Orographic Modification of Precipitation Processes in Hurricane Karl (2010)

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ABSTRACT

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 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 a strong reduction in the upper-level cloud coverage; 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.

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 Hemisphere 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 1993). 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 caused the convection to dissipate.

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 12 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). In the case of shallow terrain, if the hydrometeor advection time scale is greater than the particle fallout time scale 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, which 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 wind speed and precipitation. 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 topography 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 terrain 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 h (Geerts et al. 2000; Yu and Cheng 2013, 2014; Yu and Tsai 2017), Karl presents an opportunity to examine the orographic modification processes in a tropical cyclone completely deteriorating during landfall. In this context, our study assesses the hypothesis that warm-rain processes dominate the orographic precipitation enhancement process. 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. 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 multifaceted 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 with an along-beam resolution of 30 m. Although the radar scans a narrow cross-track swath of 24 beams ± 25° from nadir, we restrict our study to the beam closest to vertical. The 1.8 s of scan time comprises 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). For the high frequencies at which APR-2 operates (35.6 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 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 Hitschfeld 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 (Hitschfeld 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

<table>
<thead>
<tr>
<th>Frequency</th>
<th>13.4 GHz (Ku-band)</th>
<th>35.6 GHz (Ka-band)</th>
</tr>
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<tr>
<td>Antenna effective diameter</td>
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<td>1.4 m</td>
</tr>
<tr>
<td>Antenna gain</td>
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<td>33 dBi</td>
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<td>Antenna sidelobe</td>
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<td>−30 dB</td>
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<tr>
<td>Antenna beamwidth</td>
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<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⁻¹</td>
<td>1 m s⁻¹</td>
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nearby nonraining 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).

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-h, 3-h, and 10-min measurements. The 24-h measurements were the most numerous, with far fewer stations reporting at higher frequencies.

3. Hurricane Karl background

a. 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. 1). Karl weakened only slightly and reemerged in the Bay of Campeche as a weak tropical storm (maximum wind 20 m s⁻¹) early on 16 September. Between 0000 UTC 16 September and 1200 UTC 17 September Karl rapidly intensified to an intense category 3 hurricane (maximum
wind of $57 \text{ m s}^{-1}$) during passage over the warm sea surface (between $28^\circ$ and $30^\circ C$, not shown) in the southern Gulf of Mexico (Fig. 1). Infrared satellite imagery from 0615 UTC 17 September in Fig. 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 1993). The center of Karl made its second landfall at 1645 UTC 17 September, 16 km northwest of the city of Veracruz, Mexico, while the storm was at an intensity of $50 \text{ m s}^{-1}$.

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 (Figs. 2b,c). By 0000 UTC 18 September, Karl had weakened to tropical storm intensity with few deep clouds remaining (Fig. 2d). Despite continued deterioration of Karl, convection occurred intermittently, and ultimately a large cloud shield developed (Figs. 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 toward the ocean in central Veracruz (Fig. 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-h precipitation ending at 1300 UTC 18 September was greatest in two regions (Fig. 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.

b. 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 three locations within Karl’s inner core (Fig. 4): Jalapa, Orizaba, and Córdoba, in Veracruz, Mexico. The National Climatic Data Center (NCDC) provided 3-hourly data for Jalapa and Orizaba, although these stations did not report data at 0000 UTC 18 September. We subtracted the sum of the 3-hourly measurements available from the 24-h totals to estimate the value for the missing 3-h interval. The 3-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, 92, and 261 mm of rain, respectively (Fig. 4a). Despite widespread upper-level clouds over Veracruz before Karl’s landfall (Fig. 2a), precipitation was minimal prior to 1800 UTC 18 September. Without radar data, we cannot determine if clouds were present only at upper levels or if downslope northwesterly flow was suppressing rainfall. 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 3-h 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. 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 northeast–southwest while the mountains west of Orizaba and Córdoba are oriented northwest–southeast. Thus, the circulation surrounding Karl provided different flow directions with
rect 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. 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.

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 Fig. 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 ($\sim -1 \times 10^{-5}$ s$^{-2}$), indicating weak conditional instability. Between 1.8 and 2.5 km, values of $N_m$ are $\sim 6 \times 10^{-3}$, $9.5 \times 10^{-3}$ s$^{-1}$ and the wind speed is $\sim 22$–25 m s$^{-1}$; assuming a mountain height of 2.5 km, the Froude number ($Fr = U/N_m H$; $U$ is the wind speed and $H$ is the mountain height) between 1.8 and 2.5 km is $\sim 1$–2. These calculations suggest that any low-level buoyant motions would have been modest and the flow below 2.5 km 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 topography 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 south-eastern flank of the high terrain that juts northeastward toward the coastline.

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. 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 and 1906 UTC 17 September are shown in Figs. 6b and 6c, 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 (Figs. 6b,c). Given the broad nature of these differences and extension of the shallow echo over the ocean (e.g., 1845–1846 UTC), we suspect that 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 Fig. 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 reflectivity toward the surface, at times increasing by 10 dB from 4.5-km altitude to the surface (Fig. 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 2017). In their study of Typhoon Xangsane (2000), Yu and Cheng (2008) demonstrated that reflectivity intensity over terrain increases with wind speed (cf. their Figs. 11–13). In Karl, upstream wind speeds 1 km above sea level were between 20 and 25 m s$^{-1}$ (Fig. 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 of 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 not always connected to the 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 nonuniformity 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. 6c). Instead, the precipitation was organized into fallstreaks connected to and emanating from the bright band. The intensity of the fallstreaks remained approximately constant toward 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 Córdoba beginning at 2049 UTC 17 September and again encountered deep stratiform rain (Fig. 6d). Consistent with the earlier leg, sloping fallstreaks were present. The nearly constant reflectivity intensity with height and frequent connection with the bright band in the downslope legs suggests that precipitation originating aloft was not noticeably enhanced by collection of cloud water or generation of tiny raindrops.

Midway through the flight, the DC-8 completed a long straight leg from the northwest to the southeast (Fig. 7).
The APR-2 captured a wide range of precipitation types across the storm, culminating in deep, strong stratiform rain that causes substantial attenuation of the Ku-band data between 1952 and 1953 UTC on the southeast side of Karl (Fig. 7b). Low-level enhancement similar to Fig. 6b exists between 1943 and 1948 UTC. North of the highest terrain (e.g., 1944–1946 UTC), low-level enhancement is concentrated in two small regions below a strong bright band. This connection with the bright band is muted south of the high terrain (e.g., 1946–1948 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 the 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 18 September (Fig. 2e); however, 5 h passed between Fig. 7b and Fig. 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 and 2100 UTC 17 September indicated in Fig. 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 3-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).

To confirm that the low-level structure is different between the legs in Figs. 6 and 7, we isolated portions of the flight track where we would expect upslope and downslope (or flat) flow (Fig. 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 Fig. 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 km$^{-1}$, stretching from 1 to $-2$ dB km$^{-1}$. In contrast, the middle upslope beams have reflectivity slopes between $-1$ and $-3$ dB km$^{-1}$, with a median slope of $-2$ dB km$^{-1}$. This difference is consistent with the picture given by the sample cross sections shown in Figs. 6b and 6c, showing that this structure is consistent throughout the flight.

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.

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 3 hours as the aircraft flew over land (Fig. 1a). As shown in Fig. 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 3 hours following the flight and in Córdoba between 0300 and 0600 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 westsouthwestward, 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 Fig. 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 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 Cordoba (Figs. 2d–f). More formally, Fig. 9 shows how the distribution of IR brightness temperatures surrounding Karl changed through its life cycle. 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 1.5° radius of the calculated center. Although the coldest cloud tops disappeared as Karl decayed through 0000 UTC 18 September, there was a resurgence
of the coldest temperatures after 0200 UTC 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 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-θ_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 the top of 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 (Figs. 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 the warm ocean, whereas Karl was completely over land with no chance to maintain tropical cyclone structure. Whenever mechanism occurred, low-level high-θ_e air remained in the area to feed convection. A coastal sounding from Veracruz at 0000 UTC 18 September shows instability due to a warm, moisture-rich layer of easterly flow at low levels (Fig. 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 as follows: 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.

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 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 neither 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, 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 topography as well as the specific region of the storm encountering the terrain. Despite rapid deterioration of 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 under way 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.

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FIG. 10. Sounding launched from Veracruz, Mexico, at 0000 UTC 18 Sep.
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