Orographic modification of precipitation processes in a tropical cyclone moving over a continental mountain range

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Airborne radar reflectivity data and numerical simulations are examined to determine how tropical cyclone precipitation processes are impacted by landfall over a continental mountain range. Analysis of the high-resolution radar data collected within Hurricane Karl (2010) during the Genesis and Rapid Intensification Processes (GRIP) shows that radar reflectivity enhancement in regions of upslope flow is constrained to low-levels. Reflectivity enhancement is not uniform and discrete regions of enhanced precipitation are embedded within a broad echo. In conjunction with an upstream dropsonde that exhibits weak instability, the radar data suggest a mix of gentle ascent and shallow convection occur. Regions of downslope flow are characterized by precipitation originating further aloft with little modification near low levels. Satellite data further indicate that deep convection develops after the high clouds dissipate, indicating that the evolving
thermodynamic environment favors orographic modification processes beyond collection of orographically-generated cloud water.

Numerical simulations examine how modification processes controlling precipitation are affected by the height of an idealized plateau. When terrain is minimal, the tropical cyclone decays slowly, the upper-level warm core remains robust, the moist neutral environment persists, and precipitation processes are largely concentrated within the eyewall and rainband. Movement over a tall topographic barrier induces rapid decay, which erodes the warm core and moist neutral environment. A mix of forced ascent and buoyant motions contribute to enhanced warm rain processes over the terrain. Overall, all microphysical quantities are greater for the tall plateau storm, but concentrations within the innermost core decay rapidly along with the storm. It is shown that the simulated tropical cyclone precipitation is heavily influenced by overestimated graupel production, which is a common problem of microphysical schemes. Surface precipitation is comparable between the two experiments, suggesting that strong decay of the storm affects the upper limit of precipitation. Similar precipitation patterns between the observations and tall plateau simulation suggest that the model obtains realistic precipitation through incorrect microphysical processes, but a lack of microphysical observations prevent full assessment of that hypothesis. Overall, this dissertation demonstrates that decay due to landfall over complex terrain affects the inner core thermodynamic and kinematic environment, which in turn affects the type and organization of precipitation processes that occur.
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DEDICATION

To my family.
Chapter 1. INTRODUCTION

Tropical cyclones have long been a subject of interest in the meteorological community due to their ability to cause widespread damage. Destructive storm surge, damaging winds, and intense rainfall lead to human fatalities, property destruction, and agricultural losses. Near complex terrain, rainfall can intensify and increase the risk for devastating flooding and landslides. The high potential for loss motivates the need for accurate forecasting of multiple facets of tropical cyclones moving over steep terrain: where a storm will go, how strong a storm will be, and where and how much rain will reach the surface.

Research studies have historically emphasized the impact of terrain on the tropical cyclone track because slight deviations can dramatically affect the location of hazards. Taiwan is a frequent setting for studying the orographic modification of tropical cyclones since approximately three typhoons make landfall each year on that small mountainous island (Central Weather Bureau). But regions on land near much larger continental mountain ranges are also affected by hurricane landfalls; for example, the landfalls of Hurricanes Karl (2010) and Patricia (2015) in Mexico. The topography affects the environmental flow and shifts the storm track of a tropical cyclone approaching to the east to the left of the track (Bender et al. 1985; Zehnder et al. 1993a). The flow over the terrain also affects the storm’s precipitation, but a smaller fraction of the literature has been devoted to the effect of the topography on the precipitation. Both observational and modeling studies demonstrate that precipitation often intensifies around elevated terrain (Bender et al. 1985; Huang et al. 2014; Misumi 1996; Smith et al. 2009; Yang et al. 2008; Yu and Cheng 2008; Yu and Cheng 2013, 2014; Yu and Tsai 2017),
An even smaller fraction of the available literature is dedicated to examining how precipitation is created, modified, or redistributed as a tropical cyclone moves over terrain. This dissertation focuses on these mechanisms. In general, the small number of studies stems from the difficulty of obtaining high-quality observations within tropical cyclones, especially over complex terrain. The manner in which precipitation processes are modified depends on the thermodynamic and kinematic environment in which a cloud is embedded (Smith 1979; Roe 2005; Houze 2012, 2014). Given the strong variability of wind speed, direction, and thermodynamic environments over short distances within a tropical cyclone, precipitation processes vary accordingly, and high density observations are necessary. Surface instrumentation (e.g., rain gauges, disdrometers) can observe local precipitation conditions, but they are point measurements, restricted to accessible locations, limited in high wind speeds, and only represent the precipitation after it has formed and consequently say nothing regarding the processes that produced the raindrops.

Instrumented aircraft are necessary to obtain in situ measurements in clouds. The high mobility of aircraft allow them to obtain measurements in targeted locations in-cloud. However, strong winds and intense turbulence, especially at lower altitudes, make maintaining safe conditions for pilots a challenge even when a storm resides over the ocean (Marks et al. 2008); this concern is more serious near complex terrain, where turbulence, mountain waves and the high terrain itself can create dangerous flight conditions. Consequently, most observations by aircraft in tropical cyclones are obtained over ocean prior to storm landfall. In addition, many aircraft have altitude limitations that preclude observations at the highest (coldest) altitudes in cloud. Even in situations where aircraft can fly, in situ microphysical observations only sample the immediate environment surrounding the plane. Dropsondes can sample the thermodynamic environment below a plane, but distinct storm sectors with wildly different thermodynamic and kinematic
conditions lie close to one other; strong wind speeds can transport a dropsonde over great distances, complicating their interpretation (Stern et al. 2016).

Radar is an effective remote sensing capability that enables observations of hydrometeor characteristics from a great distance. However, radars can only determine bulk statistical properties of cloud and precipitation particles, and these statistics are dominated by large particles, which is an issue in regimes dominated by small drops observed in other warm, moist neutral synoptic situations (Blanchard 1953; Martner et al. 2008; Friedrich et al. 2016; Zagrodnik et al., submitted). Dual-polarization radars could partially circumvent this issue, but they are not currently deployed on any research aircraft. Most radars sacrifice high-resolution capabilities to extend the range of observations, which is especially problematic for ground-based radar since orographic enhancement in certain synoptic situations can occur in the lowest kilometer above the surface (Hobbs 1975). Terrain blocking can restrict the view of ground-based radars such that valleys are under-sampled. Airborne radars are mobile and can scan into valleys; however, aircraft motions (pitch, roll, and yaw) complicate the retrieval of weak hydrometeor vertical velocities. Despite numerous observational challenges, it is possible to use airborne radar to advance our understanding of how passage over steep terrain affects precipitation processes in tropical cyclones.

Since rainfall is the integrated product of the instantaneous rain rate and precipitation duration, rain rates do not have to be large to cause dangerously large accumulations, especially over mountains, where runoff is funneled into narrow valleys. Observational studies of tropical cyclones moving past mountainous islands have demonstrated that intense rainfall is a secondary contributor to the total precipitation behind long-lasting moderate rainfall (Misumi 1996; Smith et al. 2009). The precipitation itself is related to the upstream wind speed, terrain height, and
precipitation rate (Yu and Cheng 2008, 2013, 2014). The three-dimensional terrain shape, wind speed, and stability affect the location of maximum precipitation. When the terrain is sufficiently tall and wide, precipitation maximizes along the windward slopes (Yu and Cheng 2008, 2013).

Due to the typically moist-neutral environment of a tropical cyclone eyewall, cloud water generated by simple upslope ascent is often invoked as the dominant modification mechanism in a landfalling tropical cyclone. This hypothesis has not been adequately tested, as microphysical observations alongside surface rain gauge data do not exist, the horizontal radar maps used to infer this process do not reveal the full precipitation structure, and the vertical resolution of previously analyzed radar reflectivity data is not sufficiently high. Furthermore, the eye, eyewall, and rainband do not support the same types of precipitation processes, as evidenced, for example, by deep convection that developed in the eye of Hurricane Georges (1998) over Dominica (Geerts et al. 2000). Finally, distinguishing between tropical cyclone features that are unrelated to terrain and the orographic modification of those features is critical in determining the potential for rainfall enhancement (Smith et al. 2009; Yu and Cheng 2013, 2014).

The precipitation mechanisms of interest to this study cannot be determined from observations alone. Numerical simulations are needed to assess the role of topography through experiments with and without terrain. Such experiments can be checked for consistency with observations and compared with each other to deduce the mechanisms of orographic modification of the cyclone’s precipitation. Full fields of thermodynamic, kinematic, and microphysical properties can be subjected to analysis in these comparisons.

Most modeling studies to date find that precipitation is greater when terrain is present. However, most studies are set on Taiwan or are highly idealized (Bender 1985; Huang et al. 2014; Yang et al. 2008). A study of Tropical Cyclone Larry (2006), which is the only tropical cyclone to
be studied that made landfall over a continental mountain range, finds that while precipitation is
greater over the terrain, precipitation downstream is reduced (Ramsey and Leslie 2008). Neither
this study nor the others inferred or analyzed the specific precipitation processes that occurred, and
although the simulated Larry was found to decay more rapidly when terrain was present, the
average terrain height in northeast Australia is only around 800-1000 m altitude.

The only landfalling storm to have been thoroughly studied is Typhoon Nari (2001), which
tracked parallel to Taiwan’s Central Mountain Range (Yang et al. 2008; Yang et al. 2011a, b, c).
Those studies examined the kinematic, thermodynamic, and microphysical changes as Nari moved
over Taiwan. A suite of topographic modification experiments found that a taller Central Mountain
Range resulted in a weaker storm, increased precipitation, increased latent heat released by upslope
flow that induces stronger primary and secondary circulations, and more dominant cold rain
processes after landfall. However, these studies did not consider how the nature of the precipitation
diffs between the experiments: the changing thermodynamic environment was not investigated
with respect to the specific precipitation mechanisms over the terrain and precipitation processes
farther from the eyewall were not examined in detail.

This dissertation aims to address gaps in our knowledge of how passage over a continental
barrier alters the nature and evolution of precipitation processes in a landfalling tropical cyclone.
Since tropical cyclones are easily disrupted by external forces, loss of the moist neutral inner core
environment and decreasing wind speeds during passage over terrain are expected to affect the
type of modification processes that occur. Since the rate of storm decay varies with terrain height,
it is hypothesized that the height of a continental barrier will affect the type of precipitation that
occurs. Hurricane Karl (2010), which made landfall in Mexico, serves as a case study to test these
hypotheses. The goal of this dissertation is thus twofold:
1. The nature of the orographically modified precipitation processes (i.e., whether modification proceeds through collection of orographically-generated cloud water) during the landfall of Hurricane Karl (2010) will be determined insofar as possible from the radar and dropsonde data collected in that storm.

2. Model simulations of Karl’s landfall with idealized topography of different heights and guided by the data obtained in Karl will test the hypothesis that the height of a continental barrier will affect the degree to which moist neutral precipitation modification processes occur.

The dissertation is organized as follows: Chapter 2 analyzes high-resolution airborne radar data collected in Karl during the NASA Genesis and Rapid Intensification Processes (GRIP) field campaign to examine the vertical structure of precipitation in regions of upslope and downslope flow. This chapter uses airborne radar data to assess whether warm rain processes, specifically cloud water generated by forced ascent, dominate the precipitation processes in the modification of Karl’s eyewall during its passage over the mountainous terrain of Mexico, and satellite data after the DC-8 flight will be used to indicate how a different type of precipitation also contributed to rainfall as the storm lost its coherence as a tropical cyclone. Chapter 3 will use model simulations featuring two orographic modification experiments where numerical simulations of Karl move over idealized plateaus with peak altitudes of 0.5 and 2.5 km. The idealized terrain allows for a controlled examination of how orographic modification processes are impacted by the terrain height and how the rate of decay affects the type of precipitation processes. Furthermore, the experiments provide an opportunity to assess the accuracy of the simulated processes in the model. Together, these chapters provide a comprehensive examination of the evolution and organization of precipitation processes in a landfalling tropical cyclone. Chapter 4 summarizes the results of
Chapters 2 and 3, discusses the overall conclusions, and provides areas of future research motivated by the results of this dissertation.
Airborne radar data collected within Hurricane Karl (2010) provide a high-resolution glimpse of variations in the vertical precipitation structure around complex terrain in eastern Mexico. Widespread precipitation north of Karl’s track traced the strong gradient of terrain, suggesting orographic enhancement. Although the airborne radar did not sample the period of peak precipitation, time series of surface rainfall at three locations near the inner core show greater precipitation where flow was oriented to rise over the terrain. In regions of upslope flow, radar observations reveal reflectivity enhancement within 1-2 km of the surface. The shallow nature of the enhancement points to orographically-generated cloud water accreted by falling drops as a mechanism consistent with prior studies, while the heterogeneous nature of the enhancement and nearby low-level instability suggests shallow convection was playing a role. In contrast, regions of downslope flow were characterized by uniform reflectivity above the ground and fallstreaks originating above the melting level. Unlike most previously studied tropical cyclones passing over topography, Karl made landfall on a mountainous continent, not an island. As Karl weakened and decayed over land, the vertical structure of the radar echo deteriorated north of the storm center, and infrared satellite imagery revealed complete destruction of the cloud field; however, a small region of intense convection appeared and produced locally heavy rainfall as Karl was close to dissipation. These results indicate that orographic modification processes in a landfalling tropical cyclone are not static, and surface precipitation is highly sensitive to the changes.

2.1 INTRODUCTION

In addition to strong winds and storm surge, landfalling tropical cyclones can produce intense rainfall, catastrophic flooding, and/or landslides when they pass over hills and mountains. Understanding the spatial structure and evolution of the precipitation processes around terrain in tropical cyclones is therefore critical to improving precipitation forecasts. Furthermore, as numerical simulations employ increasingly higher resolutions, handling the details of terrain effects on tropical cyclones is becoming increasingly important.

Historically, investigation of the orographic modification of tropical cyclones has focused on the response of the storm track to the evolving environmental flow, which is sensitive to the downstream three-dimensional terrain shape. Upstream of mountainous islands (e.g., Taiwan), disruption of the low-level flow shifts the storm track, where the resulting path depends on the relative positions of the cyclone and the island (Chang 1982; Bender et al. 1987; Roux et al. 2004; Wu et al. 2015). Upstream of continental mountain ranges, a westward-moving Northern Hemispheric tropical cyclone is deflected southward by anomalous anticyclonic and cyclonic flow driven by column compression and stretching on the northern and southern flanks of the storm, respectively (Bender et al. 1985; Zehnder et al. 1993a). In terms of precipitation processes, most research has focused on surface rainfall patterns rather than the cloud structures and mechanisms through which precipitation is enhanced or redistributed. This paper examines the full vertical structure of the precipitation layer to gain insight into these mechanisms.

Orographic modification of cloud microphysics can occur through numerous pathways intrinsically tied to the kinematic and thermodynamic environment in which a cloud is embedded (Smith 1979; Roe 2005; Houze 2012, 2014). Tropical cyclones are unique phenomena with strong radial and azimuthal variations in kinematic and thermodynamic characteristics. With such
diversity in background conditions present over short distances, distinct precipitation types exist in close proximity to one another. Even so, orographic enhancement of the precipitation processes in tropical cyclones is typically attributed to the accretion of cloud water, because the near neutral stability of the inner core prevents blocking of the flow (Rotunno and Ferretti 2001, 2003) and the release of strong buoyant motions triggered by upslope flow (Misumi 1996; Smith et al. 2009; Yu and Cheng 2008). But this attribution is typically inferred from horizontal radar echo patterns, coarse vertical sampling by radar, or surface rain gauge data. Far less analysis has been done with detailed examination of radar data of high enough resolution in the vertical to assess this hypothesis.

Unblocked flow and accretion of cloud liquid water are not the only orographic effects that can occur during a tropical cyclone's landfall. The specific portion of a tropical cyclone that moves over terrain determines the type of modification processes that occur. For example, when the eye of Hurricane Georges (1998) passed over the central mountain range of Hispaniola, high-resolution X-band radar data collected by high altitude aircraft showed intermittent intense convection erupting within the eye of the storm for several hours (Geerts et al. 2000). Since the convection developed rapidly, Geerts et al. (2000) speculated that topographically-forced ascent enabled air to break through the convective inhibition that characterizes the eye; however, depletion of CAPE and reduced surface fluxes could not sustain the convection.

Total precipitation depends on both intensity and duration of the rainfall at a given point. In their analysis of precipitation during Hurricane Dean (2007) over the mountainous island of Dominica, Smith et al. (2009) showed that an outer rainband at one time produced rain rates in excess of 100 mm h\(^{-1}\) at peak elevations. Ultimately, that intense rainfall was short-lived, and a larger fraction of the total precipitation came from weak rain rates within a stationary rainband
aimed at Dominica for twelve hours. Similarly, Misumi (1996) found that weak orographically-enhanced precipitation within Typhoon 9037 between a rainband and the eyewall contributed the largest fraction of observed surface precipitation, not the intense rainband nearby.

The location of rainfall in a tropical cyclone passing over complex terrain is dictated by the orographic geometry, microphysical processes and cyclone kinematics. Upstream of mountain ranges with sufficient height and width in the direction of the flow, surface rainfall can maximize on the windward slopes (Yu and Cheng 2008, 2013). Precipitation can increase in concert with the elevation, although precipitation typically maximizes at high elevations when the mountains do not exceed 2 km altitude (Geerts et al. 2000; Smith et al. 2009). Finally, in the case of lower mountains, if the hydrometeor advection timescale is greater than the particle fall speed the maximum precipitation can shift downwind into the lee (Misumi 1996; Yu and Cheng 2008). The combination of strong gradients in wind direction, storm motion, and stationary topography means that these factors are continually changing and highly variable.

Finally, precipitation processes occurring as a landfalling storm passes over terrain are subject to factors related to the structure and intensity of the storm itself. The precipitation intensity at a given point on land is sensitive not just to modification by the terrain, but also to the background precipitation intensity, wind speed, and the relative positioning of tropical cyclone features (e.g., eye, eyewall and rainbands) with respect to terrain determines the path through which modification proceeds.

In previously studied tropical cyclones, these factors play out in different but significant ways. For example, as Typhoon Xangsane (2000) moved past Taiwan, the peak radar reflectivities over high terrain occurred downstream of the most intense radar reflectivities over the ocean (Yu and Cheng 2008). Yu and Cheng (2013, 2014) showed that the circulation strength of a tropical
cyclone can be a critical factor, the maximum precipitation is often collocated with the strongest wind, and under certain terrain configurations the orographic enhancement of precipitation is proportional to the product of the upstream precipitation and wind speed. Smith et al. (2009) showed that the orientation of the stationary rainband in Hurricane Dean (2007) responsible for the deluge over Dominica was crucial in supplying the moisture and precipitation but was unrelated to the terrain itself. These studies imply that distinguishing between features inherent to a tropical cyclone and the specific orography encountered by a tropical cyclone is critical to evaluating the storm’s potential for rainfall and flooding during its landfall.

The present study is the first to explore how landfall over a broad continental mountain range impacts a tropical cyclone. While the aforementioned studies have focused on the impact of island orography on rainbands (Misumi 1996; Smith et al. 2009; Yu and Cheng 2008) or the inner core in storms that are over land for only 12 hours (Geerts et al. 2000; Yu and Cheng 2013; Yu and Cheng 2014; Yu and Tsai 2017), Karl presents an opportunity to examine the orographic modification processes in a tropical cyclone completely deteriorating during landfall. Our study assesses the hypothesis that warm-rain processes are the dominant orographic precipitation enhancement mechanism during landfall. We achieve this objective by examining the vertical structure of precipitation revealed by airborne radar data collected during the landfall of Hurricane Karl (2010) near Veracruz, Mexico.

First, we introduce the meteorological history of Hurricane Karl and the dataset used in this study. We then discuss the surface precipitation patterns in Mexico during Karl’s landfall, and follow that with a comparison of the structures from airborne radar reflectivity measurements around the terrain and discuss what they reveal about orographic modification. Finally, we
examine the broader applicability of these patterns through an analysis of the storm structure after
data collection had ceased.

2.2 DATA AND METHODOLOGY

Hurricane Karl was investigated as part of the NASA Genesis and Rapid Intensification Processes (GRIP) field experiment, carried out during August-September 2010 to document the inner core processes at the heart of two important questions in tropical meteorology: genesis and rapid intensification. GRIP coordinated efforts with the National Science Foundation’s PRE-Depression Investigation of Cloud-systems in the Tropics (PREDICT) field experiment. Working together these programs employed several aircraft with a variety of instruments (Montgomery et al. 2012; Braun et al. 2013). These flights provided a multi-faceted look at the structure of several tropical systems from birth to decay. Although the investigation of orographic modification processes was not an explicit goal of the campaign, flight planners took advantage of the unique opportunity presented by Karl to obtain high-quality data over complex terrain. The aircraft sampled three stages of Karl: just after genesis on 14 September, rapid intensification on 16 September, and landfall on 17 September. This study focuses on the final flight, which collected data over the nearby water, where dropsondes were launched, followed by a series of inlands legs around the complex terrain and remnant eye.

The key instrument onboard the NASA DC-8 was the Airborne Precipitation Radar 2 (APR-2), developed and operated by NASA’s Jet Propulsion Laboratory (JPL). Instrument specifics are provided in Table 1, which was designed to emulate the Global Precipitation Measurement (GPM) satellite. The radar obtains high-quality data with an along-beam resolution of 30 m. Although the radar scans a narrow cross-track swath of 24 beams ± 25 degrees from nadir, we restrict our study to the beam closest to vertical. The 1.8 s of scan time is comprised of 1.2 s of
data collection and 0.6 s to return the radar beam to the original starting point. The horizontal resolution when flying around 12 km is 800 m at the surface (Sadowy et al. 2003). At the high-frequencies at which APR-2 operates (35.6 GHz and 13.4 GHz at Ka- and Ku-band, respectively), there is potential for strong attenuation in heavy rain; for reference, theoretical estimates of specific attenuation are 0.1-1 dB km\(^{-1}\) (g m\(^{-3}\))\(^{-1}\) and 0.7-7 dB km\(^{-1}\) (g m\(^{-3}\))\(^{-1}\) for Ku- and Ka-band, respectively (Battaglia et al. 2016). Since attenuation is less at Ku-band, we therefore focus our analysis on the Ku-band data. Although dual-frequency retrievals can be useful under the proper circumstances, this study relies heavily on near-surface reflectivity, which often suffered substantial attenuation at Ka-band in the dataset used.

Although Ku-band data suffer less degradation from attenuation than Ka-band data, heavy precipitation still poses a problem. Methods exist to estimate attenuation, which rely on either the measured reflectivity or surface returns. Two such techniques include the Hitschfield and Bordan (1954) and Dual-Wavelength Surface Reference Technique (DSRT) methods (Meneghini et al. 1987). Without belaboring the specifics of each technique, a simplified comparison follows. The former estimates the path-integrated attenuation through an integrated measurement of reflectivity at each gate; however, this method is unstable when the attenuation is large and estimates of radar sensitivity and microphysical properties are inaccurate (Hitschfield and Bordan 1954). The DSRT method circumvents the instability by comparing the ratio of surface return measurements from two frequencies to reference measurements from a nearby non-raining region to constrain estimates of the attenuation (Meneghini et al. 1987). Unfortunately, the DSRT method typically used for APR-2 data is only reliable over the ocean, where the intrinsic variability of the surface return is assumed to be small (Durden et al. 2012).
Table 2.1. Parameters for the APR-2 adapted from Sadowy et al. (2003) and Tanelli et al. (2006).

<table>
<thead>
<tr>
<th>Parameter</th>
<th>13.4 GHz (Ku-band)</th>
<th>35.6 GHz (Ka-band)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Frequency</td>
<td>13.4 GHz (Ku-band)</td>
<td>35.6 GHz (Ka-band)</td>
</tr>
<tr>
<td>Antenna effective diameter</td>
<td>0.4 m</td>
<td>1.4 m</td>
</tr>
<tr>
<td>Antenna gain</td>
<td>34 dBi</td>
<td>33 dBi</td>
</tr>
<tr>
<td>Antenna sidelobe</td>
<td>-30 dB</td>
<td>-30 dB</td>
</tr>
<tr>
<td>Antenna beamwidth</td>
<td>3.8 °</td>
<td>4.8 °</td>
</tr>
<tr>
<td>Peak power</td>
<td>200 W</td>
<td>100 W</td>
</tr>
<tr>
<td>PRF</td>
<td>5000 Hz</td>
<td>5000 Hz</td>
</tr>
<tr>
<td>Range gate resolution</td>
<td>30 m</td>
<td>30 m</td>
</tr>
<tr>
<td>Doppler precision</td>
<td>0.4 m/s</td>
<td>1 m/s</td>
</tr>
</tbody>
</table>
Without a reliable method to correct for attenuation in the flight segments of interest over land, we must exclude compromised data from our analysis. We identified affected radar reflectivity beams on 17 September through a comparison with data from 16 September when Karl was over the ocean and attenuation correction is available. Radar beams on 16 September were considered problematic when the surface attenuation correction in a radar beam exceeded 5 dB; this attenuation typically occurred when at least 50 radar reflectivity range gates (30 m spacing) within a beam, excluding the melting layer, surpassed 35 dBZ. Based on this experience, we removed all data points below the location of maximum reflectivity in beams from the 17 September dataset in which 50 range gates exceeded 35 dBZ. Overall, 38 out of the 580 beams shown in this study (7%) were affected, and our results are not sensitive to the chosen thresholds (within ± 5 dBZ and ± 10 gates).

Finally, the airborne data are complemented by rain gauges operated by Mexico’s Servicio Meteorológico Nacional (SMN). Rain gauge data exist in three time intervals: 24-hour, 3-hour and 10-minute measurements. The 24-hour measurements were the most numerous, with far fewer stations reporting at higher frequencies.

2.3 Hurricane Karl Background

2.3.1 Meteorological history

After developing into a tropical storm on 14 September 2010, Karl first made landfall in the Yucatan peninsula at 1245 UTC 15 September 2010 (Fig. 2.1). Karl weakened only slightly and reemerged in the Bay of Campeche as a weak tropical storm (maximum wind 20 m s$^{-1}$) early 16 September. Between 0000 UTC 16 September and 1200 UTC 17 September Karl rapidly intensified to an intense Category 3 hurricane (maximum wind 57 m s$^{-1}$) during passage over the warm sea surface (between 28 and 30°C, not shown) in the southern Gulf of Mexico (Fig. 2.1).
Figure 2.1. (a) Central pressure (gray line) and maximum wind speed (black line) for Hurricane Karl during 14-18 September 2010. The vertical gray line marks the time of the DC-8 flight. (b) Best track of Hurricane Karl where the filled circles denote the 0000 UTC positioning on each day. Hollow diamonds indicate the 6-hourly positions. All data come from the NHC. The thick black line outlines the state boundary of Veracruz and the thin black lines indicate other state and national boundaries. The dot indicates the location of the city of Misantla and the star indicates the capital city of Veracruz (location of the sounding in Fig. 2.10).
Infrared satellite imagery from 0615 UTC on 17 September in Figure 2.2a depicts Karl’s classic tropical cyclone structure while it was over water. Approaching Mexico, Karl's track shifted southwestward, consistent with theory of a tropical cyclone approaching a large barrier with a northwest-southeast orientation (Zehnder 1993a). The center of Karl made its second landfall at 1645 UTC 17 September, 16 km northwest of the city of Veracruz, while the storm was at an intensity of 50 m s$^{-1}$. When the DC-8 carried out its flight over land, the cloud shield was deteriorating quickly, and the coldest cloud tops were being sheared off to the southwest (Fig. 2.2b, c). By 0000 UTC on 18 September, Karl had weakened to tropical storm intensity with few deep clouds remaining (Fig. 2.2d). Despite continued deterioration of Karl, convection occurred intermittently, and ultimately a large cloud shield developed (Fig. 2.2e, f).

Notwithstanding Karl’s status as a major hurricane at landfall, the winds were not the major impact of the storm; rather, heavy rainfall and devastating landslides contributed significantly to the total damage (Stewart 2011). Eastern Mexico, encompassing the states of Veracruz, Puebla, Hidalgo, Tlaxcala, and Oaxaca, is notable for elevated terrain located in close proximity to the Gulf of Mexico. Mountains extend from northwest to southeast, with a triangular protrusion jutting out towards the ocean in central Veracruz (Fig. 2.1b). Just north of this feature, Misantla, Veracruz, received 452 mm of rain, which was the recorded maximum for the storm (Stewart 2011). Overall, the 24-hour precipitation ending at 1300 UTC 18 September was greatest in two regions (Fig. 2.3): near the coast immediately surrounding the track of the eyewall and along the steep terrain slopes. Around the high terrain, precipitation maximized on the windward slopes, consistent with Yu and Cheng’s (2008) conceptual model of a tropical cyclone passing over a tall, wide barrier.
Figure 2.2. 4-km infrared satellite imagery of Hurricane Karl. White circles denote the 1.5º radius from the interpolated best track center at the time of the image. White dots indicate the locations of Jalapa, Córdoba, and Orizaba. (a) 0615 UTC on 17 September. (b) 1845 UTC on 17 September. (c) 2015 UTC on 17 September. (d) 0015 UTC on 18 September. (e) 0215 UTC on 18 September. (f) 0445 UTC on 18 September.
2.3.2  *Surface rainfall during Karl*

Daily rainfall totals alone are insufficient for determining how movement over terrain impacted precipitation processes; the evolving rainfall intensity more accurately reflects how Karl’s structure and dynamics changed as the storm moved inland. Higher frequency precipitation observations are only available for only three locations within Karl’s inner core (Fig. 2.4): Jalapa, Orizaba and Córdoba. The National Climatic Data Center (NCDC) provided three-hourly data for Jalapa and Orizaba, although these stations did not report data at 0000 UTC 18 September. We subtracted the sum of the three-hourly measurements available from the 24-hour totals to estimate the value for the missing 3-h interval. Three-hourly measurements were available during the DC-8 flight and the estimated data values do not affect the interpretation of our radar analysis. Córdoba had continuous data collection, which we converted from 10-min to 3-h data.

During the 36 h that Karl impacted Mexico, Jalapa, Orizaba, and Córdoba received 182 mm, 92 mm, and 261 mm of rain, respectively (Fig. 2.4a). Even though less than 100 km separates Jalapa, Orizaba, and Córdoba, their respective precipitation time series exhibited notable differences. At Jalapa, precipitation peaked between 1800 UTC 17 September and 0300 UTC 18 September. After a three-hour break, modest precipitation resumed. In Orizaba, rainfall was weak throughout Karl’s landfall, but persisted through 1200 UTC 18 September. In Córdoba, which lies east of Orizaba at a lower elevation, precipitation gently increased before skyrocketing after 0000 UTC when nearly 140 mm fell in a 3-h period. Differences in the relative positions of the three cities are consistent with the precipitation's evolution (Fig. 2.4b). Jalapa is the easternmost city and might be expected to be the first to receive precipitation. However, the complex three-dimensional shape of the mountain range is another factor: the mountains surrounding Jalapa are oriented NE-SW while the mountains west of Orizaba and Córdoba are oriented NW-SE. Thus, the circulation
Figure 2.3. 24-hour precipitation totals (mm) ending at 1300 UTC 18 September. Black line denotes the best track of Hurricane Karl from the NHC. Thick black line indicates the state boundary of Veracruz, Mexico. Dots outlined in red indicate the locations of Jalapa, Córdoba, and Orizaba.
Figure 2.4. (a) Time series of 3-hourly precipitation measurements at Jalapa, Córdoba, and Orizaba. Measurements in orange were missing and calculated from the 24-hour totals in Fig. 3. (b) Track of the DC-8 from 1752 to 2109 UTC on 17 September. The DC-8 began its first inland leg at 1847 UTC on 17 September.
surrounding Karl between 1800 UTC and 2100 UTC provided different flow directions with respect to the terrain (i.e., upslope vs downslope) at the two sites, and the precipitation structure would be expected to differ accordingly. Fortunately, the DC-8 flew close to these locations on 17 September (Fig. 2.4b), enabling a comparison of the radar reflectivity structure between the two regions so that we can infer aspects of the operative precipitation mechanisms in the two locations.

2.4 Inner Core Structure

Before delving into the radar data, we note characteristics of the nearby air mass, because orographic modification is intimately tied to the surrounding thermodynamic and kinematic environment. Dropsonde data in Figure 2.5 show a deep, moisture-rich layer. Despite low-level departures from moist-neutrality near the surface, the instability is minor. For reference, the square of the moist Brunt-Väisälä frequency ($N_m^2$), as derived by Durran and Klemp (1982), is weakly negative below 1.8 km (~$1\times10^{-5}$ s$^{-2}$), indicating weak conditional instability. Between 1.8-2.5 km, values of $N_m$ are $6\times10^{-3}$ s$^{-1}$-$9.5\times10^{-3}$ s$^{-1}$ and the wind speed is ~22-25 m s$^{-1}$; for terrain below 2.5 km, the Froude number ($Fr = U/N_mH$; $U$ is the wind speed and $H$ is the mountain height) is ~1-2. These calculations suggest that any buoyant motions would have been modest and the flow would not have suffered blocking by the terrain. As a result, we expect precipitation enhancement where low-level orographic upslope flow was occurring and generating cloud water to be accreted by falling hydrometeors (Smith et al. 2009; Yu and Cheng, 2008). Any embedded weak convection would have further aided the process. The relative orientation of the flow with the orography, given the easterly component of the winds measured by the dropsonde and expected circulation around Karl, suggests air would have been flowing up over the gently sloping terrain on the southeastern flank of the high terrain that juts northeastward towards the coastline.
Figure 2.5. Sounding from a dropsonde launched from the DC-8 at 1846 UTC. Inset shows the DC-8 track (black line) and the position of the dropsonde (red line).
Following the dropsonde launch, the DC-8 tracked west along the southern flank of the triangular terrain feature near Jalapa before returning to the northeast near Córdoba and Orizaba (Fig. 2.6a). Karl's west-southwestward track sandwiched the storm center between the two legs, as indicated by the 15-min positions calculated from a cubic spline fit to the NHC best track data. As a result, conditions favored upslope flow near the northern segment and downslope or nearly flat flow near the southern segment. Cross sections of radar reflectivity obtained along the segments beginning at 1845 UTC and 1906 UTC on 17 September are shown in Figs. 2.6b and c, respectively, which display data from the radar beams closest to vertical. The surface reflection and small echoes (less than 75 contiguous pixels in the full radar swath) have been removed; however, because of the issues discussed in Section 2, these plots have not been corrected for attenuation. We expect that attenuation was modest in the situations shown, which had modest reflectivity values. Attenuation notwithstanding, striking reflectivity differences are revealed by the cross sections. First, storm-scale differences are present. In the northern leg, the minimum detectable echo top is lower, extending up to only 7 km altitude, whereas on the southern leg the echo was deep, extending up to 10 km altitude (Fig. 2.6b, c). Given the broad nature of these differences and extension of the shallow echo over the ocean (e.g., 1845 through 1846 UTC), we suspect storm deterioration caused by landfall and vertical wind shear is the main determinant of these differences, as opposed to the mountainous terrain. These large-scale changes underscore the importance of recognizing the evolving nature of the storm during landfall seen in Figure 2.2. We address these changes in Section 5.

Analysis of small-scale features in the radar data reveals noticeable differences in the reflectivity structure between the first two flight segments. In the northern leg beginning at 1845 UTC, although the echo is shallow with a broken bright band, there is a strong increase in
Figure 2.6. (a) DC-8 flight segments on 17 September and the corresponding center of Karl as interpolated from the NHC best track data. Times next to the hurricane symbols indicate the time to which the best track data were interpolated. Colors denote the matching flight track segment and center, where black indicates the first pair and light gray indicates the last pair. Times next to the flight track indicate the starting point of each track. White dots indicate the position of the plane at every minute along the track. Locations of Jalapa, Córdoba, and Orizaba are labeled with their first initial. (b) Reflectivity cross section for the segment beginning at 1845 UTC. Thick red line corresponds to the red line of leg 1 shown in Fig. 8a. (c) As in (b), but for the segment beginning at 1906 UTC. Thick blue line corresponds to leg 2 shown in Fig. 8a. (d) As in (b), but for the segment beginning at 2049 UTC. Thick blue line corresponds to leg 4 shown in Fig. 8a.
reflectivity towards the surface, at times increasing by 10 dB from 4.5 km altitude to the surface (Fig. 2.6b). The low-level enhancement is more frequent over the terrain where upslope flow is favorable, lending credence to the idea that passage over the terrain causes this feature. Overall, the most intense reflectivity values reside within 1-2 km of the surface. Slight enhancement of a similar character evident between 1845 and 1847 UTC, upstream of the sloping terrain, might have resulted from lifting ahead of the downstream terrain since lifting induced by a terrain barrier can be felt at a considerable distance upstream (Grossman and Durran 1984). The features over the terrain are consistent with vertical cross-sections of radar reflectivity data from typhoons near Taiwan (Yu and Cheng 2008; Yu and Tsai 2014). In their study of Typhoon Xangsane (2000), Yu and Cheng (2008) demonstrated that reflectivity intensity over terrain increases with wind speed (c.f. their Figs. 11-13). In Karl, upstream wind speeds 1 km above sea level were between 20 and 25 m s$^{-1}$ (Fig. 2.5), but the reflectivity was weaker than the data for the same range of wind speeds in Yu and Cheng (2008). The weak reflectivities in Karl are likely a result of weak background precipitation and reduced hydrometeors aloft.

An advantage of the APR-2 over a ground-based radar is that its high resolution combined with the aircraft speed allows for observation small-scale spatial structures and essentially continuous vertical resolution. The latter is especially important for inference of precipitation mechanisms, which vary strongly in the vertical dimension. Despite smoothly increasing terrain, the reflectivity increase was not uniform, with pockets of more intense reflectivity embedded along the flight track. In the upslope flow (1848–1852 UTC), intense reflectivity maxima lay in pockets beneath and disconnected from bright band. Although microphysical measurements do not exist for this case, the shallow nature of the enhancement is consistent with accretion of orographically-generated cloud by existing hydrometeors as a likely mechanism, as has been hypothesized in prior
studies that lacked high-resolution vertically continuous data (Misumi 1996; Smith et al. 2009; Yu and Cheng 2008). The non-uniformity in the near-surface echo could result from variable hydrometeor concentrations aloft seeding the low-level precipitation field or embedded shallow convection, which cannot be discounted without accurate velocity information. Most importantly, deep convection was not present.

In contrast, data collected during the southern leg beginning at 1906 UTC showed no low-level reflectivity enhancement (Fig. 2.6c). Instead, the precipitation was organized into fallstreaks connected to and emanating from the bright band. The intensity of the fallstreaks remained approximately constant towards the surface. The fallstreaks descend away from the highest terrain, providing evidence that the ambient flow carried hydrometeors down the sloping terrain. This reflectivity structure southwest of the storm center was persistent throughout the flight. Immediately prior to departing Karl, the DC-8 flew northeastward past Orizaba and Cordoba beginning at 2049 UTC on 17 September and again encountered deep stratiform rain (Fig. 2.6d). Consistent with the earlier leg, sloping fallstreaks were present. The nearly constant reflectivity intensity with height and connection with the bright band in the downslope legs suggests that precipitation originating aloft was not modified by cloud water or tiny raindrops.

Midway through the flight, the DC-8 completed a long straight leg from the northwest to the southeast (Fig. 2.7). The APR-2 captured a wide range of precipitation types across the storm, culminating in deep, strong stratiform rain on the southeast side of Karl (Fig. 2.7b). Low-level enhancement similar to Figure 2.6b exists between 1943 and 1948 UTC. North of the highest terrain (e.g., 1944 to 1946 UTC), low-level enhancement in concentrated in two small regions below a strong brightband. This connection with the brightband is muted south of the high terrain (e.g., 1946 to 1948 UTC), where the overall echo is shallow, indicative of the lack of falling
hydrometeors accreting water below. Finally, a region of intense, convective reflectivities in what
was likely the remnant eyewall exists between 1948 and 1949 UTC. Many beams in this region
were flagged by the attenuation algorithm, which is noticeable where weak reflectivities lay below
reflectivities surpassing 40 dBZ. It is possible that movement over the higher terrain released these
buoyant motions, but deep upward motion already existed in the eyewall; we lack the observations
to understand how it evolved. Given the position of the northern eyewall and Karl’s track south of
Córdoba, it is possible that this convection was a precursor of the convection that developed around
0200 UTC on 18 September (Fig. 2.2e); however, 5 h passed between Figure 2.7b and Figure 2e
and later segments of the flight track suggest that the convection was deteriorating during the flight
(not shown). The dearth of observations of the eyewall evolution prevents further interpretation of
this aspect of Karl’s landfall.

The differing reflectivity structures in regions of upslope and downslope flow indicate
disparate precipitation processes, which had strong implications for surface precipitation. This
difference is reflected in the measured rainfall between 1800 UTC and 2100 UTC on 17 September
indicated in Figure 2.4: Jalapa (40 mm), near the path of the northern leg, received more
precipitation than Orizaba (7 mm) and Córdoba (20 mm), which lie close to the southern legs.

Because only three-hourly rainfall measurements are available at Jalapa and Orizaba, a detailed
comparison of rainfall between the flight legs is not possible. We further note that the rainfall rates
during the flight are unimpressive, especially when considering orographic rainfall. Weak rainfall
totals do not, however, signify a lack of orographic enhancement or modification, but rather that
the rainfall intensity before encountering terrain is an important determinant of the final rainfall
amount, as demonstrated by Yu and Cheng (2008).
Figure 2.7. (a) DC-8 flight segment on 17 September beginning at 1943 UTC and the corresponding center of Karl as interpolated from the NHC best track data. Time next to the hurricane symbol indicates the time to which the best track data were interpolated. Time next to the flight track indicates the starting point of the track. White dots indicate the position of the plane at every even numbered minute along the track. Locations of Jalapa, Córdoba, and Orizaba are labeled with their first initial. (b) Reflectivity cross section for the segment shown in (a). Thick red line corresponds to the red line of leg 3 in Fig. 8a.
To confirm that the low-level structure is different between the legs in Figures 2.6a and 2.7b and Figures 2.6b,c, we isolated portions of the flight track where we would expect upslope and downslope (or flat) flow (Fig. 2.8a). The first and third legs lie north of the storm center where an easterly wind component is expected, in the same direction as the gradient in terrain (i.e., upslope flow). The second and fourth leg lie southwest of the storm center, suggesting a northwesterly wind component, in either the opposite direction of or perpendicular to the gradient in terrain (i.e., downslope or flat flow). We excluded the convective eyewall portion of leg 3 as the precipitation type is expected to be different. Restricting our analysis to those radar gates below the bright band, we removed any data likely affected by substantial attenuation (recall Section 2). We excluded any beams where the radar echo is too shallow (mean altitude above surface less than 1 km or depth of echo less than 1.5 km), too far removed from the surface (mean altitude above surface greater than 2.5 km), or too weak (mean reflectivity less than 10 dBZ). Remaining are radar reflectivity from liquid precipitation echoes that were deep enough, while still connected to the ground, to calculate the reflectivity slope above the surface. We use linear regression to calculate the reflectivity slope for each beam. The slopes have been segregated based on whether we expect downslope or upslope flow. The distributions are displayed in Figure 2.8b. Although there is large spread for beams found in each flow regime, the medians are significantly different at the 95\% level by the Wilcoxon-Mann-Whitney test (p < 0.001). Reflectivity slopes in the downslope flow are modest, with a median value just below 0 dB per km, stretching from 1 to -2 dB per km. In contrast, the middle upslope beams have reflectivity slopes between -1 and -3 dB per km, with a median slope of -2 dB per km. This difference is consistent with the picture given by the sample cross sections shown in Figures 2.6b,c, showing that this structure is consistent throughout the flight.
Figure 2.8. (a) DC-8 flight segments on 17 September from Figs. 6 and 7. Red lines indicate flight track segments in upslope flow and blue lines indicate flight track segments in downslope flow. (b) Box and whisker plots of radar reflectivity slope for radar beams in upslope and downslope segments shown in (a). Horizontal colored lines indicate the 25th, 50th and 75th percentiles. Numbers at the top of the figure indicate the number of radar beams included in each plot.
In light of these results, it is important to recall that radars respond to both the number and size of liquid hydrometeors, with the latter usually dominating. Unfortunately, no microphysical data were collected from either airborne or ground-based platforms. Prior orographic modification studies in a variety of synoptic environments indicate that changes in the number of small drops, medium drops, large drops, and combinations thereof, are possible (Blanchard 1953; Martner et al. 2008; Friedrich et al. 2016). Moreover, changes to the drop-size distribution may manifest in different ways in upslope and downslope flow. In a tropical cyclone where we expect warm rain processes dominate, we speculate that upslope enhancement consists of droplet growth by accretion and creation of new tiny drops, whereas adiabatic warming during downslope flow causes the smallest drops to evaporate. Since airborne radar measurements are more sensitive to changes in large particles, the inability to properly document changes to the full distribution of drop sizes could impact estimates of liquid water content. Given the importance of the drop-size distribution to surface rainfall, this issue needs to be addressed in future microphysical studies.

2.5 CONVECTION AT LANDFALL

Although airborne observations provide valuable data in remote locations inaccessible by ground instrumentation, fuel and crew limitations prevent continuous data collection. In the investigation of Hurricane Karl on the day described here the DC-8 obtained data around the inner core of the storm for only three hours as the aircraft flew over land (Fig. 2.1a). As shown in Figure 2.4, impacts were longer-lasting, with rain occurring in the 24 h before 1200 UTC 18 September, at which time NHC declared that Karl had dissipated. The peak precipitation in Jalapa occurred during the three hours following the flight and in Córdoba between 3 and 6 UTC 18 September, when Karl was rapidly deteriorating. Several evolving factors determined the delayed arrival of the peak rainfall: the wind direction, region of the storm encountering terrain, and storm structure.
As Karl progressed west southwestward, the ground-relative wind direction shifted, increasing the easterly component of the wind (i.e., upslope flow) in the vicinity of Córdoba. At the same time, continued progression of Karl brought the eye closer to the terrain; compared with the eyewall, the eye undergoes different modification processes, possibly of the type described by Geerts et al. (2000). The convective eyewall shown in Figure 2.7b would have also been brought closer to the higher terrain, which may have potentially been amplified. As the storm weakened, the inner core lost its structural integrity and efficiency of vertical transport of high-$\theta_e$ air. Finally, the wind speed weakened and potentially limited the ability for air to flow up terrain instead of being blocked. Without continued measurements of precipitation and flight data, determining which factor contributed most to the delayed rainfall is difficult.

As Karl weakened to a tropical storm and then a tropical depression, convection developed along the flank of the terrain near Córdoba (Fig. 2.2d, e, f). More formally, Figure 2.9 shows how the distribution of IR brightness temperatures surrounding Karl changed through its lifecycle. This comparison was accomplished by interpolating the NHC best track to the time of each satellite image using a cubic spline scheme and calculating the distribution of brightness temperature within a 1.5º radius of the calculated center. Although the coldest cloud tops disappeared as Karl decayed through 0000 UTC on 18 September, there was a resurgence of the coldest temperatures after 0200 UTC on 18 September despite the greater frequency of warm temperatures. The convection was deep but horizontally limited. It coincided with the deluge at Córdoba, where 115 mm fell in 80 min. Even though the cloud shield continued to expand as rain subsided in Córdoba, the general sequence of events is consistent since convective rainfall can evolve quickly. Rain gauges do not blanket Mexico and precipitation could have moved outside the region sampled by ground instrumentation. Nonetheless, the coincident timing of the emerging cloud shield and intense
Figure 2.9. Time evolution of infrared brightness temperature frequency within 1.5º of the center of Hurricane Karl.
rainfall suggests convection played a role in accumulation totals, despite a heavily weakened Karl. Two potential drivers of the convection are forced lifting of the eye as in Geerts et al. (2000) or destruction of the structure of Karl, reducing upward transport of high-\(\theta_e\) air and modifying the thermodynamic profile.

Unfortunately, no radar data exist to resolve this question. We cannot ascertain whether an eye, with potential instability present in the warm, dry deep layer residing over top the moist boundary layer air, was still present. The final DC-8 pass through Karl revealed a remnant echo-free eye just before 2100 UTC (not shown), but the cloud presentation no longer resembled a tropical cyclone at the time of peak rainfall (Fig. 2.2d, e). In contrast, Hurricane Georges (1998) analyzed by Geerts et al. (2000) weakened just slightly and maintained a classic satellite presentation when the convection occurred. However, that storm passed over an island and could maintain contact with warm ocean, whereas Karl was completely over land with no chance to maintain tropical cyclone structure. Whichever mechanism occurred, low-level high-\(\theta_e\) air remained in the area to feed convection. A coastal sounding from Veracruz at 0000 UTC on 18 September shows instability due to a warm, moisture-rich layer of easterly flow at low levels (Fig. 2.10). Even though the sounding was removed from the convection, the surrounding environment would have supported buoyant motions. The practical lesson from this convective event is: modification processes other than accretion of enhanced cloud water over the terrain exist, and strong transient local impacts can be caused even by weak tropical cyclones.

2.6 Conclusions

We have examined airborne radar data to understand how the vertical reflectivity structure in a landfalling tropical cyclone varied around complex terrain. While prior studies have documented the surface precipitation patterns and horizontal radar structures, most have not had
Figure 2.10. Sounding launched from the city of Veracruz at 0000 UTC on 18 September.
access to high-quality data in the vertical dimension. A notable exception is the study of Geerts et al. (2000), but that study documented only processes associated with the eye of the storm. The high-resolution APR-2 data in the vertical dimension obtained in Hurricane Karl (2010) facilitated a test of the hypothesis that warm-rain processes dominate orographic modification of a tropical cyclone.

The radar analysis of Karl after the storm's landfall has revealed substantial differences in regions of upslope and downslope flow. Where upslope flow is favorable, modification of Hurricane Karl manifests in the form of low-level reflectivity enhancement. Since the enhancement is shallow, accretion of orographically-generated cloud water is the likely primary mechanism, consistent with suggestions from prior studies (Misumi 1996; Yu and Cheng 2008; Smith et al. 2009). However, the enhancement is not spatially uniform, either a result of variations in the background precipitation, embedded shallow convection, or a combination of the two. In contrast, in downslope or flat flow, the presence of fallstreaks that have constant intensity with height indicates that hydrometeors originating aloft do not grow from accretion of cloud water nor do their numbers increase through the creation of new, small raindrops. The contrasting precipitation types were reflected in the surface rainfall, for although measurements were infrequent, precipitation was greater in the vicinity of the upslope flow. Despite orographic enhancement present during the flight, the resulting precipitation rates were muted compared with other tropical cyclone orographic studies (Smith et al. 2009; Yu and Cheng 2013; Yu and Cheng 2014). Weakening of Karl, deterioration of the vertical precipitation structure, and suboptimally oriented flow all prevented larger accumulation.
Orographic modification processes are influenced by a variety of factors in time and space. They occur within an evolving storm structure and depend on the relative orientation of the flow and orography as well as the specific region of the storm encountering the terrain. Despite rapid deterioration of the Karl as it made landfall, a short period of very heavy rain fell along the southern terrain near Córdoba. Satellite imagery indicates that a sudden intense convective development was responsible. The driving mechanism could have been lifting of remnant high-$\theta_e$ air from the decayed storm or the potentially unstable eye as in the case studied by Geerts et al. (2000).

This work, made possible by a unique aircraft investigation, has examined a few hours within a single storm interacting with a unique terrain configuration. The broader applicability of our results remains unknown. Numerical simulations currently underway will determine in a more generic manner the sensitivity of modification processes investigated here to storm intensity, storm track, complex three-dimensional terrain shape, and storm sector.
Chapter 3. THE IMPACT OF TERRAIN HEIGHT ON PRECIPITATION PROCESSES IN A LANDFALLING TROPICAL CYCLONE

This chapter analyzes numerical simulations to investigate how the height of a continental mountain range, which affects the rate of decay of a landfalling tropical cyclone, impacts the processes controlling precipitation formation. Simulations of Hurricane Karl (2010) are generated by the Weather Research and Forecasting (WRF) model, where the Mexican terrain is replaced with an idealized plateau. Two experiments, each comprised of a control simulation and nine ensemble members, are designed with plateau heights of 0.5 and 2.5 km altitude. The control runs are first utilized to document the changing horizontal and vertical spatial patterns of the thermodynamic and microphysical structures. Next, the ensemble members are used to quantitatively analyze the temporal evolution of the thermodynamic, kinematic, and microphysical structures of each experiment. The two experiments reveal that the rate of decay determined by the plateau height induces disparate evolutions of the kinematic and thermodynamic structures. When the terrain is shallow, the weak rate of decay enables the simulated storm to retain a deep warm core, moist neutral environment within the robust eyewall and rainband for a longer period of time. When the terrain is tall, rapid decay during landfall causes the warm core to shrink, removing the environment suitable for moist neutral processes. The simulations indicate shallow buoyant motions are more frequent than previously suggested near the remnant eyewall, whereas widespread convection develops in the broad region of upslope flow north of the storm. During a

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nine-hour period as the storms pass over the sloping edge of each plateau, the maximum precipitation values for each experiment are comparable between the two experiments, but the horizontal organization differs. From a microphysical standpoint, cloud water mixing ratios are greater during landfall over the sloping terrain in the tall plateau experiment. The enhanced cloud water is due to a mix of gentle ascent and buoyant motions near the inner core and convective motions at larger radii. Enhanced cloud water is located near locally enhanced rain mixing ratios, indicating a strong role of warm rain processes in precipitation enhancement. However, rain mixing ratio values and the precipitation intensity are correlated with graupel mixing ratios aloft, which exceed observational values by an order of magnitude. This result indicates that graupel overestimation in microphysical schemes needs to be addressed and that realistic precipitation patterns may be obtained through incorrect processes. This chapter concludes that the height of a continental mountain range dictates the rate of storm decay, which determines the manner in which precipitation is created.
3.1 INTRODUCTION

It is well-known that passage over complex terrain affects a tropical cyclone in myriad ways, but great emphasis in the literature is placed on changes to tropical cyclone track and intensity. Modification of the surrounding environmental flow by terrain shifts the track of a tropical cyclone. For example, a westward-moving Northern Hemispheric tropical cyclone is deflected southward by anomalous mesoscale circulations that develop upstream of the terrain along the flanks of the storm. Column compression and stretching drives anticyclonic and cyclonic circulations that develop on the northern and southern flanks of the storm, respectively (Bender et al. 1985; Zehnder et al. 1993a). Additionally, movement over terrain induces storm decay due in part to intrusion of low-level air dehydrated by adiabatic descent, loss of access to the warm, moisture-rich boundary layer over the ocean, and increased surface friction (Ramsay and Leslie 2008; Yang et al. 2008; Yang et al. 2011c).

Passage over terrain also affects tropical cyclone precipitation, but greater attention in the literature is given to the changing surface pattern and not the specific processes that determine where and how much precipitation accumulates. In general, precipitation is more intense when mountains are taller in both observational and modeling studies (Bender et al. 1985; Huang et al. 2014; Misumi 1996; Smith et al. 2009; Yang et al. 2008; Yu and Cheng 2008; Yu and Cheng 2013, 2014; Yu and Tsai 2017). Observational studies show that the three-dimensional terrain shape and wind speed dictate the location of maximum rainfall, but precipitation in a landfalling tropical cyclone typically maximizes along the windward slopes when the mountain range is sufficiently tall and wide (Yu and Cheng 2008, 2013).

In terms of precipitation processes, the orographic modification mechanisms that occur depend upon the thermodynamic and kinematic environment of the air mass passing over a
mountain range (Smith 1979; Roe 2005; Houze 2012, 2014). Tropical cyclones are notable for regions with disparate horizontal and vertical stabilities that exist in close proximity to one another. Each tropical cyclone region favors different precipitation types (Houze 2010). The eye is stable with no precipitation and weak wind speeds. The eyewall is characterized by intense rain where the upward branch of the conditionally symmetrically neutral secondary circulation is located. At larger radii, the rainbands exhibit a mix of stratiform and convective rainfall. Furthermore, an external force like the environmental vertical wind shear systematically organizes stratiform and convective processes in different azimuthal sectors around the storm, especially in the rainbands (Corbosiero and Molinari 2002, 2003; Didlake and Houze 2013a,b; Hence and Houze 2012). As a result, the orographic modification of each tropical cyclone region would not manifest in the same precipitation processes. Despite these structural differences, the orographic modification of tropical cyclones is often attributed to collection of orographically-generated cloud water by falling raindrops, since strong wind speeds in the often-assumed moist neutral environment would favor laminar flow over buoyant motions and blocked flow (Misumi 1996; Rotunno and Ferretti 2001, 2003; Smith et al. 2009; Yu and Cheng 2008). Nonetheless, evidence exists that deep convection within a lifted eye (Geerts et al. 2000), shallow convection outside of the eyewall, and deep convection in the wake of a deteriorated storm (Chapter 2) are possible. Since landfall modifies the three-dimensional thermodynamic and kinematic environment within a tropical cyclone, the modification processes should not be assumed to be static.

In general, the studies that analyze the changing precipitation processes within tropical cyclones moving over terrain examine storms that do not fully decay. The observed and simulated tropical cyclones either move past mountainous islands where only the rainbands encounter elevated terrain (Misumi 1996; Smith et al. 2009; Yu and Cheng 2008), the inner core exits a
mountain range quickly (Geerts et al. 2000; Yu and Cheng 2013; Yu and Cheng 2014; Liu et al 2016; Yu and Tsai 2017), or the storm skirts along a coastline removed from elevated terrain (Liu and Smith 2016; Liu et al. 2016). Only one study examines a tropical cyclone making a complete landfall over a continental mountain range: an analysis of Tropical Cyclone Larry (2006), which made landfall over mountainous northeastern Australia (Ramsay and Leslie 2008). In a comparison of two topographic modification experiments, surface rainfall is initially larger when the terrain is taller, but as Larry decays the precipitation that accumulates further inland is weaker than in the experiment where the mountain range is eliminated. However, the three-dimensional processes that control precipitation formation and redistribution are unexamined.

Several studies examining the changing structure of Typhoon Nari (2001) perform a thorough analysis of the changing kinematic and thermodynamic conditions that dictate the type of precipitation processes over terrain (Yang et al. 2008; Yang et al. 2011a, b, c). Nari was a moderately strong typhoon that made landfall in northern Taiwan and tracked parallel to Taiwan’s Central Mountain Range over a 48-h period. Despite weakening during the initial landfall, Nari maintained a weak, but constant intensity even as the inner core remained over land. Consistent with other studies, several terrain modification experiments revealed that precipitation generally increased as the height of the mountain range increased. Additionally, the studies found that the intrusion of midlevel low-\(\theta_e\) air into the inner core leads to storm decay, the presence of high terrain induced a stronger secondary circulation that enhanced horizontal vapor transport, and a mixture of warm and cold rain processes contribute to precipitation formation above terrain. But the studies focus heavily on the initial landfall, do not thoroughly examine the processes contributing to precipitation formation away from the eyewall, and do not thoroughly examine how the processes evolve as the warm core is disrupted.
In light of this general knowledge, some outstanding questions remain. First, how does landfall over a continental mountain range affect the precipitation processes in a tropical cyclone? Second, since the height of a topographic barrier affects the rate of decay, does terrain height affect the manner in which precipitation processes evolve? Third, what are the relative roles of warm and cold rain processes in the topographic modification of tropical cyclones? Fourth, how realistic are the structures simulated by numerical models when compared with observations?

In an attempt to answer these questions, this chapter analyzes numerical simulations of Hurricane Karl (2010) to examine the evolving conditions that control precipitation formation and to compare the simulated processes with the observations presented in Chapter 2. Chapter 2 describes the meteorological history of Karl in detail, but a short discussion follows. Karl made landfall as a weak Category 3 hurricane, deteriorated quickly, and produced a wide swath of precipitation over the mountains in eastern Mexico (Stewart 2011). Despite steep terrain, the precipitation totals were muted in comparison with rainfall observed in other tropical cyclones that moved past small islands such as Taiwan or Dominica, suggesting that storm decay limits precipitation accumulation in a landfalling tropical cyclone. The prevalence of low-level enhancement observed by airborne radar is consistent with the hypothesis that warm rain processes dominate precipitation enhancement as a tropical cyclone moves over elevated terrain. However, shallow convection suggested by a heterogeneous radar reflectivity echo and upstream low-level instability along with the deep convection that developed in Karl’s wake indicate that the changing three-dimensional structure within a decaying storm favors alternative modification processes. Since the rate of storm decay is related to terrain height, Chapter 3 analyzes two terrain modification experiments to assess the hypothesis that terrain height impacts the type of precipitation processes that occur.
3.2 Methodology

Simulations were performed using version 3.8.1 of the National Center for Atmospheric Research (NCAR) Advanced Research Weather Research and Forecasting (WRF-ARW) model (Skamarock et al. 2008). The model was initialized with the ERA-Interim analysis at 0000 UTC 15 September 2010 and was run for 72 h through 0000 UTC 18 September. This time period encompasses Karl’s development from a weak tropical storm to a major hurricane, second landfall over Mexico, and subsequent weakening back down to a tropical storm (Stewart 2011). The model configuration consisted of four nested grids with horizontal resolutions of 54, 18, 6, and 2 km. The two inner domains spanned 210 grid points in the north-south and west-east directions and followed the vortex at 750 hPa. The results of this chapter are derived from the model output from the innermost domain. For the physical parameterizations, this study utilized the Goddard microphysics scheme (Tao et al. 1989) and the Mellor–Yamada–Janjić Scheme (MYJ) boundary layer scheme (Janjic 1994). In the initial testing of microphysical parameterizations, the more sophisticated Thompson microphysics scheme (Thompson et al. 2008) was included, but the simulated storm consistently moved too quickly across the Gulf of Mexico to be considered in this study. Additional details regarding the model configuration and references are described in Table 3.1.

This chapter examines the effect of terrain height on the precipitation processes in a landfalling tropical cyclone through its influence on the rate of storm decay; however, the Mexican terrain exhibits strong horizontal variations in Mexico where Karl made landfall (Fig. 2.1b). Since large changes to the vertical dimension of a continental mountain range induce systematic shifts in the tropical cyclone track (Zehnder et al. 1993), simply multiplying the height of the existing terrain by a scale factor would cause the simulated storms to encounter distinct horizontal patterns
Table 3.1. WRF Parameterizations and References.

<table>
<thead>
<tr>
<th>Model Process</th>
<th>Scheme</th>
<th>Reference</th>
</tr>
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<tbody>
<tr>
<td>Microphysical Parameterization</td>
<td>Goddard</td>
<td>Tao et al. (1989)</td>
</tr>
<tr>
<td>Cumulus Parameterization</td>
<td>Kain-Fritsch, none on domains 3 and 4</td>
<td>Kain (2004)</td>
</tr>
<tr>
<td>Shortwave Radiation</td>
<td>Dudhia</td>
<td>Dudhia (1989)</td>
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<tr>
<td>Longwave Radiation</td>
<td>Rapid Radiative Transfer Model</td>
<td>Mlawer et al. (1987)</td>
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<td>Land Surface</td>
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<td>Tewari et al. (2004)</td>
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<tr>
<td>Surface Layer</td>
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of terrain. Additional complexity due to different patterns of convergence, varying distance of the terrain from the ocean, and gradients in terrain height would add unwanted signals to the model data that are not a focus of this study. To better ensure that the changing precipitation processes are due the different terrain heights, the Mexican mountain ranges are replaced with the idealized plateau shown in Figure 3.1a.

The curved edges of the plateau are described by the same equations that Menchaca and Durran (2017) use to construct an isolated ridge (see their equation 1), except the terrain slope is restricted to $0.8 \leq r \leq 1$; when $r < 0.8$, the terrain height is equal to the maximum plateau height. For the purposes of this study, the plateau was rotated 32.5° counterclockwise from due north to run parallel to the eastern coastline. To fit within the geographical boundaries of Mexico, a similar half-ridge of sloping terrain rotated 80° counterclockwise from due north replaced the southeastern portion of the plateau. The idealized terrain was then lightly smoothed to remove discontinuities and blend in with the existing terrain. Finally, all oceanic data points were set equal to 0 km altitude. This study uses plateau heights of 0.5 and 2.5 km altitude and the two experiments are hereafter referred to as the short plateau and tall plateau experiments, respectively. Although the short plateau experiment requires the interpolation of reanalysis data on pressure levels that exist below the ground, 48 hours pass before the storm approaches the terrain for environmental changes to dissipate. Additionally, the thermodynamic profiles above 2.5 km for experiment are similar and the low-level thermodynamic profile over the short plateau would not prevent upslope flow.

3.3 THE IMPACT OF TERRAIN ON HURRICANE KARL’S TRACK AND INTENSITY

The different plateau heights have far-reaching impacts on the track and intensity evolutions for each experiment (Fig. 3.1). Reducing the plateau height results in a systematic northward shift in the storm track between the tall and short plateau experiments, consistent with
Figure 3.1. (a) Simulated storm tracks for the short plateau (northern track) and tall plateau (southern track) experiments. The control members are denoted by grey lines and the observed track of Karl is denoted by a thick black line. Filled circles indicate 0000 UTC points and the open diamonds indicate 0600, 1200, and 1800 UTC points. All tracks go through 0000 UTC 18 September. The colors indicate the elevation of the idealized plateau in the tall plateau experiment (the short plateau is the same shape, with a maximum elevation of 500 m). (b) Mean sea level pressure for the tall plateau and short plateau experiments and Karl. Blue line denotes the tall plateau experiment and the red line denotes the short plateau experiment. Filled circles indicate the best track intensity of Karl from the NHC.
theory (Zehnder et al. 1993). The simulated track for the tall plateau experiment is closest to the observed track provided by the National Hurricane Center (NHC), lying just to the south of Karl’s track. In contrast, the storm simulated in the short plateau experiment makes landfall three degrees north of Karl’s observed track. Each simulated storm track was calculated using a simplex algorithm calculates the center location that maximizes the tangential wind speed at 2 km altitude within a 5-km radius (Nelder and Mead 1965). The algorithm is most effective when the vortex is well-defined on a constant altitude surface. When the storm was weak or near the steep terrain of the tall plateau experiment, the center had to be determined manually. When the track was noticeably off, the storm track was subjectively edited to identify the approximate center of the circulation at 2 km altitude. When the tall plateau storm moved over the 2.5 km tall plateau, the storm center was subjectively identified at 3 km altitude.

The two simulated storms reach similar peak intensities of approximately 976 hPa before landfall (Fig. 3.1b). Both simulated storms fail to reproduce the peak intensity of Karl, due to a combination of initial condition errors and faster than observed storm motion across the Gulf of Mexico. The resolution of the global reanalysis data (0.75°) is likely insufficient to initialize the vortex. Additionally, the tropical cyclone simulated by the tall plateau experiment made landfall four hours before the observed storm. A reduction in time spent over the ocean would limit the intensification of the simulated storm. These issues aside, the primary objective to determine the effect terrain height has on precipitation processes is still achievable through a sensitivity study comparing the two simulations. Even though the tall plateau experiment fails to adequately replicate Karl’s intensity, several similarities between the tall plateau storm and Karl exist, which are discussed later in this chapter.
Landfall is known to be a negative factor to storm intensity and the two experiments experience different rates of decay after moving over Mexico. The tall plateau storm experiences rapid decay after moving inland as the minimum surface pressure increases by approximately 25 hPa over 12 hours (Fig. 3.1b). The rate of decay is weaker for the short plateau experiment, where the minimum surface pressure increases just 12 hPa over 12 hours. The different rates of decay are further reflected in Figure 3.2, which shows the time evolution of the mean tangential wind speed on all model levels that lie between 0.5 and 1.0 km above the plateau. As the two storms approach the eastern edge of the plateau (model hours 60 and 64, for the tall and short plateau experiments, respectively), the azimuthal mean tangential wind speeds are characterized by similar intensities between 35 and 40 m s\(^{-1}\). The tall plateau storm is slightly stronger than the short storm initially, partially due to channeling of the wind field along the terrain (not shown). But the mean wind speed for the tall plateau storm plummets from 40 to 17.5 m s\(^{-1}\) over 8 h. Meanwhile, the mean wind speed for the short plateau storm gently decays from 35 to 27 m s\(^{-1}\) over 8 h. A quicker rate of decay during passage over higher terrain is consistent with prior tropical cyclone orographic modification studies (Huang et al. 2014; Lin et al. 2002; Ramsay and Leslie 2008; Yang et al. 2008).

3.4 Evolution of the Three-Dimensional Structures

To ensure an equitable comparison of the three-dimensional structures simulated by the two experiments, this dissertation isolates the nine-hour period surrounding the hour when each tropical cyclone passes over the middle of the sloping terrain. The nine-hour period captures both the approach and departure of the storms with respect to the sloping terrain. Additionally, the innermost domain of the model only spans 420 km in the north-south and east-west dimensions and a nine-hour period ensures sufficient horizontal overlap of the first and last domain locations.
Figure 3.2. (a) Time evolution of the mean tangential wind from the tall plateau experiment on model levels between 0.5 and 2.0 km above the surface between model hours 60 and 72. (b) As in (a), but from the short plateau experiment.
The middle of the sloping terrain and the hour the storm center reaches that location are hereafter referred to as the midpoint and midpoint time, respectively.

Reflecting the different rates of decay, the two simulated storm structures undergo distinct evolutions as inferred by the simulated outgoing longwave radiation (OLR), which is used to approximate cloud height (Fig. 3.3). Two hours prior to the midpoint time, both storms exhibit cloud shields typical of tropical cyclones, with well-defined eyes visible in the center of each storm (Figs. 3.3a,d). At the midpoint time, the tall plateau storm has lost its eye, the cloud shield weakens, and the overall structure is less coherent (Fig. 3.3b). By the end of the nine-hour analysis period, no recognizable tropical cyclone features exist as deeper clouds near the remnant center have moved west quickly leaving scattered clouds over the terrain (Fig. 3.3c). Although the cloud shield is disrupted for the short terrain simulation, key structural features are retained: an eye is visible at the midpoint hour (Fig. 3.3e) and a robust rainband persists through the end of the analysis period (Fig. 3.3f).

Despite divergent storm evolutions as indicated by the cloud shields, the maximum precipitation accumulated at any location for each experiment over the nine-hour analysis period is equivalent (Figs. 3.4a,b). The precipitation patterns simply exhibit different horizontal organizations for the two terrain heights. Precipitation from the tall plateau experiment covers a broad region and the influence of no single feature is visible. Meanwhile, precipitation from the short plateau experiment is concentrated within two bands of heavy rain that are located closer to the storm center. In contrast, other studies indicate that unlike the patterns presented in Figure 3.4, inner core precipitation increases when high terrain is present (Ramsay and Leslie 2008; Yang et al. 2008). But this precipitation variability is likely due to the lower terrain height compared to Mexico (800 m vs. 2500 m) in the case of Tropical Cyclone Larry, continued access of Typhoon
Figure 3.3. (a) Simulated outgoing longwave radiation for the tall plateau experiment 2 hours before midpoint time. (b) As in (a), but at the midpoint time. (c) As in (a), but 4 hours after the midpoint time. (d, e, f) As in (a, b, c), but for the short plateau experiment.
Figure 3.4. (a) Cumulative precipitation (shading) for the tall plateau experiment for the nine hours surrounding the hour when the storm crosses the midpoint of the terrain. The terrain height is indicated by contours. The storm track is shown in black and portion of the track corresponding to the nine hours is overlaid in yellow. (c) As in (a), but only for data that lie within 75 km of the storm center. (e) As in (a), but only for data that lie outside 75 km of the storm center. (b, d, f) As in (a, c, e), but for the short plateau experiment.
Nari to the warm, moisture-rich boundary layer air above the ocean that helped maintain a constant intensity, or the limitations of generalizing the results of a single simulation.

Segregating the precipitation by distance from the storm center reveals different contributions of precipitation within and beyond 75 km of the storm center for each experiment. Precipitation within the innermost core of the tall plateau storm contributes a smaller fraction to the total precipitation over the analysis period, both in terms of the maximum amount and the lack of precipitation further inland (Fig. 3.4c). Ramsay and Leslie (2008) found similar results in their orographic modification study of Tropical Cyclone Larry (2006): precipitation from the experiment with flat terrain was weaker during the initial landfall, but extended further inland than the precipitation from the tall plateau experiment due to the different rates of weakening. A larger fraction of the simulated precipitation exists beyond the 75-km radius. In contrast, precipitation in the short plateau experiment is more balanced between the two regions and the inner core precipitation is relatively constant during the nine-hour period (Fig. 3.4d). The simulated precipitation pattern for the tall plateau experiment is similar to the horizontal distribution of precipitation observed in Hurricane Karl shown in Figure 2.3, where a broad swath of precipitation extends far north of Karl’s center. Unfortunately, without additional cross sections of radar reflectivity or horizontal velocity data, it is unknown whether precipitation in Karl was due to widespread convection or an organized rainband.

Surface precipitation is insufficient to determine exactly how orographic modification of a tropical cyclone proceeds. The three-dimensional thermodynamic, kinematic, and microphysical structures are required to diagnose the type of precipitation. The next three sections examine horizontal maps and vertical cross sections of several variables to characterize how the nature of the precipitation evolves in each experiment. The specific times analyzed are two hours prior to
the midpoint, the midpoint time, and four hours after the midpoint time. Since this section relies heavily on the microphysical output from the model, it is important to recall that simulated hydrometeor categories can vary between microphysical schemes. The goals of this section are to investigate how structural changes manifest between the two experiments and to assess the realism of the simulated structures.

3.4.1 Two hours prior to the midpoint time

Noticeable structural differences between the two storms are present even as they approach the plateau (Fig. 3.5). The horizontal pattern of rain mixing ratios at 3 km altitude is already disrupted for the tall plateau storm (Fig. 3.5a). The simulated rain preferentially occurs in the immediate vicinity of the increasing terrain. What appears to be a deteriorating eyewall is noticeable on the inner edge of the circulation, but it merges with a broad region of rain further downstream on the southwest side of the storm. Two additional bands of enhanced rain begin upstream of the terrain to the northwest of the storm center. Huang et al. (2014) presented similar evidence of enhanced reflectivity upstream of terrain in an outer rainband, but the precipitation was concentrated in a single intense band embedded within a region of broad, weak precipitation. In this case, rain is not organized within a distinct rainband. The simulated storm in the short plateau experiment exhibits the typical structure of a mature tropical cyclone. Although the intense rain mixing ratios that define the eyewall are asymmetric, the eyewall is distinct from other precipitation features (Fig. 3.5b). Outside of the eyewall, rain is concentrated in a rainband on the north side of the storm. The different rates of decay are further reflected in the horizontal extent of the warm core. The warm core in the tall plateau storm is already shrinking, whereas the short plateau storm maintains a larger warm core (Figs. 3.5c,d). Additionally, the eyewall rain mixing ratios for the short plateau experiment coincide with the strong gradients of $\theta_e$, whereas the rain
Figure 3.5. Kinematic, thermodynamic, and microphysical fields two hours before the midpoint time. (a) Horizontal map of selected wind barbs in shading and rain mixing ratios at 3 km for the tall plateau experiment. Contours indicate the terrain height. (c) Horizontal map of equivalent potential temperature at 3 km for the tall plateau experiment in shading. Contours indicate the terrain height. (e) Cross section of vertical velocity along line 1-2 in (a, c) in shading. Blue contours indicate the graupel mixing ratio every 0.5 g kg\(^{-1}\) and black contours indicate the rain mixing ratio in every 0.5 g kg\(^{-1}\). Grey shading indicates data below the terrain. (g) Cross section of equivalent potential temperature along line 1-2 in (a, c) in shading. Black contours indicate the cloud water mixing ratio every 0.25 g kg\(^{-1}\). (b, d, f, h) As in (a, c, e, g), but for the short plateau experiment where line 3-4 in (b, d) shows the location of cross sections (f, h).
within the remnant eyewall in the tall plateau storm lies radially outward of the most intense $\theta_e$ gradient.

A vertical cross section through the remnant eyewall of the tall plateau experiment that displays the vertical velocity, rain, and graupel along line 1-2 is shown in Fig 3.5e. The vertical motion is disorganized, but a region of strong upward motion associated with what remains of the upper-level eyewall is located at the eastern edge of the plateau. Along the increasing terrain lies a broad region of rain mixing ratios greater than 0.5 g kg$^{-1}$. Directly above the rain lies a robust layer of graupel, where mixing ratios exceed 1 g kg$^{-1}$ over much of the cross section. These mixing ratio values are problematic since the peak mixing ratio ever observed in a tropical cyclone is only 0.2 g kg$^{-1}$ (McFarquhar et al. 2006). Additionally, the rain mixing ratios below the melting level is well-correlated with the graupel mixing ratios directly above. Although melting graupel is an important source of rain, unphysically large graupel concentrations have been previously linked to unrealistic precipitation amounts (McFarquhar et al. 2006). Yang et al. (2011a) finds similar vertical structures in their simulations of Nari (2001) where graupel mixing ratios likewise exceed 1 g kg$^{-1}$ over ocean and land, indicating that unphysical graupel concentrations are a systematic problem in numerical simulations of tropical cyclones. Furthermore, Yang et al. (2011a) note that cold rain processes are elevated above the terrain, thereby increasing the surface rainfall; since graupel exceeded observations in their study, the model may be overestimating the role of cold rain processes in the orographic modification of tropical cyclones.

Consistent with the lack of overlap between the eyewall rain mixing ratios and the strong horizontal gradients in $\theta_e$ in Figures 3.5a,b, the thermodynamic environment along line 1-2 indicates that a moist neutral assumption is not valid at low levels over the sloping terrain (Fig. 3.5g). The retreating warm core is restricted to upper levels between 50 and 70 km along the cross
section. Below 4 km altitude, positive $\theta_e$ lapse rates exist between 40 and 60 km due to a mid-level minimum of $\theta_e$ laying on top a thin layer of high-$\theta_e$ air. This thermodynamic environment would support buoyant motions, but the resulting instability would be weak. The simulated low-level instability is consistent with the dropsonde presented in Chapter 2, where weak low-level instability was observed north of Karl. Although the dropsonde sampled the environment outside of the eyewall, it provides evidence that weak instability is present in a landfalling tropical cyclone. Other studies such as Yu and Cheng (2008) and Misumi (1996) found that nearby thermodynamic profiles were moist neutral at low levels and stable further aloft. However, the inner cores of Typhoon Xangsane (2000) and Typhoon 9307 remained offshore, maintaining the moist neutral environment over the terrain. Throughout line 1-2, cloud water is directly tied to regions of upward motion. Three shallow regions of cloud water coincide with pockets of upward motion that exceed 2 m s$^{-1}$. Below the cloud water maxima lies enhanced rain indicating warm rain processes contribute to the growth of precipitation that originates aloft.

The eyewall of the short plateau experiment reveals different precipitation processes. The vertical cross section through line 3-4 reveals a broad and relatively uniform precipitation structure (Fig. 3.5f). Unlike the tall plateau storm, there are no small, near-surface pockets of elevated rain mixing ratios. But similar to the tall plateau experiment, a layer of graupel is located just above the rain. Although graupel mixing ratios are not as large as the tall plateau experiment, they still exceed observational values. Additionally, large graupel mixing ratios extend up to 12 km altitude. Even though Black et al. (2003) observed graupel at 12 km altitude in the convective eyewall of Hurricane Bonnie (1998), it is unclear how frequently high-altitude graupel occurs in tropical cyclones. Given the problematic graupel mixing ratios and location in both experiments, graupel formation mechanisms must be addressed in future model validation studies. The robust inner core
depicted along line 3-4 supports moist neutral processes more than the tall plateau experiment, as \( \theta_e \) values above 350 K air extend from the surface to the upper troposphere (Fig. 3.5h). In contrast to the frequent low-level occurrence in the tall plateau experiment, cloud water in the short plateau experiment covers a broad region of the deep eyewall updraft between 40 and 60 km. There is an additional narrow region over the weakly sloping terrain near 35 km. It is possible that the narrow region of cloud water is triggered by lifting over the weakly sloping terrain, but a weak banded structure is present between the eyewall and rainband in Figure 3.5d, which could be the source. Nevertheless, enhanced cloud water is not as frequent at low levels along the terrain.

3.4.2 The midpoint time

As each tropical cyclone reaches the midpoint, the manner in which the different rates of decay affect the storm structure for each experiment is ever more apparent. The tall plateau storm has lost all distinct precipitation features as the increasingly ragged remnant eyewall is barely distinguishable from the widespread convection located northwest of the storm center (Fig. 3.6a). The warm core is miniscule at 3 km altitude as it retreats upward (Fig. 3.6b). Widespread convection dominates the vertical cross section along line 1-2 where pockets of intense rain and graupel are embedded within a large region of hydrometeors (Fig. 3.6c). Intense upward motion extends up past 10 km altitude, beginning immediately as the terrain increases between 55 and 70 km. Weak low-level downdrafts are present further up the plateau, indicative of the convective nature of the of the banded precipitation (Fig. 3.6f). Upstream of the terrain between 60 and 75 km, \( \theta_e \) values exceeding 350 K reside in a 2-km deep layer. The horizontal wind vectors in Figure 3.6a indicate that the air mass would have recently passed over the ocean, supporting latent heat fluxes to the boundary layer. Positive lapse rates and the favorable orientation of the low-level
Figure 3.6. Kinematic, thermodynamic, and microphysical fields at the midpoint hour for the tall plateau experiment. (a) Horizontal map of selected wind barbs and rain mixing ratios at 3 km in shading. Contours indicate the terrain height. (b) Horizontal map of equivalent potential temperature at 3 km altitude in shading. Contours indicate the terrain height. (c) Cross section of vertical velocity along line 1-2 in (a, b) in shading. Blue contours indicate the graupel mixing ratio every 0.5 g kg\(^{-1}\) and black contours indicate the rain mixing ratio every 0.5 g kg\(^{-1}\). Grey shading indicates data below the terrain. (e) Cross section of equivalent potential temperature along line 1-2 in (a, b) in shading. Black contours indicate the cloud water mixing ratio in g kg\(^{-1}\). Grey shading indicates data below the terrain. (d, f) As in (c, e), but along line 3-4 in (a, b).
wind would support the release of buoyant motions. Reduced upper-level stability enables the updrafts to extend further aloft. Cloud water is less frequent near the surface, indicating that near-surface warm rain processes are not as dominant in this region.

Within the remnant eyewall along line 3-4, graupel and rain mixing ratio values have decreased, but hydrometeors still cover a broad region of the curved plateau edge (Fig. 3.6d). Vertical motion is weaker and the strong upper-level motion from the upper-eyewall in Figure 3.5e is no longer present. A shallow layer of weak to moderate updrafts still exists between 20 and 50 km. It is less clear whether the updrafts are due to forced ascent or shallow convection as the velocities are weaker. Collocated with the upward motion is a broad region of cloud water and the largest rain mixing ratios are located directly below the elevated cloud water concentrations. Graupel mixing ratios continue to exceed observations, despite the ragged appearance of the storm, and melting graupel dictates the location of the rain. The cross section indicates that the model still predicts contributions from warm and cold rain processes to rain production, although the cold rain contribution has decreased. The thermodynamic environment surrounding the remnant eyewall has no warm core and upper-level stability has decreased (Fig. 3.6f). But even though low-level instability is still present, it has weakened as the maximum near-surface $\theta_e$ value decreased by 2.5 K from the near-surface values shown in Figure 3.5g. Wind vectors plotted in Figure 3.6a indicate that the air may not have recently passed over the ocean, causing a reduction in latent heat fluxes. This reduction in instability is consistent with the weaker vertical velocities in this cross section.

Meanwhile, the short plateau storm has undergone modest changes over the two hours (Fig. 3.7). Although the storm remains asymmetric, the eyewall and rainband still stand out from the broad region of rain (Fig. 3.7a). The warm core at 3 km altitude remains robust and the eyewall continues to coincide with the area of strong $\theta_e$ gradients (Fig. 3.7b). In the rainband along line 1-
Figure 3.7. As in Figure 3.6, but for the short plateau experiment.
2, deep upward motion, excessive graupel mixing ratios, and intense rain are concentrated within the feature (Fig. 3.7c). The nearly constant $\theta_e$ vertical profile would not support the convective motions that characterized the tall plateau storm at larger radii. Additionally, larger upper-level stability exists in comparison with the tall plateau experiment; assuming a similar pattern of upper-level stability further north, this upper-level environment would restrict upward vertical motions north of the rainband. Cloud water is most frequent above 2 km altitude (Fig. 3.7e). The vertical cross section along line 3-4 through the eyewall strongly resembles the structure seen in Figure 3.5. Variations in the rain mixing ratio values along the cross section are gentle and there is strong vertical coherence between the mixing ratio values just above the surface and immediately below the melting layer (Fig. 3.7d). The warm core remains robust and the moist neutral assumption is still valid (Fig. 3.7f). Although elevated cloud water mixing ratios exist below 3 km altitude that could result from weak upslope flow, the elevated location and concentration within the eyewall updraft suggests low-level enhancement is minimal.

3.4.3 *Four hours after the midpoint time*

At the end of the analysis period, the precipitation structure of the tall plateau storm has completely deteriorated as what remains of the inner core progresses west. The inner core of the tall plateau storm has minimal rain water and the warm core has eroded (Figs. 3.8a,c). Disorganized convection is the only precipitation that remains over the sloping terrain and its organization resembles the widespread convection shown in Figure 3.6a. Along line 1-2, convective blobs are discrete, characterized by strong upward motion, artificially strong and deep graupel, and intense rain, located both upstream of and above the sloping terrain (Fig. 3.8e). Near-surface $\theta_e$ values are similar in magnitude to the low-level $\theta_e$ shown in Figure 3.6c and positive $\theta_e$ lapse rates would support convective motions due to reduced $\theta_e$ around 4 km altitude (Fig. 3.8g).
Figure 3.8. As in (Fig. 3.5), but 4 h after the midpoint hour.
Upper-level stability has weakened further, allowing convection to grow taller and extend above 10 km altitude. Although the short plateau storm is finally beginning to exhibit signs of decay, particularly in the eyewall, the warm core and rainband remain intact (Figs. 3.8b,d). The vertical structure of upward motion, rain, and graupel along line 3-4 is convective in nature like the precipitation along line 1-2, except that the precipitation entities are not discrete, instead blending together (Fig. 3.8f). The $\theta_e$ structure is noticeably different from the tall plateau storm, as the upper-level stability is stronger, thereby restricting the maximum height of the updrafts to 9 km altitude (Fig. 3.8h).

3.4.4 Type of vertical motions

Vertical cross sections through the tall plateau storm in the previous section suggested shallow convective motions contributed to low-level enhancement. In contrast, observational studies have previously inferred gentle forced ascent as the dominant mechanism due to local neutral to stable environments (Misumi 1996; Smith et al. 2009; Yu and Cheng 2008); however, the inner cores of the studied tropical cyclones remained offshore, likely enabling maintenance of a thermodynamic structure that would favor gentle ascent. To evaluate the type of vertical motion produced by the model, this section compares calculated forced ascent with the vertical velocity simulated by the model. Forced ascent is calculated with the following equation:

$$w = \vec{u} \cdot \nabla h$$  \hspace{1cm} (3.1)

where $u$ is the horizontal velocity vector and $h$ is the height of the terrain. The calculation is performed on the model levels that follow the terrain 0.5, 1, and 2 km above the ground. Horizontal maps of the simulated and forced ascent vertical velocities for each simulation are plotted for the midpoint time on the model level 2 km above the plateau in Figure 3.9. For emphasis, only the
Figure 3.9. (a) Horizontal map of vertical velocity from the tall plateau experiment on the model level that 2 km above the terrain surface in shading. Data is only plotted where the terrain gradient (m km$^{-1}$) is 0.5% of the plateau height (m) and white shading indicates no data. Wind barbs indicate the horizontal wind on the same model level. Circle indicates the 75-km radius. (b) As in (a), but the forced ascent calculated from the horizontal wind is in shading. (c, d) As in (a, b), but for the short plateau experiment.
vertical velocities over the sloping terrain are shown, where the sloping terrain is defined by the horizontal gradient in terrain height (m km$^{-1}$) exceeding 0.5% of the plateau height (m). Overall, both experiments exhibit substantial deviations from forced ascent. Inside the 75-km radius of the tall plateau storm, a vertical velocity couplet spans the storm center that is moderately correlated with the pattern of forced ascent, but the simulated vertical wind speeds exceed the predicted forced ascent (Fig. 3.9a). Otherwise, the outer region of the tall plateau experiment and both regions of the short plateau experiment do not experience forced ascent: the outer region of the tall plateau experiment is strongly convective, the azimuthal distribution of upward and downward motion in the inner 75 km of the short plateau storm is out of phase with the forced ascent, and the simulated vertical motions in the outer region of the short plateau storm are the opposite sign as the forced ascent (Figs. 3.9b,c,d). These structures need to be compared with vertical velocity observations to assess how realistic these patterns are, but forced ascent is not occurring 2 km above the terrain in these experiments.

3.4.5 Summary

The three-dimensional structures presented in this section provide compelling evidence that terrain height impacts the type of precipitation processes. Differing rates of storm decay reflect strong changes to the overall thermodynamic structure, which controls the type and organization of precipitation processes. But this case study consists of only two experiments, which limits the generality of the findings. Furthermore, differences in the upstream storm structure, precipitation organization, and environmental flow could partially determine how precipitation processes evolve throughout passage over the plateau. To ensure the structures from the two model runs are due to orographic modification processes and no other factors, a small ensemble is created to
examine the evolution of the kinematic, thermodynamic, and microphysical characteristics of each experiment. The ensemble configuration and findings are discussed in the next section.

3.5 **Statistical Analysis of the Evolving Kinematic, Thermodynamic, and Microphysical Fields**

3.5.1 *Ensemble configuration*

Two small ensembles were created using the same model configuration as the control simulations, but with the addition of stochastic perturbations. The small amplitude perturbations are produced by the stochastic kinetic-energy backscatter scheme (SKEBS: Berner et al. 2011) that is available in WRF. The perturbations are applied to the horizontal wind and temperature fields at the end of each time step to the largest domain. The perturbations are then interpolated down to each nest. As a result, the horizontal resolution of the perturbations is restricted to twice the horizontal resolution of the parent domain (i.e., 108 km). The maximum perturbation scale is 500 km, similar to the mesoscale predictability experiment of Judt et al. (2015). Finally, the perturbations have a westward tilt in the vertical structure. Nine ensemble members are generated for each experiment, resulting in ten simulations for each plateau height and twenty simulations total.

The storm track and intensity for each ensemble member is shown in Figure 3.10. Every ensemble member for the tall plateau experiment was run for the same 72 h period as the control simulation. Since the nine-hour period of analysis for the short plateau experiment included the final hour of the simulation, the short plateau ensemble members were run for an additional 24 hours to ensure each ensemble member encompassed the full period of analysis. The tracks through model hour 84 (out of 96) for each short plateau ensemble member are shown in Figure 3.10. Exactly like the control experiments, the storm tracks are systematically affected by the height of
Figure 3.10. (a) Simulated ensemble storm tracks for the short plateau (northern tracks) and tall plateau (southern tracks) experiments. The ensemble members are denoted by grey lines, the control members are denoted by blue lines, and the observed track of Karl is denoted by a thick black line. Filled circles indicate 0000 UTC times and the open diamonds indicate 0600, 1200, and 1800 UTC times. All tracks go through 0000 UTC 18 September, except the northern ensemble members, which go through 12000 UTC 18 September. The colors indicate the elevation of the idealized plateau in the tall plateau experiment on the first nested domain (the short plateau is the same shape, with a maximum elevation of 500 m). (b) Mean sea level pressure for the simulated storms for the tall plateau experiment and Karl. Thin grey lines denote the ensemble members, the thick blue line denotes the control member, and the thick black line indicates the ensemble mean intensity. Filled circles indicate the best track intensity from the NHC. (c) Same as (b), but for the short plateau experiment.
the plateau, consistent with Zehnder (1993). Slight differences in timing exist for each simulation, but track deviations within each experiment are minor when compared to the systematic track shift between the two experiments.

Each ensemble spans a wide range of storm intensities (Figs. 3.10b,c). Overall, the short plateau ensemble members are stronger, as evidenced by the peak ensemble mean intensity. The increased storm intensity is due to longer residence time over the warm sea surface of the Gulf of Mexico in the simulations and the storms that tracked westward the slowest were the storms that intensified the most. To ensure the different distributions of storm intensity do not affect the results, the statistics were compared with a subset of each ensemble (8 members each) that have similar ensemble mean intensities. Deviations from the figures presented in this chapter were minimal, demonstrating that the findings of this chapter are robust.

3.5.2 Methodology

To compare the structures between the different ensemble members, the midpoint time when a storm is midway through the sloping terrain is calculated for each ensemble member. Within the nine-hour period surrounding the midpoint time, distributions and area averages are calculated for all ensemble members at each hour with respect to the midpoint time. As in the previous section, the data are separated based on distance with respect to the 75-km radius, which serves as an approximation for the innermost core. For data beyond the 75-km radius, only data points that lie within 0.5° to the south and 2.0° to the north of the storm center are included. Figure 3.9 illustrates this range. Due to the moving nests, the latitude range occasionally covers a region beyond the domain boundaries as shown in Figure 3.9. The number of non-missing data points is far smaller than the total number and based on the subsampling procedure discussed in the ensemble set up section, the results are not sensitive to this issue. The same sloping terrain
definition as Section 4 is utilized where only the data points where the terrain gradient is 0.5% of the plateau height are analyzed. It is important to recall that there will be less overlap between the inner 75 km of each storm and the gradient in terrain for the hours as the storm approach and depart the sloping terrain. Due to a variety of translation speeds within each experiment, the statistics are only calculated when at least five ensemble members have more than 900 horizontal data points intersecting the region of sloping terrain. The statistics are calculated on all data points that adhere to these criteria at each hour relative to the midpoint time. The next four sections examine the time evolution of the kinematic (vertical velocity and deviation from forced ascent), thermodynamic (vertical $\theta_e$ profiles), surface rainfall, and microphysical (rain, cloud, and graupel) fields that determine precipitation in the model. Unless stated otherwise, each field was interpolated to constant altitude levels every 0.5 km and only data points located between 0.5 and 2.0 km above the surface were included.

3.5.3 Kinematics

The evolution of the ensemble mean and distribution of low-level vertical velocities is shown in Figure 3.11. Regardless of the tropical cyclone region, the ensemble mean vertical velocity for the tall plateau experiment either equals or exceeds its counterpart from the short plateau experiment at all hours relative to the midpoint time. However, the maximum difference between the ensemble mean vertical velocities is only 0.5 m s$^{-1}$. The similar ensemble mean vertical velocities mask contrasting distributions. The vertical velocity distribution for the tall plateau experiment is broader than the short plateau ensemble distribution at each time step, where strong upward and downward motion occur more frequently. Enhanced upward motion directly above the surface would support enhanced warm rain processes through greater cloud water production, as demonstrated by Figure 3.5. Additionally, the different rates of storm decay have a
Figure 3.11. (a) Time evolution of near-surface vertical velocity frequencies within the 75-km radius for all ensemble members in the tall plateau experiment. The hours are relative to the hour when the tall plateau storm is midway through the sloping terrain. The line denotes the average mixing ratio at each simulation hour. Portions of the figure that are colored black indicate where fewer than 5 ensemble members have 900 horizontal data points intersect with the sloping terrain. (b) As in (a), but for all members of the short plateau experiment. (c) As in (a), but for data points farther than 75 km from the center and between -0.5 and 2.0º latitude of the storm center. (d) As in (c), but for all members of the short ensemble experiment.
strong impact on the evolution of the vertical velocity distributions. Rapid weakening of the tall plateau storms limits the occurrence of intense vertical motions within the 75-km radius to the three hours before the midpoint time (Fig. 3.11a). After the storms pass the midpoint, the inner core vertical velocity distributions for the tall and short plateau experiments are similar (Figs. 3.11a, b). Beyond 75 km, the broad distribution of vertical velocities confirms that the widespread convective motions shown in Figure 3.6 are a robust feature for the tall plateau ensemble members. Meanwhile, the constant distribution of vertical velocity in both regions for the short plateau ensemble indicates that the minimal disruption of the eyewall and rainband shown in Figures 3.5, 3.7, and 3.8 is robust.

To thoroughly examine the role of forced ascent that was discussed in Section 4.5, it is necessary to perform a bulk comparison of the calculated forced ascent and simulated vertical velocity. This comparison was performed on the three model surfaces that are 0.5, 1, and 2 km above the terrain, again limiting the analysis to points below the melting level. The squared difference between the calculated forced ascent and the simulated vertical velocity is calculated at each point. Summing the squared differences and dividing by the total number of points calculates the bulk deviation of vertical velocities from the expected forced ascent on each model level. A value of zero would indicate a strong correspondence between the simulated vertical velocity and the forced ascent for all data points. For both experiments, the lowest model level at 0.5 km exhibits the closest adherence to the forced ascent, while high variance further aloft indicates that forced ascent is not valid further above the surface (Fig. 3.12). As the tall plateau ensemble members decay, the variance decreases suggesting that forced ascent may be more valid as the eyewall circulation decays and low-level instability weakens. The outer region dominates at later times as widespread convection is triggered by the upslope flow (Figs. 3.12a, c). Vertical velocities in the
Figure 3.12. Time evolution of the ensemble vertical velocity deviation with from the expected forced ascent on the model levels 0.5, 1.0, and 2.0 km above the sloping terrain. Time is relative to the hour when each ensemble member crosses the midpoint of the sloping terrain. (a) Variance for the tall plateau experiment for data points within 75 km of the storm center. (b) As in (a), but outside 75 km of the storm center and between -0.5 and 2.0° latitude of the storm center. (c, d) As in (a, b), but for the short plateau experiment.
short plateau ensemble in both regions consistently deviate from forced ascent, but the degree to which it deviates is reduced, due to the small fractional area covered by the robust eyewall and rainband (Figs. 3.12b,d).

3.5.4 Equivalent Potential Temperature

The time evolution of the area-averaged vertical $\theta_e$ profile is shown in Figure 3.13. It is necessary to point out that the vertical profile of $\theta_e$ in the inner 75 km will smear together the different stability profiles of the eye, eyewall, and region immediately outside the eyewall. Nonetheless, the mean profile is the simplest way to quickly compare the two environments. Throughout the analysis period, the inner core of the tall plateau ensemble is consistently less stable and characterized by lower $\theta_e$ values than the short plateau ensemble (Figs. 3.13a,b). The thin layer of high $\theta_e$ that was present in the tall plateau cross sections disappear through averaging since the terrain height ranges from 0 to 2.5 km altitude. Nonetheless, the decrease of $\theta_e$ below 5 km altitude indicates that slightly unstable motions would be favorable (Fig. 3.13a). The consistently high $\theta_e$ values for the short plateau ensemble indicates that the warm core is strong, ensuring continuation moist neutral processes in the eyewall. As in the velocity distributions, the different decay rates result in disparate evolutions of the stability profiles. The tall plateau ensemble mean $\theta_e$ decreases rapidly throughout the troposphere in Figure 3.13a as the storms weaken, consistent with the shrinking of the warm core shown in Figures 3.5 and 3.6 and the results of Yang et al. (2011b). In contrast, weakening of the warm core in the short plateau ensemble is gentle and the slight instability at -4 and 3 h is partially due to less of the warm core overlapping with the sloping terrain. At radii larger than 75 km, the upper-level stability is greater in the short plateau simulation, consistent with a weaker storm (Figs. 3.13c,d). Although the ensemble mean low-level instability for the short plateau experiment is greater, examination of Figs. 3.6 and 3.8
Figure 3.13. (a) Time evolution of the mean $\theta_e$ vertical profile within 75 km of the storm center for the tall plateau experiment. Portions of the figure that are colored black indicate where fewer than 5 ensemble members have 900 horizontal data points intersect with the sloping terrain. (b) As in (b), but for the short plateau experiment. (c) As in (a), but for data beyond 75 km of the storm center and between -0.5 and 2.0º latitude of the storm center. (d) As in (b), but for data beyond 75 km of the storm center and between -0.5 and 2.0º latitude of the storm center.
suggests this difference occurs because convection realizes the instability in the tall plateau experiment.

3.5.5 Precipitation

The evolution of the rainfall frequencies is displayed in Figure 3.14. Broadly speaking, the precipitation mimics the evolution of the low-level vertical velocity distributions shown in Figure 3.11, although precipitation is not always greater for the tall plateau experiment, especially within the inner 75 km. Once again, the different evolutions of precipitation frequencies reflect the distinct storm evolutions. Within the inner 75 km, a sharp decrease occurs in the ensemble mean and spread of precipitation values for the tall plateau ensemble before the midpoint time (Fig. 3.14a). In contrast, both regions of the short plateau ensemble exhibit relatively consistent distributions with time. A slight peak in the inner core ensemble mean and distribution width occurs at the midpoint time for the short plateau ensemble members, which likely occurs due to a maximum number of eyewall data points intersecting the sloping terrain at that time (Fig. 3.14b). In the outer region, the ensemble mean and spread for the tall plateau ensemble maximize at the midpoint time when the wind vectors are parallel to the gradient of the plateau height and wind speeds remain sufficiently strong to flow over the plateau. Overall, the ensemble frequencies demonstrate that although passage over the tall plateau initially enhances precipitation processes, deterioration of the storm structure discourages precipitation production near the center of the storm, despite a similar kinematic distribution as the short plateau ensemble as shown in Figure 3.11. Sustained upward motion and cloud water production over the terrain cannot overcome the reduced background precipitation simulated by the model, consistent with the finding that the background precipitation is a critical factor in determining the final precipitation intensity over terrain (Yu and Cheng 2008). Meanwhile, the increased occurrence of precipitation in the outer region broadens
Figure 3.14. As in Figure 3.11, but for surface precipitation.
the areal coverage of the rainfall. For the short plateau, the constant precipitation distributions reflect the slow evolution of the eyewall and rainband that concentrate precipitation closer to the storm center. It is important to recall that, as shown in Section 4, the simulated precipitation is highly sensitive to the microphysical fields, whose evolutions are discussed in the next section.

3.5.6 Microphysical fields

The evolving near-surface rain mixing ratios are displayed in Figure 3.15. Not surprisingly, the patterns shown in Figure 3.15 replicate the trends shown in Figure 3.14. The distinct storm evolutions affect the time evolution of rain mixing ratio frequencies, where the tall plateau ensemble experiences greater variations than the short plateau ensemble, particularly in the innermost region of analysis. Storm decay once again induces a decrease of the ensemble mean and distribution width for the inner core of the tall plateau ensemble at the end of the nine-hour period (Fig. 3.15a). There is less variation with time at larger radii, but a slight peak exists around the midpoint time when the wind direction and speed would both be most favorable to flow directly over the terrain (Fig. 3.15c). Less variation exists for rain mixing ratios in both regions of the short plateau ensemble, with only a slight peak in the ensemble mean value and distribution width during the first half of landfall in the inner core (Figs. 3.15b,d). Overall, the rain is highly correlated with the precipitation that reaches the surface.

Figure 3.16 shows the frequencies of cloud water mixing ratios. Although near-surface vertical motion, rain mixing ratios and precipitation initially exhibit a strong peak in the tall plateau ensemble, the cloud water frequencies exhibit little variation over the analysis period. In fact, the ensemble mean and distribution of the cloud mixing ratios for the inner core of the tall plateau experiment exhibit the opposite trend of the rain mixing ratios (Fig. 3.16a). Since the cross sections presented in Section 4 showed regions of enhanced rain were found below regions of enhanced
Figure 3.15. As in Figure 3.11, but for rain mixing ratios.
Figure 3.16. As in Figure 3.11, but for cloud water mixing ratios.
cloud water, cloud water production is likely balanced with collection by falling raindrops. It seems unlikely that cloud water production would remain constant in time, especially as the distribution of vertical velocities narrows (Fig. 3.11a). Beyond 75 km, large cloud water mixing ratios are consistently more frequent, though cloud water is generated by ubiquitous convection, not ascent restricted to low levels, as was seen in the cross section in the previous section (Fig. 3.16c). When compared with the tall plateau ensemble, the short plateau ensemble has a smaller ensemble mean cloud mixing ratio and a narrower distribution at all times and radii (Figs. 3.16b,d).

Finally, the graupel mixing ratio frequencies are shown in Figure 3.17. Since graupel does not exist within 2 km of the terrain surface, data for this hydrometeor category came from the altitude levels that are 5, 6, and 7 km above sea level. As demonstrated in the cross sections of Section 4 (Figs. 3.5—3.8), rain mixing ratios are well-correlated with graupel mixing ratios, indicating the importance and overestimation of melting graupel to rain production in the Goddard microphysical scheme. As a result, the evolution of graupel strongly resembles the evolution of the rain and precipitation categories in the narrowing distributions for the inner core of the tall plateau experiment, the constant distributions for the short plateau experiment in both regions, and a broader distribution at larger radii for the tall plateau experiment. The close correspondence of the low-level vertical motion and the graupel distributions in Figures 3.11 and 3.17 is puzzling since the two layers should not be related. Closer examination of the vertical velocity field in Figure 3.5e reveals that low-level upward motion is ubiquitous across the sloping terrain whereas the maximum upper-level motion occurs just at the initial increase in terrain height, east of the peak graupel mixing ratios. Although not shown in this dissertation, further examination of each tall plateau ensemble member reveals that strong upper-level upward motion and graupel are initially concentrated upstream and at the beginning of the plateau. In contrast, the short plateau
Figure 3.17. As in Figure 3.11, but for graupel mixing ratios.
ensemble members have regions of enhanced upward motion and graupel scattered around the storm with no systematic preference for an azimuthal sector. A comparison of the ensemble graupel mixing ratio frequencies for all data within the inner 75 km, not just data that intersect with the sloping terrain, indicates that graupel mixing ratio frequencies before landfall are similar, but the location of graupel differs between the experiments.

3.5.7 Discussion

Across all ensemble members, the model consistently simulates upper-level lifting and cold rain processes that are a strong factor in rain enhancement ahead of the tall plateau terrain that were not observed in Chapter 2. By itself, the lack of upper-level graupel observations does not mean that upper-level enhancement processes do not occur in tropical cyclones, but additional observations of microphysical processes aloft and validation studies to test the realism of this process are critical. Moreover, the excess graupel in tropical cyclone simulations suggests that the model overestimates the importance of cold rain processes in the orographic modification of tropical cyclones (McFarquhar et al. 2006; McFarquhar et al. 2012). Nonetheless, this potential source of model error does not invalidate the result that warm rain processes are enhanced when terrain is higher, since low-level cloud water production should occur independently of graupel production and collection of cloud water is a simpler microphysical process. In particular, the correspondence between larger rain mixing ratios in the previous section with shallow upward motion and heightened concentrations of cloud water in the tall plateau control simulation suggests that existing rain is being enhanced by ubiquitous cloud water. The issue going forward is determining the relative contributions of warm and cold rain processes in generating precipitation both over the ocean and near complex terrain.
3.6 CONCLUSIONS

It is well-known that a tropical cyclone moving over a continental barrier will decay due to disruption of the inner core and removal from its oceanic energy source. The rate of decay and change in storm structure is related to the three-dimensional shape of the topographic barrier. This chapter has investigated how the height of a large continental mountain range impacts the orographic modification of precipitation processes present in a landfalling tropical cyclone. This objective is accomplished through analysis of numerical simulations carried out with the Weather Research and Forecasting (WRF) model, in which the Mexican terrain is replaced with an idealized plateau that fits within the geographic boundaries of Mexico. Two experiments with plateau heights of 0.5 and 2.5 km have been analyzed, where the tall plateau most closely resembles the real topography of Mexico. Several cross sections through each simulated storm reveal the kinematic, thermodynamic, and microphysical structures as the idealized storms move over the sloping terrain. Two small ensembles have been used to generate statistics on the evolution of these same fields.

Throughout this chapter, unrealistic graupel concentrations are apparent and preclude full acceptance of the relative role of cold microphysical processes in the orographic modification of tropical cyclones. In both experiments, the rain is intimately related to the graupel aloft, whose simulated mixing ratios exceed observations (McFarquhar et al. 2006; McFarquhar et al. 2012). This result does not mean the model is not capturing different low-level processes in each experiment, but the relative contributions of warm and cold rain processes may not match reality. Additionally, the observations showed no evidence of increased upward motion and graupel aloft, although the decreased echo on the northern side of Karl could have rendered such motions invisible to the radar. It is also possible that increased upward motion aloft could have occurred
outside of the time the aircraft spent on station. Nonetheless, if the model cannot properly represent microphysical processes, errors will propagate through the microphysical scheme and affect both the simulated precipitation and latent heating patterns.

Despite concern about unrealistic graupel mixing ratios, consistencies between the tall plateau experiment and the observations presented in Chapter 2 were present. These consistencies include:

- In Chapter 2, thermodynamic data and radar reflectivity data suggested that shallow convection may accompany forced ascent in upslope flow due to weak instability and variable reflectivity intensity that was not correlated with the brightband intensity. The simulations show similar structures, where the simulated vertical velocities only match the forced ascent predictions in the lowest model layers; strong deviations from forced ascent exist above the lowest levels.

- Warm rain processes are contributors to orographic modification processes, although excess graupel concentrations may be overshadowing any near-surface processes in the numerical simulations. Further work is needed to test sensitivity to the choice of microphysical parameterization.

- In the higher-plateau experiment (more similar to reality), Karl’s precipitation extended far beyond the inner core where upslope flow was favorable. However, it remains unclear whether the precipitation on the outskirts of Karl’s circulation was due to a well-defined rainband or widespread convection.

Analysis of the two experiments reveals that several differences in the simulated structures arise due to the different plateau heights. These differences include:
Consistent with prior orographic modification experiments (Lin et al. 2001; Ramsey and Leslie 2008; Yang et al. 2008), the simulations show that a tropical cyclone passing over a tall continental barrier weakens more rapidly than a tropical cyclone passing over a barrier of lesser height.

The two plateau heights result in different horizontal distributions of precipitation. When the plateau height is lower, precipitation is concentrated within the eyewall and rainband, which remain intact for a longer period resulting in approximately constant precipitation during landfall. When the plateau is tall, the storm structure deteriorates and the inner core precipitation weakens as the storm marches westward. Farther from the storm core center, precipitation follows an opposite trend, maximizing when the storm is at its midpoint. Despite different horizontal patterns, the precipitation in the two experiments is similar. As mentioned previously, the precipitation is heavily dependent upon graupel in the Goddard scheme, so further work is needed to determine if this result is robust.

The differing rates of decay during landfall induce different changes to the vertical $\theta_e$ structure. When the plateau is tall, the upper-level warm core retreats upward quickly and upper-level stability is weaker over the full extent of the storm, consistent with Yang et al. (2011b). The changing stability favors different patterns of vertical motion between the experiments. In the short plateau experiment, moist neutral ascent in the eyewall and rainband is maintained for a longer period of time. In the tall plateau experiment, the innermost 75 km experiences forced lifting right above the surface and increased shallow convective motions, whereas widespread deep convection erupts farther from the center.

From a microphysical standpoint, the three-dimensional distribution of cloud liquid water exhibits substantial variation between the two experiments. When the plateau is tall, cloud
water is enhanced at low-levels along the sloping terrain; like the vertical velocity, the cloud mixing ratios are not uniform, suggesting a mix of forced ascent and shallow convection. Outside the inner 75 km, the enhanced cloud water is produced through convective processes, not forced ascent. In contrast, when the plateau is not as high, cloud water is more frequent at altitudes above 2 km throughout the storm.

- Graupel, rain, and surface precipitation peak in the inner core initially for the tall plateau experiment due to upward motion occurring more frequently over the terrain at upper levels, whereas these variables remain constant with time for the short plateau experiment. All three variables are highly correlated, indicating that melting graupel is a key mechanism of rain production in the Goddard microphysical parameterization. This model characteristic is a concern since excess graupel in tropical cyclones has been shown to produce artificially intense rainfall (McFarquhar et al. 2006; McFarquhar et al. 2012).

Overall, this chapter demonstrates that the vertical extent of a continental topographic barrier affects the rate of storm decay, the thermodynamic environment throughout the storm, the manner in which precipitation is created, and the horizontal organization of precipitation. Concerns about the microphysical scheme exist and accurately simulating the microphysical processes in tropical cyclones undergoing orographic modification is critical to improving precipitation forecasts. To address this problem, microphysical observations in landfalling tropical cyclones and model validation studies are critical.
Chapter 4. CONCLUSIONS

The airborne radar data and numerical simulations analyzed in this dissertation show how movement over a continental barrier affects the precipitation processes in a tropical cyclone. Due to the possibility for devastating damage to life, property, and agriculture, gaining a better understanding of the processes responsible for the location and intensity of rainfall is highly motivated. As numerical models are capable of resolving increasingly smaller scales, understanding and representing precipitation processes near complex terrain is becoming ever more important. Prior studies have made strides in inferring the specific orographic modification mechanisms that occur during landfall over a mountainous region, but have generally lacked the observational tools necessary to adequately confirm whether specific processes such as the warm rain (drop coalescence) mechanism. Overall, this study uses the vertical structure of radar reflectivity (i.e. the vertical distribution of precipitating hydrometeor concentration) to expand upon the information revealed by the horizontal mapping of precipitation and radar reflectivity used in previous studies. Both observations and numerical simulations have been employed.

High-resolution radar reflectivity data from Hurricane Karl (2010), which made landfall over a tall mountain range and plateau in eastern Mexico, allowed for a novel examination of precipitation processes in a tropical cyclone undergoing orographic modification. The data have revealed dissimilar vertical structures of precipitation processes in regions of upslope and downslope flow and indicated that inner-core reflectivity enhancement in upslope flow was confined to low-levels. In contrast to the stable ascent often invoked as an orographic modification mechanism, thermodynamic data revealed weak instability near the inner core and reflectivity was organized into discrete regions of enhanced reflectivity that were independent of the bright band
strength. Thus, shallow convection embedded in the upslope flow is hypothesized to have played a role as air was forced upslope. The low stability precluded blocking, and the upslope flow produced cloud liquid water, which was collected by precipitation particles generated in the convective elements. Unfortunately, vertical velocity measurements were subject to too much error to confirm this hypothesis. In the future, increased spatial and temporal sampling of the thermodynamic environment and air motions are critical to make further progress on the evolution and horizontal distribution of precipitation processes.

Despite being an intense tropical cyclone moving over steep terrain, rainfall accumulation during Karl was modest compared to other instances of orographic modification in places such as Taiwan and Dominica, where the maximum 24-h precipitation can often exceed 500 mm and occasionally exceed 1000 mm (Huang et al. 2014; Smith et al. 2009; Yang et al. 2008; Yu and Cheng 2013). Satellite data helped place the reflectivity structures captured by the airborne radar into the context of Karl’s life cycle; mere hours after the DC-8 departed, Karl’s satellite presentation deteriorated entirely at which point a large convective cloud developed in its wake. This convection produced high rain rates over a small region near Córdoba, which constituted a large fraction of rain measured locally. To my knowledge, convection in the wake of a decayed tropical cyclone has not been previously documented. Deep convection developed in the eye of Hurricane Georges (1998), but had very recently exhibited a classic tropical cyclone satellite presentation (Geerts et al. 2000). In summary, rainfall and airborne radar data suggest that Karl’s landfall over steep terrain limited the accumulation of intense precipitation, although this study does not consider the impact of storm translation speed on precipitation totals. Nevertheless, Chapter 2 provided observational support for the warm-rain enhancement hypothesis as long as the inner core structure is robust, but highlighted alternative mechanisms through which
precipitation processes are modified by passage over terrain in a deteriorating storm. Precipitation processes are not static and storm decay during landfall can dramatically alter the thermodynamic environment to support a wide variety of processes.

Analysis of numerical simulations demonstrated that the height of the continental mountain range in eastern Mexico affected the processes controlling precipitation formation. The height of the terrain systematically affected the rate of decay; storms moving over a tall plateau fill and weaken more quickly than storms that encounter a lower-elevation plateau. In the tall plateau experiment, the warm core shrank horizontally while retreating upward and the upper-level $\theta_e$ values decreased near the inner core and at larger radii, consistent with the results of Yang et al. (2011b). As the moist neutral environment in the eyewall eroded and upper-level stability weakened, the nature of the precipitation changed. In the remnant eyewall, a mix of forced ascent and shallow buoyant motions were present, which is consistent with the observational results in Chapter 2. At larger radii, widespread convection was triggered by lifting ahead of and over the terrain that released buoyant motions tapping into the high-$\theta_e$ boundary layer. Similar to the precipitation observed in Karl, the precipitation for the tall plateau experiment during landfall weakened closer to the storm center, but widespread convection at larger radii broadened the area affected by rainfall; overall, total precipitation near the center and at larger radii was comparable. In contrast, the slow weakening rate associated with the short plateau simulation allowed the tropical cyclone to retain the eyewall and rainband features for a longer period of time. Precipitation processes progressed independent of the terrain, concentrated within the eyewall and rainband and constant in time. The two simulations suggest that the maximum precipitation for each experiment was similar in amount but was organized differently and the orographic mechanisms differed.
Consistent with the observational results, model cross-sections of microphysical quantities revealed cloud liquid water was collocated with regions of upward motion. Rain mixing ratios just below regions of elevated cloud water were greater, indicating that the model washout processes were collecting this cloud water. However, the overall magnitudes of the rain mixing ratios and surface precipitation were strongly controlled by the graupel mixing ratio aloft. Previous microphysical validation studies have demonstrated that microphysical schemes of varying complexity consistently overestimated graupel in tropical cyclones (McFarquhar et al. 2006; McFarquhar et al. 2012).

In the tall plateau experiment, graupel was concentrated over the terrain initially; this feature was not present in the observations. It is possible cold rain processes were present prior to the aircraft sampling Karl or that upward motion regions aloft were not apparent in the radar data due to a lack of hydrometeors, but model error cannot be discounted. Since the precipitation pattern for the tall plateau experiment and Karl observations were similar in spite of graupel errors, it is likely that the model obtained a realistic result through incorrect microphysical processes. In summary, possible graupel overestimation is a serious issue for future research and operational forecasts alike.

It is clear that results from the numerical simulations are subject to the strengths and weaknesses of the selected microphysical parameterization. Since low-level enhancement processes in the tall plateau experiment are broadly consistent with the observations collected in Karl, the differences in thermodynamic structures and low-level cloud and rain patterns seem well-captured by the model. The challenge is determining how realistic the relative contributions of warm and cold rain processes are, which is limited by incomplete nature of aircraft microphysical and thermodynamic observations with which to compare these structures. Obtaining high-quality
microphysical and kinematic data is critical for further testing of the hypotheses that warm rain processes dominate and shallow convection occurs either in place of or alongside forced ascent, and also assessing the accuracy of simulated microphysical processes near complex terrain. Dual-polarization radar has been successful in identifying microphysical processes in tropical mesoscale convective systems (Barnes and Houze 2017) and assessing microphysical scheme performance in tropical cyclones over the ocean (Brown and Bell 2016); dual-polarization radar would be an asset in tropical cyclones making landfall over complex terrain. Targeted field research utilizing ground-based and airborne dual-polarization radars in combination with in-situ airborne and surface microphysical observations would be a critical next step.

In summary, this dissertation presents new results on the precipitation mechanisms in a tropical cyclone making landfall over a mountainous region and on how deterioration of the tropical cyclone affects the manner in which the precipitation processes proceed. This research is far from complete and numerous questions remain regarding how orographic modification in a tropical cyclone unfolds. Future research should consider the role of vertical wind shear, initial storm intensity, storm translation speed, and distance of the terrain from the coast in affecting precipitation processes. Additionally, this study has demonstrated the need for accurate, long-lasting, and widespread observations of hydrometeors to further test these hypotheses. Finally, this study examined the processes within a single storm over a short period of time; additional case studies and statistical analyses are required to generalize the processes described in this dissertation.
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VITA

Jennifer C. DeHart was born and raised in Ann Arbor, Michigan. As a young girl, she was interested in many facets of math and science, but her initial love of weather began as she was glued to the Weather Channel for a full week to see Hurricane Andrew make landfall in Florida and reintensify over the Gulf of Mexico.

Jennifer attended the University of Michigan where she received her B.S.E from the Atmospheric, Oceanic, and Space Spaces department with a concentration in meteorology. While she was a student, she did research with Professor Allison Steiner. She had the opportunity to participate in three field experiments: a supercellular tornado campaign with faculty and graduate students from Texas Tech University, an atmospheric chemistry campaign through the NASA Student Airborne Research Program where she was mentored by Professor Don Blake of UC Irvine, and VORTEX2 where she rejoined researchers from Texas Tech University. These campaigns instilled her with a love of observational data and field work.

Jennifer joined the group led by Professor Robert A. Houze Jr. at the University of Washington in the fall of 2010. Her research throughout her studies focused on the influence of external factors on the distribution and nature of precipitation in tropical cyclones due to environmental wind shear and complex terrain. Jennifer spent three summers on Chincoteague island as a student forecaster for the NASA HS3 campaign and served as a lead forecaster for the NASA OLYMPEX campaign. She received her Masters of Science in 2013 and graduated from the University of Washington with her Doctor of Philosophy in 2017.