rapidly, reaching 67% by 1834 UTC. The overall pattern of decline in relative humidity is mirrored rather well by an increase in temperature during the same period. The initial mean surface temperature (averaged in the first 5 minutes) was recorded at 21.7 °C. Temperature increased rather slowly.
after sunrise, reaching only 23 °C by 1800 UTC. By 1834 UTC temperature had reached 27 °C. A change in thermodynamic parameters at the radar site occurs after 1830 UTC. Relative humidity suddenly declines at a much slower rate, reaching 55% by 1934 UTC. Temperature increases at a slower rate during the same period, reaching 29 °C by 1934 UTC. Dewpoint temperature remained more consistent during the 4-hour period, hardly fluctuating very far from 20-22 °C. Only a small but sudden 1 °C drop from 1830-1834 UTC seems to indicate a drying of the air. Otherwise, actual moisture content of the air appears to have remained relatively consistent throughout the period, despite the reduction in relative humidity.

Figure 3.7 shows a time series plot of wind vector separated into U- and V-components, recorded
from the mast, bounded within the same 4-hour time interval as Figure 3.6. Initial winds were recorded out of the west-northwest at 1.4 m s\(^{-1}\) and continued to be light and generally from the west, occasionally fluctuating between southwesterly and northwesterly, until abruptly dying off at 1722 UTC. At some time near 1730 UTC, the mast recorded a sudden weak surge of southeasterly winds, which gradually became highly variable through 1831 UTC. In the figure, the southeasterly surge is indicated by negative U-component and positive V-component winds after 1730 UTC, followed by a period of variable readings. The magnitude of the surge was no more than a 1-2 m s\(^{-1}\) increase in overall wind speed, which gradually subsided. The period from 1730 to 1831 UTC was highly variable in terms of both wind speed and direction, but soon became more consistent thereafter. The magnitude of the overall wind speed abruptly increased to 3-5 m s\(^{-1}\) after 1831 UTC, moving in steadily from the east-southeast. The figure shows that the V-component of the winds became consistently positive after 1831 UTC, while the U-component became consistently negative. Turbulence also appears to have increased after 1831 UTC, as evidenced by a more varied signal. Hourly mean wind speed and direction recorded from nearby surface wind stations on O'ahu show a similar pattern (Fig. 3.8). Wind speeds are generally light and variable early in the morning at 1700 UTC (Fig. 3.8b), but increase and become easterly by 2100 UTC (Fig. 3.8f). It is surmised that the period after 1831 UTC marks the abatement of the nocturnal flow regime and land-breeze circulation following sunrise at the radar site, and a transition to a daytime trade wind flow regime. This is an expected result for a mountainous tropical island under weak and variable synoptic flow. In the hours immediately following the transition at the radar site, a significant reduction in convective activity was observed by the DOW. While an early morning maximum in rainfall is expected due to nocturnal cloud-top radiative cooling, it is also possible that the interaction between land-breeze and trade wind may have had an influence.

### 3.1.4 Radar Observations

A number of storms formed offshore and were advected inland against the Koʻolau Range during the scanning period, including one storm that is the subject of this analysis. The storm passed directly over the radar, resulting in heavy precipitation at the radar site. The timing of the storm was contemporaneous with the period immediately following the transition to a daytime flow regime at the radar site mentioned earlier. The storm was observed from 1835-1906 UTC. The overall scanning strategy for the DOW was to alternate between PPI and RHI scans, with casual adjustments made periodically to the scanning angles to capture newfound areas of meteorological activity. During the period of interest, the DOW was operating with a 200 ns pulse width and 5000 Hz pulse repetition frequency, with 30 m range gates. Eleven degrees of elevation were scanned from 2° at the lowest, to 35° at the highest. The duration of one set of PPI scans was approximately 3 minutes, before the following RHI scans were initiated. Occasionally, a set of vertical scans was initiated following a set of PPI scans, if there was a storm present above the radar.
Figure 3.8: Hourly (1600-2100 UTC) surface winds (kts) for east O‘ahu on 24 October 2013. Wind speeds and direction are mean values computed from 30 minutes before to 30 minutes after each time. Data was collected from National Weather Service/Federal Aviation Administration (NWS/FAA) network, National Ocean Service Water Level Observation Network (NOS-NWLO), and from NWS Honolulu Weather Forecast Office (WFO) stations.
The relevant volumes and scanning angles that captured the storm are given in detail in Table 3.1. Following the intensive observation period, the quality control and interpolation detailed in the previous section was performed on the raw sweep files. Through the post-processing, whole radar volumes were interpolated to 3-dimensional Mercator projection grids with 100 m horizontal and vertical resolution. Figure 3.9 shows vertical cross-sections of the gridded horizontal reflectivity (ZH), cut through latitude 21.457 °N, nearly the exact latitude of the radar. The figure is divided into six panels, denoting the six time periods of interest. The topography at latitude 21.457 °N is delineated by a magenta line near the surface. The exact position of the radar is in the middle of each plot, at longitude 157.842 °W. The physical size of the storm, including its horizontal and vertical extent, and also its maximum reflectivity, are analyzed where possible.

Figure 3.9a shows the storm in its infancy, when it was approximately 6 km east of the radar. At this point, the storm is little more than a small pocket of cumulus extending from 1-3 km, with maximum reflectivity no greater than 6 dBZ. A comparison of different radar volumes showed that the system was moving westward toward the radar, with Doppler velocities ranging from 5-8 m s⁻¹ (not shown). Unfortunately, for unknown reasons, the next set of PPI scans from 1842-1845 UTC were corrupted (not shown), and no discernable radar echoes were gleaned from the volume. Furthermore, RHI scans before and after that time were directed towards other areas of interest against the mountains, so following 1838 UTC, the cell was not observed again until 1850 UTC.

By 1850-1852 UTC, nearly 12 minutes later, the situation is quite different; the storm has grown in physical size, and its position has advanced immediately upon the radar site (Fig. 3.9b). The west-east extent of the storm at the surface is at least to the middle of Kāne‘ohe Bay, a distance of about 3.5 km. The vertical extent of the storm is harder to gauge, but did not exceed 3 km. As a consequence of the scanning strategy in use at that time, angles greater than 35° were not scanned by the DOW in PPI, with the exception of vertical scans. As a result, much of the storm above 35° is missing from the gridded volume. The same can be said for other gridded volumes formed

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</tr>
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<td>10—35</td>
<td>19:05:05</td>
<td>19:06:52</td>
<td>1:47</td>
</tr>
</tbody>
</table>
Figure 3.9: Altitude-longitude diagrams of gridded $Z_H$ from the DOW, cut through 21.457° N, showing the evolution of the trade wind storm on 24 October 2013 at each of six different scanning periods. An arrow in each panel points to the radar site.

from PPI sets, namely the volumes formed from the 1857-1900 UTC set and from the 1905-1907 UTC set (Figs. 3.9d & 3.9f). By 1855-1856 UTC, the core of the storm had just passed the radar, as shown in Figure 3.9c. The maximum reflectivity detected within the storm at this time was 38 dBZ. The maximum height of the storm at this time did not exceed 3 km, the approximate elevation of the trade wind inversion. A vertical velocity of this particular timeframe is provided later in this analysis. By 1857-1900 UTC, the storm appears to have encountered the mountains to the west, and is tilted in the vertical (Fig. 3.9d). The tilt could be due to the encounter with the mountains, but could also be an artifact of the PPI compositing process. The lowest scanning angles are aged at least 2 minutes ahead of the highest scanning angles, which could result in a westward tilt with height of a vertical storm column. The maximum reflectivity detected during this period was 41 dBZ. Finally, by 1903 UTC, the storm crossed the top of the Ko’olau ridgeline (Fig. 3.9e). By 1905-1907 UTC, it has moved into the leeward valley (Fig. 3.9f).

In general, the storm followed a westward track as it moved toward the mountains. Until the storm reached the mountains, radar analysis showed that it travelled approximately 2.5 km every 6 minutes, a speed of roughly 7 m s$^{-1}$. This speed is higher than the 3-5 m s$^{-1}$ upwind trade wind speed analyzed by the GFS. It is also higher than the 3-5 m s$^{-1}$ winds observed by the DOW during the storms passage, or the 2.6 and 3.6 m s$^{-1}$ winds observed at two nearby surface stations.

Figure 3.10 shows plan view horizontal cross-sections of horizontal reflectivity from the DOW during the same six time periods, isolated at 1.5 km elevation. The black line in each panel
Figure 3.10: Plan views of gridded $Z_H$ from the DOW, cut through 1.5 km, showing the evolution of the trade wind storm on 24 October 2013 at each of six different scanning periods.

represents the windward coast of O‘ahu. The magenta lines are topographic contours, in 250 m intervals; the 500 m contour is bolded. The position of the radar is exactly in the middle of each panel.

The plan views give a better idea of the total horizontal extent of the storm, as well as its physical location. From 1836-1838 UTC, the storm is first observed to the northwest of Kāne‘ohe MCBH (refer to Fig. 3.5), with a horizontal extent no greater than 1 km in any direction (Fig. 3.10a). The horizontal cross-section in Figure 3.10a shows only the very lowest altitude boundary of the storm; the actual vertical extent is through 3 km. By 1850-1852 UTC, the storm has grown substantially in size, reaching almost 4.5 km in north-south extent (Fig. 3.10b). Although much of the convection column is hidden from view by the scanning angle limitation, it is suspected that the area of highest reflectivity would be over land just east of the radar by this point. By 1855-1856 UTC, much of the storm has moved west of the radar (Fig. 3.10c). Limitations in the range of RHI scanning angles precluded an analysis of the entire storm, although the main convective column was observed. By 1857-1900 UTC, the size of the storm appears to not have changed much since 1852 UTC – the north-south extent is still roughly 4.5 km (Fig. 3.10d). A large field of radar returns are visible along the Ko‘olau range in both 1903 UTC and 1905-1907 UTC (Figs. 3.10e & 3.10f), and by 1905-1907 UTC not much remains of the storm that is visible over the ridgeline.
Differential reflectivity (\(Z_{DR}\)) from the DOW was not useful as an analysis variable for this case because the analyzed storm passed directly over the radar site. To overcome this issue and others already mentioned, returns from a nearby national weather radar are also included in this analysis. The Next Generation Radar (NEXRAD) is a network of 160 dual-polarization S-band weather radars distributed across the United States and its territories. The closest of these is PHMO on Moloka‘i, 77.4 km to the southeast of Kahalu‘u Regional Park. The NEXRAD radars can only operate in PPI mode. During the time period of interest, PHMO was operating under volume coverage pattern (VCP) 11 — 14 different elevation angles are scanned every 5 minutes. Only the lowest three of these angles (0.5°, 1.5° and 2.4°) were able to capture the storm as it moved to the west; each had beam heights of 1.44 km, 2.79 km and 4.01 km respectively at the distance of Kahalu‘u Regional Park, including the altitude of the radar dish. The relevant scanning times and durations of each set of PPI scans are listed in Table 3.2.

Figures 3.11 and 3.12 respectively mirror Figures 3.9 and 3.10, only with PHMO substituted for the DOW. Figure 3.11 is the vertical cross-section of the gridded results cut through 21.457 °N, and Figure 3.12 is the plan view of the gridded results cut through 1.5 km. At first glance, the storm in either figure appears larger than its counterpart in Figures 3.9 and 3.10. The apparent growth of the storm is more likely a consequence of the way that the interpolation process handles beam filling. At the much larger distance of PHMO from the storm, lateral spreading of the beam is greater. Widely distributed radar detections are treated by the interpolation process as though they originate from the mean position of the beam, so the storm may appear larger than it actually is. Resolution is often an obstacle when observing small-scale phenomenon, which serves to highlight the usefulness of the DOW radars for mesoscale studies. Regardless, in this particular case, PHMO does provide some additional insight to the analysis, as well as verification.

The scan times of PHMO do not always match perfectly with those of the DOW, in fact some tend to lie between sets of DOW scans. The first two sets of PHMO scans were recorded from

<table>
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<td>18:54:50</td>
<td>1:39</td>
</tr>
<tr>
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<td>19:00:52</td>
<td>1:40</td>
</tr>
<tr>
<td>231</td>
<td>PPI</td>
<td>0.5—2.4</td>
<td>19:05:11</td>
<td>19:06:50</td>
<td>1:39</td>
</tr>
<tr>
<td>232</td>
<td>PPI</td>
<td>0.5—2.4</td>
<td>19:11:10</td>
<td>19:12:49</td>
<td>1:39</td>
</tr>
</tbody>
</table>
Figure 3.11: Altitude-longitude diagrams of gridded $Z_H$ from PHMO, cut through 21.457 °N, showing the evolution of the trade wind storm on 24 October 2013 at each of six different scanning periods.

1841-1843 UTC and 1847-1849 UTC and lie between the first two sets of DOW scans, conveniently filling the gaps left by missing data. The third set of PHMO scans occur at 1853-1856 UTC, when the storm was immediately over the DOW. The fourth set of PHMO scans at 1859-1901 UTC match reasonably well with the fourth set of DOW scans from 1857-1900 UTC. The fifth set of PHMO scans from 1905-1907 UTC match very well with the final set of DOW scans, and the sixth set of PHMO scans is much later. Exactly the same post-processing that was applied to the DOW data was also applied to the PHMO, including the quality control and interpolation to a 3-dimensional Mercator projection grid with 100 m horizontal and vertical resolution.

Maximum reflectivity is greater than 15 dBZ in the earliest set of scans (Fig. 3.11a), which surpasses the 6 dBZ maximum reflectivity detected by the DOW 3-5 minutes prior (Fig. 3.9a). By 1859-1901 UTC the storm surpasses 45 dBZ (Figs. 3.11d & 3.12d). By 1905-1907 UTC, the core of highest reflectivity is directly above the Ko‘olau ridgeline (Figs. 3.11e & 3.12e), unlike the returns from the DOW during the same time period (Fig. 3.9f). By 1911-1913 UTC, the storm is visible on the other side of the mountains (Figs. 3.11f & 3.12f).

Figure 3.13 shows the vertical cross-section of gridded differential reflectivity cut through latitude 21.457 °N from PHMO. By 1841-1843 UTC, differential reflectivity has already reached 1.5 dB in a small area at the base and center of the cloud (Fig. 3.13a). By 1847-1849 UTC, $Z_{DR}$ values of about the same magnitude are concentrated at the head of the storm, much closer to the coast (Fig. 3.13b). Figure 3.13c shows much of the same for 1853-1855 UTC, though the 1.5 dB values extend through a 4 km depth. Up to this point, the locations of the highest values of $Z_{DR}$ have been
Figure 3.12: Plan views of gridded $Z_H$ from PHMO, cut through 1.5 km, showing the evolution of the trade wind storm on 24 October 2013 at each of six different scanning periods.

well-correlated with those of $Z_H$ from Figure 3.11. By 1859-1901 UTC however, the highest values of $Z_{DR}$ (Fig. 3.13d) are just to the west of the $Z_H$ core (Fig. 3.11d). Furthermore, these values exceed 4 dB in the half kilometer just before the Ko‘olau ridgeline. It is reasonable to assume that the convection here is orographically enhanced. By 1905-1907 UTC, $Z_{DR}$ values have sunk to 1 dB (Fig. 3.13e).

The period from 1852 to 1856 UTC marked the westward passage of the storm directly over the radar site, presenting a unique opportunity for high-resolution study. A series of vertical scans from 18:52:47 to 18:53:46 UTC were conducted during the passage of the storm, when precipitation at the radar site was heavy. In vertically scanning mode, the DOW operates in a fashion similar to a PPI scan, with an $88.7^\circ$ elevation angle of the rotating radar dish. Besides the change in elevation angle, the overall scanning strategy remained the same.\footnote{The DOW\textsuperscript{7} is a dual-frequency radar, meaning that its transmission frequency alternates between a higher (9.50 GHz) and a lower (9.35 GHz) value. The receiver separates either return into separate outputs. Up to this point, radar returns from the high frequency receiver have been used in this analysis. Due to an unknown error, a significant percentage of the data from the higher frequency receiver for the vertical set of scans was corrupted, preventing its use in this analysis. The lower frequency receiver output was used instead, with a new calibration constant applied to horizontal reflectivity (refer to Section 2.2 for information on the calibration constant for the high frequency returns). To calculate the constant, low frequency returns were subtracted from calibrated, high frequency returns of the same space and time that were also not corrupted. A new calibration constant of -9.43 dBZ was calculated from the}
Figure 3.13: Altitude-longitude diagrams of gridded $Z_{DR}$ from PHMO, cut through 21.457°N, showing the evolution of the trade wind storm on 24 October 2013 at each of six different scanning periods.

Figure 3.14 shows a times series of horizontal reflectivity from the DOW during the vertical scan, with the first 5 seconds of data after the start of the scan eliminated while the antenna was in transition. Three areas of reflectivity greater than 20 dBZ are identified: at the cloud base, between 1-2 km, and above 2.2 km. Values as high as 35 dBZ are detected just above 1.5 km. Within one minute of scans, reflectivities greater than 20 dBZ and as high as 39 dBZ are detected from the cloud base through 3 km. Figure 3.15 shows a time series of fall speed-corrected Doppler velocity during the vertical scan. The terminal fall speed of a particle is proportional to its size, and can thus be approximated from horizontal reflectivity to within a 1 m s\(^{-1}\) error. The unique fall speed corrections were calculated from calibrated linear-form horizontal reflectivity using the method of Atlas, et al. (1973):

$$V_T = 2.6 * (z_H)^{0.107}$$  \hspace{1cm} (3.1)

They were then added to Doppler velocities of the same space and time to give a time series estimate of vertical velocity as the storm passed. Positive fall speed-corrected Doppler velocities in the figure thus represent updrafts, and negative values represent downdrafts. From the figure, a bimodal updraft structure is evident through the depth of the cloud. Positive values are indicated at cloud top, and also in the middle of the cloud above 1 km. At cloud top, the mean vertical velocity in the positive region is 0.5 m s\(^{-1}\), and 1.2 m s\(^{-1}\) for the middle region. The magnitude of vertical median value and applied to the low frequency horizontal reflectivity. Following the calibration, the quality control was applied.
velocity in either structure proportionally increases with time, while the downward velocity below 1 km appears to slightly decrease. By the end of the scanning period, the mean vertical velocity in the positive region at the top of the cloud had increased to 3.1 m s$^{-1}$, and 3.9 m s$^{-1}$ for the middle of the cloud. The positive velocity structure in the middle of the cloud also appears to grow in elevation with the time, increasing by roughly 200-300 m.

Following the passage of the strongest convection just to the west of the radar, a number of RHI scans covered the storm in 10° intervals from 1855-1856 UTC. Gridded horizontal reflectivity from the scans is shown in Figure 3.9c, cut through latitude 21.457 °N. The figure reveals a pattern of high reflectivity through the depth of the storm, with a maximum of 38 dBZ observed in the upper-middle part of the cell. The proximity of the storm to the radar during these scans offers a unique opportunity to observe the physical structure of the updraft. Figure 3.16 shows the gridded storm-relative Doppler velocity during this period, cut again through latitude 21.457 °N. To calculate the storm-relative Doppler velocity, the relative contributions of the vertical profile of horizontal

![Figure 3.14: Time series of horizontal reflectivity from vertically pointing scans, taken during passage of the trade wind storm over the DOW.](image)

![Figure 3.15: Time series of fall speed-corrected Doppler velocity from vertically pointing scans, taken during passage of the trade wind storm over the DOW.](image)
Figure 3.16: Storm-relative and fall speed-corrected gridded Doppler velocity from the DOW at 1855-1858 UTC, cut through 21.457 N, showing a large updraft just above 2 km for the trade wind storm. The dotted lines represent the percent confidence that the contribution is from the vertical motions within the storm.

Wind speed to Doppler velocity were subtracted from the base velocity and thus removed. The horizontal wind profile was calculated using a velocity-azimuth display (VAD) technique. A fall speed correction was also applied to the storm-relative Doppler velocity. Radial Doppler velocity measures only the component of wind velocity parallel to the direction of the radar beam. Therefore, the relative contribution to Doppler velocity from vertical velocity is proportional to the angle of the beam. The contribution to Doppler velocity from an elevation angle near 90° for instance would be 100%. Confidence lines of 80% and 90% have been added to Figure 3.16, above which the contribution is thought to be mostly from vertical velocity. The figure reveals a core of positive velocity which peaks above 2 km, between the 80% and 90% confidence intervals for vertical velocity. The results are comparable to those of Figure 3.15, in which a core of high velocity was detected moving upwards from an equivalent altitude. Unlike Figure 3.15 however, the structure is no longer bimodal; the updraft detected aloft in Figure 3.15 appears to have disappeared, or perhaps has merged with the other updraft structure in the middle of the cloud. The magnitude of the positive storm-relative Doppler velocity structure in the figure likely exceeds 7 m s⁻¹.

Finally, the post-storm environment near the radar site was examined for changes in atmospheric conditions. Wind vector and temperature recorded by the DOW7 mast during and after the storm...
are shown in Figure 3.17, where wind vector is again divided into U- and V-coordinates. The figure reveals a small but noticeable 0.2 °C dip in temperature from 1857 to 1905 UTC, corresponding to a change in wind vector. At first, U- and V-coordinate winds become more chaotic after 1857 UTC but gradually settle, finally falling to 1 m s⁻¹ by 1900 UTC. At the same time, wind direction shifts to southerly, before changing back to east-southeasterly shortly thereafter. It is likely that this time, from 1857 to 1905 UTC marks the passage of the primary downdraft near the radar site. Recalling the vertical cross-section shown in Figure 3.9b, the full horizontal extent of the storm was revealed as being quite large – nearly 3.5 km in west-east width. Had the full extent of the storm been revealed in Figures 3.9c-e, elements of the storm may well have been observed above the radar site. Furthermore, because the actual center of the storm passed just to the south of the radar, a southerly wind would be detected at the passage of the downdraft. A slight evaporative cooling signature is consistent with this interpretation.

### 3.1.5 Summary

The IOP on 24 October 2013 was characterized by light easterly trade winds and limited boundary layer moisture. The shift from a nocturnal flow regime to a daytime flow regime at the radar site
occurred approximately 2 hrs after sunrise, and was preceded by a period of convective activity. A trade wind shower that passed immediately over the radar site evolved from an area of pre-existing trade wind cumulus, and reached the Ko‘olau ridgeline (a 9 km distance) within 26 minutes after first detection. The storm approached the mountains from due east, at a speed of approximately 7 m $s^{-1}$, greater than any speed observed at the surface, or predicted by the GFS. The total height of the storm never exceeded 3 km, the approximate elevation of the trade wind inversion. As the shower passed overhead, vertically-pointing scans revealed a bimodal positive vertical velocity structure in the core of the storm which gradually became unimodal within one minute of scans. RHIs of storm-relative Doppler velocity taken 3 minutes later also reveal a single structure, with velocities likely exceeding 7 m $s^{-1}$. The trade wind shower grew greatly in both size and horizontal reflectivity as it approached the mountains from the east. Differential reflectivity however remained fairly constant until the moment right before the storm passed over the ridgeline, where it then greatly increased. As the storm passed over the mountains, both horizontal and differential reflectivity decreased somewhat, and then increased again in the leeward valley.
3.2 Sea-breeze Storms on 27 October 2013

3.2.1 Synoptic Conditions

The second IOP, conducted 27 October 2013, saw some of the most significant convective activity of the entire HERO campaign. The overall forecast for the day called for light southerly winds, abundant moisture, and a conditionally unstable atmosphere, in stark contrast to the previous IOP only three days before. In that time, a surface ridge that had been situated to the northwest of Hawai‘i had migrated to the southeast, as shown in the 1800 UTC GFS 1000 hPa analysis for 27 October 2013 (Fig. 3.18). A decaying trough further to the northwest associated with a low pressure system near the Aleutian Islands likely contributed to the migration of the ridge, which resulted in light and southerly winds over Hawai‘i. The magnitude of the surface wind was no greater than 5 m s\(^{-1}\), and abundant equatorial boundary layer moisture was advected northward into the region as a result of the southerly flow. At 600 hPa, a cold air mass associated with a nearby upper-level low was present over Hawai‘i (Fig. 3.19). Winds at this level were no greater than 5 m s\(^{-1}\), although their direction was primarily from the east. At 250 hPa, a large trough embedded within the subtropical jet stream was situated over Hawai‘i, and cold air was being advected into the region by northerly 15-18 m s\(^{-1}\) winds. With the cold air mass aloft and the forecast for a warm, moist surface layer, greater instability than normal was present throughout the Hawaiian

![Figure 3.18: 1800 UTC GFS 1000 hPa temperature (°C), wind vector, and contours of geopotential height (m) for the North Pacific region on 27 October 2013.](image)
Figure 3.19: 1800 UTC GFS 600 hPa temperature (°C), wind vector, and contours of geopotential height (m) for the North Pacific region on 27 October 2013.

Figure 3.20: 1800 UTC GFS 250 hPa temperature (°C), wind vector, and contours of geopotential height (m) for the North Pacific region on 27 October 2013.
Islands. The 1200 UTC sounding from Lihu‘e on 27 October 2013 showed a large increase in CAPE compared to the previous sounding on 24 October 2013, from 787 J/kg to 1918 J/kg (Fig. 3.21). A shearing zone at 6-7 km separated weak southeasterly winds from strong northerly winds aloft. The LCL was found to be at 494 m.

3.2.2 Deployment Strategy

Although greater instability than normal was forecast throughout the region due to abundant surface moisture and cold weather aloft, the large-scale winds at the surface were still relatively weak. In the absence of a strong synoptic forcing pattern, sea-breeze convergence and upslope flow play a more dominant role in initiating convection over tropical islands such as O‘ahu (Schroeder, 1993). When combined with significant instability, the diurnal circulation can help to trigger deep convection. In such a situation, the strongest convection over O‘ahu is expected to occur in the central valley, between the islands two north-south mountain ranges.

The decision was made to deploy the DOW in the morning to a pre-chosen site near Wahiawa, which is central to O‘ahu. The specific site chosen was a dirt parking lot near Kaukonahua Road just north of Wahiawa and east of Schofield Barracks. The central valley of O‘ahu is well-known for its expansive pineapple fields and limited habitation. Thus, ground clutter at the site was minimal,
and sweeping views of the surrounding mountains and plains made the site a prime location to observe convection throughout the central valley, including against the slopes of the Wai‘anae Range to the west, and the Ko‘olau Range to the east (Fig. 3.22).

3.2.3 Local Weather Conditions

The DOW began receiving its first radar telemetry by 2023 UTC (10:23 a.m. HST). At that time, light southwesterly winds (about 0.6 m s$^{-1}$) were recorded by the radar trucks mast anemometer. Towering cumulus were visible upon arrival and light, scattered showers were already present in the central valley. Conditions were mostly overcast, and deep clouds were observed along the Wai‘anae Range, particularly in the valley behind Schofield Barracks, near Mt. Ka‘ala. During the late morning hours, deeper convection was observed along the Wai‘anae Range with thunder being heard at the radar site by 2208 UTC. In Hawai‘i, the trade wind inversion typically prevents the development of deep convection, so thunder and lightning are a rare occurrence. In addition to the deeper, orographically-enhanced convection that was observed against the Wai‘anae Range early in the deployment, clusters of shallower, single and multi-cell storms were observed throughout the region. Many of these storms formed near Malama Bay to the south, where they were advected inland by the southwesterly flow.

A radiosonde was successfully launched at 2112 UTC, though the humidity sensor failed during the flight. The sounding is shown in figure 3.23 with only a red line for temperature, and the interpolated vertical profile of the wind to the right. Extremely light and variable winds were detected through 5 km, with magnitudes no greater than 3 m s$^{-1}$. Above 5 km, wind speeds gradually increased to 20 m s$^{-1}$ by 9.5 km, and blew consistently from the north-northeast. Evidence
of a small, 1 °C inversion was detected above 4.8 km, near the approximate shearing zone between the upper-level and lower-level winds. The freezing level was detected just below the inversion, at 4.3 km.

The transition from land- to sea-breeze regime during the late morning was detected from surface wind stations. Figure 3.24 shows hourly mean winds at 1700 and 2100 UTC for a number of stations around O‘ahu. At 1700 UTC, stations along the north shore of O‘ahu, the south coast near Pearl Harbor, in the Wai‘anae valley along the west coast, and near Kailua/Kāne‘ohe east of the Ko‘olau Range all show wind directions away from the island interior (Fig. 3.24a). By 2100 UTC, quite the opposite pattern is observed for nearly all of the stations (Fig. 3.24b). Along the western and southern coasts of the island in particular, wind directions are consistent with southwesterly winds analyzed by the GFS.
Figure 3.24: Hourly 1700 and 2100 UTC surface winds (kts) for O‘ahu on 27 October 2013. Wind speeds and direction are mean values computed from 30 minutes before to 30 minutes after each time. Data was collected from NWS/FAA network, NOS-NWLOK, NWS Honolulu WFO, Interagency Remote Automatic Weather Stations (RAWS), and Automatic Position Reporting System WX NET/Citizen Weather Observer Program (APRSWXNET/CWOP) stations.

3.2.4 Radar Observations

A wide range of convective phenomena was observed by the DOW throughout the IOP, including a warm rain, sea-breeze convective storm that is the subject of this study. The storm was already active when the radar began scanning the valley, however its most convective period occurred shortly thereafter. The storm was first detected 12 km south-southeast of the radar (Fig. 3.22). Although wind speeds were not directly observed at the storm location, an analysis of nearby surface stations showed that the storm was located within an area of 3-5 m s\(^{-1}\) southwesterly winds, which were influenced by sea-breeze and synoptic forcing (Fig. 3.24b). At this range, RHI scans were too far apart in degree separation to get much horizontal resolution on the storm, so PPI scans are primarily used in this analysis. Again, the overall scan strategy for the DOW was to alternate between RHI and PPI scans, with occasional adjustments made to the scanning angles. During the period of interest, the DOW was operating with a 400 ns pulse width and 2500 Hz pulse repetition frequency, with 60 m range gates. Twenty degrees of elevation were scanned from 0.5° at the lowest, to 38.5° at the highest, every two degrees. The duration of one full set of PPI scans was approximately 4 minutes, before the following set of RHI scans were initiated. The azimuthal resolution of the beam at this distance was approximately 200 m.

The relevant times and volumes that captured the storm are detailed in Table 3.3. Eight relevant scanning periods are chosen to reflect the evolution of a single convective cycle in the life of the storm. As with the previous case, the intensive observation period was followed by the quality control and interpolation post-processing of the raw radar sweep files detailed in Chapter 2. The gridded results highlighting the storm of interest are shown in Figure 3.25 for each of the eight
Table 3.3: Description of the relevant volume numbers, scan types, angles, times, and volume durations of the DOW that pertain to the sea-breeze storm on 27 October 2013

<table>
<thead>
<tr>
<th>Vol. Number</th>
<th>Scan Type</th>
<th>Scan Angles</th>
<th>Start Time (UTC)</th>
<th>Stop Time (UTC)</th>
<th>Duration (min)</th>
</tr>
</thead>
<tbody>
<tr>
<td>165</td>
<td>PPI</td>
<td>0.5—20.5</td>
<td>20:44:22</td>
<td>20:46:21</td>
<td>1:59</td>
</tr>
<tr>
<td>166</td>
<td>PPI</td>
<td>0.5—22.5</td>
<td>20:48:38</td>
<td>20:50:48</td>
<td>2:10</td>
</tr>
<tr>
<td>168</td>
<td>PPI</td>
<td>0.5—20.5</td>
<td>20:54:55</td>
<td>20:56:51</td>
<td>1:56</td>
</tr>
<tr>
<td>170</td>
<td>PPI</td>
<td>0.5—24.5</td>
<td>21:01:08</td>
<td>21:03:29</td>
<td>2:21</td>
</tr>
<tr>
<td>172</td>
<td>PPI</td>
<td>0.5—26.5</td>
<td>21:07:22</td>
<td>21:09:56</td>
<td>2:34</td>
</tr>
<tr>
<td>174</td>
<td>PPI</td>
<td>0.5—30.5</td>
<td>21:13:37</td>
<td>21:16:35</td>
<td>2:58</td>
</tr>
<tr>
<td>176</td>
<td>PPI</td>
<td>0.5—26.5</td>
<td>21:19:30</td>
<td>21:22:04</td>
<td>2:34</td>
</tr>
<tr>
<td>178</td>
<td>PPI</td>
<td>0.5—26.5</td>
<td>21:25:22</td>
<td>21:27:57</td>
<td>2:35</td>
</tr>
</tbody>
</table>

scanning periods, as plan views of horizontal reflectivity ($Z_H$) cut through 2 km.

A qualitative attempt has been made to delineate the rough boundaries between individual cells embedded within the storm; cells are labeled numerically in order of their maturity. The divisions between cells in the 3-dimensional volume are idealized as imaginary vertical planes cut through the storm roughly from west to east, and from bottom to top. The location and orientation of each the planes were adjusted to cut through the low reflectivity troughs present between the high reflectivity convective cores unique to individual cells. This was accomplished through a careful trial-and-error analysis, by holding the longitude and elevations of each of the four corners of each plane to be constant, and then adjusting the latitude of three corners (two at the bottom, one at the top) as needed. When convenient, multiple stacked planes were used at varying elevations, particularly when it was evident that one convective cell had overridden another. In Figure 3.25, the delineations between cells are denoted as black lines, where the horizontal cross-sections cut through the vertical planes at that level.

Figure 3.25 reveals a small multi-cell storm, roughly 3-7 km in any direction, with cores of highest reflectivity roughly 1-2 km across. The storm appears to grow significantly during the observation period, from about 3 km across at 2044-2046 UTC (Fig. 3.25a), to 7 km across the major axis nearly 40 minutes later at 2125-2128 UTC (Fig. 3.25h). The storm grows from an almost unicellular state with only a single high-reflectivity core to a larger storm with multiple high-reflectivity cores in later scanning periods. New cells tended to form upwind (to the south) of older cells, and advanced to the north as they aged. The center of cell 2 travelled approximately 5.5 km to the north-northeast during the analysis period, for a velocity of about 2.3 m s$^{-1}$. No radar bright band was detected, and maximum reflectivity values were all found below the freezing level, indicating that the storm was predominantly warm rain.

Figure 3.26 shows the longitudinally-averaged vertical cross-sections of the gridded horizontal
Figure 3.25: Plan views of gridded $Z_H$ from the DOW, cut through 2 km, showing the evolution of the sea-breeze storm on 27 October 2013 at each of eight different scanning periods. Unique convection cells are numbered chronologically and separated by black lines representing cross-sections at 2 km of imaginary vertical planes dividing each cell. 1 km² boxes represent the cross-sections of vertical columns used as a spatial partition for the CFAD analysis.
Figure 3.26: Longitudinally-averaged altitude-latitude diagrams of gridded $Z_H$ from the DOW, showing the evolution of the sea-breeze storm on 27 October 2013 at each of eight different scanning periods. Unique convection cells are numbered chronologically and separated by black lines representing the mean vertical cross-sections of imaginary vertical planes dividing each cell.
reflectivity for each of the eight sets of scans. Rather than a simple cross-section through a specific longitude, in this case all grid elements with equivalent latitude and elevation were averaged to create a single projection. Because the motion of the storm was mostly to the north, the evolution of each of the individual cells within the storm were easier to track using this method, in lieu of a more descriptive 3-dimensional projection. To calculate the longitudinal average, horizontal reflectivity values were first converted to linear form, and then back to log form when the averaging was complete. This largely helped to preserve the intensity of the high-reflectivity convective cores imbedded within the storm, regardless of any spatial contrasts in the storms width. In Figure 3.26, the average latitudinal positions of each of the vertical planes separating the convection are denoted by black lines.

From Figure 3.26, the height of the storm rarely exceeds 6 km at cloud top, just above the freezing level of about 4.5 km estimated from the sounding in Figure 3.21. No bright band was detected near the freezing level, indicating that the storm has few stratiform characteristics and is predominantly of a warm rain convective type. As stated before, the eight volumes of interest were chosen to reflect the evolution from birth to maturity of a single convective cell within the storm. In each of the figures, this cell is labeled as cell 2. Over the course of its lifetime, cell 2 became one of the most intense and expansive cells to inhabit the storm, becoming the dominant convective cell by 2101-2103 UTC (Fig. 3.26d). The horizontal structure of the cell exhibits a distinctly elongated west-east region by this time, formed into an arch-like shape (Fig. 3.25d). A similar feature is observed at 3 and 4 km (not shown). Through the next four sets of scans, the structure grows in size but shrinks in intensity, indicating a gradual weakening of the cell thereafter (Figs. 3.25e-h). The vertical structure of cell 2 roughly follows the idealized model of Houze (1981) (Fig. 1.2), who described the convective precipitation process as exhibiting well-defined cores of maximum reflectivity following the convection through its vertical evolution.

In the earliest scanning periods, cell 2 is thought to be a small nodule of low reflectivity extending south of the dominant convective cell at those times. The nodule is visible near latitude 21.395 °N by 2044-2046 UTC (Fig. 3.26a) and also during 2049-2051 UTC (Fig. 3.26b), and can also be seen at the same times and latitude in the plan views, Figures 3.25a-b. The nodule appears to show distinct upper-level and lower-level features in Figure 3.26a, although the upper-level feature is likely a remnant of the first convective cell, and is counted as such by this analysis. In any case, by 2049-2051 UTC, only a feature mid-level of the storm remains, which is counted as cell 2 (Fig. 3.26b).

By 2055-2057 UTC, a distinctive area of high reflectivity apart from the older convection is discernable for cell 2 near latitude 21.405 °N (Fig. 3.26c). The core itself is most intense in the mid-levels from 1-3 km, reaching a maximum of 38 dBZ at 2.2 km. The highest values of horizontal reflectivity for cell 1 are near the cloud base, indicating a period of precipitation fallout. The maximum reflectivity for cell 1 at this time is 42 dBZ, still higher than cell 2. By 2101-2103 UTC,
cell 2 has grown substantially in both size and intensity, largely dwarfing cell 1 as it continues to
decay (Fig. 3.26d). Reflectivity values as high as 52 dBZ are found in the convective core at 1.6 km.
The reflectivity core (values greater than 35 dBZ) itself extends from cloud base to 4 km altitude,
and displays a distinctive northward tilt with height, which also broadens in west-east extent with
height (Fig. 3.25d). A small shaft of high reflectivity to the north of the storm appears to be an
independent cell, probably formed from the same downdraft and gust front that spawned cell 2.
Otherwise, virtually nothing remains of cell 1 above 2 km in Figure 3.26d, the cell having largely
shed much of its liquid water content since the previous set of scans.

With the almost total dissipation of cell 1 by 2107-2110 UTC, cell 2 has come to largely dominate
the storm (Fig. 3.26e). A small area of high reflectivity is present to the north of the primary cell,
but is thought to be likely the remnant of the small, independent cell observed in the previous set
of scans (Figs. 3.25d and 3.26d). Reflectivity values as high as 50 dBZ are detected at 1.6 km, with
values greater than 40 dBZ observed from the cloud base through 2 km. Though the maximum
reflectivity is detected at the same level as the previous set of scans, it is apparent from Figure
3.26e that the center of the vertical core is now much closer to the cloud base. The core appears to
have moved about a kilometer (roughly 0.01°) to the north, and is now centered near 21.420 °N.

It is clear that by 2114-2117 UTC, the cell has begun to dissipate (Fig. 3.26f). Reflectivity
values greater than 35 dBZ are no longer common in the core, and when they exist, are concentrated
near the cloud base. In general, the highest reflectivity values seem concentrated near the cloud
base for the next two sets of scans as well (Figs. 3.26g-h). The maximum reflectivity within cell 2
at 2114-2117 UTC was 44 dBZ, detected at 0.4 km. The downward trend in maximum reflectivity
continues through the next two sets of scans, where respectively, values of 41 and 36 dBZ are
observed. The cell continues its northward trajectory, reaching 21.440 °N by 2125-2128 UTC (Fig.
3.26h). Very little remains of cell 2 by this time, however cell 3, which is actually likely two separate
cells, has intensified.

Longitudinally-averaged vertical cross-sections of gridded differential reflectivity (Z\(_{\text{DR}}\)) for each
of the eight scanning periods are shown in Figure 3.27. Again, all grid elements with equivalent
latitude and elevation were averaged to create a single projection, and again, the average latitudinal
positions of each of the vertical planes separating the convection are denoted by black lines. As
noted previously, Z\(_{\text{DR}}\) is the difference between the horizontal and vertical logarithmic reflectivity
of a sample volume — essentially a measure of the oblateness of the mean drop size. Positive values
indicate that drops are larger, and thus more oblate due to aerodynamic drag. The greater area
of the storm in the earliest set of scans appears to be exhibiting extremely low Z\(_{\text{DR}}\), even in the
inner core of cell 1, which would seem to indicate prolate drops (Fig. 3.27a). Negative values are
typically indicative of ice particles, which clearly cannot dominate the storm below the freezing
level. Thus, Z\(_{\text{DR}}\) values are likely skewed by an anomalous radar error in the first set of scans.

By 2055-2057 UTC, the highest Z\(_{\text{DR}}\) values within cell 1 are found near 3 km (Fig. 3.27b). The
Figure 3.27: Longitudinally-averaged altitude-latitude diagrams of gridded $Z_{DR}$ from the DOW, showing the evolution of the sea-breeze storm on 27 October 2013 at each of eight different scanning periods. Unique convection cells are numbered chronologically and separated by black lines representing the mean vertical cross-sections of imaginary vertical planes dividing each cell.
maximum reflectivity for cell 1 at this time is 42 dBZ, with an average ZDR of 1.8 dB in the same area (within a 0.3 km³ box encompassing the detection). The high ZDR in the core of cell 2 is in sharp contrast to cell 1. Average ZDR values as high as 2.4 dB were observed in the immediate area of the horizontal reflectivity maximum.

Differential reflectivity values vary wildly across cell 2 by 2102-2103 UTC, but are concentrated between 1.5-2.5 dB in the convective core (Fig. 3.27d). Near the area of maximum reflectivity at 1.6 km, the average ZDR was 1.5 dB. Three areas of particularly high ZDR are identified: the first, sporadically distributed across cell 2 near 1 km, the second, in the center of the storm from 2-4 km, and the third, on the outer periphery of the storm above 4 km. It is surmised that the last of these is likely due to the steady evaporation of localized cloud pockets at the cloud edge; smaller drops within the pockets evaporate first due to entrainment of dry air, leaving behind a DSD heavily weighted toward larger drops. Such a DSD would have a high ZDR signature, and would be located along the outer cloud boundary. High ZDR values in the center of the storm from 2-4 km correspond well with high values of ZH in Figure 3.26d.

By 2107-2110 UTC, ZDR is generally 1-2 dB below 3 km (Fig. 3.27f), and as high as 3 dB in a few isolated areas. High ZDR values generally do not extend above 3 km by this time, except in a few small pockets on the periphery of the storm aloft (not shown). At 1.6 km, where the maximum reflectivity was detected, an average ZDR of 1.3 dB was calculated in the immediate area. By 2114-2117 UTC, average ZDR values of 1.4 dB were detected in the immediate area of the reflectivity maximum. The downward trend in local ZDR at the area of maximum reflectivity continues through 2119-2122 UTC and 2125-2128 UTC, where respectively, values of 0.6 and 0.2 dB are observed (Figs. 3.27g-h). The extremely high values of ZDR that were present along the arch-like feature of high reflectivity are absent in this scanning period (Fig. 3.27f). The cell does however still retain its 1-2 dB core at the lower levels, although this gradually fades to nearly 0 dB in later volumes.

While horizontal and differential reflectivity are particularly useful for microphysical examinations, high-resolution gridded Doppler velocity allows a unique inspection of the fine-scale dynamics of storms. Figures 3.28-3.31 show plan views of gridded Doppler velocity, each cut through horizontal levels from 1-4 km in 1 km intervals, respectively, for each of the eight time periods. Warmer colors show positive Doppler velocity toward the radar, and cooler colors show negative Doppler velocity away from the radar. An arrow plotted on each panel points the direction to the radar. For the first two time periods of interest, Doppler velocity plots show a weak convergence signature below cell 1 at 1 km (Figs. 3.28a-b). Doppler velocities for cell 2 at these times and this level are still predominantly negative, however by 2055-2057 UTC and through 2107-2110 UTC, Doppler velocities at 1 km are a mix of near-zero and negative value areas (Figs. 3.28c-e). By 2114-2117 UTC and through 2125-2128 UTC a negative area at the center of cell 2, surrounded by more positive Doppler velocities generally moves to the north (Figs. 3.28f-h). A small anticyclonic
Figure 3.28: Plan views of gridded Doppler velocity from the DOW, cut through 1 km, showing the evolution of the sea-breeze storm on 27 October 2013 at each of eight different scanning periods. Unique convection cells are numbered chronologically and separated by black lines representing cross-sections at 1 km of imaginary vertical planes dividing each cell. An arrow points the direction to the radar.
Figure 3.29: Plan views of gridded Doppler velocity from the DOW, cut through 2 km, showing the evolution of the sea-breeze storm on 27 October 2013 at each of eight different scanning periods. Unique convection cells are numbered chronologically and separated by black lines representing cross-sections at 2 km of imaginary vertical planes dividing each cell. An arrow points the direction to the radar.
Figure 3.30: Plan views of gridded Doppler velocity from the DOW, cut through 3 km, showing the evolution of the sea-breeze storm on 27 October 2013 at each of eight different scanning periods. Unique convection cells are numbered chronologically and separated by black lines representing cross-sections at 3 km of imaginary vertical planes dividing each cell. An arrow points the direction to the radar.
Figure 3.31: Plan views of gridded Doppler velocity from the DOW, cut through 4 km, showing the evolution of the sea-breeze storm on 27 October 2013 at each of eight different scanning periods. Unique convection cells are numbered chronologically and separated by black lines representing cross-sections at 4 km of imaginary vertical planes dividing each cell. An arrow points the direction to the radar.
A couplet is detected directly beneath one of the newly formed cells at 2119-2122 UTC (Fig. 3.28g), and continues through the next set of scans (Fig. 3.28h). Velocity couplets are a common feature of Doppler radar returns that indicate rotation within a storm. Couplets are either cyclonic or anticyclonic, and can be identified by parallel areas of positive and negative Doppler velocity along the radial axis of the radar beam.

At 2 km and from the beginning through 2114-2117 UTC, Doppler velocities are mostly negative across the storm, likely indicative of northward advective flow (Figs. 3.29a-f). This is especially true within cell 2, where Doppler velocities are detected as low as -6 m s\(^{-1}\). By 2125-2128 UTC however the situation is reversed — a large area of positive Doppler velocities is detected within the center of cell 2 (Fig. 3.29f). The anticyclonic couplet that was visible at 1 km is also visible at 2 km in the core of cell 3 during this time. At 3 km, a large region of deeply negative Doppler velocity is present at the center of cell 2 by 2101-2103 UTC (Fig. 3.30d). This area directly corresponds with a horizontal region of high reflectivity at 3 km (not shown), at the northern end of which is the arch-like feature described previously. Just to the north of the negative Doppler velocities, more positive values are present, indicating some amount of convergence, or perhaps a nudging of the cloud mass to the north by the cell. Additionally, evidence of counter-rotating vortices is detected on either side of the feature at latitude 21.415 °N. Specifically, an anticyclonic couplet is observed to the right at longitude 158.04 °W at 3 km (Fig. 3.30d), and a cyclonic couplet is observed to the left at longitude 158.02 °W at 4 km (Fig. 3.31d). Neither of the couplets are more than 1.5 km in diameter, and each is also observable at latitude 21.420 °N in the next set of scans, 2107-2110 UTC (Figs. 3.30e & 3.31e). At 4 km, Doppler velocities are still largely negative across the center of cell 2 by 2101-2103 UTC, although the negative region is much more to the north (Fig. 3.31d).

Beyond the basic observational analysis of the storm, which utilized plan views and cross-sections, both contoured frequency-altitude diagrams (CFADs) and probability density functions (PDFs) were utilized. First developed by Yuter and Houze (1995), CFADs are the contoured statistical distributions of various storm properties with each level of height. They have the advantage over observational plots that they leave the full distribution of the data intact, although the horizontal geographic information is lost. Ideally, they are used together with an observational analysis to effectively understand the full vertical evolution of the convective lifecycle of a storm. In this analysis, CFADs are plotted for two different radar variables with two different spatial partitions of the storm for each of the eight scanning periods. The first of the two specific spatial partitions used as data sources for the analysis is cell 2 by itself with the other cells removed from the dataset. To isolate cell 2, the qualitatively-deduced imaginary planes described before in the analysis were used as dividers. As there is no way to guarantee the accuracy of this method of separating individual cells within the volume, it is simply regarded as an intuitive approach. CFADs of Z\(_{DR}\) and Z\(_{H}\) are plotted in Figures 3.32 and 3.33 respectively, for each of the eight scanning periods using this method. Probability distributions were constrained between -2 and 5 dB with 0.5 dB interval bins.
for differential reflectivity, and between -20 and 55 dBZ with 5 dBZ interval bins for horizontal reflectivity. Finally, each plot was normalized by its maximum frequency.

The most striking feature of the cell-isolated CFADs in Figure 3.32 is the clear vertical evolution of the maximum $Z_{DR}$ frequency center throughout the convective lifecycle of the storm. Furthermore, each of the probability distributions are essentially unimodal for each of the eight scanning periods. The location of the peak rises from 1.7 km at 2044-2046 UTC (Fig. 3.32a), eventually cresting at 4.6 km by 2101-2103 UTC (Fig. 3.32d), before falling again to 1.7 km in the final scanning period (Fig. 3.32h). The actual $Z_{DR}$ bin of the peak frequency never seems to venture far from 0 dB, only occasionally becoming marginally negative (with the exception of 2044-2046 UTC (Fig. 3.32a), which as mentioned before, is probably due to an anomalous radar error). This result seems to imply that a large mass of smaller drops are updraft-following. This begs the question, where are the larger drops? For each of the volumes where they exist, the probability distribution shifts towards larger $Z_{DR}$ bins at an elevation immediately beneath the peak frequency. This trend begins by 2049-2051 UTC (Fig. 3.32a), where 0-1 dB drops are evidently formed at 1-2 km, and ends at 2119-2122 UTC (Fig. 3.32g), when the last of the large drops greater than 1 dB are emptied from the cell. By 2055-2057 UTC (Fig. 3.32c), drops as large as 3 dB are seen to form near 2 km, however by 2101-2103 UTC, these large drops are mostly near the cloud base (Fig. 3.32d). By 2107-2110 UTC, many of the 3 dB drops have disappeared, leaving 2.5 dB and lower drops at the cloud base (Fig. 3.32e). A similar pattern of high $Z_{DR}$ signatures falling out quickly from the core after the updraft peak has been observed in a continental hail storm (Zeng et al., 2001) and in trade wind showers (Minor et al., 2011).

The $D^6$ dependence of horizontal reflectivity on drop size diameter implies that some similarities between Figures 3.32 and 3.33 should be immediately apparent. Indeed, to a certain degree, this is the case. The large 2-3 dB $Z_{DR}$ spike at 2 km at 2055-2057 UTC of Figure 3.32c is well-correlated with a 20-35 dBZ $Z_H$ spike at the same level in Figure 3.33c. A similar pattern is found from 2101-2103 UTC to 2119-2122 UTC, where large values of $Z_H$ relate rather well to the observed spikes in $Z_{DR}$ (Fig. 3.33d-g). The observed frequency pattern of $Z_H$ however is not typically unimodal as it was with $Z_{DR}$. Instead, it is often quite uneven, and typically decreasing with height in a streak-like pattern. On occasion, multiple high-frequency patterns of $Z_H$ exist. While this could be due to some microphysical process, it could also be the product of a cloud mass that is geospatially diverse.

The second of the two specific spatial partitions used as data sources for the analysis were 1 km$^2$ vertical columns extending through the strongest convection in cell 2, with the rest of the cell, and the storm cropped from the analysis. To isolate vertical columns through the strongest convection, the location of highest average horizontal 1 km$^2$ reflectivity was first identified in each gridded volume, and the column extending through the depth of the grid at that location was cropped. Ideally, such a vertical column would be quasi-Lagrangian — that is, it would track the same 1
Figure 3.32: Contours of normalized CFADs of gridded $Z_{DR}$ from the DOW, showing the vertical evolution of the sea-breeze storm on 27 October 2013 at each of eight different scanning periods, with cell 2 isolated from the rest of the storm by using imaginary vertical planes dividing the convective cells as guides.
Figure 3.33: Contours of normalized CFADs of gridded $Z_H$ from the DOW, showing the vertical evolution of the sea-breeze storm on 27 October 2013 at each of eight different scanning periods, with cell 2 isolated from the rest of the storm by using imaginary vertical planes dividing the convective cells as guides.
km$^2$ parcel around the grid, following its horizontal motion as it evolved vertically. The columns identified with the 1 km$^2$ averaging technique do appear to follow a consistent path that travels with the convection, and are shown as 1 km$^2$ boxes in Figure 3.25 for each of the eight scanning periods. The only exception was the column for 2107-2110 UTC, which was adjusted 0.006° (0.6 km) to the west. This method of parcel-finding does not however take into account the possibility of a tilted vertical parcel, or its overall expansion and contraction. CFADs of $Z_{DR}$ and $Z_H$ using this method are plotted in Figures 3.34 and 3.35 respectively, for each of the eight scanning periods. The probability distributions of $Z_{DR}$ and $Z_H$ were constrained to the same size bins and range as in Figures 3.32 and 3.33, and each plot was again normalized using its maximum frequency.

When the spatial partition of the data is restricted to vertical columns through the strongest convection in cell 2, the $Z_{DR}$ and $Z_H$ frequency distribution is predictably far narrower with height. A wide distribution in this case would represent a large diversity of particle sizes in a small area. This is occasionally the case, as in 2101-2103 UTC up to 3 km in the $Z_{DR}$ distribution of Figure 3.34d, and up to 1.5 km in $Z_H$ distribution of Figure 3.35d. This is also observed at 2107-2110 UTC in Figure 3.34e, until the $Z_{DR}$ distribution suddenly narrows above 1.5 km. Later volumes show similarly narrow distributions, which also decrease in magnitude in both height and time (Figs. 3.34f-h and 3.35f-h). The much wider $Z_{DR}$ distribution at 2101-2103 UTC is therefore quite unique in its consistency – the vertical profile through 3 km consistently peaks in the vicinity of 2 dB, and never far exceeds 3 dB or goes below 0.5 dB (Fig. 3.34d). The vertical profile of $Z_H$ for the same scanning period also displays an interesting feature – there appears to be a hard cutoff just above 50 dBZ from 1-3 km (Fig. 3.35d). Barring an unconsidered source of error, there appears to be some microphysical process at work. It is speculated that this is evidence of collisional breakup within the column. A wide array of drop sizes might be observed in such a scenario, coinciding with a widening of the $Z_{DR}$ distribution. The abrupt constancy of the peak $Z_H$ and $Z_{DR}$ with height provides some credence to the idea that a threshold on drop size has been reached, and lends to the argument.

Another useful statistical method utilized in this analysis are contoured joint probability density functions (PDFs) of horizontal and differential reflectivity. The utility of bivariate probability distributions in $Z_H$-$Z_{DR}$ space is primarily that they enable a unique microphysical description of the convective lifecycle, while ignoring its geographic aspects. Changes in the PDF from one set of scans to the next are thought to describe the handful of unique microphysical processes that lead to precipitation (Kunjian and Prat, 2014). The PDFs generated for this analysis are shown in Figure 3.36 for each of the eight periods of interest, and have been spatially partitioned by isolating cell 2 from the rest of the storm, in the same manner as the CFADs of Figures 3.32 and 3.33. PDFs of the vertical columns through the convection for each of the eight scanning periods were also generated but are not shown. Horizontal reflectivity was constrained between -20 and 60 dBZ, with 2 dBZ bins, while differential reflectivity was constrained between -2 and 4 dB with 0.2 dB bins. Similar
Figure 3.34: Contours of normalized CFADs of gridded $Z_{DR}$ from the DOW, showing the vertical evolution of the sea-breeze storm on 27 October 2013 at each of eight different scanning periods, with vertical columns through the strongest convection in cell 2 isolated from the rest of the storm.
Figure 3.35: Contours of normalized CFADs of gridded $Z_H$ from the DOW, showing the vertical evolution of the sea-breeze storm on 27 October 2013 at each of eight different scanning periods, with vertical columns through the strongest convection in cell 2 isolated from the rest of the storm.
to the CFADs, each plot was normalized by its peak frequency.

Beginning from 2049-2051 UTC and continuing through 2101-2103 UTC, the joint probability distribution increases towards higher $Z_H$ and higher $Z_{DR}$ a clear indication of collision-coalescence (Figs. 3.36b-d). Even as early as 2055-2057 UTC, $Z_{DR}$ values as high as 3 dB are correlated with $Z_H$ values as high as 40 dBZ (Fig. 3.36c). By 2101-2103 UTC however, the $Z_H$ values have increased beyond 50 dBZ, while the $Z_{DR}$ distribution has widened substantially (Fig. 3.36d). As a result, the joint probability distribution takes a distinctive crowbar-shape that is also observed at 2107-2110 UTC (Fig. 3.36e). The widening of the $Z_{DR}$ distribution in either scanning period is comparable to the widening observed earlier in the vertical column-isolated $Z_{DR}$ CFADs of Figure 3.34b. In that figure, it was concluded that drop breakup was responsible for the wide distribution and hard cutoffs of $Z_H$ and $Z_{DR}$. Following 2101-2103 UTC, the joint probability distribution in Figure 3.36 decreases towards lower $Z_H$ and lower $Z_{DR}$ — this time, an indication of precipitation fallout (Figs. 3.36e-h). It is also observed that where the distribution preceded the convective peak in 2155-2157 UTC, $Z_{DR}$ is quicker to rise than $Z_H$ (Fig. 3.36c). In 2114-2117 UTC following the convective peak, $Z_{DR}$ is quicker to fall (Fig. 3.36f). The faster drop in overall $Z_{DR}$ seems to be an indication of the faster fallout of larger drops.

The extremely high reflectivity observed at this time is particularly noteworthy, because it seems to indicate the presence of very large raindrops. Such drops are common in continental supercells, where ice-cloud processes dominate, but are generally considered to be unusual for warm rain storms. Beard et al. (1986) however did discover evidence of very large raindrops in Hawaiian storms during the Joint Hawaiian Warm Rain Project. The work of Brandes et al. (2004) established an empirical relation between $Z_{DR}$ and median drop diameter ($D_o$) as shown in equation 1.1. Values of median drop diameter calculated from the equation are 1.6 mm, 2.1 mm, and 3.2 mm for $Z_{DR}$ values of 1 dB, 2 dB, and 3 dB respectively. From Figure 3.36d the highest values of $Z_H$ were correlated with $Z_{DR}$ from 1-2 dB. While it is therefore possible that the large 4-5 mm drops observed in Beard et al. (1986) exist in this particular storm, they are greater than the median drop diameter in this case and would exist on the far end of the drop size spectrum.

A considerable quantity of very high $Z_{DR}$ values are clustered near -10 dBZ in each panel of Figure 3.36. By 2101-2103 UTC however, the frequency of detection of these values increases substantially (Fig. 3.36d), but then gradually declines thereafter. While it is conceivable that such a signal could be attributed to random noise, it is also possible that much of the signal is due to evaporation at the cloud edge. Recalling from the observational analysis conducted earlier, a few large pockets of high $Z_{DR}$ were detected aloft, near the upper cloud boundary (Fig. 3.27d). Dry air entrainment of the updraft along the boundary is thus suspected to be the cause.
Figure 3.36: Contours of normalized PDFs of gridded $Z_H$ vs. gridded $Z_{DR}$ showing the evolution of the sea-breeze storm on 27 October 2013 from the DOW at each of eight different scanning periods, with cell 2 isolated from the rest of the storm by using imaginary vertical planes dividing the convective cells as guides.
3.2.5 Summary

The IOP on 27 October 2013 was characterized by light southwesterly winds, abundant boundary layer moisture, and a conditionally unstable atmosphere. Convective development in the central valley of O‘ahu was primarily triggered by sea-breeze convergence and anabatic flow. A single cell from an entirely warm rain, convective storm which appeared shortly after deployment was analyzed exclusively. Overall, high-reflectivity convective cores observed within the storm followed the standard model presented by Houze (1981) rather well. The strongest reflectivity within the cell was detected at 2101-2103 UTC — about 15-19 minutes after convective initiation began. This time represented the peak of the updraft, at the moment of overturning. The following scanning period showed that the convective core had descended to the cloud base, indicating that heavy precipitation was occurring at the surface. The total descent time of the downdraft was no greater than 6 minutes over a distance of about 2 km. Cell-isolated CFADs showed that a single mass of low ZDR radar returns stayed just above the highest ZDR radar returns in each scanning volume, and also above the high reflectivity core. The highest value ZDR returns were also removed first from the storm, following the convective peak. Joint PDFs show a distinctive "crowbar"-like feature at the convective peak, likely due to drop breakup. ZH and ZDR were infrequent above 52 dBZ or 3 dB in any of the scanning periods, indicating a drop size cutoff prior to breakup.
A high-resolution, dual-polarization radar study of two common Hawaiian warm rain convective types was conducted using data collected during HERO on the island of O‘ahu in the fall of 2013. The cases analyzed represent two of the most frequently observed precipitation systems on O‘ahu, the most populous island in the state. Doppler radar is one of the few research tools that can explore the complex structure of convection at high spatial and temporal resolution. High-resolution Doppler radar observations of Hawaiian storms are not commonly available however due to the cost-prohibitive distance from the mainland United States. A detailed radar analysis of common convective storms throughout the lifecycle of convection could lead to improved forecasts.

On 24 October 2013, the DOW was deployed to the windward side of O‘ahu to observe trade wind convective development upwind of the Ko‘olau mountain range. Mast data collected at the radar site showed an abrupt change in environmental conditions two hours after sunrise. The wind direction shifted from a mean westerly direction at the radar site, to east-southeasterly. At the same time, relative humidity, which had been falling rapidly since 30 minutes prior, suddenly began falling at a much slower rate. It is suggested that this period marks the abatement of the nocturnal land-breeze circulation at the radar site, in favor of daytime flow regime. Evidence of a relatively weak diurnal circulation on the windward side of O‘ahu was previously observed by Hartley and Chen (2010). Radar observations recorded throughout the day revealed that most of the early morning convection occurred before the transition. While nocturnal cloud-top radiative cooling is expected to play a role in producing an early morning maximum in rainfall, it is also likely that the interaction between land-breeze and trade wind had an influence.

A trade wind shower that passed directly over the radar site is a subject of this study. A small pocket of shallow cumulus no larger than 1 km in west-east width was first observed 6 km to the east of radar site. The shower grew larger as it approached, spanning nearly 3.5 km in west-east width by the time it passed over. Cloud tops never reached beyond 3 km, the approximate height of the trade wind inversion. The horizontal motion of the storm was approximately westerly at 7 m s\(^{-1}\), faster than the 3-5 m s\(^{-1}\) wind speed observed by the DOW or nearby surface wind stations, or predicted by the GFS. The exact reason for the increased speed is not known. Horizontal reflectivity recorded from PHMO increased as the trade wind shower approached the mountains, reaching a maximum of 50 dBZ just before encountering the ridgeline. After crossing the ridgeline, the storm descended into the leeward valley where it continued to precipitate. Differential reflectivity remained fairly consistent at 1.5 dB until the storm reached the mountain range, but then peaked above 4 dB within 500 m of the ridgeline, indicating a shift in the median DSD towards much larger drop sizes.

The passage of the shower over the radar site presented a unique opportunity to observe the core structure of the updraft at high resolution. High resolution vertical scans of Doppler velocity from
the DOW were corrected for raindrop fall speed, revealing the fine-scale vertical velocity structure of the convective core. Positive velocity was observed both at the top of the storm above 2.5 km, and in the middle from 1-2 km. To the authors knowledge, this is the first time that a bimodal positive vertical velocity structure has been observed in a landfalling trade wind shower. The overall magnitude of positive vertical velocity increased by about 2.6 m s$^{-1}$ in both regions during one minute of scans. The elevation of the positive velocity structure in the middle of the storm also appears to increase by 200-300 m during the same period. It is not clear from the scans however if the increase in either elevation or magnitude during that period is due to a physical change in the updraft, or the result of the migration of the storm. In either case, storm-relative Doppler velocity from gridded RHIs in the following set of scans reveal a single mode positive velocity updraft structure at the core of the storm, with vertical velocities likely exceeding 7 m s$^{-1}$. Following the passage of the storm, evidence of a downdraft was recorded by surface observations. A small but noticeable drop in temperature was felt at the radar site, likely the result of cold air descent. The drop in temperature directly corresponded with a change in wind vector from easterly to light and southerly, consistent with the interpretation that the core of the downdraft passed narrowly to the south of the radar site following the passage of the updraft.

Although trade wind convection on the windward side of tropical islands has been studied to some extent in the past, sea-breeze storms occurring within island interiors have received comparably little attention. On 27 October 2013, the DOW was deployed to the center of the island to observe sea-breeze induced convective development in the central valley. A shallow multi-cell storm that appeared to the southwest of the radar soon after deployment was studied extensively. The storm was judged to be predominantly warm rain and convective-type. Eight scanning periods of interest captured the evolution of a single cell within the storm. The core of highest reflectivity within the cell ascended above 2 km after 15-19 minutes of development, with maximum reflectivity values detected as high as 52 dBZ. Differential reflectivity values detected in the core of the storm during the same period exceeded 3 dB. The decent of the core to the cloud base occurred after another 6 minutes, and the storm slowly disappeared thereafter.

Statistical descriptions such as contoured frequency-altitude diagrams (CFADs) and joint probability density functions (PDFs) are shown to reveal additional information beyond the direct analysis. While cross-sections through the gridded radar data necessarily eliminate or obscure a significant portion of the storm, frequency distributions leave the full volume of data intact, instead eliminating some of the geographic information. Because they preserve the vertical coordinate, CFADs are particularly good descriptors of the convective lifecycle. The cell-isolated differential reflectivity CFADs showed that a large quantity of 0 dB drops ascend to 4.5 km after 15-19 minutes and descend thereafter. The largest drops, which reach their peak elevation after 11 minutes, are more frequent beneath the mass of 0 dB drops and fall out quickly following the updraft peak. The smaller 2 and 1 dB drops are slower to fall from the storm, but they too are
eventually removed in turn. A similar pattern has been observed in a continental hail storm (Zeng et al., 2001) and trade wind showers (Minor et al., 2011).

Joint PDFs differ from CFADs in that they sacrifice the vertical coordinate in exchange for a bivariate frequency distribution. The result in $Z_H$-$Z_{DR}$ space is a statistical description of convection that includes a complex microphysical analysis. Positive increases in both $Z_H$ and $Z_{DR}$ through the first 15-19 minutes indicate that collision-coalescence is occurring within the updraft. The magnitudes of $Z_H$ and $Z_{DR}$ however are infrequent above 3 dB or 52 dBZ during the collision-coalescence phase, suggesting a hard cutoff for droplet growth. Instead, a drop breakup signature becomes prominent at the convective peak — the distinctive “crowbar”-shape after 15-19 minutes.

At the same time, a larger proportion of low $Z_H$, high $Z_{DR}$ drops signals some degree of evaporation somewhere within the storm. Extremely high $Z_{DR}$ detected aloft at this time is likely the product of evaporation. The protrusion of the updraft into the upper cloud periphery and subsequent mixing with environmental air leads to evaporation of the smallest drops. Finally, negative changes in both $Z_H$ and $Z_{DR}$ in the last few sets of scans signal precipitation fallout, and the convective cycle of updraft and downdraft begins again.

In the future, a modeling component may be necessary to connect microphysical processes to radar observations. Additionally, an extended analysis of the HERO dataset may yet provide further information concerning trade wind showers and sea-breeze storms through the analysis of additional cases. A variety of weather phenomena were observed during HERO, including an anabatically-forced deep convection case and a cold frontal case. An analysis of these cases may yet provide a more well-rounded representation of Hawaiian weather.
BIBLIOGRAPHY


