Large-scale influences on the pre-genesis of tropical cyclone Karl (2010)

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LARGE-SCALE INFLUENCES ON THE PRE-GENESIS OF TROPICAL CYCLONE KARL (2010)

by

Kyle S. Griffin

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ABSTRACT

The genesis of Tropical Cyclone (TC) Karl (2010) in September 2010 provided a unique opportunity to examine the continuing problem of understanding tropical cyclogenesis. The precursor disturbance to Karl originated from a cluster of showers east of the Windward Islands and was well sampled by ongoing field campaigns, particularly the PRE-Depression Investigation of Cloud-systems in the Tropics (PREDICT), as the targeted disturbance progressed westward. While traditional genesis theories focusing on moisture and mass fields (e.g. top-down showerhead method) can explain the initial spin-up of the disturbance several days prior to its official genesis, additional perspectives are examined in concert with more traditional methods in order to provide a more complete analysis of the synoptic-scale patterns that influenced the pre-Karl disturbance.

A surge of westerly winds from northern South America aids the initial spin-up of the pre-Karl disturbance on 8-9 September, leading to the formation of a nearly closed earth-relative circulation. It can be shown that these anomalous westerly winds are tied to the convectively active phase of a convectively coupled Kelvin wave (CCKW). The observed formation of the nearly closed circulation on 10 September is well timed with the passage of this convectively active phase, a relationship that has been shown to hold true in cases of CCKW-TC interactions around the globe. Physically, the CCKW increases deep convection and aids in the generation of low-level relative vorticity on the cyclonic shear side of the low-level westerly wind anomalies, both of which serve to help organize the pre-Karl disturbance.

Finally, the passage of the CCKW coincides with an equatorward surge of cold air and southerly winds in the lee of the Andes, triggered by a passing mid-latitude
disturbance on 31 August. As the surge passes the equator on 7 September, little
temperature perturbation remains with the surge, but terrain-channeled low-level flow
acts to turn southerly flow into westerly flow south of the pre-Karl disturbance. By 8-9
September, anomalous westerly winds with the surge merge with and enhance the
anomalous westerly winds associated with the CCKW passage, strengthening the low-
level vorticity generation in the cyclonic shear of the merged CCKW-wind surge on the
pre-Karl disturbance. However, despite this increase in vorticity and convection, the
environment surrounding the pre-Karl disturbance remained unfavorable for several more
days, with increased vertical wind shear and the convectively suppressed phase of a
CCKW inhibiting further development before TC genesis occurred on 14 September.
Despite this delayed development, the interplay of both mid-latitude and equatorial wave
precursors likely contributed to the eventual genesis of TC Karl.
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1. Introduction

a. Motivation

The relatively rare phenomenon of tropical cyclogenesis remains poorly understood and difficult to forecast, as repeatedly emphasized by Gray (1998) in his summary work on the topic. Gray also stated that “it seems unlikely that the formation of tropical cyclones (TCs) will be adequately understood until we more thoroughly document the physical differences between those systems which develop into TCs from prominent tropical disturbances which have a favorable environment, look very much like they will develop, but still do not.” This call for additional observations of tropical disturbances prior to the time of genesis was answered, in part, by the PRE-Depression Investigation of Cloud-systems in the Tropics (PREDICT) field campaign (Montgomery et al. 2012, Evans et al. 2012). In the field campaign stage of PREDICT, which lasted from 15 August – 30 September and was based from St. Croix in the U.S. Virgin Islands, a number of both developing and non-developing tropical disturbances were sampled by aircraft in order to attempt to better understand the dynamics and thermodynamics of these disturbances.

The operational forecasting necessitated by the field campaign also allowed researchers (who doubled as forecasters) to dedicate more of their time than usual to real-time observational studies of these disturbances. As a result, researchers were able to gain a much better grasp on the interesting or unusual aspects of the evolution of a system before beginning the post-analysis of the tropical disturbances. The disturbance that became TC Karl was one of these closely observed disturbances, and real-time
forecasting notes from the researchers opened the doors to some additional research avenues. Although TC Karl itself did not undergo genesis until 14 September, this thesis will explore the formation of the precursor tropical disturbance during the period 8-12 September.

While the understanding of TC genesis is one well-known shortcoming in the field of tropical meteorology, the observation and understanding of equatorial wave modes is of increasing importance to the field of tropical meteorology. Recent studies have shown that the Madden-Julian Oscillation (MJO; Zhang et al. 2005) and various types of convectively coupled equatorial waves (CCEWs; Kiladis et al. 2009) modulate TC activity through dynamic and thermodynamic processes (e.g. Camargo et al. 2009, Bessafi and Wheeler 2006, Schreck and Molinari 2009). These CCEWs include convectively coupled Kelvin waves (CCKWs) and equatorial Rossby (ER) waves, as well as others discussed in Wheeler and Kiladis (1999). Focus is placed on the literature discussing CCKWs and their interactions with TCs, as this is one of the larger and better established bodies of work and proves to be the most relevant to the study of the genesis of TC Karl.

The goal of this thesis is to provide a detailed analysis of the environment surrounding the tropical disturbance that became TC Karl (hereafter, pre-Karl disturbance) and attempt to understand how the dominant large-scale features interacted to allow the formation of this disturbance. Emphasis will be placed on: (1) identifying the role of mid-latitude interactions with the formative tropical disturbance, (2) identifying interactions, if any, between CCEWs and the tropical disturbance, and (3) analyzing the possible interactions between CCEWs and mid-latitude features.
The thesis is organized as follows: the remainder of Chapter 1 contains a literature review of relevant topics. Chapter 2 highlights some of the techniques and data sources used in this thesis. Chapter 3 contains a discussion of synoptic conditions over the Atlantic Ocean during the genesis of the pre-Karl disturbance. Chapter 4 discusses the evolution of a CCKW, its interactions with the pre-Karl, and its interactions with the mid-latitudes. Finally, Chapter 5 will contain a discussion of the interactions between these features and a discussion of the key topics discussed previously.

\textit{b. Tropical cyclogenesis theories}

Over the past fifty years, a number of connected theories have been presented in an attempt to explain the physical processes that surround TC genesis. One of the original theories was termed Conditional Instability of the Second Kind (CISK), as presented by Ooyama (1964) and popularized by Charney and Eliassen (1964). Ooyama (1964) focused on the hypothesized cooperation between an initial vortex and a developing field of cumulus clouds to release latent heat and further saturate the boundary layer, allowing for continued frictionally induced growth of the vortex and the associated ascent at the top of the planetary boundary layer (PBL; Ekman pumping). Charney and Eliassen (1964) stressed the importance of low-level moisture convergence (as implied by Ooyama 1964) in enhancing convection with the developing vortex. This moisture convergence occurs with the assumption of an ever-present “source of near-saturated air in the boundary layer,” although Charney and Eliassen (1964) also assumed heat fluxes from the ocean surface to the boundary layer air to be negligible – a major downfall in light of more recent TC genesis theories. Interestingly, Smith (1997) pointed out that Ooyama (1964) did not ignore surface heat fluxes in his original theory of CISK, as these
fluxes were part of the driving mechanism behind intensification. Combined with other alterations and assumptions Charney and Eliassen (1964) made in their version of CISK, Ooyama (1982) criticized the way in which CISK has been popularized and highlighted that many objections to the theory were not present or would not be valid in the original form of the theory. Smith (1997), while highlighting some criticism noted above, also highlights that Ooyama (1964) was not originally a theory for the growth of disturbances, despite the mathematical work of Charney and Eliassen (1964) presenting it as such. As such, it remains useful alongside more recent theories that include surface heat fluxes.

Most notably, the Wind-Induced Surface Heat Exchange (WISHE) paradigm proposes a method for the intensification of TCs (Emanuel 1986). WISHE focuses on surface heat and moisture fluxes (also called enthalpy fluxes by Emanuel 1986 and Bell et al. 2012) that occur in response to the “disequilibrium,” or subsaturation, between the boundary layer and ocean surface. Energy, in the form of moisture, is transferred into the boundary layer and increases low-level equivalent potential temperature as wind blows over the tropical ocean towards the center of a TC. This energy is realized when the air parcels approach the center of the TC, are lifted, and release latent heat, resulting in a process that Emanuel (1986) envisioned as a Carnot heat engine (Fig. 1.1). The key to WISHE comes with the positive feedback loop created by the conversion of boundary layer moisture into latent heat. The heat release results in lower surface pressures in the disturbance, increasing the pressure gradient and the strength of the low-level convergence, and further increasing low-level winds and intensifying the WISHE process. Smith (1997) highlighted that both WISHE from Emanuel (1986) and CISK from Ooyama (1964) are primarily driven by surface fluxes of
moisture from the tropical ocean leading to the release of latent heat via convection. However, WISHE is considered a TC intensification theory while CISK is a TC genesis theory. The key difference between WISHE and CISK originates from the role of cumulus convection – in WISHE it is deemphasized, while in CISK it is an important part of the feedback process. This difference suggests that the development of deep cumulus convection is important to the subsequent development of a low-level circulation. However, neither of these theories accounts for the initial formation of the low-level circulation, an issue that was not addressed until more recent TC genesis theories.

One of the first theories attempting to explain the development of a circulation is generally referred to as the top-down “showerhead” theory, proposed by Bister and Emanuel (1997). Stemming from the intensive observations of Hurricane Guillermo (1991) in the eastern Pacific Ocean during the Tropical Experiment in Mexico (TEXMEX), top-down theory suggested that an intense convective system resulted in the formation of a mid-level vortex, similar to a mesoscale convective vortex. With strong convection in and around this mid-level vortex, stratiform rain aloft would result in the cooling and moistening of the low-level environment and would advect mid-level vorticity into the lower levels (hence “showerhead”, conceptualized in Fig. 1.2). This was theorized to aid in spinning up a low-level circulation that eventually moistens through convective mixing and the development of increasingly deep convection. In the real atmosphere, this vertical advection of vorticity is of a negligible order of magnitude on the spatio-temporal scales of TC genesis. Ritchie and Holland (1997) presented an alternate top-down method, focusing on the merger of vortices as observed in Typhoon Irving (1992) during the Tropical Cyclone Motion experiment (TCM-92). Ritchie and
Holland (1997) proposed that several smaller-scale mid-level vortices are generated in the smorgasbord of deep convection and eventually amalgamate into a single larger vortex (Fig. 1.3). The potential vorticity (PV) with each vortex also amalgamates and allows for an increasingly greater penetration depth until eventually the circulation reaches the surface. This top-down “vortex merger” theory remains credible in the field and shares some features with still newer theories of TC genesis.

More recently, bottom-up theories of TC genesis have been proposed by various authors (e.g. Montgomery and Enagonio 1998, Hendricks et al. 2004), with the vortical hot tower (VHT) theory as first presented by Hendricks et al. (2004) becoming the most popular. VHTs are defined as localized regions of intense convection, often responsible for latent heat release and the creation of vorticity. Hendricks et al. (2004) observed VHTs in TC Gustav (2002) in proximity to mesoscale vortices visible in satellite imagery, suggesting these VHTs create small-scale vortices that can grow upscale via a process similar to the vortex merger process proposed by Ritchie and Holland (1997). Hendricks et al. (2004) also presented other observational cases where pairs or groups of vorticity maxima are observed to exist before merging with each other and resulting in more intense vorticity maxima on time scales that remain relevant to TC genesis (e.g. Davis and Bosart 2001, see Fig. 1.4). Montgomery et al. (2006) present VHT theory via a more numerical perspective, and showed that an individual VHT could develop a dipole of cyclonic and anticyclonic vorticity in as little as 20 minutes. In turn, this promotes continued deep convective growth and the generation of low-level PV, leading to further development of low-level vortices and subsequent vortex mergers as proposed in previous theories.
An extension of VHT theory is found in Dunkerton et al. (2009), where the “marsupial” paradigm is first presented. The name “marsupial” stems from its focus on the closed “pouch” that often presents itself with tropical disturbances examined from the Lagrangian perspective. This pouch roughly represents a region of recirculating air in the horizontal (Fig. 1.5), effectively protecting air within the recirculation region from the potentially less favorable environmental conditions. At the same time, Dunkerton et al. (2009) suggested air parcels within the recirculation region slowly moisten, allowing deep convection to develop and promoting the aggregation of low-level PV, akin to the vortex merger and VHT theories presented by Ritchie and Holland (1997) and Hendricks et al. (2004). The marsupial paradigm also provides objective methods (if given the phase speed) for identifying the center of the pouch at the intersection of the critical line and the trough axis associated with the tropical disturbance. Termed the sweet spot, Dunkerton et al. (2009) defined this intersection as the mostly likely location of TC genesis within the pouch. Numerous additional papers have been published since Dunkerton et al. 2009 in support of the theory, but all fail to explain some crucial elements of TC genesis. Specifically, there is no discussion of how the region of recirculation first forms, a problem that has existed with TC genesis theories since CISK nearly fifty years ago. In addition, Dunkerton et al. (2009) and subsequent works focusing on the marsupial paradigm fail to answer the key questions of if and when a tropical disturbance will undergo genesis. Regardless of these gaps in the marsupial paradigm, the concept of the pouch provides a useful and potentially objective way of tracking tropical disturbances throughout the pre-genesis period. This method was utilized for real-time operations.
during the PREDICT field campaign and will be discussed further in Chapter 2 and is used to track the pre-Karl disturbance later in this thesis.

c. Equatorial wave theories and observations

Modern study of equatorially trapped waves is based on solutions of the shallow water equations presented by Matsuno (1966), which served to outline the dynamical basis for a spectrum of dry equatorial wave modes. Using shallow water solutions, Matsuno (1966) presented a clean and robust set of equations for all equatorial waves, including Kelvin, equatorial Rossby (ER), mixed Rossby-gravity, and westward and eastward inertio-gravity waves, but with the notable exceptions of African easterly wave type disturbances (hereafter, TD-type waves) and the MJO. Matsuno also presented the theoretical pressure and wind anomalies associated with each mode, schematics of which can be found in his 1966 work. For a Kelvin wave, these anomalies are simply alternating westerlies and easterlies along the equator and are associated with high and low pressure, respectively (Fig. 1.6a, his Fig. 8). With no meridional component to the anomalous winds, the creation of relative vorticity is dynamically impossible, and PV creation is also impossible with dry dynamics on and near the equator. In contrast, the ER wave schematic portrays a well-defined pair of cyclonic circulations exhibiting equatorial symmetry, with enhanced westerly wind anomalies along the equator (Fig. 1.6b, his Fig. 4c). Finally, Matsuno shows that these anomalies are equatorially trapped and reduce quickly with increasing latitude. If these shallow water equations are solved for the phase speed of a Kelvin wave, theory suggests eastward propagation at ~40-50 m s$^{-1}$, values not often observed in the real atmosphere. This apparent inconsistency stems from the presence of moisture and, in turn, deep convection that is commonly observed with
equatorial waves and effectively slows the propagation of the convectively coupled form of waves. As a result of this coupling to convection, the observed phase speed of a CCKW is typically \( \sim 15 \text{ m s}^{-1} \) (Straub and Kiladis 2002). Even with convection, the direct impact of Kelvin waves is generally confined to within \( \sim 10^\circ \) of the equator, but generally shifts northward in the Western Hemisphere (Kiladis et al. 2009; see their Fig. 5).

The convection coupled to equatorial waves also aids in the generation of low-level cyclonic circulations (Roundy 2008) and PV (Schreck and Molinari 2011), something not dynamically possible in theoretical dry wave modes. Fig. 1.7 is a set of lagged composites of CCKW events in the Indian Ocean categorized by MJO phase taken from Roundy (2008) and shows the presence of low-level circulations poleward of the convection associated with a CCKW, similar to those expected with the presence of an ER wave, but without the physical requirement of equatorial symmetry. The circulations persist for more than two days after the CCKW passage, suggesting a significant impact on flow patterns in the wake of a CCKW. Schreck and Molinari (2011) further expanded this work by showing that CCKW-induced westerly wind anomalies near the equator can aid in the development of cyclonic relative vorticity when in a background of climatological easterlies due to the presence of cyclonic shear strips. Schreck and Molinari showed that low-level PV is also created in this region poleward of the CCKW but equatorward of the strong easterly trade winds. In their case study, the passage of a CCKW leaves a strip of 850 hPa PV \( \sim 3000 \) km long, out of which two distinct TCs form (Fig. 1.8, their Fig. 6). This strip of vorticity forms on the poleward edge of the Kelvin-filtered Tropical Rainfall Measuring Mission (TRMM; see Kummerow et al. 2000) rainfall anomalies and is associated with the CCKW passage.
In addition to case studies like that of Schreck and Molinari (2011), broader studies of the interactions between formative TCs and CCKWs have been performed recently. In the Indian Ocean, Bessafi and Wheeler (2006) explored the modulation of TCs by various equatorial wave modes via a principal component analysis of wave-filtered outgoing longwave radiation (OLR). Both the MJO and ER waves were shown to exert a strongly significant influence over TC genesis, while CCKW activity was shown to exert a weaker (but still statistically significant) influence on TC genesis. For a global perspective, Schreck et al. (2012) provided an acute, objective study of the frequency of TC genesis events attributable to common equatorial wave modes. Schreck et al. (2012) attributed a TC genesis event to an equatorial wave mode if the mode’s filtered TRMM rainfall exceeded a set threshold on the day of TC genesis. For the Atlantic, this attribution method showed that TD-type waves are by far the most common wave to which TC genesis can be attributed (~80%), with all wave types other than CCKWs are significantly responsible for a smaller portion (~20% each) of TC genesis events (Fig. 1.9). However, as shown previously by Roundy (2008), low-level cyclonic circulations are only beginning to form in association with a CCKW on the day it passes, leaving open the possibility that TC genesis could occur in association with these gyres formed due to a CCKW in the days subsequent to the CCKW passage. Ventrice et al. (2012) also suggested this temporal lag in TC genesis post-CCKW passage and showed that TC genesis frequency in the Atlantic main development region peaks at nearly double the climatological mean ~2 days after the passage of the leading edge of a CCKW (Fig. 1.10). Combined with the Schreck and Molinari (2011) case study, these observed lags in TC genesis events suggests the approach of Schreck et al. (2012) might be improved by
accounting for lagged attribution cases, which could in turn bring out a stronger signal of potential CCKW attribution (Schreck, personal communication).

A growing body of literature exists that focuses on the nature of interactions between equatorial wave modes and mid-latitude features (e.g. Moore et al. 2010, Straub and Kiladis 2003). Liebmann et al. (2009) discussed the relationship between South American cold surges and the in-situ creation of CCKWs over equatorial South America, near the northern end of the Andes. Composites are presented based on Kelvin-filtered OLR anomalies at 60°W on the equator and subsequently separated into cases with either Pacific precursors (pre-existing CCKWs propagating eastward through South America) or South America precursors (an Andean cold surge, usually creating a “new” CCKW over South America). Following a method comparable to that of Garreaud and Wallace (1998), Liebmann et al. separate the South America precursors utilizing a criteria of OLR anomalies at least 1.5 standard deviations below the mean at 20°S, 60°W (central Bolivia) 3 days prior to CCKW passage at 60°W. A total of 48 events meet their high-amplitude criteria over the November-May period (Southern Hemisphere [SH] warm season) of 1979-2006, and produce synoptic-scale composites (Fig. 1.11, their Fig. 8) similar to those presented in Garreaud (2000). Perhaps interestingly, only 4 of these 48 events overlapped with the 53 CCKW passages found to have high-amplitude origins in the Pacific, suggesting that it is relatively rare for a cold surge to interact with or amplify an already strong CCKW as it propagates across northern South America. It should be noted that being in the SH warm season, these cold surges are likely of lesser magnitude and frequency than those observed during the SH cold season, yet also easier to identify via OLR criteria. While the work of Liebmann et al. (2009) cannot be directly applied to TC
activity in the North Atlantic due to the seasonal shift, it serves to highlight the possible role of South American cold surges as a source of equatorial wave variability. This study of the pre-Karl disturbance will highlight one such case and establish a potential need for further research into these interactions.
FIG. 1.1. Visualization of a tropical cyclone as a Carnot heat engine. [From Fig. 13, Emanuel 1986.]
Fig. 1.2. A schematic for top-down genesis of a tropical cyclone. [From Fig. 12, Bister and Emanuel 1997.]
Fig. 1.3. Streamline analyses of circulations at (a) 500 hPa, (b) 700 hPa, and (c) 850 hPa near the pre-Irving disturbance on 1 August 1992. [From Fig. 10, Ritchie and Holland 1997.]
Fig. 1.4. Vorticity during a simulation of the pre-genesis stage of Hurricane Diana (1984). Two vorticity maxima (left) merge and strengthen (right) over a time period of ~30 minutes. [From Fig. 10, Hendricks et al. 2004.]
FIG. 1.5. Schematic for a “pouch” developed from the “marsupial” paradigm. [From Dunkerton et al. 2009.]
Fig. 1.6. Idealized schematic showing the wind field (vectors) and pressure anomalies (contours) expected from the (a) dry Kelvin wave-like and (b) $n=1$ equatorial Rossby wave solutions of the shallow water equations. [From Fig. 8 and Fig. 4c, Matsuno 1966.]
Fig. 1.7. Maps of composite Kelvin-filtered OLR and 850 hPa winds, where lag 0 is defined as dates when Kelvin-filtered OLR was minimized at 80°E and peaked at an amplitude less than -1 standard deviation. Winds only plotted if statistically significant from zero. OLR shaded every 5 W m$^{-2}$. [From Fig. 6, Roundy 2008.]
Fig. 1.8. Unfiltered 850 hPa PV (shaded) and 850 hPa wind vectors, with Kelvin-filtered rainfall (red contours, only 4 mm day$^{-1}$) and MJO filtered rainfall (green contours, only 4 mm day$^{-1}$) at 0000 UTC daily for 22-28 Jun 2002. Kelvin waves are identified in order of appearance across the top of each panel. Genesis locations of TCs labeled in panel (g). [From Fig. 6, Schreck and Molinari 2011.]
FIG. 1.9. Percentages of TC genesis events in the Atlantic basin that are attributable to each type of wave using wave-filtered TRMM rainfall thresholds of 2 mm day$^{-1}$ (white bars) and 4 mm day$^{-1}$ (gray bars). Red lines indicated 99% significance level. [From Fig. 5, Schreck et al. 2012.]
Fig. 1.10. TC genesis events relative to the leading (eastern) edge of the Kelvin-filtered negative OLR anomalies, defined to be Day 0. The “Climo” category represents the event count for an average daily lag. Error bars indicate the 90% confidence interval. [From Fig. 9, Ventrice et al. 2012.]
FIG. 1.11. Unfiltered rainfall (shaded, mm day$^{-1}$), 1000 hPa height anomalies (contoured every 10 m, dashed when negative, zero line omitted), and 1000 hPa wind anomalies (vectors). Offset described in text, with base points defined at offset 0 and offset -3. [From Fig. 8, Liebmann et al. 2009.]
2. Data and Methods

To focus on the environmental conditions surrounding the development of the pre-Karl disturbance, gridded reanalyses from the National Centers for Environmental Prediction (NCEP) Climate Forecast System Reanalysis (CFSR) are utilized (Saha et al. 2010; available from http://nomads.ncdc.noaa.gov/data.php?name=access#cfs-reanal). The CFSR provides a high-resolution analysis in horizontal spatial (0.5° grid spacing, ~55 km), vertical spatial (37 isobaric levels from 1000 hPa to 1 hPa), and temporal (analyses performed every 6 h) senses and contains a sufficiently long period of record (1979-2010) to calculate a reliable climatology. Extra effort has been taken with calculating the climatology to account for the diurnal cycle, as many basic properties of the atmosphere experience significant modulation by diurnal forcing in the tropics. As a result, a climatology was calculated for each synoptic hour (0000 UTC, 0600 UTC, etc.) using a 21-day centered running mean of values from only the given hour. While these calculations do not completely eliminate the diurnal cycle, as the diurnal cycle itself can have day-to-day variations, it does serve to mitigate its effects. The case study of pre-Karl will focus on examining mass, momentum, and moisture analyses from the CFSR, in both raw and anomaly form. Specific emphasis is placed on the utility of low-level vorticity and moisture in Chapter 3 and temperature and wind in Chapter 4.

As PREDICT was a field campaign, acquisition of observational data was a key component of the efforts in the field (e.g. Montgomery et al. 2012). Reconnaissance flights were made across the Caribbean and much of the western and central Atlantic Ocean (Fig. 2.1) throughout the 15 August – 30 September field campaign. While this thesis does not deal with any of these data directly, numerous dropsondes from the
National Science Foundation (NSF) Gulfstream-V (G-V) high-altitude reconnaissance jet made it into some real-time numerical guidance models (Roger Smith, personal communication), but was rejected from all analyses and reanalyses created by the National Centers for Environmental Prediction (NCEP), including the CFSR data used here. Despite this lack of assimilation by NCEP, gridded analyses of the pre-Karl disturbance are observed to generally improve around the time of the first investigative flight, centered on 1200 UTC 10 September. (It will later be shown that this flight occurs at a crucial stage in the organization of the pre-Karl disturbance, and this improved representation in analyses could be a result of this similarly improved organization.) Each flight released ~15-20 dropsondes throughout a ~6 h period (see example flight and drop map in Fig. 2.2. The G-V flew each morning starting on 10 September and ending as TC Karl underwent genesis on 14 September, with an additional afternoon flight occurring on 10 September. PREDICT was not a solo field campaign, however, and collaborated extensively with the Genesis and Rapid Intensification Process (GRIP; see http://airbornescience.nsstc.nasa.gov/grip/), funded by the National Aeronautics and Space Administration (NASA), as well as the Intensity Forecasting Experiment (IFEX; see http://www.aoml.noaa.gov/hrd/HFP2010/IFEX.html), funded by the National Oceanic and Atmospheric Administration (NOAA) and the operational forecasters at the National Hurricane Center (NHC). GRIP also flew into the pre-Karl disturbance, with a DC-8 flying into the disturbance every afternoon (centered around 2200 UTC) starting 12 September and continuing through most of Karl’s time as a TC. Combined, PREDICT and GRIP flew a total of 7 flights into the pre-Karl disturbance in the 96 h period prior to genesis. Numerous other real-time and quality-controlled post-processed data were
produced in association with PREDICT and contributed both directly and indirectly to this thesis, including extensive satellite products produced by the Cooperative Institute for Meteorological and Satellite Studies (CIMSS). This data can be accessed at http://catalog.eol.ucar.edu/predict/index.html.

Methods utilized for tracking pre-genesis disturbances during PREDICT, GRIP, and IFEX were similar to those presented by Dunkerton et al. (2009) and discussed previously in Chapter 1. Numerous disturbances (both with and without significant genesis potential) were tracked in near-real time with these methods, utilizing data from daily 0000 UTC model analyses and subsequent short-range forecasts. By first determining the approximate phase speed of the disturbance, which is a subjective determination, features such as the critical line (along which phase speed equals zonal wind speed) and dividing streamlines (separating inside and outside of pouch at a given level) can be calculated. The intersection between the trough axis (defined where meridional winds go to zero in regions of cyclonic curvature) and the critical line defines the “sweet spot” and roughly indicates the center of the pouch and where genesis would be most likely to occur per the marsupial paradigm (real-time example in Fig. 2.3). Each summer from 2009 to present (2012), Mark Boothe of the Naval Postgraduate School tracks these disturbances by hand on a near-daily basis. Data from his efforts can be found at http://www.met.nps.edu/~mtmontgo/storms2010.html (and linked pages) along with a pouch “best track” dataset, which includes interpolated consensus tracks from each of the available models during the summer of 2010.

The National Hurricane Center (NHC) keeps official records (“best tracks”) for TCs for the North Atlantic basin prior to, during, and after each system’s life as a TC.
However, these records often go no further back than the first point at which the low pressure center is “well-defined.” Although this point can be around the time the precursor tropical disturbance initially forms, it was not the case with the pre-Karl disturbance. NHC best track data does not track the pre-Karl disturbance prior to 1800 UTC 13 September, although the post-season TC Report from NHC discusses a “broad surface low pressure system” as early as 8 September (Stewart 2011). Pouch tracking for the pre-Karl disturbance began at 0000 UTC 9 September with updated analyses from every 0000 UTC and 1200 UTC model run (in contrast to the once-daily 0000 UTC analyses performed for most disturbances). While the method of tracking used by Boothe in real-time is undoubtedly more subjective than the careful post-storm analysis from NHC, it is able to provide a complete track dataset through the pre-genesis period for the pre-Karl disturbance and is utilized accordingly in this thesis (Fig. 2.4).

Chapter 4 features equatorial wave analysis based on the utilization of filtered fields (also briefly mentioned in Chapter 1). Wheeler and Kiladis (1999) were first to utilize filtering techniques to identify equatorial wave modes, and virtually all subsequent work relating to these wave modes is based upon their techniques. Using OLR, Wheeler and Kiladis calculate a power spectrum in the wavenumber-frequency domain. This OLR power spectrum suggested broad regions of slightly increased power, but did not clearly differentiate these peaks from the background power. While the appropriate background power is nearly impossible to accurately determine, a rough approximation of this background was obtained via iterative smoothing of calculated total power spectrum. In calculating these smoothed powers, any strongly periodic signals are removed or greatly reduced, leaving a reasonable base state for the tropical atmosphere. Finally, Wheeler and
Kiladis (1999) divided the total power by this approximation of background power and obtained distinctive regions of power that roughly aligned with the corresponding shallow water solutions from Matusno (1966). Rough boundaries were drawn around the regions of highest spectral power and served to establish boundaries for wavenumber-frequency filtering of virtually all equatorial wave modes, including the MJO and CCKW (Fig. 2.5). While more recent work has suggested slightly different boundaries for the CCKW region (Kiladis et al. 2009), the Wheeler and Kiladis (1999) definitions are used here. Specifically, CCKWs were defined within a window of eastward-moving wavenumbers \(k\) from 1 to 14, periods \((1/f)\) from 2.5 to 20 days, and equivalent depths \(h\) from 8 to 90 m. This filtering window is applied to the once-daily 2.5-degree resolution OLR dataset (Liebmann and Smith 1996; available from http://www.esrl.noaa.gov/psd/data/gridded/data.interp_OLR.html ). Filtering is only applied within 25 degrees of the equator in order to avoid sampling signals more likely associated with mid-latitude variability. Once-daily OLR data represent a daily average, and as a result are plotted with 1200 UTC gridded data when possible to best represent the midpoint of the high temporal resolution data that accompanies the daily-averaged OLR data.
Fig. 2.1. Approximate domain of the PREDICT field campaign while utilizing the NSF Gulfstream-V high-altitude reconnaissance jet (shaded). Stars represent location of the operations base (St. Croix) and the evacuation location for the aircraft (Barbados).
Fig. 2.2. Flight track for the G-V aircraft for the afternoon flight on 10 Sep, including dropsonde points (red dots). Flight began at 1630 UTC and ended around 2200 UTC. Underlain satellite image valid approximately at the end of the flight.
Fig. 2.3. Plots of relative vorticity (shaded) and co-moving streamlines for a westward phase speed of 5.6 m s⁻¹ (arrows) associated with the pre-Karl disturbance using 1200 UTC 11 Sep ECMWF analysis data. Objectively identified trough axis shown as black line; objectively identified critical line shown as purple line; “sweet spot” shown by black circle.
Fig. 2.4. Tracks of tropical disturbances investigated by PREDICT G-V flights. Points along tracks are plotted every 6 h, with red tracks indicating time as a pouch and green tracks indicating time spent as an official TC. Blue squares indicate the times and disturbance locations for G-V flights, to the nearest 6 h point. Pouch numbers are labeled at the ending point of the respective track.
Fig. 2.5. Subset of the total power divided by the “background” (smoother) power, calculated between 15°N and 15°S for only symmetric waves (those with odd-numbered meridional modes \([n]\)). Contour interval every 0.1 starting at 1.1. Thick black lines represent borders of filtered regions for each labeled wave type. [From Fig. 6, Wheeler and Kiladis 1999.]
3. Synoptic overview of the near-disturbance environment

Before examining the environment surrounding the pre-Karl disturbance, a summary of Karl and other proximal TCs follows. While official best-track records for TC Karl begin at 1200 UTC 13 September, the PREDICT field campaign tracked the precursor disturbance for Karl as early as 9 September. The track (Fig. 3.1) and intensity (Fig. 3.2) of Karl and its precursor disturbance track an ill-formed and meandering disturbance on 9 September before beginning a clearer west-northwestward path on 10 September. It is around this time that the disturbance first begins to organize, as seen in gridded relative vorticity fields from the CFSR (Fig. 3.3), with some mid-level circulation appearing in satellite views of the storm on 10-11 September (not shown). The pre-Karl disturbance continued to struggle through 12 September as it traversed the central Caribbean Sea, with transient pulses of convection failing to increase the low-level relative vorticity associated with the disturbance (Fig. 3.3). By late 13 September, the disturbance developed more sustained convection (Fig. 3.4) and low-level vorticity, in turn, increased, beginning the second developmental stage of the pre-Karl disturbance. The NHC declared the disturbance a tropical storm at 1200 UTC 14 September (Stewart 2011) and TC Karl continued to intensify quickly before making landfall in Belize on 15 September. TC Karl then reemerged over the Bay of Campeche late on 16 September and reorganized before a period of rapid intensification made Karl a major hurricane with maximum 1-minute sustained winds of 110 kt early on 17 September (Figs. 3.1,b) prior to its final landfall later that day in eastern Mexico (Stewart 2011).
Many Atlantic TCs form from tropical disturbances associated with African easterly waves (AEWs) progressing westward across the main development region (MDR) of the Atlantic Ocean (e.g. Landsea 1993, Schreck et al. 2012). In the case of the pre-Karl disturbance, this common relationship is of uncertain truth. The NHC post-season TC report states that TC Karl “formed from the interaction between a westward-moving tropical wave and an elongated trough of low pressure” (Stewart 2011). An examination of limited observational data from the MDR alongside gridded analyses suggests that no such precursor wave existed. A sequence of 700 hPa stream function and relative vorticity maps show little cyclonic curvature in the vicinity of or upstream from the first identification of the pre-Karl disturbance via pouch tracking at 0000 UTC 9 September around 11˚N and 57˚W (Fig. 3.5). Based on the position and speed of the broad AEW near the coast of Africa at 0000 UTC 2 September, and extrapolating its motion across the entire Atlantic, it is plausible that this AEW could have been in the vicinity of the pre-Karl disturbance on 8-9 September. However, the MDR remained devoid of persistent deep convection between 2-9 September (Stewart 2011) and no noticeable dynamical signals were evident in the gridded wind field analyses, suggesting upstream AEW activity played little to no role in the formation of the pre-Karl disturbance. Additionally, deep convective activity across the MDR remained relatively low in early September. Average infrared brightness temperatures over four days (6-10 September, Fig. 3.6a) immediately prior to pre-Karl’s initial spin up highlights the convective signal of the ITCZ at a lower latitude (~8˚N) while little convective activity poleward of the central and western portions of the MDR is evident. In contrast, colder
infrared brightness temperatures at higher latitudes over the subsequent four days (10-14 September; Fig. 3.6b) suggest increased convective activity over the MDR, which is generally associated with increased AEW activity. This higher-amplitude AEW activity also results in a more diffuse representation of the ITCZ. Since an AEW is not associated with the formation of the pre-Karl disturbance, it falls into the minority of developing tropical disturbances in the Atlantic and this thesis will explore the conditions surrounding this atypical pathway to the formation of the pre-Karl disturbance through the 8-11 September time period.

b. Dry air surge

An examination of physical factors that influence TC genesis previously discussed in Chapter 1 will serve to highlight key elements of the process surrounding the atypical formation of the pre-Karl disturbance. Of first interest is an outbreak of dry air that spread westward across the Atlantic during the early part of September. In order to analyze the origin of this dry air, a backward trajectory analysis is performed using a program constructed by Matthew Janiga that employs bi-linear interpolation in both time and space in order to compute trajectories with custom time steps (M. Janiga, personal communication). Here, 120 h backward trajectories were calculated from 0000 UTC 1 September. Using CFSR 6-hourly gridded analyses, temporal interpolation was set to calculate the trajectory points every 10 minutes. Trajectory analysis reveals that despite its proximity to the North African coast, this region of dryness consists mainly of air with subtropical mid-latitude origins (Fig. 3.7). Some parcels from the western Atlantic are associated with the ridge enhanced by the upper-level outflow from TC Danielle around 28 August as well as an upstream upper-level trough exiting the United States (Figs. 3.7c-
d; parcels undergo an anticyclonic loop in the western Atlantic). While the backward trajectory analysis in Fig. 3.7 only includes parcels that end with relative humidity values below 20%, a complete run was performed and other, moister regions nearby are made up of air parcels from distinctly different sources (e.g. the Sahel and farther south and west over the Atlantic).

As this region of dry air begins moving southwestward at ~10 kt on 2 September, an enhancement of westerly flow occurs along the northern periphery of a high-amplitude, non-developing AEW (Fig. 3.8a-b and Fig. 3.5a) and serves to hasten the westward movement of the dry air. By 4-5 September, this surge of dry air begins to interact with the remnant of TC Gaston (Fig. 3.9), which was “very close to becoming a tropical depression again” (Blake 2010). However, organized convection did not persist with ex-Gaston, and PREDICT G-V flights show dry air wrapping into the circulation of the ex-Gaston disturbance on 5 September (Blake 2010). The surge of dry air continued to envelope ex-Gaston over the next three days as it progressed quickly to the west with short-lived and intermittent bursts of convection. While the presence of the dry air ensures that any organized convective activity remained limited, the general persistence of convective activity (as described by Blake (2010) and visible in Fig. 3.8c) acts to slowly moisten the dry air as it continues westward. By 0000 UTC 8 September, the leading edge of the dry air becomes poorly defined (around 50°W, Fig. 3.9) and consists of a loose gradient of moisture in a region of relatively light mid-level flow (Fig. 3.8d). Over the next ~48 h, a region of high relative humidity to the southwest of the dry air begin shifting progressively eastward, serving to somewhat slow or temporarily halt the westward motion of the dry air during 9-10 September (Fig 3.8e). Pouch tracking
initiated on the pre-Karl disturbance at 0000 UTC 9 September, with organization quickly increasing throughout 10 September (Figs 3.5e-f). During the 9-10 September period, dry air became progressively closer to the developing pre-Karl disturbance (Fig. 3.8d-e, Fig. 3.9), with the distance from the 20% relative humidity contour to the pouch track decreasing from ~1000 km to ~500km. However, the dry air does not fair well as it begins to impinge on the expansive region of high relative humidity values to the south and west, leaving much of the already-diffuse leading edge to continue to mix with the moisture ahead of it and become less of a hindrance for the pre-Karl disturbance. A more detailed analysis of this interaction between the dry air surge and the pre-Karl disturbance is presented in the following section.

c. Analysis of air mass interactions

While the proximity of the low relative humidity values to the pre-Karl disturbance is useful information, it is hard to conclusively analyze the interaction between the two moist and dry air masses by looking at a single time or single pressure level. The utilization of a Lagrangian perspective allows for a further analysis of interactions between dry air and the developing disturbance. Specifically, backward trajectories are computed from the vicinity of the pre-Karl disturbance and forward trajectories are computed from the region of dry air, calculated using previous described methods (M. Janiga, personal communication). Backward trajectories were calculated for a 72 h period ending at 1200 UTC 10 September from the vicinity of the pre-Karl disturbance ending at the 600 hPa level. It should be noted that while this interpolation represents an improvement over linear interpolation, smaller-scale diabatically driven radiative and sensible and latent heat flux processes are not accounted for, leaving the
finer details of the timing, locations, and values of variables along each trajectory trustworthy to only first order without deeper inspection via a high-resolution model with short time steps, an option not pursued here.

In order to facilitate a clear understanding of an otherwise dense set of trajectories, a small set of representative trajectories were selected to broadly represent three regions of air parcels which undergo a unique and physically meaningful time evolution in the 72 h prior to 1200 UTC 10 September. In order to better understand these processes, a progressive series of time-stepped plots of the trajectories can be seen in Fig. 3.10 with relevant gridded analyses of relative humidity and satellite infrared brightness temperature as underlays. Fig. 3.10a, valid at 1200 UTC 8 September, highlights the distinct origins of each trajectory cluster. The first cluster of trajectories (numbered 0, 3, 4, and 5) appears to be at levels near the top of the boundary layer (roughly 800 hPa) and embedded deep within a region of high relative humidity at 600 hPa. This region of high relative humidity roughly encompasses the clusters of deep convection over northern South America while the parcels within it move mainly westward. The second cluster of trajectories (numbered 1, 2, 6, 7, 8, and 9) consists of parcels that are already located in the mid levels (~700 hPa) or are ascending towards that level as they progress westward in the trade winds. These parcels are located east of the Leeward Islands and just off the northern coast of South America and remain just outside of the envelope of high relative humidity that exists at 600 hPa. Finally, the third cluster of trajectories (numbered 10, 11, 12, and 13) are located near the 600 hPa level and farther north in a region with low relative humidity values. This region of dry air is the same one discussed in the previous section 3b and these parcels are located near the increasingly diffuse leading edge of the
dry air mass (Fig. 3.10a). While the third cluster of trajectories continues level on a southwestward path toward generally moister air over the next 24 h, parcels in the first two clusters of trajectories have either slowed or turned back eastward by 1200 UTC 9 September (Fig. 3.10b). These parcels have also become further enveloped by both the 80% relative humidity contour at 600 hPa and the general envelope of convective activity. In addition to the aforementioned zonal movements, the trajectories of the first cluster begin to move northward while the parcels in the second cluster move somewhat southward, implying some weak confluence in the region immediately surrounding the deepest convection north of Trinidad.

By 1200 UTC 10 September, the parcel trajectories from the first and second clusters indicate that a region of zonal shear has become established, with parcels south of roughly 12°N moving quickly eastward and parcels north of 12°N moving westward with some cyclonic curvature (Fig. 3.10c). While the trajectories themselves cannot be used to infer the spin-up of a low- or mid-level vortex, the observed zonal shear pattern combined with the increase in relative vorticity as previously noted in Fig 3.3 suggests that the pre-Karl disturbance is beginning to spin up by 1200 UTC 10 September. The region of high relative humidity values at 600 hPa has expanded northward while progressing eastward and is beginning to encompass the region surrounding the convection associated with the pre-Karl disturbance, helping to establish a moister and more favorable environment for development. Further, the third cluster of parcel trajectories associated with the dry air has turned westward by 1200 UTC 10 September, suggesting that the drier parcels have perhaps been somewhat turned away from the moister environment near the broad center of the developing disturbance by the
strengthening mid-level circulation, lending some credence to the marsupial paradigm discussed in Chapter 1 and Dunkerton et al (2009). This spin up can also be observed via the evolution of the component terms of low-level vorticity. The ratio of shear vorticity to total relative vorticity is presented in Fig. 3.11 and illustrates the low-level portion of the pre-Karl disturbance is initially dominated by shear vorticity and becomes increasingly associated with curvature vorticity by 0000 UTC 11 September.

d. South American moisture and vertical wind shear

   As discussed in conjunction with the backward trajectory analysis, an expansive region of high (> 80%) relative humidity values at 600 hPa can be traced back to its development over northwestern South America around 0000 UTC 6 September (Fig. 3.8e). This region of high moisture can then be tracked moving eastward over subsequent days through 10 September (Fig. 3.8e). Using infrared brightness temperature as a proxy for deep convection at 0000 UTC (near the diurnal maximum of convection over northern South America), a semi-coherent region of convection, or convective envelope, exists between 10°S and 10°N in the lee of the Andes Mountains on 6 September that did not exist the previous day (Figs. 3.12a-b). This convective envelope generally progresses eastward (Figs. 3.12b-f) in good agreement with the eastward progression of the region of high relative humidity at 600 hPa, suggesting the moisture and convective envelopes are related and likely tied to each other. While the relation between the mid-level moisture and deep convection seems apparent, it is impossible to say at this point in the analysis whether the convection begets the mid-level moisture or vice-versa or if perhaps some exterior forcing could be the cause of both.
In addition, the convective development and associated spin-up of low-level vorticity on 10 September seem with the pre-Karl disturbance occurs in conjunction with a favorable superposition in the deep-layer vertical wind shear fields. For this analysis, vertical wind shear is calculated as the difference between the average wind in an upper-level layer (specifically 250-150 hPa) and the average wind in a lower-level layer (specifically 900-800 hPa). This technique allows for the analyses to better account for changes in the depth of convection with the pre-Karl disturbance. Prior to the initial formation of the pre-Karl disturbance, a strip of low vertical wind shear values exists across most of the MDR between 15°N-20°N (Fig. 3.13a) and likely aids in allowing intermittent convective activity with the remnants of TC Gaston despite being enveloped in an environment of dry air (Blake 2010). Lagging behind the westward-moving low-level remnants of Gaston is the center of a region of light, anticyclonic shear (Fig. 3.13a), likely enhanced by the sporadic short-lived convective bursts associated with the system. This region of low shear (< 10 kt) shifts south and west during 8-9 September (Figs. 3.13a-b) toward the Leeward Islands and the convection associated with the small pre-Karl disturbance. By 1200 UTC 10 September, the center of anticyclonic vertical shear crosses the southern Leeward Islands and is almost directly above the pre-Karl disturbance as it begins to quickly organize (Fig. 3.13c). This low-magnitude vertical wind shear oriented in an anticyclonic manner around the storm indicates the general warm-core nature of the convection and improved upper-level convective outflow with pre-Karl and likely both contributes to and represents the effects of the low-level spin-up and associated increase in convective organization of the tropical disturbance.
e. Summary of Chapter 3

The North Atlantic served as a “battleground” of sorts between a large outbreak of westward-moving mid-latitude dry air over northern portions of the MDR and a convectively active region of high moisture progressing eastward off of northern South America during the early portion of September 2010 (Figs. 3.8, 3.12). As the dry air continued westward across the MDR, it smothered most deep convection associated with TC Gaston, which in turn served to moisten the dry air as it continued westward (Fig. 3.8a-c). Further interaction between the dry surge and the region of high moisture over northern South America resulted in additional moistening of the dry air surge, lessening its distinct leading edge and possible potency for the pre-Karl disturbance.

The region of high relative humidity over northern South America developed on 6 September, progressing eastward to slowly encompass the developing pre-Karl disturbance by early 9 September. Deep convection associated with the region of high relative humidity also progressed eastward and likely served to increase the previously minimal convective activity with the pre-Karl disturbance. In turn, increased deep convection near the disturbance can serve to generate low-level vorticity and help to spin up the developing disturbance. A backward trajectory analysis (Fig. 3.10) shows this is the case, indicating pronounced low- and mid-level cyclonic shear and suggestions of the presence of cyclonic vorticity, which is confirmed via gridded analyses (Fig. 3.3). Finally, an upper-level anticyclone responsible for a region of low-magnitude anticyclonic vertical wind shear drifts slowly westward before becoming juxtaposed with the developing convection around the pre-Karl disturbance, enhancing upper-level outflow
while promoting and/or indicating the development of a warm core with the disturbance (Fig. 3.13).

While this confluence of generally favorable environmental parameters served the pre-Karl disturbance well, it seems optimistic to think that all of these events were able to come together by pure chance and, if this were the case, tropical cyclogenesis would likely be a much more rare event. Further analysis shows that many of these features were driven by larger-scale physical processes located to the south and west of the disturbance, directions not usually monitored in the deep tropical Atlantic due to being generally downstream of the mean flow pattern in the region. However, as already observed with the region of high relative humidity at 600 hPa, broad synoptic-scale regions of moisture and relative dryness do interact with the developing pre-Karl disturbance from this direction. With this in mind, Chapter 4 further investigates the role of South American weather events in the evolution of the pre-Karl disturbance and the region of high relative humidity over northern South America.
Fig. 3.1. Two tracks of TC Karl (top) and the disturbance tracked as PGI44L during the PREDICT field campaign, which eventually became TC Karl (bottom). Red and black circles mark the 0000 UTC positions along both tracks. [Note: PGI designation represents the PREDICT, GRIP, and IFEX collaboration during the field campaign phases of the experiments).
Fig. 3.2. Intensity of TC Karl in terms of (top) maximum wind and (bottom) minimum central pressure from NHC best track. Official values follow the black line, with other intensity estimates indicated by markers per the respective legends. Some estimates track almost as far back as the beginning of the PGI44L track, allowing a rough estimate of the pouch intensity to be inferred as well. From Figs. 2 and 3 of Stewart (2011).
Fig. 3.3. 850 hPa relative vorticity averaged within a 2˚x2˚ box following the center of PGI44L track (shown in Fig. 3.1).
Fig. 3.4. Time series of tropical overshooting tops (TOTs; see Monette et al. 2012) associated with PGI44L and TC Karl. Note the strong diurnally-aligned convective bursts nightly through 13 September. (From Montgomery et al. 2012.)
Fig. 3.5. Plots of 700 hPa relative vorticity (shaded, units of $10^{-5}$ s$^{-1}$) and stream function (contoured every 2, dashed where negative, units of $10^6$ m$^2$ s$^{-1}$) for 0000 UTC (a) 4 Sep, (b) 6 Sep, (c) 8 Sep, (d) 9 Sep, (e) 10 Sep, and (f) 11 Sep 2010. Black ‘x’ in panels c-f tracks the appx. location of the vorticity maximum with the pre-Karl disturbance.
Fig. 3.6. Average infrared brightness temperatures over the four-day periods of (a) 0000 UTC 6 Sep – 0000 UTC 10 Sep and (b) 0000 UTC 10 Sep – 0000 UTC 14 Sep. Averages constructed using data from six-hourly time steps (0000 UTC, 0600 UTC, etc.) within each period from NASA’s Merged IR dataset.
Fig. 3.7. (a-c) 120 h backward trajectory analysis originating from region of dry air over the subtropical eastern Atlantic at 0000 UTC 1 September. Parcels only plotted if the terminal point has a relative humidity value below 20%. Colors along the parcel traces correspond to the pressure levels indicated by the color bar on the right. Numbers are plotted at the origin of each trajectory (at 0000 UTC 27 August) with directional arrows plotted at the end point of the trajectory. Terrain elevation is contoured in yellow/brown/black coloring every 500 m. Regions of stippling represent the regions of relative humidity exceeding 80% (in green) and below 20% (in brown) valid at the end time of the trajectories, 0000 UTC 1 September. Parcels in panel (a) is divided into two subsets: panel (b), which contains parcels ending west of 46˚W, and panel (c), which contains parcels ending east of 46˚W. Panel (d) shows potential temperature and wind on the dynamic tropopause (defined as the +2 PVU surface) along with layer-average 925-850 hPa cyclonic relative vorticity contoured every $2.5 \times 10^{-5} \text{s}^{-1}$ at 0000 UTC 27 August, the origin time of the trajectories in panels (a-c).
Fig. 3.8. Plots of 600 hPa relative humidity (shaded, %) and wind (arrows where greater than 4 m s\(^{-1}\)) for 0000 UTC (a) 2 Sep, (b) 4 Sep, (c) 6 Sep, (d) 8 Sep, (e) 10 Sep, and (f) 1200 UTC 12 Sep 2010. Red ‘x’ in panels d-f indicates appx. location of pre-Karl vorticity maxima, as indicated in selected panels of Fig. 3.5.
Fig. 3.9. Continuity diagram documenting the progression of dry air as portrayed in Fig. 3.8. The line at each time represents the leading edge of the dry air surge, or where the leading edge can be estimated after becoming diffuse. Colors represent progression of the leading edge with time.
Fig. 3.10. Backward trajectories calculated over the 72 h period prior to 1200 UTC 10 Sep terminating at the 600 hPa level. Only selected trajectories are plotted from a set whose locations at 1200 UTC 10 Sep were spaced every 1° of latitude and longitude between 9°N - 17°N and 58°W - 66°W. Trajectory traces plotted between the times of 1200 UTC 7 September and 1200 UTC (a) 8 September, (b) 9 September, and (c) 10 September. Each trajectory is color-coded by its pressure level along the track. Numbers corresponding to each parcel are placed at the beginning of the respective trajectory, with arrows placed at the end in the approximate direction of the trajectory at the respective ending time point. Terrain elevation is contoured in yellow/brown/black coloring every 500 m. Regions of stippling represent the regions of relative humidity exceeding 80% (in green) and below 20% (in brown). Faded solid fill colors represent satellite infrared brightness temperatures colder than 273K. Both relative humidity fields and satellite infrared brightness temperature are plotted at the respective ending points of the trajectory segments in each panel.
Fig. 3.11. Ratio of shear vorticity term of vorticity equation to the total relative vorticity (shaded above 0.2, only plotted where total relative vorticity exceeds $2.5 \times 10^{-5} \text{ s}^{-1}$), contoured cyclonic relative vorticity (black, every $2.5 \times 10^{-5} \text{ s}^{-1}$), and 850 hPa winds (barbs, kt) at (a) 1200 UTC 9 September and (b) 0000 UTC 11 September.
Fig. 3.12. Infrared brightness temperature data from NASA’s Merged IR dataset at 0000 UTC (a) 5 Sep, (b) 6 Sep, (c) 7 Sep, (d) 8 Sep, (e) 9 Sep, and (f) 10 Sep. Colored regions indicate progressively cooler infrared brightness and generally represent cold cloud tops and active convection.
Fig. 3.13. Vertical wind shear (wind barbs, kt) and vertical wind shear magnitude (shaded between 10 and 50 kt) for (a) 0000 UTC 8 Sep, (b) 1200 UTC 9 Sep, and (c) 1200 UTC 10 Sep. Vertical wind shear calculated via method described in text.
4. Equatorial waves and mid-latitude interactions

a. Convectively coupled Kelvin wave history and passage

The role of equatorial waves in the modulation of tropical convection and mass and moisture fields may significantly impact the evolution of a TC or tropical disturbance. Previous discussion in Chapter 1 highlighted the role of the wind field associated with CCKWs in enhancing low-level moisture and convergence, supporting upper-level divergence, and creating low-level cyclonic relative vorticity (Fig. 1.8). In order to trace the propagation of large-scale global intraseasonal signals, a time-longitude (or Hovmöller) diagram is commonly utilized. While zonally oriented Hovmöller diagrams require the data to be averaged across a specified latitude band, this technique works well for CCEW modes due to the predominantly zonal dispersion observed. CCKW-filtered OLR anomalies are seen passing over northern South America during 7-10 September (Figs. 4.1, 4.2) in filtered OLR data (Wheeler and Kiladis 1999). This CCKW appears to have origins over extreme eastern portions of the tropical Indian Ocean around 21 August (Figs. 4.1, 4.2). As the CCKW progressed eastward, the negative CCKW-filtered OLR anomalies associated with this Kelvin wave reached a peak in excess of -30 W m\(^2\) (averaged between 5-15°N) on 25-26 August as the wave approached 180° longitude (Figs. 4.1, 4.2). During this ~5 day transit of the western Pacific a number of westward-propagating disturbances with cyclonic relative vorticity are observed to either first appear or be enhanced within the convectively active phase of the CCKW. Subsequently, these disturbances progress westward away from the CCKW and maintain their cyclonic relative vorticity (Fig. 4.1), an evolution that is consistent with previous observations of cyclonic relative vorticity generation in the wake of CCKWs (e.g. Roundy 2008, Schreck
By 28 August, the aforementioned CCKW began to accelerate across the central and eastern Pacific Ocean as its CCKW-filtered signature began to decrease (peak intensity falls to roughly -20 W m\(^{-2}\) by 3 September). Despite this apparent decrease in convective activity, the dynamical signature of the CCKW remains somewhat evident. The low-level zonal wind convergence associated with the convectively active phase of a CCKW is difficult to observe in unfiltered anomalous zonal wind fields (Fig. 4.2a) and is better illustrated using CCKW-filtered zonal wind anomalies (Fig. 4.2b). Although only sometimes evident in the unfiltered zonal wind anomalies, a general shift from CCKW-filtered anomalous easterlies to anomalous westerlies is easily observed across the convectively active phase of the CCKW and is best represented as the CCKW traverses the western and central Pacific (24-29 August). Low-level convergence calculated from the CCKW-filtered anomalous wind is shown in Fig. 4.3 and highlights the relatively strong pattern of convergence associated with the CCKW for much of its life over the Pacific and northern South America. These sequential shifts of zonal wind anomalies and resulting convergence agree well with the wind pattern of the Kelvin wave as originally described by Matsuno (1966).

The eastward propagation of the CCKW can also be observed with horizontal maps, allowing a better understanding of the meridional extent of the convective and dynamic signals associated with the CCKW. Fig. 4.4 presents instantaneous infrared brightness temperatures and 850 hPa anomalous wind vectors to support the CCKW-filtered OLR anomalies (Figs. 4.1, 4.2) and zonal wind anomalies (Fig. 4.2) previously referenced in Hovmöller diagram format. Over the western Pacific, CCKW filtered OLR anomalies (in red, Fig. 4.4a-b) are present over a broader latitudinal band (−0-20°N) that
narrows with continued eastward propagation, eventually spanning a band of only 5-15°N over the tropical Atlantic (Fig. 4.4f). This latitudinal narrowing of convective anomalies is supported by the infrared brightness temperature data, with widespread convection generally observed only within and in the immediate vicinity of the convectively active phase of the CCKW (Fig. 4.4a,d-f). This latitudinal narrowing is likely a result of the increased stability associated with the South Pacific cold tongue and its representation near the equator, generally making the atmosphere less prone to deep convection along the equator (Fig. 4.5). Anomalous 850 hPa wind vectors are also observed with the convectively active phase of the CCKW beginning around 2 September (Fig. 4.4c). Prior to this point, strong background anomalous easterly flow superimposed with the CCKW results in net anomalous easterly flow through the convectively active phase of the CCKW (Fig. 4.4a,b), but a general reduction in the anomalous easterly flow can be noted in association with CCKW-filtered westerly wind anomalies as seen in Hovmöller format (Fig. 4.2b). This dominance of the environmental background flow suggests the CCKW wind anomalies are of a lower magnitude than the background anomalies in the tropical eastern Pacific.

The CCKW is of greatest relevance to the study of the pre-Karl disturbance as it transits northern South America, roughly between 6 – 11 September (Figs. 4.4d-f). During this time period, convection with the CCKW increases and the CCKW-filtered negative OLR anomalies increase from values around -8 W m⁻² (6 September) to values in excess of -18 W m⁻² (8-9 September, averaged from 5-15°N; Fig. 4.1, 4.2). This amplified convective signal is maintained as the wave exists northern South America and continues across the Atlantic on 11 September (Fig. 4.4d-f). In the low-level wind field, a
broad region of anomalous westerly winds can be observed generally within a couple degrees of latitude of 10°N. While somewhat difficult to follow across Central America, this region of anomalous westerly wind progresses eastward behind the convectively active phase of the CCKW (Fig. 4.4d-f) and intensifies as the magnitude of CCKW-filtered OLR anomalies increase (Fig. 4.2). A more detailed examination of the low-level wind field associated with the CCKW and its interaction with the pre-Karl disturbance will be presented later in this chapter.

While much in the way of equatorial wave research focuses on the application of statistical techniques such as spatial-temporal filtering (e.g., Wheeler and Kiladis 1999), observational studies have been performed as well. Since there are relatively few opportunities to observe equatorial waves using in-situ radiosonde data, a single set of radiosonde data taken every 4 hours from a ship deployed with the Tropical Eastern Pacific Process Study (TEPPS; Yuter and Houze 2000) in August 1997 provided an excellent opportunity to observe the vertical structure of a CCKW removed from the effects of terrain as in Indonesia and equatorial South America. The vertical time series observed in TEPPS provided data that was consistent with regressions from ECMWF reanalysis data (Fig. 4.6, from Straub and Kiladis 2002) and has proven consistent with studies calculating CCKW-based regressions and composites in other regions of the tropics (e.g. Kiladis et al. 2009, Ventrice et al. 2012). Specifically, the vertical structure of zonal wind in the convectively active phase of a CCKW consists of shallow westerly winds at the surface beneath a deep region of strong easterly winds and a shallow lens of westerly winds at and above the tropopause. Knowledge of this vertical structure allows
for validation of the statistical filtering methods when CCKWs of relatively high amplitude cross over land, as the CCKW associated with the pre-Karl disturbance did.

In order to perform such a validation, vertical wind profiles from once-daily soundings at Bogota, Colombia (station SKBO) and Boa Vista, Brazil (SBBV) suggest that the zonal wind signatures associated with a CCKW did pass over the stations during the 8-10 September period. Fig. 4.7 compares time series of the observed vertical wind profiles (plotted in reverse time order to increase similarities to a zonal cross section) and the cross section of a composite Atlantic CCKW (from Ventrice et al. 2012, their Fig. 2). A pattern of low-level westerlies that increase in depth with time is observed underneath a column dominated by deep easterly flow. In addition, upper level winds (above ~150 hPa) switch back to westerly in a descending layer. Although the comparison figure from Ventrice et al. (2012) is a cross section of composited grids (Fig. 4.7a), a portion of the cross section that aligns with the convectively active phase of the CCKW does appear to be consistent with the simulated cross sections constructed from the radiosonde stations. In addition, the observed vertical wind profile at the two South American stations also matches the time series of TEPPS observational data shown in Fig. 4.6. This observational evidence of a CCKW passage strongly supports the published statistical filtering methods utilized to track equatorial waves in this case.

b. South American cold surge

Despite having their origins in mid-latitudes, equatorward surges of anomalously cold air do sometimes extend into the subtropics and tropics, resulting in both circulation and convective responses. Often, these equatorward surges of anomalously cold air (hereafter, “cold surges”) are channeled equatorward via some flow barrier to their west,
which is almost always high terrain (e.g. Mailler and Lott 2010). Equatorward-flowing surges of primarily meridional wind serve as the primary mechanism behind these cold surges. When air behind a cold surge in the Southern Hemisphere flows equatorward, it experiences a westward torque via the Coriolis effect, which in turn results in a build-up and acceleration of the mass along the barrier – in this case, equatorward. One of the primary modes of temperature and meridional wind variability in South America stems from the year-round occurrence of cold surges in the lee of the Andes (Garreaud 2000). Several detailed case studies (e.g Garreaud 1999, Lupo et al. 2001) and composite studies (e.g. Kiladis and Weickmann 1997, Liebmann et al. 1999) have aided in the development of a better understanding of the synoptic patterns that lead to cold surge events in South America. Lupo et al. (2001) present a composite diagram of 1000 hPa height anomalies for Andean surge events, seen in Fig. 4.8. A distinct region of anomalously high 1000 hPa heights is concentrated along the lee slopes of the Andes in northern Argentina and Bolivia as the surge begins its equatorward push. A conceptual model derived from compositing these surges from Garreaud (2000; Fig. 4.9 here).

Further work has been done in tying these Andean cold surges to convective activity over the Amazonian basin. While a number of regional climate studies have been performed, Liebmann et al. (1999) and Kiladis and Weickmann (1997) present global perspectives on potential forcing mechanisms for composited tropical convection. Both of these studies contain composites that are similar to those presented in Lupo et al. (2001) and other South American cold surge climatologies, indicating cold surges can lead the development of convection near the equator by ~2-4 days. The importance of cold surges in the tropics was further emphasized in the more recent work of Liebmann et
al. (2009), in which a direct relationship between Andean cold surges and CCKWs is established. Utilizing time-lagged composites, Liebmann et al. (2009) identified two potential sources for high-amplitude CCKWs (<=-1.5 sigma OLR anomalies) over northern South America – precursor CCKWs moving eastward from the Pacific and cold surges moving equatorward east of the Andes. Of the events producing CCKWs with OLR anomalies exceeding -1.5 sigma (53 from the Pacific and 48 from cold surges), only 4 events were common to both sets of cases. However, the Liebmann et al. (2009) study only examined cold surges in the Southern Hemisphere (SH) cold season (defined to be November – May) and thus loses some relevance to the study of CCKW and cold surge events with regard to TCs during the North Atlantic hurricane season (June – November). Regardless, Liebmann et al. (2009) does highlight the dynamical background for an interaction between a CCKW and South American cold surge and subsequent amplification of convection associated with the CCKW over northern South America and the tropical Atlantic Ocean. Fig. 4.10 contains a terrain map highlighting topographic features that play a role in interactions between CCKWs and South American cold surges.

Examination of the Andean cold surge will utilize low-level temperature and wind anomalies to follow the progression of both the cooler temperatures and anomalous southerly winds that constitute the primary dynamical signals of South American cold surges. Figs. 4.11 and 4.12 contain plots of 850 hPa standardized temperature anomalies with anomalous 850 hPa vector winds portraying the equatorward progression of the early September 2010 cold surge. Initially, weak temperature anomalies and near-climatological flow dominates the Patagonia region at 0000 UTC 31 August (Fig. 4.11a). This changes as a strong cyclone passing south of Cape Horn works with a strong
anticyclone (~1040 hPa central seal level pressure) over the southeastern Pacific Ocean to push a cold front across the southern Andes and into southern and central Argentina by 1 September. To the west of the Andes, cold air surges southward with no impediments, while cool air east of the Andes pools in the immediate lee of the mountains and creates a ridge of higher sea level pressures against the terrain (Fig. 4.11b). Anomalous flow around the eastward extension of the Pacific anticyclone helps to pile mass against the eastern slopes of the mountains, further amplifying the shallow low-level ridge and enhancing the southerly winds. These winds result in the ridge of high pressure and anomalous cold (standardized to appx. -2 sigma) reach 20°S by 0000 UTC 2 September (Fig. 4.11b), but appear to stall near this latitude for ~1-2 days during which some warming of the cold anomalies can be noted. This stagnant pattern is quickly changed with the development of a surface cyclone north of Uruguay on 3 September. The developing cyclone subsequently moves southward to a location along the Uruguayan coast at 0000 UTC 4 September where a tight sea level pressure and implied low-level height gradient between the cyclone and the ridge associated with the cold surge farther west induces 850 hPa wind anomalies that exceed 30 m s⁻¹ (Fig. 4.11c). Farther north, surface isobars are less tightly packed and curve away from the terrain with the flow around the developing cyclone, implying diffluence in the wind field. This diffluence appears to aid in the dispersion of the high pressure ridge associated with the cold surge, leaving the cold air to continue equatorward with a significantly weakened sea level pressure signature, but with significantly enhanced southerly winds driving the surge (Fig. 4.11c).
The cold air surge initially maintains its previously described structure with temperature anomalies around -2 sigma and highly anomalous terrain-following southerly flow approaching 10°S by 1200 UTC 5 September (Fig. 4.12a). Over the subsequent 48 hours, the cool standardized temperature anomalies weaken significantly and blend into the environmental field by 1200 UTC 7 September (Fig. 4.12b). Meanwhile, the terrain-following wind field, having crossed the equator around 0000 UTC 7 September, has progressed to ~5°N by 1200 UTC 7 September. It should be noted the terrain the wind follows during this ~48 h period has changed direction significantly, with the ridgeline of the Andes shifting from a NNW-SSE orientation along 10°S to an orientation along a NE-SW line at 5°N – approximately a 70° turn (Fig. 4.10). As a result, the south and southeasterly flow observed with the cold surge south of the equator is being forced increasingly westward. For air parcels that cross the equator into the Northern Hemisphere, the direction of the Coriolis torque changes from toward the Andes (enhancing equatorward flow via barrier jet-type dynamics) to away from the Andes. However, the proximity of the region of interest to the equator makes it unlikely that the effects of Coriolis torques are large.

At 1200 UTC 7 September, the remaining vector wind anomalies associated with the cold surge (peak anomalies of ~10 m s⁻¹ at this time) are located between the northern Andes and the elevated terrain of the Guiana Highlands of southern Venezuela, both of which extend above the 850 hPa level in some locations (Fig. 4.12c). As a result, the low-level (850 hPa and below) winds are funneled between the two elevated terrain features in much the same way wind is funneled between skyscrapers in a large city. Stronger (nearly 15 m s⁻¹) wind anomalies are observed by 1200 UTC 9 September between the
mountains of northern Venezuela and the Guiana Highlands to the south. These anomalies have become almost pure westerly at a latitude of ~8˚N, forced by the continued eastward turn of the ridgeline of the northern Andes, and have merged with a zonally elongated region of westerly wind anomalies spanning much of Central America and the Atlantic (Fig. 4.12c). The merged westerly wind anomalies over northern Venezuela intensify throughout 10 September and appear to aid in the creation of an anomalous vortex on the cyclonic shear side of the anomalous westerlies (near Barbados), which is directly associated with the pre-Karl disturbance. By 1200 UTC 11 September, this vortex is well defined in the anomalous wind field and has warm 850 hPa temperature anomalies in excess of +1 sigma (Fig. 4.12d). This corresponds with a doubling of the low-level relative vorticity associated with the disturbance as discussed in Ch. 3 (Fig. 3.3).

Finally, a backward trajectory analysis was performed in order to verify the mid-latitude origin of the cold surge. Using calculation settings similar to those discussed in Chapter 3, backward trajectories were taken 120 hours back from 0000 UTC 8 September originating at 950 hPa. These trajectories were chosen to roughly encompass the leading portion of the anomalous southerly wind surge, being placed every 2˚ of latitude and longitude between 0-10˚S and 66-76˚W. A total of 17 (of 33) trajectories originate south of 20˚S (Fig. 4.13), with 11 of those 17 also originating at a higher altitude – generally above 700 hPa at 0000 UTC 3 September, indicating strong subsidence in the lee of the southern Andes as the developing coastal cyclone reinforced the cold surge with stronger cold, southerly flow. These trajectories confirm the initial role of the strong ridge over the
southeast Pacific and the general continuity of the air mass as it moves equatorward and
warms.

c. Kelvin wave-cold surge interaction

While the CCKW passage and South American cold surge discussed in the
previous two sections can be considered as separate and distinct factors that contribute to
the initial spin up of TC Karl, it is important to note that they share a key common feature.
Anomalous westerly winds associated with the cold surge along the Andes reached
central Venezuela around 1200 UTC 8 September, approximately the same time that the
convectively active phase of a CCKW also entered the region with its anomalous
westerly winds. An analysis of both events via the anomalous 850 hPa winds reveals that
the CCKW and cold surge begin to interact on 8 September, with the anomalous westerly
winds from each becoming indistinguishable from each other on 10 September. Fig. 4.14
demonstrates this interaction, with wind anomalies from both the CCKW and cold surge
annotated in addition to the longitude of the pre-Karl disturbance throughout the
interaction. Initially (1200 UTC 6 September; prior to interaction) the individual events
are associated with 850 hPa wind anomalies $< 10$ m s$^{-1}$ in magnitude located in the Gulf
of Tehuantepec and against the eastern slopes of the Andes at 5˚S (Fig. 4.14a). Over the
next two days, the CCKW westerly wind anomalies traverse the relatively high terrain of
Central America and move toward the central Caribbean while the southerly wind
anomalies in South America continue across the equator and turn eastward into
Venezuela (Fig. 4.14b). At this point, the anomalous winds from the two events separated
only by the far northern portion of the Andes along the north coast of Venezuela at 1200
UTC 8 September. With continued eastward movement of these anomalous westerly
winds, the two streams merge smoothly and become increasingly anomalous (>10 m s⁻¹) as they exit northern South America around 60°W at 1200 UTC 10 September (Fig. 4.14c) and result in almost doubling 850 hPa relative vorticity associated with the pre-Karl disturbance (Fig. 3.3). Both satellite imagery and gridded analyses alike suggest that the pre-Karl disturbance has a well-defined circulation beginning late 10 September and lasting through 11 September, likely spinning up due to the enhanced westerly winds to the south of the disturbance.

While the focus of this work is on the development of the pre-Karl disturbance, it is important to note that the interaction between the cold surge and CCKW is also significant and may actually feed back onto the development of the pre-Karl disturbance via the enhancement of convection. At 1200 UTC 6 September, the CCKW-filtered OLR anomalies exceed 15 W m⁻², a value which increases to ~30 W m⁻² by 1200 UTC 8 September and is maintained as the CCKW begins to traverse the open Atlantic Ocean by 1200 UTC 10 September (Fig. 4.14c). Two possible physical explanations can explain this increase in convective activity. First, CCEWs generally remain trapped at or near the equator as they propagate eastward or westward. With a cool tongue of sea surface temperatures in the equatorial eastern Pacific (Bjerknes 1966), the environment is generally more stable and thus less favorable for convection. As an eastward-moving CCEW exits this less favorable environment into one with a generally warmer lower boundary or otherwise decreased stability (such as land, especially high terrain), it is possible that convective activity increases due to this implied decreased stability alone (Fig. 4.5), although it is not known how significant the convective response to this change in stability may be.
The second and likely more significant factor supporting the enhancement of CCKW convection between 6 – 8 September comes from the enhancement of the anomalous westerlies. When juxtaposed with a region of near-zero anomalies to the east (ahead of the convectively active phase of the CCKW) in a region of climatological easterlies, there is an implication of low-level convergence driven by the strength of the westerly anomalies in the convectively active phase. As the westerly wind anomalies within the convectively active phase strengthen to peak values >10 m s$^{-1}$, the average intensity of convection within the envelope of the CCKW increases as well (infrared brightness temperature in Figs. 4.14a-b). The strengthening of the winds during this period is associated with the addition of westerly flow from the remains of the South American cold surge, as shown in Fig. 4.14b. In turn, the CCKW-filtered OLR anomalies also strengthen over this same period, implying at least a portion of this increased convection is associated with the CCKW passage. Combined, the enhanced convection and increases in vorticity associated with anomalous westerly winds were enough to spin up a well-defined persistent vortex around 1200 UTC 10 September that persisted for approximately 24 hours. While important to note that the pre-Karl disturbance did not meet the NHC definition of a TC at that time, it likely remains an example of an initially weak disturbance enhanced by the dynamics of a cold-surge enhanced CCKW.
Fig. 4.1. Time-longitude (or Hovmöller) diagram of shaded 850 hPa relative vorticity (warm colors where cyclonic) averaged between 7.5-17.5°N with CCKW-filtered OLR data (black contours starting at +/- 10 W m⁻² and every 5 W m⁻²; dashed where negative). Red dashed lines highlight disturbances enhanced by the CCKW; green dashed line represents track of the pre-Karl disturbance with an ‘X’ marking the time and longitude of TC genesis.
Fig. 4.2. As in Fig. 4.1, but with shaded (a) zonal wind anomalies calculated from a 31-year CFSR climatology and (b) zonal wind anomalies averaged between 5-15°N filtered for the CCKW band.
Fig. 4.3. As in Fig. 4.2b with CCKW-filtered divergence (warm colors indicate negative values/convergence).
Fig. 4.4. 850 hPa anomalous winds (vectors; reference vector in upper right of panels), infrared brightness temperature (shaded; plotted only below 270 K), and CCKW-filtered OLR anomalies (contoured only at +/- 8 W m^-2; red and dashed used for negative/convectively active). Panel (a) begins at 1200 UTC 27 August and steps in 3-day increments to 1200 UTC 11 September at the bottom in panel (f).
Fig. 4.5. Lapse rates calculated from 1980-2009 climatological mean temperatures and heights during the months of August and September for (a) the 950-500 hPa layer and (b) the 950-200 hPa layer. Shading in units of K km$^{-1}$. 
Fig. 4.6. (a) Observed zonal wind anomaly (contour every 1 m s$^{-1}$, dark shading where positive/westerly) calculated from 4-hourly radiosondes launched as a part of TEPPS. Observations taken at 7.8°N, 125°W (TEPPS base point). (b) Regressed daily ECMWF temperatures from day -3 to day +3, based on a -125 W m$^{-2}$ anomaly in OLR at the TEPPS base point on day 0. (From Fig. 11, Straub and Kiladis 2002.)
Fig. 4.7. (a) Vertical cross section through a CCKW composite using a base point along the west African coast (adapted from Ventrice et al. 2012). (b) Location of upper air sounding stations of Bogota and Boa Vista used in panels (c) and (d) (Note: background image courtesy of the University of Wyoming). (c) A reverse time series of the observed vertical wind profile at Bogota, Colombia (SKBO). Time is reversed in order to mimic the structure of a cross section for an eastward-propagating disturbance. (d) As in (c) for the station in Boa Vista, Brazil. Annotations in (a), (c), and (d) intended to mimic each other in order to highlight similarities between the vertical wind profiles. Arrows indicate general direction of the zonal components of the wind, while red lines indicate the level at which the zonal wind switches sign.
Fig. 4.8. Composite of strong Andean cold surges centered on the time of the greatest equatorward extent of the 1000–850-hPa thickness contours, defined as Day 0. From this base, a lifecycle of the 1000 hPa height anomalies can be seen at (c) Day-2, (f) Day 0, and (i) Day+2. Shading represents regions of >95% significance. (From a portion of Fig. 3, Lupo et al. 2001.)
Fig. 4.9. Conceptual diagram of an Andean cold surge, where dark (light) thick arrows represent low-level wind advecting cold (warm) air. Thin contours represent surface isobars. High pressure centers and cold front at surface shown conventionally. (From Fig. 10, Garreaud 2000.)
Fig. 4.10. Significant terrain features of South America that play a role in the pre-genesis evolution of TC Karl, with terrain shaded every 200 m above sea level. A indicates the Guiana Highlands of Venezuela, Brazil, and Guyana; B indicates the Andes Mountains (note the large north-south extent and generally wavy pattern of the ridgeline); C indicates the elevated terrain of Central America.
Fig. 4.11. Standardized anomalies of 850 hPa temperature (shaded), anomalous 850 hPa wind (vectors), and MSLP (black lines; contours every 2 hPa) for 0000 UTC (a) 31 August, (b) 2 September, and (c) 4 September.
Fig. 4.12. 850 hPa wind and standardized temperature anomalies as in Fig. 4.7, with thin black contours representing terrain plotted every 1000 m starting at 500 m at 1200 UTC (a) 5 September, (b) 7 September, (c) 9 September, and (d) 11 September.
Fig. 4.13. Backward trajectory analysis of the leading portion of the cold surge at 0000 UTC 8 September. Trajectories were run backward for 120 h from the 950 hPa level between 0-10˚S and 66-76˚W. Trajectories are numbered the same at the beginning and end points, with the color of the trajectory representing the height of the trajectory at that time. Shaded fill represents topography between 0 and 2500 m.
Fig. 4.14. 850 hPa anomalous wind (vectors), infrared brightness temperatures (shaded below 270 K), and CCKW-filtered OLR anomalies (black contours, starting at +/-10 W m$^{-2}$ and every 5 W m$^{-2}$ beyond; negative values dashed). Purple arrow annotates winds with the CCKW; blue arrow annotates winds with the cold surge; red arrow annotates the longitude of the pre-Karl disturbance. Panel (a) at 1200 UTC 6 September, (b) at 1200 UTC 8 September, and (c) at 1200 UTC 10 September.
5. Summary and discussion

This study of the synoptic patterns leading to the initial spin-up of the pre-Karl disturbance has demonstrated the physical connection between the South American cold surge, the CCKW passage, and the development of the disturbance itself. A summary schematic of the synoptic patterns and their evolution can be found in Fig. 5.1. First, the cold surge develops as the South Pacific subtropical ridge builds eastward into Argentina and establishes southeasterly flow toward the Andes by 2 September (Fig. 5.1a). The resultant buildup of mass east of the Andes allows a surge to develop on 4 September, with the leading edge of the surge reaching the equator by 7 September (Fig. 5.1b). Also at this time, a CCKW can be seen crossing ~90°W as it passes over the far eastern tropical Pacific. Anomalous southerly winds from the cold surge continue to push northward and then turn eastward, following the terrain and merging with the anomalous westerly winds of the CCKW by 8 September (Fig. 5.1c). These additional westerly wind anomalies also enhance the CCKW-filtered OLR anomalies over northern South America (Fig. 5.1c). Enhanced convection and a broad region of anomalous westerlies associated with the CCKW progress eastward, helping to increase both the low-level vorticity and convective activity with the pre-Karl disturbance. The CCKW and its associated anomalies then continue eastward over the tropical North Atlantic and leave behind a disturbance (marked with an ‘X’ in Figure 1) with doubled low-level vorticity and greater convection than prior to the CCKW interaction by 11 September (Fig. 5.1d).

The schematic presentation of the synoptic evolution prior to the spin-up of the pre-Karl disturbance (Fig. 5.1) demonstrates that the spin-up was enhanced by a number of events constructively interfering prior to their interaction with the disturbance.
However, this evolution needs to be put into climatological context in order to establish its relevance to other cases beyond the pre-Karl disturbance. While a broader climatological perspective is not directly utilized in this study, some climatologies and general results from other literature are discussed below in the context of the pre-Karl disturbance. Since a detailed study examining the interaction of cold surges, CCKWs, and tropical disturbances combined (as presented here) is unlikely to prove simple nor identify many other similar cases, it is more practical to examine the role of each interaction on an individual basis. Such work would focus on the role of CCKW-TC interactions as well as the independent role of cold surge-CCKW interactions, both of which have been previously explored to various degrees in the literature. These additional perspectives should serve to further understanding of the pre-Karl disturbance as examined here and highlight additional pathways for research following these perspectives and perhaps further detailing physical processes that may lead to or otherwise force these interactions.

a. CCKW-TC interaction

As discussed in Ch. 3, the environment surrounding the initial development of the pre-Karl disturbance consisted of a surge of mid-latitude dry air from the north and east and a large region of high moisture and deep convection moving eastward off northern South America (Figs. 3.8, 3.12). This dry air was shown to remain on the fringes of the pre-Karl disturbance as the air closer to the core of the disturbance moistened during the 8-10 September period, feeding from a moisture source to the south and west of the disturbance (Fig. 3.10). As discussed in Ch. 4, the passage of the convectively active phase of a CCKW that moved eastward over northern South America with a phase speed
comparable to that observed with this region of high moisture (~15 m s$^{-1}$), suggesting the two are related as shown in composite studies (e.g. Kiladis et al. 2009, Ventrice et al. 2012). The combination of a slow westward motion by the cluster of thunderstorms that is to become the pre-Karl disturbance and the eastward shift of moisture with the CCKW passage results in the disturbance becoming enveloped in the larger region of convective activity with the CCKW.

The environment within the CCKW convective envelope is generally more favorable for a potential tropical disturbance to undergo TC genesis than its surrounding environment, with higher levels of moisture, enhanced low-level convergence, and upper-level divergence, all of which favor increased deep convection and the resulting low-level potential vorticity generation (discussed in Chapter 1). Forcing for the dynamical response comes from anomalous westerly flow in the convectively active phase and anomalous easterly flow in the convectively suppressed phase (shown from Matsuno 1966). In background easterly flow, both the westerly and easterly anomalies can result in regions of enhanced cyclonic shear (Fig. 5.2, adapted from Matsuno 1966). However, due to the proximity of the Kelvin wave to the equator, any cyclonic shear equatorward of the anomalous easterlies is in a less favorable environment of lower background planetary vorticity and generally drier air with only isolated deep convection and thus not favorable to the organization or intensification of a TC (Ventrice 2012). The most favorable location for TC genesis or intensification in a CCKW is therefore poleward of the convectively active phase, a fact that has been previously documented in both case studies and composites (e.g. Schreck and Molinari 2011, Ventrice et al. 2012). With the case of a CCKW centered off-equator to the north, such as with the pre-Karl case (also
demonstrated in Fig. 5.2), this favorable region for TC genesis is only to the north of the convectively active phase of the CCKW.

In order to put the CCKW-disturbance interaction observed with the pre-Karl disturbance into perspective, a broader climatological analysis must be conducted. Initial climatologies of CCKW-TC interactions indicate that a distinct link exists globally between CCKWs and TC genesis events. Schreck et al. (2012) take a simple approach in an attempt to attribute TC genesis to various equatorial wave modes. They calculate the percentage of TC genesis events that occur when within the convective envelope of a particular equatorial wave mode while allowing for attribution to multiple wave types when equatorial wave modes are superposed on the developing TC. CCKWs were found to be associated with some 15-35% of TC genesis events in each of the global TC basins at zero lag (Schreck et al. 2012). A time-lagged study of this relationship by Ventrice et al. (2012) shows that, for the North Atlantic, TC genesis events are nearly twice as likely to occur ~2 days after the passage of the peak of CCKW convective activity (Fig. 1.10).

Physically, this lagged relationship can be explained with the development of deep convection and its positive feedback cycle with increasing low-level vorticity – these processes can, in a favorable environment, eventually lead to the disturbance organizing into a TC. While the passage of a CCKW can enhance both deep convection and low-level cyclonic vorticity, neither will necessarily reach levels of sufficient organization to be classified a TC while the CCKW directly interacts with the disturbance. In the case of the pre-Karl disturbance, a significant increase in vorticity (Fig. 3.3) and convection (Fig. 4.14) can be noted in the 1-2 days following the CCKW passage, yet the disturbance is still not classified a TC at that time. (This is discussed further in section
5c.) Disturbances that evolve akin to the pre-Karl disturbance also weaken the results of a lagged attribution study based on the subjective designation of a disturbance as a TC (Karl developed at Day -1 and Day +5 in relation to its two closest CCKW passages). A CCKW passage does serve to make the environment more favorable for the development of a TC, but does not appear to be either a necessary nor a sufficient condition for the broader genesis process. Additional work is still needed in order to fully understand the physical processes occurring in CCKW-disturbance interactions in addition to the broader significance of these interactions on the climatology of TC genesis.

b. Cold surge-CCKW interaction

Interactions between mid-latitude Rossby waves and equatorial waves have been previously documented throughout the tropical Pacific, suggesting that a breaking Rossby wave either directly or indirectly (via a cold surge) forces a tropical convective response (e.g. Lau 1982, Kiladis 1998, Straub and Kiladis 2003). A number of studies have focused on the role of breaking Rossby waves in exciting convection over tropical South America (e.g. Kiladis and Weickmann 1997, Liebmann et al. 1999, Liebmann et al. 2009), including Rossby waves originally excited by tropical convection over the western Pacific (Straub and Kiladis 2003). Physically, tropical convection can be excited via a low-latitude upper-level trough inducing a westerly low-level flow in a region of background easterlies, although other mechanisms may exist as well. Fig. 5.3 (Fig. 2 from Straub and Kiladis 2003) highlights two mid-latitude Rossby wave breaking events in the Southern Hemisphere based on the existence of CCKW-filtered convection located at the dateline at day 0 (Fig. 5.3c). The first Rossby wave break occurs over Australia (Fig. 5.3a) while the second occurs about 12 days later over South America (Fig. 5.3d).
The Australian wave break excites a region of negative OLR anomalies along the equator that propagate eastward with a phase speed comparable to that of a CCKW while also re-exciting the mid-latitude Rossby wave. The mid-latitude Rossby wave disperses eastward faster than the CCKW-like convective anomalies, reaching South America as the tropical convective anomaly crosses 180˚W (Straub and Kiladis 2003).

While not examined in this study, it is possible that a mid-latitude Rossby wave dispersion event similar to the regression analysis presented in Straub and Kiladis (2003; Fig. 5.3) serves as a precursor to the cold surge and CCKW that interact with the pre-Karl disturbance. Fig. 4.4b shows CCKW-filtered OLR anomalies located around 180˚W ~3-4 days prior to the initial development of the low-level continental anticyclone over Argentina (Fig. 4.11a). This pattern implies that the CCKW reaching northern South America around 8 September evolves in a way that is consistent with a CCKW from the regression analysis by Straub and Kiladis (2003), specifically that this CCKW may have played a role in the excitation of the Rossby wave train that led to the initial cold surge over southern South America. Making such a statement conclusively would be difficult in this case, as the cold surge appears to pause between 2-4 September before being driven equatorward by a developing cyclone near the Uruguayan coast (Fig. 4.11c). The cyclone’s enhancement of what may have otherwise been a weak South American cold surge further complicates the timing of the interaction between the cold surge and the CCKW several days later. Taking these different factors into consideration and noting the rich spectrum of physical processes that can be associated with cold surges, it appears unlikely that this case of CCKW-cold surge interaction would appear frequently in a composite or regression based on any one of the key synoptic-scale precursors.
Considering the pre-existing composites of mid-latitude Rossby wave dispersion, it is possible that the regressions from Straub and Kiladis (2003) could contain the signal of a cold surge due to their similarity with upper-level composites found in Garreaud (2000). Based on these considerations, a more thorough effort to identify possible remote origins of South American cold surges could prove fruitful in establishing physical connections and preferred temporal patterns between CCKW activity and the generation of South American cold surges.

Liebmann et al. (2009) investigate the role of South American cold surges on Western Hemisphere CCKW activity during the November – May 1979 – 2006 period. Liebmann et al. (2009) show a relationship between cold surges that trigger convection around 20°S and the subsequent creation or amplification of CCKW-like convective anomalies near the equator ~3 days later. Of equal importance to these Liebmann et al. (2009) findings is the role of pre-existing CCKW events that propagate eastward from the tropical eastern Pacific Ocean and result in the development or enhancement of CCKW-like convective anomalies over northern South America (likely the result of the eastward propagation of the dynamically induced CCKW circulation over/around the northern Andes). Liebmann et al. (2009) also state that only 4 of the top ~50 CCKW-producing events (defined to be OLR anomalies with either CCKW or cold surge precursors in excess of -1.5 sigma) were common to each set of precursors.

While such a statistic would tend to indicate a relatively small probability of this coincidence occurring, a few caveats should first be considered. The Liebmann et al. (2009) study only looks at November – May because these months have the highest frequencies of CCKW-filtered OLR anomalies in excess of -1.5 sigma over northern
South America. A more extensive analysis is necessary in order to determine potential differences in the frequencies of cold surge-CCKW interactions during the North Atlantic tropical cyclone season, when CCKW activity is ~60% lower than during the boreal winter months over northern South America (Liebmann et al. 2009). The presence of easterly waves and other tropical disturbances in the vicinity of northern South America during this season could allow for cold surge-CCKW interactions to play a more important role in the evolution of TCs and tropical disturbances. Additional work is needed in order to establish the climatological context of the cold surge-CCKW interaction observed in the case of the pre-Karl disturbance.

c. Additional considerations

Much of the work presented both in Chapter 1 and in previous sections of this chapter has discussed the time-lagged relationship between CCKWs and TC genesis (Figs. 1.9, 1.10). Although this relationship holds up well in the case of the pre-Karl disturbance and the pre-Karl disturbance did experience an increase in convective organization and low-level relative vorticity during 10-11 September, TC Karl did not actually undergo genesis until 14 September (per NHC best track data; Stewart 2011). Despite having many of the characteristics of a TC (e.g. warm core, organized convection, well-defined low-level circulation), a closed surface circulation was not observed with the disturbance on 10-11 September, likely preventing the storm from being upgraded to TC status (Stewart 2011, Davis and Ahijevich 2012). After this initial spin-up, the pre-Karl disturbance struggled for the next 3 days (11-13 September) as its convection became intermittent and vertical wind shear increased across the system. The low-level tangential velocities associated with the circulation of the storm (as measured by
dropsondes from reconnaissance flights, some of which were a part of PREDICT) decreased slightly during 10-11 September, dropped off considerably during 12-13 September, and finally ramped up again by late on 14 September (Fig. 5.4; Davis and Ahijevich 2012). Davis and Ahijevich (2012) attribute much of the lack of development to general easterly vertical wind shear across the system, as evidenced by calculations from research aircraft-derived dropsonde data and the horizontal displacement between circulation centers at 900 hPa and 500 hPa. While Davis and Ahijevich do not speculate as to the cause of this vertical wind shear, it appears a small trough over the western Atlantic moves southward over the Greater Antilles on 10 September (Fig. 3.13c) and may prove at least partially responsible for this increased shear. However, this increase in westerly shear does not correspond with the increased easterly shear noted from the PREDICT dropsonde data. Regardless of the cause, this unfavorable shear (as calculated from dropsonde observations) abated by late 13 September (Davis and Ahijevich 2012) and allowed the pre-Karl disturbance to gain an official designation as a TC at 1200 UTC 14 September.

Emerging work focusing on the suppressed convective phase of a CCKW suggests that equatorial wave dynamics could also play a role in the suppression of convection and generally making the environment less favorable for TC genesis in the vicinity of the pre-Karl disturbance. Ventrice (2012) highlights the negative impact on TCs Danielle and Earl (2010) when the convectively suppressed phase of a CCKW passed through the TCs. Ventrice (2012) also presents a composite analysis of convectively suppressed phases of CCKWs utilizing a base point of 45°W to show that suppressed phases of CCKWs are generally associated with decreased precipitable water
and an increase in the magnitude of vertical shear (his Figs. 5.14 and 5.15). Although the physical mechanisms behind this increase in vertical wind shear magnitude is unclear, initial ideas point to forcings from the zonal wind anomalies associated with the CCKW. However, it should be noted that the favorable or unfavorable nature of the shear is dependent on both the direction of the CCKW-induced shear and the environmental shear values. Future work exploring the role vertical shear modulation by CCKWs will need to emphasize variability between background synoptic environments in addition to more general regional differences. In addition, the general lack of convection (anomalously high OLR values) associated with the convectively suppressed phase of a CCKW results in an environment that tends to be less favorable for TC genesis. In the case of the pre-Karl disturbance, the convectively suppressed phase of a CCKW can be observed moving toward the disturbance on 11 September (Fig. 4.4f, in more detail in Fig. 5.5), resulting in the pre-Karl disturbance being located within the envelope of suppressed convection until 13 September. These dates (11-13 September) correspond well with the period of decreased organization as described by Davis and Ahijevich (2012). This pattern of the pre-Karl disturbance being first enhanced by the convectively active phase and then hindered by the convectively suppressed phase of a CCKW fits well with the pattern presented in the Ventrice (2012) study of TCs Danielle and Earl. However, the dearth of literature investigating the possible relationship between the convectively suppressed phase of a CCKW and delayed or failed attempts at TC genesis prevents a more general statement from being made at this time. Any study that investigated this relationship would first need to identify a method of determining null cases of potential TC genesis, a dataset not currently available from operational or official sources and one that could be
difficult to construct without conducting multiple case studies. It is important to understand the role of the convectively suppressed phase of the CCKW as much as the convectively active phase, as both appear to have played roles in the development and evolution of the pre-Karl disturbance.
Fig. 5.1. Schematic illustrating the evolution of the South America cold surge, CCKW, and pre-Karl disturbance. Background map contains terrain filled brown above 500 m and black above 1500 m elevation. Black lines indicate anomalous wind associated with the cold surge, while red lines indicate anomalous wind associated with the CCKW. Blue shading indicates region of anomalous cold low-level temperatures. Blue ‘H’ designates the location of high pressure centers associated with the development of the cold surge. Green dashed contours indicate levels of CCKW-filtered negative OLR anomalies. Yellow ‘X’ indicates location of the pre-Karl disturbance.
Fig. 5.2. Idealized schematic of a Kelvin wave. Anomalous low-level winds indicated by black arrows and anomalous low-level geopotential height in green (dashed when negative). Shaded orange (blue) indicates region of cyclonic (anticyclonic) shear vorticity when the Kelvin wave is propagating in background easterly flow and has a latitudinal center slightly in the Northern Hemisphere. If the Kelvin wave is centered on the equator, the southern shear vorticity regions would switch signs and leave cyclonic shear vorticity associated with both the northern and southern fringes of the westerly wind anomalies.
Fig. 5.3. Regression of OLR (shading, dark where negative, levels of +/- 6 and 15 W m$^{-2}$) and 200 hPa streamfunction (contours, every $7.5 \times 10^7$ m$^2$ s$^{-1}$) and wind vectors based on a CCKW-filtered OLR anomaly of -40 W m$^{-2}$ at the basepoint of 180°W at day 0 for (a) Day -9, (b) Day -5, (c) Day 0, and (d) Day +3. Winds only plotted where significant at 95% level. (From Fig. 2, Straub and Kiladis 2003.)
Fig. 5.4. Vertical profiles of average tangential wind with respect to the circulation center (determined independently at each level). Legend beneath plot indicates the general time of the reconnaissance mission corresponding to each profile. Final mission (14 Sep 18Z) occurred post-genesis. (From Fig. 8b, Davis and Ahijevich 2012.)
Fig. 5.5. 850 hPa anomalous wind, infrared brightness temperature, and CCKW-filtered OLR anomalies as in Fig. 4.14. CCKW-filtered OLR anomalies are in red (solid lines represent positive anomalies/suppressed convection). Plots for 1200 UTC (a) 11 September, (b) 12 September, and (c) 13 September.
REFERENCES


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