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Evolution of Frontal Structure Associated with Extratropical Transitioning Hurricanes

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EVOLUTION OF FRONTAL STRUCTURE ASSOCIATED WITH
EXTRATROPICAL TRANSITIONING HURRICANES

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To my mother, Rosemary Maue, who loves me and misses me. And to my late aunt Geraldine Adamski and late uncle Frank Adamski who were always proud of me.
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ABSTRACT

Many tropical cyclones move poleward, encounter vertical shear associated with the midlatitude circulation, and undergo a process called extratropical transition (ET). One of the many factors affecting the post-transition extratropical storm in terms of reintensification, frontal structure, and overall evolution is the upper-level flow pattern. Schultz et al. (1998) categorized extratropical cyclones according to two of the many possible cyclone paradigms in terms of the upper-level trough configuration: The Norwegian cyclone model (Bjerknes and Solberg 1922) associated with high-amplitude diffluent trough flow and the Shapiro-Keyser cyclone lifecycle (1990) with low-amplitude confluent troughs. Broadly speaking, the former category is associated with a strong, meridionally oriented cold front with a weak warm front while the latter lifecycle usually entails a prominent, zonally oriented warm front. However, as will be shown, simple antipode lifecycle definitions fail to capture hybrid or cross-lifecycle evolution of transitioned tropical cyclones.

To exemplify the importance upper-level features such as jet streaks and troughs, a potential vorticity framework is coupled with vector frontogenesis functions to diagnose the interaction between the poleward transitioning cyclone and the midlatitude circulation. Particular focus is concentrated upon the evolution and strength of frontal fracture from both a PV and frontogenesis viewpoint. The final outcome of extratropical transition is highly variable depending on characteristics of the tropical cyclone, SSTs, and environmental factors such as strength of vertical shear. Here, three storms (Irene 1999, Fabian 2003, and Kate 2003) typify the inherent variability of one such ET outcome, warm seclusion. Very strong winds are often observed in excess of 50 ms$^{-1}$ along the southwestern flank of the storm down the bent-back warm front. The low-level wind field kinematics are examined using vector frontogenesis functions and QuikSCAT winds. A complex empirical orthogonal function (CEOF) technique is adapted to temporally interpolate ECMWF model fields ($\theta$, MSLP) to overpass times of the scatterometer, an improvement over simple linear interpolation. Overall, the above diagnosis is used to support a hypothesis concerning the prevalence of hurricane-force winds surrounding secluded systems.
CHAPTER 1
INTRODUCTION

As tropical cyclones translate poleward, many undergo a process called extratropical transition (ET). Several authors have studied the lifecycle evolution of extratropical transition for the major basins including the northwest Pacific (Klein et al. 2000, Harr and Elsberry 2000), southwest Pacific (Sinclair 2002) and northern Atlantic (Hart and Evans 2001; McTaggart-Cowan et al. 2001). While considerable attention has been paid to the transition process, less attention has been focused upon the frontal and overall cyclone structure of the consequent extratropical storm. Many authors have concluded that the midlatitude circulation into which the tropical cyclone is moving plays an important part in determining whether a transition will occur as well as reintensification. Thus, this thesis concentrates on the synoptic and dynamic interactions between the midlatitude circulation and the transitioning tropical cyclone in order to diagnose the factors responsible for the development of marine extratropical cyclones. While many post-transition storms undergo warm seclusion reminiscent of the Shapiro-Keyser (1990) cyclone lifecycle, other documented categorizations of development arise including LC1- and LC2-type waves (Thorncroft et al. 1993). It will be demonstrated that a simple categorization of cyclone evolution is insufficient to diagnose the overall continuum of cyclone variety. Three distinct hurricanes, Irene (1999), Fabian, and Kate (2003) exemplify the many differences between transitioning storms but also illustrate continuity in terms of midlatitude interaction.

The first part of this study explains the advantages and limitations of QuikSCAT scatterometer wind data at various spatial and temporal resolutions. Generally speaking, individual swath-oriented oceanic wind measurements offer high spatial resolution with inherent difficulties associated with rain contamination and gaps in data coverage away from the poles associated with most polar orbiters. Bourassa et al. (2004) developed a gridded scatterometer product using a variational technique to spatially and temporally combine separate swaths into a basin-wide representation. To reconcile differences in ECMWF model times (common synoptic intervals of six hours) and the QuikSCAT overpass times, which can be off by up to 3 hours, simple linear interpolation proves insufficient. For evolving, translating cyclones, simple linear interpolation tends to smear out the
overall structure and intensity greatly. A complex empirical orthogonal function (CEOF) method developed by Zavala-Hidalgo et al. (2003) is adapted to overcome the problems associated the temporal inconsistency of ECMWF temperature and pressure fields with respect to the QuikSCAT winds. This method proves useful for analyzing surface frontal evolution and structure discussed in Chapter 6.

The location of the upper-level jet and midlatitude trough with respect to the poleward moving tropical cyclone are two important considerations for ET. Furthermore, the interaction between these features generally creates an environment conducive for baroclinic instability and regeneration of the decaying tropical system. Chapter 3 presents a summary of the prevailing conceptualizations of extratropical transition in terms of midlatitude trough interaction and overall cyclone structural changes. Many authors have referenced several well-documented paradigms of cyclone development in order to explain extratropical transition. The intricacies of the classical Norwegian cyclone lifecycle, Shapiro-Keyser (1990) lifecycle models are examined with respect to an alternative classification scheme developed by Thorncroft et al. (1993) who distinguish between LC1 and LC2-type systems. The different cyclone paradigms combined with attendant analyses of upper-level jets from a potential vorticity (PV) and quasigeostrophic viewpoint provide a useful diagnostic framework. Coupled with the cyclone phase space (Hart 2003; Evans and Hart 2003), this framework seeks to elucidate upon the thermal and frontal structure of extratropical cyclones.

The PV framework furthered by Agusti-Panareda et al. (2004) for hurricane Irene (1999) is utilized to diagnose the effects of the upper-level jet. One phase of the Shapiro-Keyser (1990) lifecycle, the frontal fracture, will be shown to consistently develop as a result of the superposition of the advection pattern of the tropical cyclone upon the midlatitude baroclinic environment. This frontal fracture is also explained in terms of the Thorncroft et al. (1993) lifecycles to better understand the consequent development of a warm seclusion. Isentropic trajectory maps along with potential vorticity cross-sections demonstrate the interplay between the different scales inherently important to cyclone development. The development of warm seclusion is still a poorly understood process. However, a coherent analysis is presented which aims at interpreting the warm seclusion process in terms of the respective cyclone paradigms.

A vector frontogenesis analysis at upper-levels (500-hPa) and the surface using QuikSCAT winds is presented to show the interaction between the tropical cyclone vortex and the midlatitude baroclinic zone. While Harr and Elsberry (2000) analyzed typhoon extratropical transition, they did not present the final outcome of ET or document warm seclusions. The northern West Pacific climatologically does not favor warm seclusion development to the same degree as the northern Atlantic basin. Thus, a coherent lifecycle of extratropical cyclogenesis and cycloysis with emphasis upon warm seclusion is presented for the first time in terms of vector frontogenesis functions. Additionally, use of QuikSCAT winds to study extratropical transition from a vector frontogenesis viewpoint has not been documented. Additionally, preliminary hypotheses about the predominance of very strong winds (>40 ms\(^{-1}\)) on the southwestern flank of the storm along the bent-back warm front are presented. The use of Advanced Very High Resolution Radiometer (AVHRR)
satellite imagery provides for practical interpretation of the processes and conclusions presented forthwith. The potential vorticity and vector frontogenesis framework combined with the cyclone paradigms present a coherent and overarching diagnosis of the frontal structure and post-ET extratropical cyclone evolution with special emphasis upon warm seclusion structure at all levels of the troposphere. For clarity throughout the subsequent thesis, the remnants of the respective hurricanes will still be referred to as their National Hurricane Center (NHC) names.
CHAPTER 2
DATA: QUIKSCAT WINDS, ECMWF, AND CEOF INTERPOLATION

The SeaWinds microwave scatterometer provides a high-resolution spatial and temporal measurement of surface winds over the Earth’s oceans. For a period of six months, dual scatterometers orbited the Earth, one on the QuikSCAT satellite launched in June 1999 and the other on the short-lived ADEOS-II platform. It has been demonstrated that SeaWinds can provide an accurate picture of global oceanic winds when compared with buoy, ship, and other observations (Bourassa et al. 2003). Many different products have been created and tested in order to most accurately depict the wind data while minimizing errors primarily from rain. Two different products, gridded and swath-aligned will be used to examine the wind fields associated with transitioning tropical cyclones and the consequent extratropical cyclones.

2.1 SeaWinds Operation

Launched on June 19, 1999, the polar-orbiting QuikSCAT satellite replaced the failed Advanced Earth Observing Satellite (ADEOS-I), which suffered a fatal malfunction attributed to the payload solar cell (Fig 2.1a). The SeaWinds scatterometer aboard QuikSCAT proved to be a considerable enhancement over the failed NSCAT instrument primarily in terms of improved spatial coverage and data density. With the launch of the ADEOS II and its SeaWinds instrument destined to replace the retiring QuikSCAT in late-2002, dual scatterometer coverage of the ocean became a reality for a period of six months the following year. Unfortunately, the 4-ton satellite developed by the Japanese space agency suffered a similar solar panel failure as its predecessor and clunked out possibly due to enhanced solar flare activity. Thus, the QuikSCAT mission has been extended beyond its predicted operational lifespan until the next generation of scatterometers slated for 2008. Nevertheless, the dual coverage during 2003 has benefited scientists with a nearly complete four-times daily coverage of ocean winds. Storms, fronts, and other propagating features are resolved much better spatially and temporally especially near the poles where the adjacent swath spatial-overlap is largest. Near the equator, considerable gaps in data coverage exist both in time and space. Another
recent study has used the temporally dense scatterometer data to elucidate a diurnal cycle over the global oceans (Lombardi 2004).

Scatterometers operate by acquiring multiple spatially and temporally co-located measurements of backscattered power from different viewing geometries. The SeaWinds radar operates in the microwave band at 13.4 GHz using a rotating dish antenna with two conically rotating pencil spot beams that sweep in a circular pattern receiving horizontal and vertical polarization backscatter at 46.25° (707 km radius) and 54° (900 km radius) respectively ahead and behind the orbiting satellite (Fig. 2.1b). Microwaves are scattered by small-scale water waves including capillary and ultragravity waves, which respond quickly to surface wind changes. The backscatter cross section, which is the fraction of transmitted energy that returns to the satellite from the different viewing geometries, is a known function of wind speed and direction relative to the orientation of the scatterometer (Bourassa et al. 2003). The Ku band scatterometers (NSCAT and SeaWinds) are calibrated to “equivalent neutral wind speeds” at a height of 10 m above the local mean water surface, which agree well with observations (Bourassa et al. 2003).

Figure 2.1: (a) Rendering of QuikSCAT satellite (NASA). (b) Seawinds measurement geometry illustrating the antenna beam locations for different polarizations. Inner beam is H-pol and outer beam is V-pol. (adapted from Weissssman et al. 2003).

This viewing geometry yields the best observations between 200 and 700 km from nadir, with increased uncertainties farthest away and closest to nadir where colocation of observations is weakest. The individual footprints are binned into 25 x 25 km cells, with up to 76 cells across the satellite swath and 812 lines of vectors pole to pole. Each swath is 1800 km wide and separated in time by 101 minutes with a recurrent period of 4 days. At the equator, two neighboring swaths are separated
by approximately 2800 km compared to 2000 km in the midlatitudes. Thus, one scatterometer can provide a twice-daily complete picture of the surface winds over the midlatitude oceans under cloudy and clear conditions at grid spacing of roughly 25 km. In rainy conditions, the shape of the water surface is modified causing ripples as well as crowns and stalks which in turn influence the microwave backscatter.

2.2 Rain contamination

If left in, rain contamination may create spurious high-wavenumber features. The antenna illumination of QuikSCAT caused greater errors in rain contaminated scatterometer wind estimates when compared to the older NSCAT. The rain attenuates the radar signal and reduces the measured radar cross-section $\sigma_0$. On the other hand, the radar cross-section $\sigma_0$ is increased due to scattering of the radar signal by raindrops. Also, splashing of raindrops increases the sea surface roughness and increases the measured $\sigma_0$. At low wind speeds, it is relatively easy to determine where observations are rain contaminated and wholly unrealistic through a comparison of the dual polarizations (vertical and horizontal $\sigma_H$ and $\sigma_V$) of backscattered energy. However, at higher wind speeds, especially in tropical cyclones, the presence of rain modifies each polarization to a certain degree. The resulting wind vectors are typically changed in magnitude and rotated towards a direction perpendicular to the satellite track (cross track direction). This situation often introduces anomalous curl and interferes with tropical cyclone center location (Mears et al., 2000).

To correct the effects of rain contamination, empirical geophysical model functions have been developed with different modicums of success. In this study, SeaWinds data is used with the updated Ku-2001 geophysical model function developed by Remote Sensing Systems (RSS). This improved vector wind retrieval algorithm provides a fully integrated stand-alone rain flag and the capability to retrieve winds up to 70 m/s (Wentz et al., 2001). RSS conducted many observational studies to determine if SeaWinds was capable of measuring hurricane force winds and assess the effects of rain contamination. Wentz et al. (2001) found remarkable agreement with coastal stations during a strong extratropical storm and the major tropical forecasting centers in terms of maximum sustained winds for tropical cyclones.

2.3 COAPS Gridded Scatterometer Product

COAPS has developed a new FSU Winds and Flux Climatology through the extension of an objective gridding technique previous used to produce daily fields of scatterometer wind observations (Pegion et al. 2000). Several considerations must be examined in order to produce a valuable gridded wind product. Near the equator,
large gaps of unobserved ocean lie between neighboring swaths. However, in the midlatitudes and poleward, swaths at different times intersect, often with substantial differences in wind patterns due to fast moving weather systems. Along with observational error such as rain contamination and ambiguity selection, a simple averaging technique is not applicable. The results of such a simple averaging would introducing spurious wind curl and divergence into the wind fields, thus translating into spurious Rossby and Kelvin waves when used in ocean model calculations (Bourassa et al., 2004).

A major difference between the new objective technique and the previous one (Pegion et al. 2000) is the independent incorporation of a vast array of data sources including volunteer observing ships, buoys, and scatterometers. To minimize the objective function in order to retrieve the fields, a variational method utilizes several constraints to maximize similarity to observations, minimize non-geophysical features in the spatial derivatives and does so with the minimum necessary smoothing (Bourassa et al., 2004). In this study, only scatterometer data was used. The weights applied to each constraint in the functional are determined independently, which contain the considerations of observational uncertainty and data coverage. The weights vary for different platforms. In this analysis, the gridded fields are produced in the domain defined between 30°-70° N and 70°-10° W where the intersection of swaths is a more important consideration than the gaps in observational coverage. The ~25 km scatterometer swaths are objectively analyzed and outputted at a regular 0.50° resolution over the northern Atlantic using predetermined weights.

An example of the COAPS gridded scatterometer product is presented for the blending of two individual swaths over the north Atlantic corresponding to extratropical Irene (Fig. 2.2). The swaths (Fig. 2.2a,b) are approximately 100 minutes apart temporally and overlap spatially several hundred kilometers near their northern edges. The variational technique provides a continuous grid at 0.5° horizontal resolutions. With this product, easy kinematic analysis of surface winds can be performed over the entire Northern Atlantic on a temporally and spatially consistent basis.

2.4 ECMWF Model Data

Operation model data used was from the ECMWF TOGA (Tropical Ocean and Global Atmosphere) Global Advanced Operational Surface Analysis and the ECMWF TOGA Global Advanced Operational Spectral Analysis. 4 times daily synoptic times since January 1985. These datasets contain a four-times daily (00z, 06z, 12z, 18z) spherical harmonic (T106) representation of the atmosphere at an approximate resolution of 1.125° at the surface and 17 vertical levels. Since polar-orbiting satellites do not provide observations over a specific area of the Earth at the aforementioned synoptic times, the issue of temporal collocation of model and satellite data arises. Patoux et al. (2005) recognized this uncertainty and for lack of a better data source relied on visual consistency to carry out his QuikSCAT front
analysis. Most often, a simple linear interpolation scheme is used to temporally interpolate six-hourly model data to short time scales with mixed results. A new method is employed to help ameliorate this inconsistency between model and satellite data based upon propagating or complex empirical orthogonal functions.

Figure 2.2: Example of gridded QuikSCAT data: (a) and (b) Two adjacent overlapping swaths (~0.25°) on 20 Oct 1999 at 0701z and 0842z above with wind speed (shaded ms$^{-1}$) and sea level pressure (contoured every 4 hPa). (c) Gridded 0.5° wind speed and unit vector direction valid at 0800z.
2.5 Methodology

Since neighboring swaths overlap by several hundred kilometers at higher latitudes, it is convenient to use the gridded product to analyze cyclones here. The closest temporal midpoint is used as a representation of the two swaths together from each satellite, which encompass a 4,000 km wide view of the North Atlantic Ocean four times a day. Moreover, neighboring swaths with times at 06:30z and 08:01z are outputted to a 07z grid as opposed to a 06z or 12z grid to reduce error. Below 40° N, the swath gaps impose increased dependency upon the background field in the functional to fill in missing observations, thus introducing error. South of 40° N, individual swaths are used individually without an objective analysis. In both cases, swath and gridded, each vector component of the wind is low-pass filtered in an effort to reduce high wavenumber noise caused by any spurious rain contamination or ambiguity selection problems that interferes with direction retrieval while still preserving the mesoscale features such as fronts and cyclones that are highly visible.

2.6 CEOF Analysis

Since polar orbiting satellites do not sample at the regular synoptic times, but at different times for different regions of the oceans, it is necessary for collocation and calculation to temporally interpolate the ECMWF surface temperature and MSLP data to one hourly intervals to match up with the gridded QuikSCAT fields produced at higher latitudes. With the hourly MSLP and temperature data, potential temperature can be computed assuming adiabatic processes with a reference isobaric height of 1000-hPa and used to analyze cyclones and fronts. The complex empirical orthogonal function interpolation technique used to achieve this one hourly data is a similar technique applied to altimetry data at COAPS (Yu et al., 2004). Typically oceanographers linearly interpolate model data such as the NCAR Reanalysis for use in geophysical models (Zavala-Hidalgo et al. 2003). This is a good approximation for physical features that behave like standing waves, i.e. change their amplitude without moving. However, when there are fast moving features, such as evolving fronts and cyclones, the interpolated fields behave like duplicated standing fronts, changing their amplitude instead of moving. Thus, the linear interpolation tends to smear fronts and in the case of moving cyclones, produce a double bulls eye feature stretched in the direction of motion. Zavala-Hidalgo et al. (2003) present a new technique for time interpolation of fields with inherent moving features. The method decomposes the fields into their propagating empirical orthogonal functions, then the information from each mode is used to interpolate in time, and finally the reconstructed.

The main goal is to have a good interpolation in time recovering moving features such as storms and fronts. The derivation of the interpolation procedure closely follows the methodology of Yu et al. (2004) who reconstructed Gulf of Mexico
mesoscale eddies from Topex/Poseidon altimeter swath data. Propagating or complex empirical orthogonal functions (CEOFs) extract information from a two-dimensional data array,

\[ H = (h_{nm}) \]  \hspace{1cm} (1)

where \( H \) is an \( N \times M \) matrix of elements \( h_{nm} \), and \( N \) and \( M \) are the number of space and time points, respectively. CEOFs are applied to \( H \) and decomposed into eigenmodes. A detailed treatment of CEOF analysis is discussed in Barnett (1983) and Shriver (1991). The eigenmodes of \( H \) consist of two complex vectors,

\[ T_i = (T_m)_{1 \times M} \]  \hspace{1cm} (2) \hspace{1cm} \text{temporal function}
\[ S_i = (S_n)_{N \times 1} \]  \hspace{1cm} (3) \hspace{1cm} \text{spatial function}

known as the temporal function (TF) and spatial function (SF) respectively. Accordingly, any complex vector \( C(x) \) may be written as

\[ C(x) = A(x) \exp[i\theta(x)] \]  \hspace{1cm} (4)

where \( A(x) \) is the amplitude or magnitude of the vector and \( \theta(x) \) is the phase. Rewriting each complex vector yields the following:

\[ T(t) = R(t) \exp(i\phi(t)) \]  \hspace{1cm} (5)
\[ S(x) = E(x) \exp(i\theta(x)) \]  \hspace{1cm} (6)

The real part of their product (ST) determines the reconstructed field and is

\[ \text{Re}\{E(x)R(t)\exp[i\theta(x)+\phi(t)]\} = E(x)R(t)\cos[\theta(x)+\phi(t)] \]  \hspace{1cm} (7)

\( E(x) \) represents the variability in amplitude associated with a given eigenmode in space, while \( R(t) \) provides information on the temporal variability. The term \( [\theta(x)+\phi(t)] \) represents the phase at a given position and time. In order to create a finer grid of ECMWF data in time but at the same locations, the amplitude \( (R) \) and phase \( (\phi) \) at the original 6-hourly times are interpolated separately to the desired hourly field. The remapped phase and amplitude of the temporal function are used along with the spatial function to finally rebuild the data set. The product of SF and the new TF are summed together and comprise the new finger grid in time (Yu et al., 2004). Only the most significant eigenmodes are chosen, represented by the largest eigenvalues, in order to capture most of the variability.

For the temporal interpolation of the ECMWF model data at the surface, a total of 20 days were used for an 81 time steps in an effort to center upon the extratropical transition time of the system. The 6 hourly data \( R(t) \) and \( \phi(t) \) were then interpolated to one hourly time steps through a cubic spline. A total of 20 modes comprising 99% of the variance were used to reconstruct the new hourly dataset. The new method does indeed identify the propagating information, and retains the
waveform and amplitude at the interpolated time better than the linear interpolation method, which results in a reduced amplitude of the wave. Further, this technique can be applied to other scalar fields and extended to vector fields, such as wind maps (Yu et al. 2004).

2.7 CEOF INTERPOLATION EXAMPLE

A simple illustration of the usefulness of the method is shown in the following short time series of mean sea level pressure (MSLP) for October 20, 1999 00z, 06z, and 12z (Fig 2.3). The ECMWF model data is plotted on the left hand column clearly showing the low-pressure system associated with the extratropical remnants of Hurricane Irene, which explosively regenerated off the Canadian maritime coast. The system is propagating at over 20 ms\(^{-1}\) while developing into a major warm core seclusion. A CEOF technique is applied to the original ECWMF MSLP time series of 20 days including 81 time steps with 20 modes retained explaining 98% of the variance. The data is reconstructed onto an hourly grid-mesh. To demonstrate the technique’s ability to capture Irene, the CEOF fields at the identical synoptic times are arranged in the right hand column. The magnitude and location of the MSLP maxima at each time step are preserved as well as the overall isobaric pattern.

The next experiment involved a comparison of simple linear interpolation of model data with the CEOF technique’s ability to resolve finer time grids. The linear interpolation compares fairly well with the CEOF technique with the most visible difference related to the central pressure. The CEOF technique maintains a lower central pressure as compared to the linear interpolation technique at both 03z and 09z with differences larger than 3 hPa near the center (Fig 2.4 e-h). In such a strong system as Irene, linear interpolation provides a fairly good approximation given that the evolution and motion of the system is linear in nature over a short time period such as 6 hours. However, if the time period is increased, linear interpolation breaks down while the CEOF technique can achieve a more realistic representation of interpolated fields. As the number of modes chosen to reconstruct the data approaches the overall number of eigenmodes, the CEOF method limits to the simple linear interpolation method. In other words, the method can do no worse than simple linear interpolation, but often is a major improvement.
Figure 2.3: Left panel corresponds to simple linear interpolation and right panel is the CEOF reconstruction of ECMWF MSLP data at (a) and (b) 10/20 00z; (c) and (d) 10/20 06z; (e) and (f) 10/20 12z; (g) and (h) 10/20 03z; (i) and (j) 10/20 09z. Case study during warm seclusion of extratropical Irene (1999).
CHAPTER 3

CYCLONE PARADIGMS AND EXTRATROPICAL TRANSITION CONCEPTUALIZATIONS

3.1 Conceptual Models of Extratropical Transition

Both Evans and Prater-Mayes (2004) and Agusti-Panareda et al. (2004) have studies the explosive reintensification of Hurricane Irene (1999) as an extratropical storm. The former approaches the ET problem through a model simulation and a largely quasigeostrophic diagnosis while the latter develops a potential vorticity (PV) conceptual model. Each study provides a unique framework with which to understand, model, and diagnose ET. Their approaches will be summarized, combined, and augmented with a vector frontogenesis function analysis (Harr and Elsberry 2001) and remote sensing products including QuikScat derived winds and satellite imagery to provide a coherent observational model of marine ET. The troposphere will be scrutinized in terms of surface winds and frontogenesis, mid-level frontogenesis and trough interaction, and upper-level PV anomalies and jet interactions with the TC. It is commonly accepted that the midlatitude circulation into which the TC moves determines the final outcome of the ET (Jones et al. 2003). In this analysis, the midlatitude circulation not only influences the intensity and outcome of ET, it also determines the structure and evolution of low-level cyclones and fronts. First, it is hypothesized that the large-scale circulation interaction during marine ET often spawns a Shapiro-Keyser (1990) type extratropical cyclone with the observed properties of frontal fracture and warm core seclusion. This extratropical cyclone may deepen significantly after ET as in the case of many storms including Irene (1999) and Kate (2003). A second observed extratropical transition does not mimic the Shapiro-Keyser lifecycle but maintains a hybrid tropical structure as evidenced by Felix (1995) (Thorncroft and Jones, 2000) and Fabian (2003) exemplified by weak frontal structure and baroclinic interaction.

3.2 Introduction to Structural Changes

As a northern hemisphere tropical cyclone moves northward and begins to interact with the baroclinic environment (westerly winds, increased vertical shear, lower sea surface temperatures (SST) or strong SST gradients), it evolves from an
axisymmetric warm core tropical cyclone into an asymmetric cold-core extratropical cyclone. In addition, the translation speed of the cyclone increases dramatically under the effects of the midlatitude steering currents. Precipitation and strong winds near the center of the tropical cyclone circulation become distinctly asymmetric and often expand greatly in area due to interaction with synoptic features including midlatitude cyclones and upper-level features such as troughs and jet streaks. Interaction with these features, especially in the presence of vertical wind shear, contributes to loss of symmetry in a tropical storm and frontogenesis (Evans and Prater-Mayes 2004). The tropical cyclone may continue to weaken under the effects of the baroclinic environment, merge with an existing extratropical cyclone, or reintensify into its own extratropical cyclone.

Klein et al. (2000) develop a conceptual model of the ET transformation stage from a composite study of 30 North Pacific typhoons. Evans and Hart (2003) prefer the
terminology “onset of transition” rather than “start of transformation” but essentially reference the same process of asymmetric TC evolution. The transformation stage begins as the tropical cyclone moves over lower SSTs with its outer circulation impinging upon a preexisting baroclinic zone (Fig. 3.1). A dipole of (cold/warm) advection of (dry equatorward / moist poleward) moving air develops (west/east) of the TC. Typically a dry-slot forms in the southwestern quadrant of the storm as deep convection decreases over the western quadrant. However, warm moist air advected poleward maintains the deep convection over the northern and eastern quadrants of the storm. The poleward advected flow turns cyclonically and ascends over the tilted isentropic surfaces inherent with the preexisting baroclinic zone producing a region conducive for warm frontogenesis (Harr and Elsberry, 2000). This asymmetry in TC cloud patterns is acknowledged as an early indicator of transformation (Klein et al., 2000).

During step 2 of the Klein et al. (2000) conceptual model (Fig. 3.1), the lower tropospheric temperature advection dipole accentuates the baroclinic zone as well as expands the dry slot over the southern quadrant. The ascending poleward flow of warm moist air turns cyclonically and subsides into the western quadrant. A vertical motion dipole exists with dry adiabatic descent west of the storm center with ascent east of the center. The increased vertical wind shear south of the upper-tropospheric jet begins to ventilate the storm by advecting the top of the upper-tropospheric warm core downstream. Deep convection may still persist in the inner-core even as mid-level westerlies envelop the weakening midtropospheric warm core.

During step 3 of the conceptual model (Fig. 3.1), the storm becomes embedded in the baroclinic zone with an increase in vertical wind shear, decrease in SSTs, and a temperature advection dipole. The dry-adiabatic descent west of the storm center continues to erode the inner-core convection with only a weak warm core at low-levels remaining. A broad multilayer cloud mass exists on the poleward side of a warm front with a weaker cloud band to the southeast that resembles a cold front (Fig. 3.2). The broad cirrus shield with a sharp edge implies confluence between the TC outflow and the upper-level jet (Bader et al. 1995). Harr and Elsberry (2000) explain the continuing convection and cirrus shield to the north and northeast as the commencement of warm frontogenesis, a more vigorous process than cold frontogenesis in most ET cases. Ascent northwest of the storm center is undercut by dry adiabatic descent originating from the poleward advected environmental flow as part of the eastern vertical motion dipole branch. This poleward flow of warm, moist air ascends over the tilted isentropes and joins a strong southwesterly jet aloft that resembles the warm conveyor belt of an extratropical storm (Carlson 1991).

It is the interaction of the TC circulation with the baroclinic zone and its associated vertical wind shear that initiates the transformation stage of ET. However, not all TCs that enter the transformation stage and may dissipate due to extreme vertical shear or vortex spin-down over cooler SSTs (Eastern Pacific hurricanes commonly dissipate quite quickly over cooler waters without experiencing unfavorable vertical wind shear). Furthermore, TCs that enter the transformation stage do not always complete transition. The definition of a “transitioned” tropical cyclone implies that it is indistinguishable from an extratropical cyclone; the system specifically develops a lower troposphere cold-core (Evans and Hart, 2003). Consequently, differences in TC transformation and ultimate completion of ET may depend upon the angle of entry of the TC into the baroclinic zone and the specific characteristics of the midlatitude circulation into which the TC translates (Hart and Evans 2004).
After completion of ET, an extratropical cyclone develops with an extensive warm frontal region but with an ill-defined cold front. The cold front is often suppressed due to a direct thermal circulation that includes the descent of cold air from upstream of the reintensifying cyclone (Jones et al. 2003). A constructive or destructive interaction with a preexisting midlatitude system may either reintensify or decay the transitioned TC. Without interaction, the transitioned TC may continue to decay in terms of MSLP but continue to have strong winds in excess of hurricane force.

Figure 3.2: Schematic of midlatitude trough/TC interaction during ET. A multilayered cloud shield in the warm frontogenetic region poleward along the trough is caused by ascent of warm, moist air over the tilted isentropes of the baroclinic zone. Cool, dry air from behind the midlatitude cold front entrains into the southwest quadrant of the storm creating a dry slot. The strongest surface winds are associated with the cool, dry air along the southern side added to the storm translation speed. (Adapted from Fogarty 2002).

3.3 Trough Interaction and Vertical Shear

As the tropical cyclone translates poleward, it often increases its forward motion, encounters lower SSTs, and vertical shear due to interaction with a midlatitude mid-level trough. The effects of vertical shear on tropical cyclones have been studied by many (Frank and Ritchie 2001; Ritchie and Elsberry 2001; Reasor et al. 2000; Hanley et al. 2001) with considerable attention paid to ET. Briefly, the TC core will weaken under strong environmental vertical wind shear because the upper-level warm core cannot be sufficiently maintained upright to support the surface low pressure. Due to the minimum in inertial stability at the top of the TC, the warm core aloft is easily advected downstream. The ventilation of the warm core under the effects of strong environmental wind shear will reduce the strength of the warm core aloft. The height of the maximum warm core is thus reduced while the lower level warm core is enhanced. Subsidence due to convergence between the environmental winds and the cyclonic circulation of the TC are responsible for the warm core enhancement at low levels along with a rise in sea-level pressure. The cyclonic circulation is weakened further aloft, which entails even lower inertial stability. This negative feedback may weaken the upper-level warm
core dramatically and erode deep convection. However, it is not well understood how midlatitude trough interactions weaken or strengthen a tropical cyclone’s circulation, yet numerical models attempt to resolve the issue (Ritchie and Elsberry 2001; 2003).

Hanley et al (2001) performed a composite analysis of TC-trough interactions and identified two instances where the trough favorably interacted with the TC and resulted in intensification. Most applicable to ET is the favorable interaction between an upper-level jet streak and the TC outflow. As the temperature gradient between the warm tropical cyclone outflow and the cold trough well to the west of the TC narrows, the intervening jet streak is enhanced, which in turn improves the upper-level TC outflow channels. Linear jet dynamics implies that ascent is favored beneath the right-entrance region of the jet streak where a conveniently located TC center would experience even greater divergence aloft and hence improved outflow. As a result of the small inertial stability aloft, the improved outflow will contribute to low-level convergence and hence cyclonic spinup of the TC (Hanley et al. 2001). This favorable interaction usually results in an increase in cyclone upper-level cirrus in satellite imagery due to the upper-level outflow enhancement. This situation is occurs when the trough is well away or ‘distant’ from the TC. As the trough nears, vertical wind shear weakens the TC by ventilating its warm core and creating an asymmetry in the cloud/precipitation patterns. Ascent is favored in the downshear left quadrant of the tropical cyclone coupled with the formation of a dry slot upshear in response to forced subsidence (Ritchie and Elsberry 2001). Weakening of the vortex usually results.

3.4 Paradigms of Cyclone Development

Harr and Elsberry (2000) note that many studies (Thorncroft et al. 1993; Evans et al. 1994; Shultz et al. 1998) have defined variability associated with the structures of maritime cyclones that depend on the dynamics of the large-scale circulation. Their examination of two transitions of Western Pacific Typhoons discusses the evolution of structural characteristics of a TC during ET with respect to two different baroclinic environmental influences. The use of vector frontogenesis functions (Chapter 5) describes the frontal evolution during ET with emphasis upon the interaction between the midlatitude baroclinic zone and the TC circulation. However, their analysis focuses on the mid-level tropospheric interactions (500hPa) where they found the largest magnitude of structural changes. Shultz et al. (1998) consider the effects of large-scale flow on low-level frontal structures in marine midlatitude cyclones through both observational and numerical approaches. Different upper-level flow regimes are found to produce variable distinct cyclone/frontal structures and evolutions, including the Norwegian and Shapiro-Keyser (Fig. 3.3) cyclone models. It is the latter of a large spectrum of different cases that is closest related and observed most often during marine ET.

The Norwegian cyclone model (Bjerknes 1919; Bjerknes and Solberg 1922) has proven to be a limited and inadequate explanation of possible cyclone/frontal structures and evolutions (Browning 1990; Shapiro and Keyser 1990; Evans et al. 1994; Bosart 1999). The model consists of a disturbance on the polar front that forms a warm/cold air advection dipole and creates warm and cold fronts. The occlusion occurs when the cold front catches up to the warm front, a fundamental difference with the Shapiro-Keyser model. A Norwegian model cold front is typically meridionally oriented and
much stronger than the warm front. This is based upon the zonal index cycle of jet stream oscillation from weak westerlies in a high-amplitude planetary wave pattern (low zonal index) to strong westerlies in a low-amplitude planetary wave pattern (high zonal index) and back again (Rossby and Collaborators 1939; Rossby and Willett 1948; Namais 1950). Rossby and Willet (1948) characterized low zonal index by deep occlusions with north-south orientation of frontal systems with maximum east-west temperature contrasts. The opposite configuration holds for periods of high zonal index with east-west oriented frontal systems and latitudinal temperature gradients. The former case usually occurs within diffluent blocking patterns downstream of the cyclone whereas the latter is associated with strongly confluent flow (Saucier 1955). Fundamentally, it is the deformation pattern associated with each pattern that favors either meridionally or zonally oriented fronts with their own respective thermodynamically forced secondary circulations. Sawyer (1950) noted that warm occlusions tend to occur within the jet stream entrance region where confluence deformation and thermal fields exist typically in high zonal index flow.

Figure 3.3: Shapiro-Keyser (1990) frontal-cyclone evolution: incipient broad-baroclinic phase (I), frontal fracture (II), bent-back front and frontal T-bone (III), and warm-core frontal seclusion (IV), Upper: sea level pressure (solid), fronts (bold), and cloud signature (shaded). Lower: temperature (solid), and cold and warm air currents (solid and dashed arrows, respectively).

3.5 Shapiro-Keyser (1990) Cyclone Model

The Shapiro-Keyser (1990) cyclone model can be partially utilized to describe marine ET (Fig. 3.3). Instead of a small-amplitude disturbance over a broad low-level baroclinic zone, ET deals with a tropical cyclone vortex impinging upon the pre-existing baroclinic zone. Warm frontogenesis occurs when warm, moist air is advected equatorward and ascended over the baroclinic zone. A weak cold front with little convection is created by the poleward advection and descent of cold-dry air around the southwest periphery of the TC. The weak cold front and strong warm front move nearly perpendicular to each other and form a frontal T-bone (II). A frontal fracture appears in the horizontal temperature gradient along the poleward portion of the cold front near the low center first associated with subsidence and frontolysis by Godske et al. (1957).
Schultz et al. (1998) note that differential adiabatic warming may weaken the low-level temperature gradient or differential rotation of the isentropes in absence of vertical motion may create frontal fracturing. The baroclinicity in the warm-front region is transported westward relative to the cyclone center forming a bent-back warm front (III). The bent-back warm front wraps around the low center enclosing a pool of relatively warmer air and forms a warm seclusion (IV) (Schultz et al. 1998). A noticeable difference between the Norwegian and Shapiro-Keyser models concerns the maintenance of a nearly perpendicular orientation of the weak cold front and strong warm front as discussed in the latter model. The end result of each observational paradigm is very different in that the Norwegian produces a cold-core occlusion as opposed to the warm-core seclusion of the Shapiro-Keyser lifecycle.

3.6 LC1 and LC2 Baroclinic Lifecycles

A complimentary view of the relationship between cyclone/frontal structure and the upper-level flow concerns the introduction of barotropic shear to the basic-state zonal flow. Thorncroft et al. (1993) (THM) simulates two distinct cyclone lifecycles LC1 (basic state with no barotropic shear) and LC2 (basic state with cyclonic barotropic shear) that roughly parallel the Norwegian and Shapiro-Keyser model lifecycle paradigm. Schultz et al. (1998) mention limitations to applying this barotropic shear model to the real atmosphere yet acknowledge its usefulness to broadly conceptualize upper-level flow and cyclone-frontal structure. This is just one of the many observational and theoretical studies that identify distinct large-scale flow patterns that lead to different modes of cyclone development (Davies et al. 1991; Sinclair and Revell 2000; Schultz et al. 1998). [For a complete analysis of cyclone life cycle characteristics discussed below, the reader is referred to Thorncroft et al. (1993).]

3.61 LC1 and LC2 Types

The LC1 is more comparable to the Norwegian lifecycle with strong temperature gradients in the cold frontal region and bent-back baroclinicity surrounding the low-pressure region (THM). The cold front eventually pinches off the warm sector, which decreases in area reminiscent of a Norwegian occlusion. Eventually, the baroclinicity in the midlatitudes is destroyed and the surface pressure elongates zonally. From a PV point of view, before the occlusion, the wave tilts westward with height in the NW-SE direction on the cyclonic side of the mean jet (THM). A pronounced cyclonic-wrap of the system continues until anticyclonic shear of the poleward isentropic flow develops. Mean anticyclonic shear untilts the wave to a NE-SW direction and barotropic decay ensues as poleward momentum fluxes are returned to the jet (THM). This process, described as “trough thinning” as cross θ-contour flow destroys the gradient, is analogous to equatorward “Rossby-wave breaking” (McIntyre and Palmer 1984).

The NW-SE tilt of LC2 is stronger than in LC1 due to the effects of stronger cyclonic mean shear. The surface pressure tilts more definitively in the NW-SE direction with the strongest temperature gradients in the warm frontal zone (Hoskins and West 1979). Similarly, a warm-core seclusion occurs as baroclinicity associated
with the extended bent-back warm front encircles the low-pressure center. THM notes that surface cyclone is elongated more zonally than the LC1 and confined more meridionally. The strongest pressure gradients exist on the northern and western sides of the storm causing strong winds. However, instead of anticyclonic shear dominating south of the mean jet, cyclonic shear continues to wrap up the wave with uniform $\theta$ inside and strong $\theta$-gradients outside the mesoscale vortex. Cyclonic wrap-up on the poleward side of the jet expands to a larger degree than LC1 with the PV-contours near the jet core remaining largely undular.

### 3.7 Jet Dynamics

Sinclair (2004) uses EOF analysis to identify significant midlatitude circulation patterns associated with ET and assess the impact upon the intensity and structure of the system. He hypothesized that a period of coupling with the divergent quadrant of an upper-level jet is a required condition for extratropical redevelopment of a tropical cyclone. McTaggart-Cowan et al. (2003) also examined the ET of Hurricane Earl (1998) and Hurricane Danielle (1998) and differentiated between two different modes of development based upon the TC remnant’s position relative to the upper-level jet. It is well known that the secondary or vertical circulation causes the typical distribution of convergence and divergence associated with a jet (Fig. 3.4). Anticyclonic (cyclonic) maximum relative vorticity is found on the anticyclonic (cyclonic) side of the jet. A maximum of positive (negative) vorticity advection is found in the left (right) exit and right (left) entrance regions. Along with the attendant ageostrophic circulation, convergence (divergence) is found in the right (left) exit and left (right) entrance regions of the jet.

Figure 3.4: Idealized jet streak model showing convergence associated with low-level horizontal ageostrophic circulation (arrows) and positive (negative) vorticity advection PVA (AVA). Shaded areas show cold (warm) temperature advection in the entrance (exit) regions. (Adapted from McTaggart-Cowan et al. 2003).
3.8 Potential Vorticity (PV) Point of View

The use of potential vorticity provides important insight into the interactions between the tropical cyclone and midlatitude circulation during ET. Agusti-Panareda et al. (2004) model and analyze the ET of Hurricane \textit{Irene} (1999) and create a PV framework/conceptual model that illustrates many important points. In order for extratropical cyclogenesis to initiate after ET, especially reintensification, the surface low and upper-level trough must tilt westward with height necessary for extraction of potential energy from baroclinic instability. The approach of the tropical cyclone to the midlatitude environment can precondition the upper-level trough and enhance the upper-level jet and thus create a favorable cyclogenetic environment. Three important anomalies associated with the moist convective processes of a TC are critical to the ET process: (1) negative PV anomaly associated with TC upper-level outflow and (2) positive PV anomaly and (3) moisture anomaly co-located with PV tower cyclonic flow and positive thermal anomaly (Fig. 3.5). Emphasis will be placed upon the features of upper-level PV anomalies responsible for the structure and evolution of a post-transition TC extratropical storm.

3.9 Upper-Level Preconditioning of Midlatitude Environment by Transitioning Tropical Cyclone

The TC outflow at upper-levels (5, Fig. 3.5) causes a large-scale negative PV anomaly that can be advected very easily into the extratropical environment where it can be identified as a region of tropopause lift (Agusti-Panareda et al. 2004). In their analysis of Supertyphoon Flo (1990), Merrill and Velden (1996) indicate that PV decreases within the isentropic layers associated with the upper-level outflow of the TC. Following Hoskins et al (1985) and Haynes and McIntyre (1987), the effect of heating due to condensation above the environmental tropopause is to produce a PV source below the heating and a sink above. Due to a decrease of heating with height, tropical cyclones are considered an upper-tropospheric sink of PV (Schubert and Altvorth 1987). PV is approximately conserved outside of the convective region for individual parcels so the outflow from tropical cyclones would therefore have relatively low PV values as the effects of the sink are spread by advection (Wu and Emanuel 1994). The effects of vertical wind shear on the TC upper-level warm core discussed in Section 3.3 provided a description from the point of view of trough interactions, which is consistent with this explanation. When this negative PV anomaly is advected into the vicinity of a positive PV anomaly associated with an upper-level trough, it can enhance the PV gradient by steepening the tropopause and hence increasing the intensity of the upper-level jet (Agusti-Panareda et al. 2004). This favorable trough/TC interaction is fundamental to determining if/when/where extratropical cyclogenesis will occur.

As the TC nears an upper-level jet, it enters into a vertical shear environment. The top of the PV tower and co-located moisture tower is advected downwind analogous to the discussion in Section 3.3. At the surface, the PV remains relatively intact allowing for intermittent bursts of convection to be maintained along the warm
frontal region. Agusti-Panareda et al. (2004) indicate that the burst of convection generated a new diabatic PV tower comparable to observations of extratropical systems by Rossa et al. (2000). This reinvigorated PV tower maintained and enhanced the negative PV anomaly aloft despite the decay of the tropical cyclone PV tower primarily due to the advection of very moist air poleward near the warm frontal region. As the TC moves poleward, the negative PV anomaly approaches the positive PV anomaly of the upper-level trough and creates a meridionally oriented PV dipole, which strengthened the upper-level jet by steepening the tropopause. As the latent heat release wanes with the decay of the PV tower, the production of negative PV ends and the upper-level negative PV anomaly is advected downstream and sheared, which in turn flattens the tropopause and reduces the upper-level jet strength (Thorncroft and Jones, 2000; Agusti-Panareda et al. 2004).

Rapid cyclogenesis may occur if the upper-level trough untilts sufficiently from the upper-level jet consistent with Petterssen-Smybe type B development (Petterssen and Smybe 1971). Thus for baroclinic instability to develop, the upper-level PV anomaly associated with the trough and the surface thermal anomaly must tilt against the shear. This is essentially a favorable trough interaction described by Hanley et al. (2001) where the surface cyclone lies immediately to the east of the upper-level trough. The upper-level outflow of a TC is highly divergent, which amplifies the upper-level jet and engineers a cyclogenetic feedback process.

Figure 3.5: (Agusti-Panareda et al. 2004) Vertical cross section schematic with potential vorticity (PV) anomalies and other anomalies associated with extratropical transition: (1) surface thermal anomaly on baroclinic zone, (2) diabatically-generated positive PV anomalies along the baroclinic zone, (3) positive PV anomaly associated with a midlatitude upper-level trough, (4) TC positive PV tower, (5) negative PV anomaly associated with the tropical-cyclone’s outflow. The funny arrow represents the upper-level jet, which is modulated by the horizontal gradient of PV at upper-levels, i.e. the steepness of the tropopause.
3.10 Cyclone Phase Space Diagnostics

The cyclone phase space (CPS, Hart 2003) is a three-dimensional continuum that describes the frontal and thermal structure of synoptic scale systems. Symmetry is defined in the phase diagnostics by the difference in 900-600 hPa thickness across the storm relative to storm motion quantified by

\[ B = (Z_{600} - Z_{900})_{\text{warm}} - (Z_{600} - Z_{900})_{\text{cold}} \]  

where a thickness difference of 10 m was determined empirically to define the onset of ET (Evans and Hart 2003). ET completion is characterized by the first point in time after the storm has become both asymmetric and cold-cored; the sign of the thermal wind in the 900-600 hPa layer defines the core structure, where a negative slope indicates a warm-cored system and a positive slope indicates a cold core (Evans and Hart 2003). Determination of the thermal structure of the storm is related to the vertical structure of the cyclone’s height perturbation, which follows from the hypsometric relationship discussed more fully by Hirschberg and Fritsch (1993). If the magnitude of the cyclone isobaric height gradient increases (decreases) with height above the surface, by thermal wind relations, the cyclone is defined to be cold (warm) core. To differentiate between tropical and extratropical systems in the continuum of possibilities, two layers of equal mass (900-600 hPa and 600-300 hPa) are used to calculate the height perturbation and thus the thermal wind. The strength and depth of the thermal structure is easily estimated from the CPS. See example below of the *Ocean Ranger* storm (Fig. 3.6).

Figure 3.6: Cyclone Phase Space (Hart 2003; Evans and Hart 2003) for the *Ocean Ranger* storm of 1982. The left plot illustrates the frontal asymmetry (B) and thermal structure (\(-V_T\)) in the 900-600 hPa layer. The right plot corresponds to the overall thermal structure of the storm from deep cold core to moderate warm core and back again. Figure obtained from http://moe.met.fsu.edu/cyclonephase.
3.11 Précis

The observational and theoretical cyclone paradigms described above are antipode discrete examples of a vast variety of cyclone structures and evolutions. The differences between the Norwegian lifecycle and the Shapiro-Keyser cyclone evolution are largely a function of observation location. Bjerknes and Solberg (1922) observed cyclones in the diffluent region of the upper-level storm-track over Western Europe. Thus, their conclusions relied heavily upon understanding the circulations inherent with primarily low zonal index flows and often associated blocking patterns. The Experiment on Rapidly Intensifying Cyclones over the Atlantic (ERICA, December 1988 – January 1989) examined many cyclones with various frontal structures and evolutionary characteristics. The Shapiro-Keyser (1990) model resulted from a perceived inadequacy in the Norwegian model to explain the observed patterns of cyclone lifecycles explained in Section 3.4. The theoretical paradigms of Thorncroft et al. (1993) model antipode cyclone lifecycles also based upon upper-level flow and found many similarities between the previously mentioned paradigms. However, as a collection, all of them still fail to define the vast continuum of cyclone lifecycle evolutions. Extratropical transition is a primary example of where the paradigms fail for a couple major reasons to be addressed in the forthcoming chapters. Primarily, the presence of a powerful tropical cyclone vortex impinging upon the baroclinic zone and its ability to precondition the midlatitude environment into which it is moving is not described in either the observational or theoretical cyclone paradigms. The Ocean Ranger storm (Fig. 3.6) is characteristic of a strong wintertime Nor’easter yet has a very similar lifecycle to the post-transition TC: a warm seclusion. Conversely, the final outcome of the extratropical transition of Floyd (1999) is a cold-core occlusion reminiscent of the Norwegian lifecycle.

Note: Schultz et al. (1998) studied the effects of different upper-level flow upon low-level frontal structure, which highlight the differences between Norwegian and Shapiro-Keyser type systems. Idealized simulations using a nondivergent barotropic model illustrate the different lifecycles to a good approximation.
CHAPTER 4
UPPER-LEVELS

Evans and Prater-Mayes (2004) examined the extratropical transition of Hurricane Irene (1999) and concluded that an upper-level jet streak and trough contributed to the cyclogenesis ahead of the storm and affected the post-transition intensification. This supposition will be explored with Kate (2003) and Fabian (2003). The interaction with the midlatitude trough is more fully explored in Chapter 5 with the aid of vector frontogenesis functions. Here, the upper-level features near jet are examined in the PV framework of Agusti-Panareda et al. (2004) as well as the cyclone-lifecycle paradigms. The ET of Irene (1999) is well documented by the authors mentioned earlier, but the extratropical phase is not. Thus, attention will be focused on the evolution of the upper-level pattern with special emphasis paid to the resulting cyclone/frontal structure.

4.1 Hurricane Irene (1999)

The potential vorticity (PV) framework above developed by Agusti-Panareda et al. (2004) discussed the extratropical transition of Hurricane Irene (1999). They concluded that indeed the upper-level outflow associated with the transformation of Irene enhanced divergent flow downstream and resulted in significant extratropical development. The analysis that follows will focus more upon the upper-level jet changes, evolution of PV and thermal anomalies near the tropopause, and interaction with the midlatitude trough. The effects on the baroclinic zone will be briefly mentioned since the cross-sections allow for easy inspection of its evolution. Additional attention will be paid to the extratropical stage of Irene and the subsequent mature stage, which is indicative of a Shapiro-Keyser (1990) warm core seclusion and LC2 baroclinic lifecycle. Isentropic PV maps on the 325 K surface and vertical cross-sections both zonally and meridionally oriented will best show the interaction between the transforming vortex and the midlatitude circulation.
4.11 Jet

The upper-level jet at 300 hPa (Fig. 4.1) is characterized as high zonal index with anticyclonic shear poleward of the strong subtropical high over the central Atlantic. Irene reaches the right entrance region of the jet at 19/00z (Fig. 4.1a) and remains in a favorable area for extratropical cyclogenesis with dynamically forced ascent. However, the strength of this jet is preconditioned by the approaching tropical cyclone, which is visible on north-south vertical cross-sections of PV and normal component winds (Fig. 4.2). This depiction nicely captures the interaction between baroclinic zone, upper-level PV anomalies, and upper-level jet all situated northward of Irene at about 50° N. At 17/12z (Fig. 4.2a), the moist convective tower of Irene is less that 1500 km away from the baroclinic zone and the associated positive PV anomaly of the trough. During the next 48 hours Fig. 4.2b,c), the upper-level jet remains strong with speeds > 70 ms⁻¹ primarily in the zonal direction (Fig. 4.1a). As the tower approaches associated with the dipping isentropes indicative of a warm thermal anomaly and high PV air, its outflow induces a negative PV anomaly downstream or poleward shown as a concave-up bump along the tropopause at about 42° N at 18/12z (Fig. 4.2b) and 46° N at 19/00z (Fig. 4.2c). As explained in the preamble to this PV analysis, a meridionally oriented PV dipole steepens the tropopause and strengthens the jet. Irene’s lateral interaction with the trough’s positive PV anomaly occurs to the northwest and is partially represented at 19/00z (Fig. 4.3a).

There are two positive potential-vorticity (PV) anomalies: one at low-levels from latent-heat release due to heavy rainfall in moist warm-core ascent and the other at upper-levels associated with a stratospheric air streamer or fold above the warm core. Hoskins et al. (1985) described this “action at a distance” baroclinic development as dominated by the mutual interaction of upper-level PV anomaly and a surface temperature anomaly, which acts like a positive PV anomaly itself. This interaction is facilitated by the latent heat release, which reduces the static stability and thus increases the Rossby penetration depth (Browning et al. 1998).

As Irene couples with the midlatitude circulation at approximately 19/00z, a definitive upstream tilt with height develops indicative of the potential for baroclinic instability. With dry air intruding from the north into the southwestern quadrant of Irene, strong convection and latent heat release continues along the baroclinic zone associated with the warm frontogenetic region. As Irene approached, the positive PV anomaly associated with the trough along the baroclinic zone increases in magnitude and spatial scale closer to the surface. Diabatically forced convection continues in this region (45°-50° N Fig. 4.2c) due to the remarkable low-level jet (>30 ms⁻¹) advection of warm, moist air (high θe content Fig. 4.3a) up and over the baroclinic zone. Irene amplifies this baroclinic zone on its own with the poleward advection of warm moist air and the analogous equatorward cold advection. As Irene superimposed its deformation pattern upon the baroclinic zone, warm and cold advection patterns were markedly accentuated. The warm conveyor belt (Carlson 1991) is an important player in
the resurrection of the PV tower, this time more identifiable as an extratropical frontal process as the tropical convection is eroded. With the upstream tilt to the NW of the new tower at 19/12z (Fig. 4.3b), rapid cyclogenesis commences.

A time series of isentropic PV maps along the 325 K surface (Fig. 4.4) illustrates the process outlined above in terms of lateral interactions near the jet level. The 325 K isentrope, as seen in the latitudinal and zonal vertical cross-sections, intersects the jet streak at around 300 hPa, a fair representation of the midlatitude circulation (Bluestein 1992). From 19/00z to 19/12z, it is clear that a LC2 cyclonic wrap begins as Irene moves onto the cyclonic shear side of a developing upstream jet streak (Fig. 4.1c). High-PV air with stratospheric origin begins to fold around the circulation. This frontal fracture, indicative of a Shapiro-Keyser cyclone lifecycle, is commensurate with the initialization of the LC2 cyclonic wrap-up. The frontal fracture in this case is associated with the dramatic intensification of the upstream jet due to a marked PV gradient. This advection of cold, dry stratospheric (high PV) air associated with the high-PV tongue trails Irene from 19/12z (Fig. 4.1c) to 19/18z (Fig. 4.1d). In this case, Irene actually enters into an anticyclonic shear environment south of the jet streak ahead of the midlatitude trough. This is not a particularly favorable configuration for the consequent LC2 cyclonic wrap, but more of a LC1 trough thinning variety. Thus, it is hypothesized that the positive PV anomaly associated with the decaying tropical PV tower and the rapidly developing extratropical tower facilitate large-scale descent upstream in the trough region. The development of a frontal fracture allows high-PV from the upper-level trough to spill into the cyclone core. This in turn drives the intensification of the jet streak along the southern edge of the high-PV intrusion. Browning et al. (1998) noticed similar dry air intrusion with Lili (1995) due to a favorable trough interaction (Hanley et al. 2001). Frontal fracture will also be addressed in terms of its impact on vector frontogenesis and evolution of frontal features at midlevels and near the surface in Chapters 5 & 6.

4.12 Frontal Fracture and Cyclonic Roll-up

Matano and Sekioka (1971), Brand and Guard (1978), and Klein et al. (2000) among others have studied extratropical transition and identified dramatic dry air intrusions from upper-levels spilling into the cyclone core characteristic of LC2 lifecycles and frontal fractures. This dry intrusion (low-wet bulb $\theta_w$ air) is representative of descending stratospheric air to about 400 hPa. Again considering the vertical cross-sections, upstream high-PV air (blue contour > 5 PVU Fig. 4.3b) descends to about the 400-hPa level. The north-south viewpoint (Fig. 4.2e) fails to capture the intrusion since it is upstream but it does highlight the intensification of the upstream jet streak, which is becoming more zonally oriented at the base of the LC2 cyclonic wrap-up (Fig. 4.4d). It is interesting to note the timing of this intrusion. Browning et al. (1998) and Agusti-Panareda et al. (2004) both indicate that the dissolution of the tropical diabatic PV tower directly preceded the intrusion of high-PV air, which reinvigorated a new tower and commenced rapid cyclogenesis. The phasing with the midlatitude circulation...
in the framework of positive trough interaction as described by Hanley et al. (2001) seems to be the key to understanding exactly when the frontal fracture will occur and subsequent LC2 cyclonic wrap-up.

As mentioned before, the right entrance and left exit regions of a jet streak are favorable for cyclogenesis. At 19/06z, *Irene* is accelerated into a superposition region where the two regions overlap both upstream and downstream. However, as the positive PV tower approaches the upper-level positive anomaly downstream, the tropopause flattens and the PV gradient decreases (Fig. 4.2e,f). The tail end of the downstream jet weakens at 19/18z with the strongest isotachs rounding the base of another far downstream trough (Fig. 4.1d). The advection of the PV tower along the tongue of the 325 K PV sheet to the northeast quickly abates and *Irene* translates on a due west course. Where the upper-level PV gradient is greatest, the equatorward jet intensifies to > 70 ms\(^{-1}\) in a zonal direction.

The advantage of the vertical cross-sections in the north-south and east-west direction is apparent as the normal winds are easily analyzed. The poleward exit region of the jet is actually advected around the PV tongue at 20/00z (Fig. 4.4f) and seen on the east-west cross-section (Fig. 4.3c) as a poleward jet flow due to a steepened tropopause associated with the cyclonic wrap-up of high-PV air. [This weak poleward flow then cyclonic wraps around the northern periphery of the broad PV tower and helps facilitate the development of a broad back-door cold front as seen on the surface analysis (Fig.5.5).] The north-south cross-sections capture the intensification of the equatorward jet streak, which increases from 60 ms\(^{-1}\) at 19/18z (Fig. 4.2f) to well over 70 ms\(^{-1}\) at 20/00z (Fig. 4.2g). At the latter time, *Irene* is at its deepest central pressure of 944 hPa, well below its minimum pressure of 960-hPa as a category 2 hurricane off the North Carolina coast at 18/12z. The maximum sustained winds were 100 knots (NHC) at this time and fairly constrained to the center of circulation. It is apparent from the N-S cross-sections that strong winds in the extratropical stage are occurring beneath the upper-level jet, which is situated along bent-back warm front baroclinic zone (Chapter 6) at about 900 hPa with >50 ms\(^{-1}\) winds. This low-level jet is addressed later with QuikScat wind observations and satellite representations.

### 4.13 Warm Core Seclusion

The PV tower in both cross-section representations is no longer tilted at its mature stage (lowest sea level pressure) at 20/00z (Fig. 4.2g, 4.3c). Baroclinic instability wanes and is replaced by barotropic decay with the maximum upper-level stratospheric PV anomaly moves directly overhead of the surface low as seen on the 325 K isentropic surface at 20/00z and 20/12z (Fig. 4.4g,h). The PV magnitude begins to slowly decrease beneath as diabatic PV creation winds finally due to the effects of the dry-air intrusion. Browning et al. (1998) describe this cyclonic wrap-up that eventually encircles the cyclone core as an extensive three-dimensional PV sheet. At upper-levels (~300-hPa), the area of maximum
PV is located in the stratosphere and extends in a curved shape past the cyclone. At lower-levels (~600 hPa), the PV sheet remains nominally in the troposphere and wraps inward around the cyclone center. The cyclonic wrap-up is able to occur due to the greater translational speed of the PV sheet as compared to the storm motion as seen with Lili by Browning et al. (1998). By 21/00z (Fig. 4.4i), the PV aloft decreases in magnitude and expands spatially and interacts with a downstream upper-level PV anomaly over the British Isles. LC2 lifecycles take a while to slowly wind down in terms of central pressure and winds. It is also interesting to point out as a counterexample of LC1-type trough-thinning or equatorward Rossby-wave breaking occurring over the southeast United States associated with a digging trough. The winds associated with the warm core seclusion and bent-back warm front responsible for its development will be addressed later.

4.2 Hurricane Kate (2003)

The upper-level jet at 07/00z was highly amplified with Kate approaching the favorable right entrance region (Fig. 4.5a). The low-zonal index is a product of a strong subtropical high located over the central Atlantic and an equatorward-digging trough over eastern Canada, which produced a high-vertical shear environment. This facilitated the rapid ET of Kate in less than 24-hours between 06/00z and 07/00z as seen from the cyclone phase diagnostics (Fig. 6.2). However, the lateral interaction with the upper-level trough did not occur until later at 07/12z with only limited PV anomaly contact. This delayed the eventual extratropical cyclogenesis that deepened Kate from its final tropical minimum SLP of 987-hPa at 07/00z to 968-hPa at 09/00z. It will be shown in the PV-upper-level framework that Kate's extratropical cyclogenesis resulted from interactions with PV-anomalies aloft associated with a strong, meridionally oriented trough. Kate's baroclinic lifecycle is more reminiscent of an LC1-type (THM) with also contains many of the same features of the Shapiro-Keyser (1990) conceptual model. The importance of the dry-air intrusion due to the stratospheric extrusion of high-PV air will be shown to have developed an upstream jet streak that aided in the extratropical development of Kate through the PV-anomaly interactions discussed by Hoskins et al. (1985).

As Kate approaches the baroclinic zone, the downstream jet streak is enhanced (Fig. 4.5a,b) due to the steepening of the tropopause by the tropical cyclone upper-level outflow outlined earlier with the Irene example. The overall convective tower associated with Kate is weak (likely poorly resolved due to the system's small size) but is resurrected by the formation of a new extratropical tower (Fig. 4.6a-c). The strong core of the jet streak is still located to the north of tower at 08/00z (Fig. 4.6c) yet strong lateral interaction with stratospheric air with high-PV from the northwest begins to spill into the middle troposphere (Fig. 4.7b). The necessary westward tilt for cyclogenesis (baroclinic instability) continues throughout this frontal fracture process clearly shown by the tropopause fold descending to about 500-hPa at 08/06z (Fig. 4.7c). The upper-level positive PV
anomaly became vertically aligned with the tropospheric tower and barotropic decay ensued at 09/00z (Fig. 4.7e). Kate entered the warm seclusion phase and weakened fairly quickly over the next two days.

On the 325 K isentropic surface, Kate does not interact with the reservoir of upstream high-PV until 08/06z (Fig. 4.8e) coinciding with the frontal fracture phase and dry air intrusion. It is at this time as well that the upwind jet-streak intensifies to over 60 ms⁻¹, with Kate located in the left-exit region beneath dynamically forced ascent. There is a weaker superposition of favorable ascent regions related to the upstream and downstream jets since the latter one weakens and moves eastward at the same time as Kate begins to feel the effects of the dry air intrusion high-PV advection. The cyclonic wrap-up indicative of dry air intrusion advects Kate along this sheet of high-PV while at the same time wrapping around the cyclone core (Fig. 4.8e,f) from 08/06z to 08/12z. However, characteristic of a LC1 baroclinic lifecycle (THM), cyclonic wrap up continues until the anticyclonic behavior of the poleward isentropic flow of PV develops. This trough thinning process destroys the potential temperature gradient and causes the system to weak quickly. The sequence from 08/12z to 09/12z shows a moderate trough thinning, yet is clearly distinct from the rapid cyclonic wrap-up associated with Irene (LC2 lifecycle). Consequently, the storm becomes more elongated along a NE-SW axis as the warm seclusion process proceeds. Even though LC1 baroclinic lifecycles favor a Norwegian type occlusion, the frontal fracture is characteristic of the Shapiro-Keyser lifecycle, which predict the formation of a bent-back warm front and warm core seclusion. Thus, the extratropical phase of Kate features properties of different lifecycles proving the inability to simply categorize storms according to a certain scheme. Nature provides much variety especially in terms of cyclone evolution.

4.3 Hurricane Fabian (2003)

After undergoing ET, the PV tower associated with Fabian remained vertically aligned and coherent until encountering a broad warm-front deformation zone that zonally deformed the system. From the sequence of IPV-325 K maps (Fig. 4.9), Fabian is involved in the formation of large-scale cut-off low over the northern Atlantic from 08/00z to 10/18z. This is another clear example of LC2 baroclinic lifecycle but in this case, Fabian does not interact directly with the midlatitude circulation or upper-level trough in terms of exchange of PV, for instance. Instead, the poleward propagation of Fabian contributes indirectly to the LC2 process along the eastern periphery of a large trough over the northwestern Atlantic. Fabian does not cross the jet axis but pushes it northward near 70°N at the apex of a broad thermal ridge associated with warm advection. Moderate upstream jet intensification is noticeable (Fig. 4.10), which in turn transports Fabian westward around the periphery of the broad PV blob (Fig. 4.9d,e) from 09/00z to 10/18z. Little upper-level PV anomaly interaction occurs except for a weak intrusion of PV at 09/00z (Fig. 4.11b) that helps maintain the PV tower (Fig. 4.12). The trough interaction is not favorable in the Hanley et al.
(2001) sense, the PV tower does not tilt westward with height, and cyclogenesis does not initiate. Instead, slow weakening occurs until Fabian is deformed and absorbed by the cutoff low to the west. Another explanation for the weakening concerns Fabian's inability to precondition an upstream jet streak with its upper-level outflow and associated negative PV anomaly. Moreover, the weak jet is already downstream, a position not favorable for baroclinic instability.

4.4 Trajectories Through the Warm Seclusions of Kate (2003) and Irene (1999)

Kuo et al. (1992) modeled the airflow and thermal structure of an occluded marine cyclone named the Ocean Ranger storm and present a clear conceptualization of the process as well as a few hypotheses about low-level frontal structure that will be addressed forthwith. The use of relative stream trajectories on isentropic surfaces allows for easy determination of vertical motion assuming adiabatic motions. Thus, the origination of various air streams, whether ascending or descending, influences the thermal structure of the system and the low-level frontal structure. Browning et al. (1998) examined the evolution of Lili as it reintensified after extratropical transition in terms of a potential vorticity framework. The goal here is to tie the two conceptualizations together to create a clear picture of the upper-level airflow and PV anomaly interactions with the low-level frontal structure during the warm seclusion process.

The trajectories labeled A, B, C, and D (Fig. 4.13) are analogous to the warm and cold conveyor-belts described by Carlson (1980) and Browning (1986) representing spreading of rising air in different directions around the storm. The storm is embedded entirely in warm advection since all four trajectories veer (turn clockwise) with increasing height. The dry intrusion discussed as high-PV, low humidity air, and high static stability of stratospheric origin subsides as it advances on the center from the west and southwest depicted by trajectories E and F. After circulating around to the east of the center, airstream E is shown to rise consistent with dynamic lifting ahead of the upper-level trough or due to frontogenetical forcing along the occluded front (Chapter 6). In both cases, the airstream’s history is one of dryness and subsidence, which explains the pronounced dry slot in the mid- to upper-level cloud pattern ahead of the occluded front. Likewise, the lack of a pronounced cold front structure with billowy convection is a product of the high static stability, dryness, and history of subsidence of the air encircling the storm. F remains along a fairly level path owing its subsidence and drying to upstream processes behind the upper-level trough. Kuo et al. (1992) note that the thermal gradient aloft is too weak to characterize as a front, but rather as a confluence of two airstreams of widely different origins (Carlson 1980).
4.41 Irene Seclusion Trajectories

The 310 K isentrope corresponds roughly with the dynamic tropopause (PV = 2) at 19/12z (Fig. 4.3b) from the zonal cross-section analyses of the frontal fracture stage. As mentioned previously, the height of the tropopause is decreased on the cold side of upper jet as expected, but a region of locally even lower tropopause exists just west of the cyclone (Browning and Roberts 1994). As Irene initially developed beneath the right entrance region of the upper-level jet streak, there was a considerable band of high cloud associated with slantwise ascent north of Irene and collocated with the jet streak. As seen, the intrusion of dry, high-PV air from upstream at < 350-hPa descended into the warm boundary layer at roughly 750-hPa (Fig. 4.14 Top) a dual jet streak structure formed (Fig. 4.1c). The slantwise ascent slowly weakened beneath the right entrance region of the downstream jet. However, a new jet strengthened upwind due to the moisture and temperature gradient associated with the high-PV stratospheric air and frontal fracture. The dynamical tropopause slowly approached the low-level PV maximum associated with moist processes and later reached a state of vertical alignment. It is the transverse ageostrophic circulation at the exit of an upper-level jet that transports warm, moist air from the warm sector in the boundary layer from right to left across the jet axis towards the cloud head and beneath the dry intrusion air (Browning and Roberts 1994). Convective potential Instability results and convective cloud formation in the cloud hook region is clear visible (Fig. 6.6). It is this same ageostrophic circulation that forces the dynamic descent and dry air intrusion or frontal fracture. A feedback occurs as an upstream jet is rapidly strengthened, with Irene now in the left-exit region of the jet favorable for cyclogenetic development. It is during this process that Irene deepened the most rapidly.

The 300 K isentropic surface intertwines Irene on a path from near the surface to upper-levels past the storm as seen in the PV N-S cross-section (Fig. 4.2f). While not exact, the trajectories in the Kuo et al. (1992) model associated with the low- to mid-level conveyor belts match observations at 19/18z (Fig. 4.15b). Ahead of Irene, a fanning out of the airstreams is observed with a southern branch characterized by gentle ascent and quick downstream movement (A in Fig. 4.13), a second with marked ascent and weak veering (B), and a third representing the cold conveyor belt wrapping cyclonically around the hook cloud and the region of dry intrusion into the center of the storm (D). This process of hook-cloud formation occurred as the upper-PV anomaly associated with the frontal fracture was wrapped around the diabatically generated low-level PV maximum and is also indicative of LC2 type baroclinic waves with barotropic shear (THM). As this cyclonic wrap-up intensifies, the superposition of the cyclonic circulations associated with frontal fracture positive PV anomaly and the low-level positive PV anomaly spin-up the system. The vertical shear along the western and south side of Irene associated with the marked PV gradient tightens and effectively forces a cyclonic low-level jet around the low-level PV maximum.
(Fig. 4.14). The source of this baroclinicty and temperature gradient is associated with the strengthening bent-back front encircling the relative warm air around center compared to the environment immediately to the west. This jet will be shown to be associated with very strong winds, especially on the southern side where the storm’s motion adds to the system-relative circulation.

A similar trajectory analysis is shown for Kate (2003) (Fig. 4.16) on the 310 K (top) and 294 K (bottom) isentropic surfaces. The former corresponds roughly to the upper-level frontal fracture and associated conveyor belts while the latter more closely represents the low-level advection patterns often observed during the seclusion process. As an example, the seclusion seen on the 294 K lower-level surface plot (Fig. 4.16, bottom) is identified by a sharp thermal ridge and wind shift line separating areas of warm and cold advection at levels primarily below 700-hPa. The pinching action of air ahead of the baroclinic zone by the occluded front is similar to that described by Bjerknes and Solberg (1922) in their original paper on the life cycle of extratropical cyclones. In comparison, it is easily seen that Kate is becoming under the influence of anticyclonic shear associated with the upper-level jet, which is not only sending Kate to the northeast, but is elongating the storm in that direction. So, a weaker incarnation of cyclonic wrap-up is apparent preceding definite anticyclonic shear associated with the trough thinning and LC1-type waves (THM). Yet, the seclusion process occurs with Kate nonetheless but it is weaker and less pronounced than Irene. It is hypothesized that the cyclonic wrap up associated with Irene is much stronger than Kate owing to a stronger frontal fracture and intrusion of dry air associated with it. The weaker frontal fracture only marginally strengthens the upwind jet, through which secondary ageostrophic circulations forces a weaker vertical motion response. Kate does not remain in the relative favorable jet streak location but instead comes under the effects of anticyclonic shear. It weakens fairly rapidly afterwards in contrast with Irene, which remains a deep, cut-off low-pressure system for nearly a week.
Figure 4.13: From Kuo et al. (1992) schematic showing airflow through the modeled Ocean Ranger storm. Fronts are in open wave and occluded stages with the usual convention; fluffy cloud boundaries at mid- to upper levels; and relative trajectories, arrows (open where rising, shaded where sinking, hatched where level); delta scale for pressure level, lower right.

Figure 4.14: Conceptualization of warm seclusion of Lili from Browning et al. (1998). The thin solid line and stippled shading represent $\theta_w > 12^\circ$C. This feature is terminated at 500-hPa and capped by a perspective view lid to give a 3-dimensional impression. $J_U$ and $J_L$ respectively denote the upper- and lower-level jet axes with the bold line representing the dynamic tropopause PV = 2 isopleth. The wavy line corresponds to the model’s RH = 80% isopleth, which is interpreted by Browning et al. (1998) as the top of the actual region of cloud.
Figure 4.15: Relative airstream analysis of Irene (1999). Top: 310 K isentropic surface at 10/19 1200z. Bottom: 300 K isentropic surface. Both with wind vectors (ms$^{-1}$) and pressure level (shaded, hPa). Storm motion at 36$^\circ$ u = 14.5 ms$^{-1}$ and v = 10.7 ms$^{-1}$. 
Figure 4.16: Relative airstream analysis of Kate (2003). Top: 310 K isentropic surface at 10/08 0600z. Bottom: 294 K isentropic surface. Both with wind vectors (ms\(^{-1}\)) and pressure level (shaded, hPa). Storm motion at 75° \(u = 11.2\) ms\(^{-1}\) and \(v = 2.5\) ms\(^{-1}\).
Figure 4.1: Irene (1999) isotachs at 300 hPa are shaded according to scale (ms$^{-1}$) with location of surface low pressure center at (a) 10/19 00z (b) 10/19 06z (c) 10/19 12z (d) 10/19 18z (e) 10/20 00z (f) 10/20 06z (g) 10/20 12z (h) 10/20 18z and (i) 10/21 00z.
Figure 4.2: Hurricane Irene (1999) north-south vertical cross-sections of potential vorticity (PV) shaded (units 1 PVU = $10^{-6}$ m$^2$ K s$^{-1}$ kg$^{-1}$) and potential temperature $\theta$ (2K intervals) black labeled contours with normal component (10 ms$^{-1}$ intervals) of wind red labeled contours (solid is eastward flow, dashed westward) at (a) 10/17 12z (b) 10/18 12z (c) 10/19 00z (d) 10/19 06z (e) 10/19 12z (f) 10/19 18z (g) 10/20 00z and (h) 10/20 12z.
Figure 4.3: Same as Figure 4.2 except for east-west cross section with normal wind (v) component (solid red contours northward flow, dashed southward) and at (a) 10/19 00z (b) 10/19 12z with blue lines indicating 300 K and 310 K isentropic surfaces c.f. Figure 4.15 (c) 10/20 00z with similar blue line convention and (d) 10/20 12z.

Figure 4.6: Hurricane Kate (2003) north-south vertical cross-sections of potential vorticity (PV) shaded (units PVU) and potential temperature $\theta$ (2K intervals) black labeled contours with normal component (10 ms$^{-1}$ intervals) of wind red labeled contours (solid is eastward flow, dashed westward) at (a) 10/07 00z (b) 10/07 12z (c) 10/08 00z (d) 10/08 12z (e) 10/09 00z and (f) 10/09 12z.
Figure 4.4: Potential vorticity on the 325 K isentropic surface for Irene (1999) in units of PVU ($10^{-6} \text{ m}^2 \text{ K s}^{-1} \text{ kg}^{-1}$) with dashed isotachs ($10 \text{ ms}^{-1}$ contours): (a) 10/18 00z (b) 12z (c) 10/19 00z (d) 12z (e) 18z (f) 10/20 00z (g) 06z (h) 12z (i) 10/21 00z.
Figure 4.5: Same as Figure 4.1 for Kate (2003) except at (a) 10/07 00z (b) 10/07 12z (c) 10/07 18z (d) 10/08 00z (e) 10/08 06z (f) 10/08 12z (g) 10/08 18z (h) 10/09 00z and (i) 10/09 12z.
Figure 4.7: Same as Figure 4.6 except for east-west cross section with normal wind (v) component (solid red contours northward flow, dashed southward) and at (a) 10/07 12z (b) 10/08 00z (c) 10/08 06z with solid blue lines indicating isentropic surfaces of 310 K and 294 K corresponding to Figure 4.16 trajectories (d) 10/08 12z (e) 10/09 00z and (f) 10/09 12z.

Figure 4.11: Hurricane Fabian (2003) north-south vertical cross-sections of potential vorticity (PV) shaded (units PVU) and potential temperature $\theta$ (2K intervals) black labeled contours with normal component (10 ms$^{-1}$ intervals) of wind red labeled contours (solid is northward flow, dashed southward) at (a) 09/08 12z (b) 09/09 00z and (c) 09/10 00z.

Figure 4.12: Same as Figure 4.11 except for north-south cross section with normal wind (u) component (solid red contours eastward flow, dashed westward) and at (a) 09/08 00z (b) 09/09 00z (c) 09/10 00z.
Figure 4.8: Same as Figure 4.4 except for Kate (2003) (a) 10/06 12z (b) 10/07 00z (c) 12z (d) 18z (e) 10/08 00z (f) 06z (g) 12z (h) 10/09 00z (i) 12z.
Figure 4.9: Same as Figure 4.4 except for Fabian (2003) at (a) 09/08 00z (b) 12z (c) 09/09 00z (d) 12z (e) 09/10 00z (f) 18z.
Figure 4.10: Same as Figure 4.1 except for Fabian (2003) at (a) 09/07 12z (b) 09/08 00z (c) 09/08 12z (d) 09/09 00z (e) 09/09 12z and (f) 09/10 00z.
CHAPTER 5

VECTOR FRONTOGENESIS MIDDLE TROPOSPHERE

Petterssen (1956) defines frontogenesis as the tendency toward formation of a discontinuity in the density field or the intensification of an existing sloping transition surface. Conversely, frontolysis is indicative of a negative tendency. This discontinuity also can be described in terms of potential temperature: frontogenesis is also defined as the Lagrangian temporal derivative of the horizontal potential temperature gradient (Keyser et al 1988). The frontogenetic function

\[ F = \frac{d}{dt} \left| \nabla_h \theta \right| = -\frac{1}{2} \left| \nabla_h \theta \right| (E \cos \beta - \nabla \cdot \mathbf{V}) \]  

relates the potential temperature gradient with the horizontal divergence \( \delta = \partial u / \partial x + \partial v / \partial y \) and the resultant deformation \( E = (E_{st}^2 + E_{sh}^2)^{1/2} \). \( \beta \) is the angle between the orientation of the axis of dilatation and the isentropes. The individual components of the deformation are the stretching deformation \( E_{st} = \partial u / \partial x - \partial v / \partial y \) and the shearing deformation \( E_{sh} = \partial v / \partial x + \partial u / \partial y \). All derivatives are carried out using centered finite differences on constant pressure surfaces. Keyser et al. (1988) devised an alternative expression of \( \mathbf{F} \) in natural coordinates \((s,n)\) so that

\[ \mathbf{F} = F_n \mathbf{n} + F_s \mathbf{s}. \]  

The \((s,n)\) system is in right handed Cartesian coordinates with the \(\mathbf{n}\) axis defined to point in the negative direction of \(\nabla_h \theta\) (towards cold air), defined as \(\mathbf{n} = -\left| \nabla_h \theta \right|^\perp \nabla \theta\). The \(\mathbf{s}\) axis is defined as \(\mathbf{s} = \mathbf{n} \times \mathbf{k}\) and is locally tangent to the isentrope and points in the direction of the thermal wind. Thus, the \(F_n\) component is called the scalar frontogenesis and describes the magnitude of the Lagrangian rate of change of the horizontal \(\theta\) gradient: \(F_n = -\frac{d}{dt} \left| \nabla_h \theta \right|\). The \(F_s\) component defines the Lagrangian rate of change of the direction of \(\nabla_h \theta\):

\[ F_s = \mathbf{n} \cdot (\mathbf{k} \times \frac{d}{dt} \nabla_h \theta). \]
To reinterpret $F_n$ and $F_s$ in terms of invariant kinematic quantities, Keyser et al. (1988) considered the frontogenetic and rotational components in parallel along with the conservation of potential temperature for horizontal adiabatic flow:

$$\frac{\partial \theta}{\partial t} + V \cdot \nabla \theta = 0. \quad (4)$$

Schultz and Doswell (1999) expanded upon the formulation of Keyser et al (1988) to include the effects of vertical motion in the thermodynamic equation in their analysis of upper-level front evolution. At the surface, the effects of vertical motion are typically small when compared to mid- and upper-levels. The three dimensional version of thermodynamic equation is differentiated by $x$ and $y$ respectively yielding the expressions for $\frac{d}{dt} \nabla_h \theta$:

$$\frac{d}{dt} \left( \frac{\partial \theta}{\partial x} \right) = -\frac{\partial u}{\partial x} \frac{\partial \theta}{\partial x} - \frac{\partial v}{\partial x} \frac{\partial \theta}{\partial y} - \frac{\partial \omega}{\partial x} \frac{\partial \theta}{\partial p}, \quad \frac{d}{dt} \left( \frac{\partial \theta}{\partial y} \right) = -\frac{\partial u}{\partial y} \frac{\partial \theta}{\partial x} - \frac{\partial v}{\partial y} \frac{\partial \theta}{\partial y} - \frac{\partial \omega}{\partial y} \frac{\partial \theta}{\partial p}. \quad (5a), (5b)$$

Expanding the vector-frontogenesis components yields the intermediate result to be combined with the above equations:

$$F_n = -1 \nabla_h \theta^{-1} \left[ \frac{\partial \theta}{\partial x} \frac{d}{dt} \left( \frac{\partial \theta}{\partial x} \right) + \frac{\partial \theta}{\partial y} \frac{d}{dt} \left( \frac{\partial \theta}{\partial y} \right) \right] \quad (6)$$

$$F_s = -1 \nabla_h \theta^{-1} \left[ \frac{\partial \theta}{\partial y} \frac{d}{dt} \left( \frac{\partial \theta}{\partial x} \right) + \frac{\partial \theta}{\partial x} \frac{d}{dt} \left( \frac{\partial \theta}{\partial y} \right) \right]. \quad (7)$$

Combination yields the final expressions for vector frontogenesis components now including the tilting term:

$$F_n = -1 \nabla_h \theta^{-1} \left[ \frac{\partial \theta}{\partial x} \left( -\frac{\partial u}{\partial x} \frac{\partial \theta}{\partial x} - \frac{\partial v}{\partial x} \frac{\partial \theta}{\partial y} \right) + \frac{\partial \theta}{\partial y} \left( -\frac{\partial u}{\partial y} \frac{\partial \theta}{\partial x} - \frac{\partial v}{\partial y} \frac{\partial \theta}{\partial y} \right) + \frac{\partial \theta}{\partial p} \left( -\frac{\partial \omega}{\partial x} \frac{\partial \theta}{\partial x} - \frac{\partial \omega}{\partial y} \frac{\partial \theta}{\partial y} \right) \right], \quad \text{F}_n\text{-div}$$

$$F_s = -1 \nabla_h \theta^{-1} \left[ \frac{\partial \theta}{\partial y} \left( -\frac{\partial u}{\partial x} \frac{\partial \theta}{\partial x} - \frac{\partial v}{\partial x} \frac{\partial \theta}{\partial y} \right) - \frac{\partial \theta}{\partial x} \left( -\frac{\partial u}{\partial y} \frac{\partial \theta}{\partial x} - \frac{\partial v}{\partial y} \frac{\partial \theta}{\partial y} \right) + \frac{\partial \theta}{\partial p} \left( \frac{\partial \omega}{\partial x} \frac{\partial \theta}{\partial x} - \frac{\partial \omega}{\partial y} \frac{\partial \theta}{\partial y} \right) \right], \quad \text{F}_s\text{-def}$$

$$F_n = -1 \nabla_h \theta^{-1} \left[ \frac{\partial \theta}{\partial x} \left( -\frac{\partial u}{\partial x} \frac{\partial \theta}{\partial x} - \frac{\partial v}{\partial x} \frac{\partial \theta}{\partial y} \right) + \frac{\partial \theta}{\partial y} \left( -\frac{\partial u}{\partial y} \frac{\partial \theta}{\partial x} - \frac{\partial v}{\partial y} \frac{\partial \theta}{\partial y} \right) + \frac{\partial \theta}{\partial p} \left( -\frac{\partial \omega}{\partial x} \frac{\partial \theta}{\partial x} - \frac{\partial \omega}{\partial y} \frac{\partial \theta}{\partial y} \right) \right], \quad \text{F}_n\text{-tilt}$$

$$F_s = -1 \nabla_h \theta^{-1} \left[ \frac{\partial \theta}{\partial y} \left( -\frac{\partial u}{\partial x} \frac{\partial \theta}{\partial x} - \frac{\partial v}{\partial x} \frac{\partial \theta}{\partial y} \right) - \frac{\partial \theta}{\partial x} \left( -\frac{\partial u}{\partial y} \frac{\partial \theta}{\partial x} - \frac{\partial v}{\partial y} \frac{\partial \theta}{\partial y} \right) + \frac{\partial \theta}{\partial p} \left( \frac{\partial \omega}{\partial x} \frac{\partial \theta}{\partial x} - \frac{\partial \omega}{\partial y} \frac{\partial \theta}{\partial y} \right) \right], \quad \text{F}_s\text{-tilt}$$

The vector-frontogenesis equations (8) and (9) are modified to include tilting effects and are presented in their invariant form for easy interpretation:
\[ F_n = \frac{1}{2} |\vec{\nabla} \theta| \left[ \vec{\nabla}_h \cdot \vec{V} - E \cos 2\beta - \frac{\partial \theta}{\partial p} \left( \vec{\nabla}_h \omega \times \vec{\nabla}_h \theta \right) \right], \quad (10) \]

\[ F_s = \frac{1}{2} |\vec{\nabla} \theta| \left( \hat{k} \cdot \vec{\nabla}_h \times \vec{\nabla}_h + E \sin 2\beta \right) \frac{\partial \theta}{\partial p} \hat{k} \cdot \left( \vec{\nabla}_h \omega \times \vec{\nabla}_h \theta \right). \quad (11) \]

Deformation is frontogenetic if the angle between the axis of dilatation and the isentropes is less than 45° while convergence is frontogenetic no matter the orientation of the isentropes. The tilting term \( F_n \)-tilt is frontogenetic when the gradient of vertical velocity is in the same direction as the gradient of \( \theta \), thus providing a negative cross product. Rotational tilting \( F_n \)-tilt frontogenesis increases as the vertical velocity and \( \theta \) gradients become more perpendicular. When perpendicular, \( F_s \)-tilt rotational frontogenesis is at a maximum implying zero contribution to scalar frontogenesis by \( F_n \)-tilt. The rotational deformation term \( F_s \)-def contributes to the counterclockwise (clockwise) rotation of the potential temperature gradient when the isentropes are oriented between 0° and 90° counterclockwise (clockwise) to the axis of dilatation. In other words, deformation tends to rotate isentropes toward the axis of dilatation assuming their orientation is neither orthogonal nor parallel to the axis (Keyser et al 1988). Regardless of orientation, cyclonic (anticyclonic) relative vorticity contributes to the counterclockwise (clockwise) rotation of the potential temperature.

Several modeling studies have examined the kinematics of the interaction between a tropical cyclone-like vortex and a straight baroclinic zone (Doswell 1984; Keyser et al. 1988). The vortex model constructed to determine the evolution of a baroclinic zone in a nondivergent steady-state vortex may also be viewed as the evolution of low-level frontal zones. The long-term evolution of the potential temperature advection pattern forced by the vortex mimics an S-shaped baroclinic zone (Fig. 5.1) characteristic of the low-level thermal field of a mature midlatitude cyclone (Keyser et al. 1988). Keyser et al. (1988) investigated the changes in the potential temperature field in terms of the vector frontogenesis function (2) that consists of scalar (10) and rotational (11) contributions to frontogenesis. This function defines the Lagrangian rate of change of direction of the horizontal potential temperature gradient in terms of scalar and rotational quantities. Scalar frontogenesis acts upon the magnitude of the \( \theta \)-gradient while rotational frontogenesis acts to rotate the \( \theta \)-gradient. Scalar frontogenesis may occur in association with divergence, horizontal deformation, and tilting whereas rotational frontogenesis may occur in association with relative vorticity, horizontal deformation, and tilting (Keyser et al. 1988; Schultz et al. 1998; Harr and Elsberry 2000).
5.1 Analysis of Two Typhoons (Harr and Elsberry 2000)

Harr and Elsberry (2000) analyzed the structural characteristics of tropical cyclones during extratropical transition over the Western North Pacific. These structural changes of the evolving extratropical cyclone are manifested in the interaction between the tropical cyclone vortex and a baroclinic zone. For the two examples of Typhoon (TY) David (1997) and TY Opal (1997), vector-frontogenesis diagnostics are employed to ascertain the developmental characteristics of extratropical transition in terms of frontal formation. David and Opal both moved poleward ahead of midlatitude troughs yet did not reintensify into strong extratropical storms. The distinction between the northeast (NE) and northwest (NW) pattern of the midlatitude circulation during extratropical transition is used as a simplifying framework to identify the possible variations in the evolution of transitioning systems. The specific nature of the midlatitude circulation (NE/NW) was found to determine the overall pattern of frontal formation and the consequent evolution of the transitioning system (Harr and Elsberry 2000).

Harr and Elsberry (2000) utilized the frontogenetic partitioning to examine the frontal-scale and wave-scale forcing during ET as it related to the overall cloud pattern. As previously mentioned, their study is based upon the hypothesis that the position of the midlatitude circulation relative to the poleward propagating TC (NE/NW) influences the position and characteristics of the frontogenesis. The analysis showed that the maximum amplitude of structural changes in terms of impact on the development of the frontal characteristics defined by \( F_n \) and \( F_s \) were at the 500-hPa level. Moreover, diagnosing the interaction between the TC and the midlatitude circulation at mid-tropospheric levels best illustrated the need for positive dynamic coupling in order to achieve reintensification. For each of the two TCs, a consistent
overall pattern of scalar frontogenesis emerged with definitive differences in the rotational frontogenesis quantity.

As expected, the horizontal deformation term dominated the scalar frontogenesis term especially north of the cyclone. Here poleward of the TC, the axes of dilatation are oriented parallel to the isentropes indicative of warm front formation. Klein et al. (2000) conceptual model describes the prevalence of warm front formation as well as Sinclair (2002) who observed warm frontogenesis poleward of transitioning southwest Pacific cyclones. The specific magnitude, areal extent, and structure of the frontogenesis depend on the characteristic midlatitude circulation pattern into which the TC is moving (Jones et al. 2003).

The maximum rotational frontogenesis was dominated by a large region of positive $F_s$ along the eastern boundary of the midlatitude thermal trough, with two regions of weaker negative rotational frontogenesis along the baroclinic zone to the NE of the center. Most importantly, the region of positive rotational frontogenesis acted to rotate the $\nabla_h \theta$ in a cyclonic direction while the negative region of $F_s$ east of the center acted to rotate the $\nabla_r \theta$ in an anticyclonic direction. Therefore, the rotational frontogenesis component provided dynamical support for the amplification of the thermal wave via cold advection to the west and warm advection to the east of the TC center. Deformation dominated along the boundary between the thermal trough and the ridge, while anticyclonic relative vorticity dominated downstream from the TC center (Harr and Elsberry 2000). This pattern of rotational frontogenesis acting to amplify the thermal wave, coupled with scalar frontogenesis acting to enhance the horizontal potential temperature gradient, is indicative of reintensification of the TC (Harr and Elsberry 2000).

5.2 Midlatitude Interactions Hurricane Irene (1999)

As Hurricane Irene is positioned off Cape Hatteras, NC on 1800 UTC 18 October, 18 hours after commencing extratropical transition according to cyclone phase diagnostics (Fig. 6.1). Evans and Prater-Myers (2004) discuss the evolution of the upper-level trough from a weak, northeast-southwest tilt to an amplified, more neutral tilt. They conclude the interaction of the approaching 500-hPa trough with Irene provided the mechanism for upper-level cyclonic development throughout the transition process. Hanley et al. (2001) describe this as a positive trough interaction in which the upper-level trough is approximately 10° to the northwest of Irene, a position conducive for rapid midlatitude cyclogenesis according to the baroclinic wave development model best described by Carlson (1991). As the trough acquired a westward tilt with height from the surface low, it was positioned at the coast and resulted in enhanced baroclinicity and reduced stability over the warm Gulf Stream SSTs (Evans and Prater-Myers 2004). From the viewpoint of absolute vorticity and geopotential height at 500-hPa (Fig. 5.2), an environment conducive for extratropical cyclogenesis is easily diagnosed.
5.21 500-hPa Frontal Evolution

As Irene approached the baroclinic zone draped over the St. Lawrence River Valley, extratropical transition defined by the cyclone phase space diagnostics occurred very rapidly (Fig. 6.1). The characteristics of the midlatitude baroclinic zone at 500-hPa and at the surface are easily diagnosed by the vector frontogenesis functions defined in (8) and (9). Following the discussion of the transformation stages of TY Opal and TY David in Harr and Elsberry (2000), a complete frontal evolution from the onset of extratropical transition through transformation into an extratropical cyclone and then maturity as a warm core seclusion for three separate hurricanes [Irene (1999), Fabian (2003), and Kate (2003)] is presented. First, analysis in the mid-troposphere at 500-hPa will help characterize the midlatitude circulation / TC interactions for each storm. Then, a similar analysis at the surface after extratropical transition using QuikScat derived winds will help aid in understanding the low-level evolution of fronts.

5.22 Extratropical Transformation

During ET, Irene approaches the coastal potential temperature gradient and associated with the thermal trough and begins interacting. At 1800 UTC 18 Oct, the scalar frontogenesis $F_n$ (Fig 5.3a) associated with Irene is weakly frontogenetic along the coast with a broad frontolytic area to the north. Along the thermal trough centered over 75° W and the thermal trough eastward into the Atlantic, strong positive $F_s$ associated with the deformation pattern forced by the narrowing trough / approaching TC couplet acts to accentuate the $\nabla^2 \theta$. In addition, with the broad delta cloud shield poleward of the TC, and cyclonic vorticity advection around the base of the 500-hPa trough, vigorous ascent over the Canadian Maritimes (not shown) is coupled with descent upstream. The gradients of vertical velocity and $\theta$ pointed in opposite directions on the poleward side of the baroclinic zone. Therefore, $F_{n-tilt}$ was frontolytic in this region, which was consistent with a thermally direct circulation implied by the dipole of vertical motion. The deformation term $F_{n-def}$ dominated both the tilting term $F_{n-tilt}$ and the divergence term $F_{n-div}$, which was only significant in a small area of divergence associated with the diffluence over New England (not shown).
Figure 5.2: 500-hPa absolute vorticity (units $10^{-5}$ s$^{-1}$) and geopotential height (units m) for the times indicated. (a) Hurricane affecting southeast US. (b) Irene and the midlatitude trough located over the Great Lakes remain separated. (c) Vorticity due to warm frontogenesis north of Irene indicates interaction; the trough weakens slightly and becomes neutral. (d) The trough has become negatively tilted and amplified with an elongated vorticity maximum oriented from the southeast to northwest. (e) A large vorticity maximum continues to strengthen as the trough amplifies. (f) Extratropical system is at maximum intensity with cyclonic wrapping of absolute vorticity and well-defined warm core seclusion. (g) The trough begins to flatten as vorticity maximum broadens and weakens (h) and (i).
The rotational frontogenesis $F_s$ is dominated by a positive/negative couplet associated with the thermal trough/ridge pattern. The strong positive $F_s$ located over 70°W (Fig. 5.3b) along the eastern boundary of the thermal trough is associated with the trough’s cyclonic vorticity. This $F_s$-vor dominated region extends southward to the transitioning TC. Negative areas of $F_s$ are found along the baroclinic zone and west of Irene over Maryland. The larger areas along the boundary of the thermal trough and ridge of the baroclinic zone were dominated by the deformation term agreeing with Harr and Elsberry (2000). With regards to the tilting term, the dipole in vertical motion along the thermal trough led to gradients in $\omega$ and $\theta$ that are perpendicular. Harr and Elsberry (2000) show that the gradient of $\omega$ acted to rotate the $\mathbf{\nabla}_h \theta$ anticyclonically east of the center and cyclonically west of the center. The end result is a flattening of the thermal trough, which is interpreted as a progression of the pattern eastward. In total, the region of positive rotational frontogenesis to the NW of Irene acted to rotate the $\mathbf{\nabla}_h \theta$ cyclonically. Conversely, the negative rotational frontogenesis regions rotated the $\mathbf{\nabla}_h \theta$ anticyclonically. The rotational frontogenesis provided dynamical support for the amplification of the thermal wave via cold advection west of Irene and warm advection to the north and east.

At 0000 UTC 19 Oct, the patterns of $F_n$ (Fig. 5.3c) and $F_s$ (Fig. 5.3d) have increased in magnitude with the increased interaction of the TC. Strong positive $F_n$ is associated with the convergence and deformation patterns forced by the poleward acceleration of Irene against the baroclinic zone. Warm frontogenesis continues along the baroclinic zone along 50°N. The pattern of rotational frontogenesis shows a more complete interaction between the vorticity maxima associated with the trough and the TC. The surface low continues to move into a region of favoring explosive cyclogenesis as explained by Carlson (1991) and exemplified by the frontal evolution. The rotational component is dynamically reinforcing the thermal wave, while the scalar component increases the magnitude of the $\mathbf{\nabla}_h \theta$. The extratropical cyclone deepens from 968 mb to 944 mb from 1200 UTC Oct 19 to 0000 UTC Oct 20 according to the Met Office surface analysis (Fig. 5.5).

5.23 Frontal Fracture

Early in Irene’s rapid intensification phase, a definitive scalar frontolytic zone developed at the base of the sharp thermal trough immediately to the west of the surface reflection of the low center along 50°N (Fig. 5.3e). The accompanying rotational frontogenesis (Fig. 5.3f) shows Irene in an extremely favorable environment to dynamically couple with the thermal wave and rapidly intensify. Harr and Elsberry (2000) first identified this frontolytic zone as a possible frontal fracture in their investigation of TY David through a trajectory analysis. Browning et al. (1997) and
others have analyzed the frontal fracture, a term first coined by Shapiro and Keyser (1990) explaining that the previously continuous front fractured near the center of the deepening cyclone and the temperature gradients at the warm and cold fronts became sharp except in the region of the fracture. The fracture precedes the T-bone phase and warm core seclusion phase in the Shapiro and Keyser (1990) frontal-cyclone life-cycle model (Fig. 3.3). It is important to note that the interaction between the decaying tropical cyclone and the midlatitude baroclinic circulation does not agree with the incipient broad baroclinic phase of the Shapiro and Keyser (1990) model. Browning and Roberts (1994) developed a conceptual model that illustrates the interaction of split fronts, conveyor belts, dry intrusion, and the cloud head. The dry intrusion of upper-tropospheric and lower-stratospheric air with characteristic high potential vorticity (PV) and low wet-bulb potential temperature ($\theta_w$) corresponds to a tropopause fold.

Several ways of easily showing frontal fracture include vertical cross-sections of potential vorticity, vertical velocity, and equivalent potential temperature ($\theta_e$). The vertical distribution of potential vorticity and vertical velocity shows a well-defined tropopause fold with modest descent at midlevels to the surface (Fig. 5.4). Conversely, an anti-cyclonic PV anomaly and strong ascent is associated with the intense cloud head and bent-back warm front development. In addition, a distinct cooling and drying with the dry intrusion is noticed in the region of the frontal fracture (Fig. 5.4,5.5). An opposite configuration is located to the northwest with strong ascent in warm, moist air concomitant with the convection in the cloud head. This configuration (Fig. 5.4) leads to strong frontolysis at all levels in the troposphere in the region of the dry intrusion and considerable warm frontogenesis in a shallow layer near the surface indicative of the region of thick clouds poleward of the center. Also noticeable is a strong upper-level positive PV anomaly with frontolytic conditions in the upper-troposphere sloping back into the cloud head region of warm frontogenesis and convective ascent.

Figure 5.3: 500-hPa Vector Frontogenesis (contours of $2\times10^{-10}$ Km$^{-1}$s$^{-1}$) with scalar frontogenesis $F_n$ (left) and rotational frontogenesis $F_s$ (right) and $\theta$ (contours, K) at times indicated (a)-(l). Cross-section from Figure 5.4 indicated in (e) and (f).
Figure 5.3: Continued.
Figure 5.4: Vertical cross sections for 10/19 12z for the slice through Irene with (a) potential vorticity (shaded, PVU) and vertical motion (colored solid descent, dashed ascent hPa/s) (b) Vertical motion (black, as in (a)) with $\theta_e$ (colored contours, 5K) and (c) Total scalar frontogenesis (colored contours, units $2 \times 10^{-10}$ Km$^{-1}$ s$^{-1}$) with $\theta$ (black, 4K).

Figure 5.5: Surface analysis for (a) 10/20 00z and (b) 10/21 00z from UK Meteorological Office www.metoffice.com.
5.24 Warm Core Seclusion

At 20/12, the formation of the bent-back warm front clearly visible in the infrared satellite imagery is also evident from the $F_n$ field (Fig. 5.3g). This warm front is characterized as a westward (in storm-relative coordinates) extension of warm frontogenesis into the polar air stream advected behind the advancing cold front (Shapiro and Keyser 1990). This T-bone phase is often correlated with deepening of the low, which in this case exceeds 24-hPa in 12-hours qualifying for recognition as a Sander’s bomb (Sanders 1980). The cold front is oriented perpendicular to the bent-back warm front and a triple point is established where the cold front meets the elongated warm boundary. The baroclinicity and winds along the bent-back warm front continue to cyclonically wrap around the storm. The $F_s$ term (Fig. 5.3h) is completely dominated by cyclonic rotation of the isentropes along the bent-back warm front axis favorable for culmination in strong warm core seclusion. A broad area of warm, moist air is advected poleward and also advances westward as a boundary around the large circulation to the north of EX Irene.

By 21/00, EX Irene has evolved into a warm core seclusion with colder temperatures cyclonically encircling the storm center along the bent-back warm front. Shapiro and Keyser (1990) describe this stage as the mature and sometimes decaying phase of an extratropical cyclone. In the case of Irene, the warm seclusion was the result of the rapid intensification and not the cause. The scalar frontogenesis has ended with only minor negative contributions along the far eastward advancing cold front. Meanwhile, the rotational frontogenesis continues to cyclonically rotate the isotherms forming a broad area of uniform potential temperature air surrounding the center. Without any midlevel support from the scalar frontogenesis and the absence of a dipole of rotational frontogenesis, steady decay of the seclusion commences (Fig. 5.6). This analysis shows the vertical profile of the warm-core seclusion and the outward sloping baroclinicity of its encircling bent-back warm front. The highest wind speeds were found near the sea surface within the low-level jet on the cold side of the secluding front (Fig. 5.7).

Figure 5.6: 700 hPa $\theta$ (shaded, K) and geopotential height (contours, m) at (a) 10/21 00z and (b) 10/21 12z.
Figure 5.7: (a) Zonal vertical cross-section of wind speed (m s$^{-1}$) and $\theta$ (°K) through latitude 50.625° at 10/20 00z where the strongest surface winds surround the central storm axis. (b) Meridional cross-section as in (a) through longitude 319°E.
Figure 5.8: 500-hPa absolute vorticity (units $10^{-5}$ $s^{-1}$) and geopotential height (units m) for the times indicated.
5.3 Hurricane *Kate* (2003) Extratropical Transition

With the same conceptual framework of frontal evolution using vector frontogenetic functions, application of the diagnostic technique to the extratropical transition and reintensification of Hurricane *Kate* shows many similarities to the ET of *Irene*. However, the distinct difference from examination of the 500-hPa geopotential heights and absolute vorticity concerns the amplitude of the trough over New England and the vast ridge encompassing the North Atlantic. While the upper-level midlatitude circulation associated with *Irene* could be described as zonal in nature, the circulation associated with *Kate* is highly amplified. It is this orientation of the trough/midlatitude circulation that leads to a meridionally oriented extratropical storm. At 07/12, the upper-level trough is positively tilted but attains a strong negative tilt and narrows 12 hours later due to the approach of *Kate*. Additional vorticity remains concentrated with a mid-level cold core low immediately to the west of *Kate* (Fig. 5.8 c,d). The geopotential height gradient narrows and the trough deepen as *Kate* approaches. This aforementioned favorable environment for extratropical cyclogenesis leads to the development of a strong vorticity center at 500-hPa at 08/12 with eventual warm core seclusion near Iceland at 09/00.

Prior to interacting with an advancing shortwave trough off the east coast of North America, *Kate* remained a small size tropical system with a compact CDO feature surrounding a small eye (Fig. 6.17a). Over the next 24 hours, *Kate* coupled with the approaching trough at a distance of about 10° and quickly underwent extratropical transition (Fig. 6.17 b,c) with contributions from a positive trough interaction in this case as described by Hanley et al. (2001). Inner-core convection erodes with the equatorward advection of cold air into the southwest quadrant and begins to wrap around the center (Fig. 6.18c). Cyclone phase diagnostics indicate ET rapidly beginning near 06/12 and completing about 24 hours later at 07/12. Infrared satellite imagery portrays a well-defined deep cloud layer along the poleward advancing trough and a weaker meridionally elongated cold front feature to the southeast of the center. *Kate* increases in thermal asymmetry until 08/06 before undergoing symmetric warm core seclusion.

5.31 500-hPa Vector Frontogenesis Functions

At the base of the deep thermal trough over Eastern Canada, strong deformation forced $F_n$ (Fig. 5.11a) strengthens the potential temperature gradient as *Kate* undergoes ET at 07/00. The rotational counterpart $F_s$ develops into the anticyclonic/cyclonic dipole required for dynamic coupling of the midlatitude circulation to the transitioned TC favorable for extratropical cyclogenesis from 07/00 (Fig. 5.11b) to 07/12 (Fig. 5.11d). The shear deformation between the 500-hPa low centered at 60°W, 50°N and *Kate* as well as the confluence between another mid-level low over Greenland provide positive $F_n$ at 08/00 (Fig. 5.11e). A pronounced rotational $F_s$ dipole (Fig. 5.11f) rotates the isentropes into the characteristic S-shape indicative of cyclogenesis (Carlson 1991).

Only six hours later, *Kate* develops a frontal fracture associated with dry air descent with the accompaniment of a high PV tropopause fold. The $F_n$ field clear
shows the frontolytic zone at the intersection of the warm and cold fronts represented by frontogenesis (Fig. 5.11f) and a broad area to the north associated with the layered cloudiness of the cloud head. A vertical cross section (Fig. 5.9) through the frontal fracture shows weak descent assumed dry adiabatic due to the intrusion of high-PV air from the lower-stratosphere. The characteristics of the frontal fracture as well as the shallow nature of the warm front are illustrated in both the distribution of $\theta_e$ and $F_n$ (Fig. 5.9). The intensification wanes due to a weakening of the $F_s$ dynamical couplet (Fig. 5.11g) in terms of the disappearance of the anti-cyclonic component. A concentrated region of negative $F_s$ exists downstream over Iceland, but it is well removed from the storm center. A bent-back warm front wraps around the center of *Kate* at 08/12 (Fig. 5.11h) and remains until 09/00 (Fig. 5.11j) as the warm core seclusion matures. *Kate* attained a low pressure of 968-hPa on 09/00 (Fig. 5.10) and decayed afterwards. A telltale signature of seclusion at 500-hPa is the superposition of the surface low and a strong circular area of frontolysis. The strongest $F_n$ remains concentrated with the bent-back warm front.

![Figure 5.9](image)

Figure 5.9: Vertical cross-sections through the slice at 08/06z (a) potential vorticity (shaded PVU) and vertical velocity (black, solid descent, dashed ascent, hPa/s) and (b) scalar frontogenesis $F_n$ (black, units) with $\theta_e$ (colored, K).

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Figure 5.10: Same as Figure 5.5 for times indicated.

Figure 5.11: 500-hPa Vector Frontogenesis (contours of $2 \times 10^{-10}$ Km$^{-1}$s$^{-1}$) with scalar frontogenesis $F_n$ (left) and rotational frontogenesis $F_s$ (right) and $\theta$ (contours, K) at times indicated (a)-(l). Cross-section from Figure 5.10 indicated in (g) and (h).
Figure 5.11: Continued.
5.5 Hurricane *Fabian* Extratropical Transition

The ECWMF cyclone phase space diagnostic (Fig. 6.3a) show a definite transition from symmetric warm core to asymmetric warm core as *Fabian* interacted with an approaching short-wave trough from 06/12 to 07/18 when asymmetry was at maximum. Thermodynamically, *Fabian* slowly lost its low-level warm core and became weakly cold-core. The lack of trough interaction and introduction of vertical shear upon *Fabian* is largely responsible for the slow decay of the warm core structure over cooler SSTs. Moreover, the primary rationale for this lack of deep cold core development may concern a lack of significant positive TC remnant/midlatitude trough coupling and the weakness of the upper-level forcing in terms of jets and PV anomalies.

The primary midlatitude circulation at 08/00 (Fig. 5.12a) steered *Fabian* around the periphery of the subtropical ridge ahead of weak, positively tilted trough. The absolute vorticity associated with *Fabian* morphs downstream, which is evidence of a weak short wave impinging and focusing likely warm frontogenesis poleward. The vorticity maximum remains coherent and separate from the upstream large-scale trough with continued weak evidence of warm front formation at 08/12 (Fig 5.12b). The trip over the subtropical ridge remains uneventful for *Fabian* (Fig. 5.12c,d) until it reaches a strong deformation zone, which is associated downstream ridge east of the cutting off low, and is elongated and sheared apart. Throughout the ET, the midlatitude trough remains weakly neutral to strongly positively tilted.

5.51 Vector Frontogenesis Analysis of *Fabian*

In the two-day sequence of scalar frontogenesis $F_n$ (Fig. 5.13), little change in the potential temperature gradient occurs evidenced by a lack of forcing in either respect, positive or negative. The rotational component $F_s$ is dominated by the vorticity contribution due to the unmolested circulation of *Fabian* until 09/12 when it is sheared apart (Fig. 5.14). The $F_s$-vor contribution remains dominant coincident with a midlevel warm pool that validates the decided lack of cold-core formation from the cyclone phase diagnostics. *Fabian* does not complete extratropical transition yet take on a more asymmetric appearance as it translates northwards. Instead, a more elucidating premise would be to define systems as asymmetrically decaying without frontal fracture or deep warm-seclusion development, tropopause/upper level influences, or midlatitude circulation intensification effects. As seen from Chapter 4, the intrusion of stratospheric high-PV into the cyclone circulation at midlevels did not occur. This seems consistent with the lack of a rotational frontogenesis dipole and rotation of the 1000-500 hPa thickness field into a defined S-shape associated with cold (warm) air advection at the base (apex) of the trough (ridge). The sea-level pressure did not drop after the commencement of extratropical transition ($B > 10$).
Thorncroft and Jones (2000) studied the extratropical transition or lack thereof *Felix* (1995), which maintained a deep PV anomaly on the 325 K isentropic surface across the Atlantic. Jones (1995) showed that strong PV anomalies are much less susceptible to the effects of vertical shear than weaker ones. *Fabian* is very similar to *Felix* in this respect, as the 325 K IPV anomaly remained intact until it moved into the warm sector of a large-scale developing baroclinic wave, where it was horizontally deformed. It was noted in the case of *Felix* that due its progression over colder water the upper troposphere became decoupled from the ocean due to the development of a stable layer near the surface (Thorncroft and Jones 2000). While this may explain the slow decay of *Fabian*, which did in fact move over chilly SSTs and decayed, their hypothesis fails to explain why *Kate* strengthened after progressing over even colder water. While *Felix* was steered poleward in a LC1-type wave breaking pattern, *Fabian*’s motion was associated with a distinct LC2-type cyclonic wrap. While each was sheared apart in a horizontal deformation zone, the final outcome of each storm’s poleward transition was achieved through a decidedly different process.

![08/00z](image)
![08/12z](image)
![09/00z](image)
![09/12z](image)

Figure 5.12: 500-hPa absolute vorticity (units 10^{-5} s^{-1}) and geopotential height (units m) for the times indicated.
Figure 5.13: 500-hPa Vector Frontogenesis (contours of $2 \times 10^{-10}$ Km$^{-1}$s$^{-1}$) with scalar frontogenesis $F_n$ (left) and rotational frontogenesis $F_s$ (right) and $\theta$ (contours, K) at times indicated (a)-(h).

Figure 5.14: Same as Figure 5.5 for times indicated.
CHAPTER 6

LOW-LEVEL FRONTAL STRUCTURE AND EVOLUTION

The evolution of the low-level frontal features is examined using QuikSCAT winds and ECMWF potential temperature derived from CEOF temporal interpolation to the satellite overpass time, as described in Chapter 2. At the same time, corresponding AVHRR Ch. 4 and infrared satellite imagery provides an easy interpretation of post-extratropical transition in the framework of the observational cyclone paradigms. Without a burdensome examination of every feature, the goal of this chapter is to diagnose the evolution of low-level frontal structure immediately after extratropical transition. The deformation pattern is the main player responsible for the cyclonic rotation of isotherms and pinching off the warm seclusion. This is consistent with the cyclonic roll-up of upper-level PV and superposition over the low-level diabatically generated PV due to condensation and latent heat release.

6.1 Hurricane Irene (1999) QuikSCAT Winds and AVHRR Satellite Imagery

After crossing Florida earlier on 10/16, Irene accelerated poleward over the warm Gulf Stream as it strengthened over the next several days. According to CPS diagnostics (Fig. 6.1a), Irene began extratropical transition (B>10m) late on 10/17 and completed transition rapidly by 06z the next day. The rapid transition (less than 24 h) continues in terms of rapid cold-core development throughout the troposphere with a marked asymmetric appearance in terms of convection and frontal appearance. The CPS uses a radius of 500 km from the storm center to compute asymmetry/thermal diagnostics, which in this case may have overestimated the rapidness and scale of transition since Irene maintained a tight, highly convective inner-core as seen on AVHRR channel 4 infrared imagery (Fig. 6.4c,d). Yet, the CPS 600 – 900 hPa thickness fields respond to the introduction of a sharp upper-level thermal gradient upon the western half of the circulation (Fig. 6.1c,d) throughout the lower troposphere and rapid deepening of a cold-core. Also notable from the satellite imagery is the tremendous outflow poleward of Irene, which is preconditioning the midlatitude environment by strengthening the upper-level jet through tightening of the PV gradient. The outline of the jet axis is easily seen as the sharp cloud edge curving anticyclonically over eastern Canada into the Atlantic (Fig. 6.4c,d).
apparent is warm frontogenesis poleward of Irene owing to bright cloud-tops associated with intense convection. A large shield of deep clouds extends along much of the baroclinic zone elongated to the east. Meanwhile, convection near the center soon begins to die down as cold air advection and subsidence intrudes the southwestern or upshear quadrant forming the characteristic dry slot.

The broad warm front/baroclinic zone at low-levels is clearly discernable from the N-S PV-\(\theta\) cross-sections (Fig. 4.2). As the positive PV tower of Irene approaches the zone, diabatically generated PV is generated along the baroclinic zone (Fig. 4.2b) at the same time as the \(\theta\)-gradient markedly intensifies. The warm frontogenetic forcing is apparent from another perspective: the zonal cross sections at 19/00z and 19/12z (Fig. 4.3a-b) show a strong advection dipole with low-level jets exceeding 30 ms\(^{-1}\) pumping warm, moist air up and over the baroclinic zone to the north. At the same time, cold dry air from behind the trough moves equatorward helping to erode the eyewall convection and dissolve the tropical PV tower at all tropospheric levels. From the QuikSCAT gridded winds (Fig. 6.5a), strong southerly flow (>40 ms\(^{-1}\)) reaches the warm front and rises and turns cyclonically showing up as a distinct zonally elongated minimum in surface winds likely associated with the warm conveyor belt (Carlson 1991). The dominance of the warm front and weak convection with the cold front fits nicely into the Shapiro-Keyser cyclone paradigm. The 600-900 hPa thickness overlain with QuikSCAT 0.5° gridded winds clearly shows the demarcation of the warm front and outline of the approaching trough. It is at this time (19/08z) that the asymmetry was largest evidenced by the large thickness gradient across the storm (B >100 m). The air reaching the cold front from the northwest was advected equatorward and had a history of subsidence and cooling, thus only isolated puffs of thunderstorms result due to a lack of potential instability along largely a wind-shift line. During the frontal fracture stage of the Shapiro-Keyser (1990) conceptual model, high-PV stratospheric air begins to intrude upon the cyclone center, which shows up very nicely as black on the infrared satellite imagery and SST on the thermal picture at 19/1003z (Fig 6.5a). Cyclonically curved bands embedded within the cloud head are commensurate with deep convection and latent heat release, which fuels the low-level diabatic PV maximum (Fig. 4.3b). Strong winds encircle the extratropical PV tower as Irene rapidly intensifies.

At 2100z 10/19, extratropical Irene has completed the T-bone phase of the Shapiro-Keyser (1990) lifecycle and is undergoing the warm seclusion phase as a bent back warm front or cloud hook cyclonically wraps-up around the center consisting of mainly dry air with some low stratus clouds possible. From the CPS, a warm-core structure mainly in the lower troposphere increases while asymmetry decreases. While air from lower-thickness values (colder) is advected into the southwestern quadrant, relative to the storm, warm sector air is advanced westward characteristic of a building thermal ridge (Fig. 6.1d). The lower-thickness values are wrapped around bent-back warm front (yellow shading) and effect the pinching off of the warm sector. The strongest winds exceeding 45-50 ms\(^{-1}\) blow along the thickness gradient and are of course strongest where the gradient is strongest. The storm motion is approximately 15 ms\(^{-1}\) to the ENE, which adds (subtracts) to the storm relative circulation along the south (north) semicircles. The relative vorticity is concentrated around a tight
circulation center, which is primarily a zonal warm frontogenetic region that extends thousands of kilometers to the east with only a weak hint of a cold front mainly from a wind shift perspective.

As noted before, it is prior to this warm seclusion phase that the cyclonic wrap-up of upper-level PV around the low-level PV maximum occurs. Previously discussed at 500-hPa, positive rotational frontogenesis cyclonically turns the isentropes around the storm center (Fig. 5.3). Deformation forcing also amplifies the temperature gradient through positive scalar frontogenesis along the western half of *Irene* (Fig. 5.3g). This corresponds with the cyclonic wrap-up of PV at upper-levels and the intensification of an upwind jet associated with the dry air intrusion familiar with the frontal fracture. The thermal signature quickly becomes warm core as *Irene* is entrenched in warm air advection. The $\theta$-contours (Fig. 4.2e-h) bow down in the central axis of the extratropical PV tower with extreme baroclinicity in all four surrounding quadrants. The strongest winds are collocated and associated with the vertical wind shear forced by the baroclinicity or in other terms, the thickness gradient surrounding the seclusion. The low-level jet in the conceptual model of Browning et al. (1998) is seen cyclonically wrapping around the storm axis in both the zonal and latitudinal cross-sections (Fig. 4.2,4.3). Summarily, the upper-level PV anomaly seen in the 325 K IPV analysis (Fig. 4.4ef) is superimposed upon the low-level diabatically generated PV secluding *Irene* and further enhancing the low-level baroclinicity (and winds).

The QuikSCAT vector frontogenesis functions for the surface (Fig. 6.7) contain the necessary ingredients for warm core seclusion. The rotational term $F_s$ is obviously aiding in the creation of the thermal ridge and bent back warm front (or occluded front) at 19/21z. At the same time, strong positive scalar frontogenesis $F_n$ amplifies the baroclinicity along the encircling bent-back front, which extends from east of the storm center back along the southwestern flank of the storm. In terms of individual dynamic forcings, the deformation pattern is the most significant in terms of maintaining and amplifying the baroclinicity as well as rotating the isotherms. The divergence and vorticity forcings are secondary in effect. Powerful winds in excess of $>50$ ms$^{-1}$ lie along the bent-back front and outside where the dry slot cyclonically curves into the circulation (Fig. 6.6a). The satellite imagery shows an arc shaped cloud pattern at almost maximum intensity in terms of sea-level pressure (Fig. 6.6cde).

At 10/20 0800z, *Irene* has completed seclusion and is just beginning to weaken in terms of minimum SLP. Satellite imagery shows a mesoscale circulation with tightly wound cyclonic band surrounding a cloud-free eye (Fig. 6.8cd). Dry air has completely circled *Irene* and a dark band of cloud-free air appears north of the eye. The strongest winds ($>50$ ms$^{-1}$) continue along the bent back warm front (Fig. 6.8a) where baroclinicity remains the strongest especially when analyzing PV-$\theta$ from 20/00z to 20/12z (Fig. 4.2gh). The normal component of the low-level jet in the zonal direction decreases from 50 to less than 40 ms$^{-1}$ at 900 hPa corresponding to a marked decrease in low-level diabatically generated PV. The convection along the occluded front begins to disintegrate and becomes sparse. The CPS (Fig. 6.1a) diagnoses *Irene* as deep warm core symmetric, similar to tropical cyclones in thermal structure, yet the method of creating and maintaining the warm cores in each type of system can be decidedly different. The 900-600 hPa thickness gradients correspond nicely
to the strongest winds. Essentially, the area of greatest thickness gradients matches with the low-level jet location, which is also the area of strongest baroclinicity. The copious wrap-up of stratospheric PV wanes (Fig. 4.4gh) as the center of Irene becomes collocated under the upper-level positive PV anomaly presaging barotropic decay.

The rotational component of vector frontogenesis is still strongly cyclonic and collocated with the center of Irene (Fig. 6.9). Deformation scalar frontogenesis continues along the cold front or boundary between the subtropical high and the dry, subsiding air flowing cyclonically around the bent-back front into Irene. Hence, little convection is created due to the lack of instability. The bent-back warm front is also maintained by frontogenetic forcings shared between the divergence and deformation terms (Fig. 6.9). The slow barotropic decay of Irene is characteristic of LC2-type waves, which may linger as cut-off lows for many days. Even at 10/20 2100z 10/21 0700z, the tight cyclonic circulation and eye-like feature are apparent on satellite imagery (Fig. 6.10, 6.12). Strong winds observed by QuikSCAT (> 40 ms$^{-1}$, Fig. 6.12a, 6.13a) exist along the bent-back front or maximum thickness gradient (Fig. 6.1f). The relative vorticity weakens and elongates except over Greenland where the back-door cold front signifies the convergence or confluence of warm sector air from ahead of Irene into cooler, drier air being drawn into the circulation along the thickness gradient (Fig. 6.12b, 6.13b, 6.15b). The frontogenesis functions also bear out this observation (Fig. 6.14, 6.16). The center becomes a rather disorganized area of low cloudiness by 10/21 2100z with weakening winds, limited convection, and negligible frontal forcing (Fig. 6.12cd, 6.13cde, 6.15cd).

### 6.2 Hurricane Kate (2003) QuikSCAT Winds and AVHRR Satellite Imagery

As it approached the meridionally oriented trough, Kate expanded in size from a compact category 3 hurricane (Fig. 6.17a-d) on 10/05 to a large but weaker category 1 on 10/07. The trough merger weakened Kate due to deleterious vertical shear causing the PV tower at upper-levels to blow-off, which resulted in weakening as the storm compensated (see Ritchie and Elsberry 2003). As with Irene, dry air intrudes into the cyclone center and erodes convection and disrupts the eyewall. According to CPS diagnostics (Fig. 6.2ab), Kate rapidly transitions (<24 hours) during the trough merger on 10/06. The introduction of Kate upon an E-W thickness gradient creates asymmetries in winds and precipitation as discussed in several papers involving vertical shear effects upon tropical cyclones (Reasor et al. 2001; Ritchie and Elsberry 2003; Hanley et al. 2001). Kate transitions into a strongly asymmetric cold core extratropical cyclone from 10/07 to 10/08 as exemplified by the CPS and 900-600 hPa thickness fields (Fig. 6.2cd).

The evolution of the QuikSCAT derived wind field is shown with respect to the high-resolution infrared satellite imagery (Figs 6.18, 6.19, 6.20). There are several important things to recognize during this time period. First, the
convection north of the center remains fairly exuberant due to warm
frontogenesis processes and the enhancement of the baroclinic zone, this time to
the west and northwest of Kate due to the meridionally oriented trough. This
helps fire up the diabatically generated low-level PV maximum that is necessary
for cyclogenesis. Similarly, a low-level jet sends warm moist air analogous to the
warm conveyor belt up and over the baroclinic zone while cold, dry air is
advected equatorward thereby eroding the convection surrounding the center.
The nearly meridionally oriented, sharp cloud edge yields the location of the
upper-level jet. The strongest winds are located east of the center of circulation,
which resurrects itself more definitively over a 3-hour period (Fig. 6.19ab). The
satellite representation is characteristic of the cloud head stage just before the
frontal fracture phase of the Shapiro-Keyser lifecycle. At 10/08 0023z (Fig.
6.19c), a broad swath of intense winds (>50 ms\(^{-1}\)) is associated with the cloud
head with a continuous band of hurricane force winds around the newly
reorganized center. These strong winds are associated with the low-level jet
along the diabatic PV maximum near the surface as shown in the PV-\(\theta\)
cross-section (Fig. 4.7b). This is likely driven as well by the tight pressure gradient
between Kate and the subtropical high (1037 hPa, Fig. 5.9b).

In terms of frontogenesis (Fig. 6.20), a warm/cold front reminiscent of an
incipient baroclinic wave develops into a warm seclusion similar to Irene. The
main difference is the sharper cold-front feature, which did not appear with Irene
to the same degree. The rotational frontogenesis is located at the apex of the
thermal ridge allowing for cyclonic wrap-up and seclusion of the system. The
frontal fracture and T-bone phase (Shapiro and Keyser 1990) occur from 10/08
0700z to 1400z followed by the warm seclusion at 2100z. These phases are
born out in the vector frontogenesis fields (Fig. 6.22, 6.24, 6.25). The frontal
fracture stage at 10/08 0700z is characterized by strong deformation forcing
along the developing bent-back warm front extending down the cold front where
convergence is the primary player. Deformation near the apex of the thermal
ridge, which drives the cyclonic pinch-off and seclusion, also dominated the
rotational frontogenesis field. Baroclinicity is thus enhanced along the occluded
front to the west and southwest of the center, which corresponds low-level PV
gradient forced low-level jet. Cold cloud tops are associated with the bent-back
front, which cyclonically wraps up by 10/08 2100z thus secluding the system
center (Fig. 6.23def). The baroclinicity is most intense at this point as seen in
PV-\(\theta\) (Fig. 4.6de) with a sharply defined zone with the PV tower as the central,
warm core axis.

The upper-level jet is responsible for the frontal decay as evidence of the
dissolution of the cold front near the British Isles (Fig. 6.23ef). The secluded
center is not as well defined as Irene or as symmetric. It is hypothesized that the
cyclonic wrap-up is weaker due to the introduction of anticyclonic shear with
respect to Kate’s low-level location relative to the upper-level jet. The CPS
maintains noticeable asymmetry (B > 10 m) and modest warm core thermal
structure, possibly due to the small scale of the storm. Nevertheless, the
tightened thickness gradient (Fig. 6.2f) around the entire occluded front is also
associated with the low-level PV gradient (Fig. 4.6e) and strong winds as
observed by QuikSCAT (Fig. 6.23abc). The frontogenesis patterns associated
with Kate’s occluded front and far-removed cold front break down over the next
13 hours from 10/09 0000z to 1300z as Kate weakens rapidly (Fig. 6.24, 6.25). The overall seclusion is weaker than Irene's and with the prevalence of a cold front, seems to be reminiscent somewhat of the Norwegian variety of extratropical cyclones. However, there are the clear indications of the stages of the Shapiro-Keyser lifecycle. The evolution from the cyclonic to anticyclonic upper-level flow regime may have affected the final seclusion makeup leading to further research on the matter.

**6.3 Hurricane Fabian (2003) QuikSCAT Winds and AVHRR Satellite Imagery**

The low-level frontal story with Fabian is less vigorous by far compared to Kate and Irene. But to offer a different vantage point, experimental 2.5 km resolution QuikSCAT swaths (personal communication, Dr. Long, BYU) are shown with corresponding AVHRR temperature images to show the evolution of the wind field just prior to extratropical transition. A series of three images (Fig. 6.26abc) definitively shows the inner-eyewall and core of strong winds and the beginning of an asymmetric wind evolution associated with a weak trough interaction. Fabian did not strengthen after extratropical transition (began 09/07 00z and completed 09/08 12z), but weakened while slowly acquiring a cold core thermal structure (Fig. 6.3a) without seclusion.

The symmetric structure of a mature category 3 hurricane with strong convective bands surrounding a well-defined eye is easily seen by the experimental QuikSCAT winds (Fig. 6.26abc). The midlatitude trough apparently enhances the upper-level divergent outflow of Fabian, as the storm remains category 2 or 3 with maximum QuikSCAT observed winds of 45 – 50 ms$^{-1}$, which matches the NHC best track data well (not shown). Hints of warm frontogenesis along the baroclinic zone to the north show up as a small band of strong winds (09/07 0856z) extending poleward to the northeast of Fabian. The intrusion of dry air due to equatorward cold advection associated with the low-level jet erodes the inner eyewall convection by 09/07 2110z (Fig. 6.27a) commensurate with a more asymmetric appearance on satellite imagery. Strong convection continued in the warm frontogenesis region with the strongest winds observed along the low-level jet associated with the ascending warm conveyor belt (Fig. 6.27de). As Fabian moved northward, it encountered a deformation/shear zone akin to a long-wave thermal ridge. The strongest winds split into two narrow bands surrounding the deformation zone (Fig. 6.28cd) as the cloud signature elongated zonally (Fig. 6.28ef). The frontogenesis functions, both scalar and rotational, are do not indicate extensive front formation (Fig. 6.29). Cyclonic vorticity continued to rotate the isotherms counterclockwise to a small degree with little change in magnitude during the time series presented (Fig. 6.29). Winds overall decrease below 20 ms$^{-1}$ (Fig. 6.30ab).
Figure 6.1: Hurricane Irene (1999) (a) and (b) Cyclone Phase Space (Hart 2003). 600-900 hPa CEOF interpolated ECMWF thickness fields overlain with QuikScat derived surface velocity vectors for (c) 10/19 08z (d) 10/19 21z (e) 10/20 08z (f) 10/20 2100z.
Figure 6.2: Same as Figure 6.1 for Hurricane Kate (2003) (a) and (b) CPS for (c) 10/07 1400z (d) 10/07 2100z (e) 10/08 0700z (f) 10/09 0000z.
Figure 6.3: Same as Figure 6.1 except for Hurricane Fabian (2003) with (a) CPS and (b) 09/07 2100z (c) 09/08 1400z (d) 09/09 1300z.
Figure 6.4: NOAA-AVHRR satellite images. Channel 4 Infrared: (a) 10/16 2353z NOAA-15 (b) 10/17 2331z NOAA-15 (c) 10/18 1026z NOAA-12 (d) 10/18 1906z NOAA-14.
Figure 6.5: (a) 0.5° QuikScat derived surface winds (contoured speed, vector direction) for 10/19 0800z and (b) as in Figure 6.4 for 10/19 0730z NOAA-14 (c) same as (b) for 10/19 1003z NOAA-12 and (d) AVHRR Temperature retrieval according to colorbar °C, same time as (c).
Figure 6.6: Same as Figure 6.5 at (a) and (b) 10/19 2100z; (c) Ch.4 10/19 1723z NOAA-17 (d) and (e) 10/19 2115z NOAA-15.
Figure 6.7: Vector frontogenesis functions (contours of $2 \times 10^{-10} \text{Km}^{-1} \text{s}^{-1}$) with QuikScat derived winds 0.5° and ECMWF derived potential temperature from CEOF (Chapter 2) for 10/19 2100z: (a) $F_n$ (scalar frontogenesis) (b) $F_s$ (rotational frontogenesis) with components (c) $F_n$-def (d) $F_n$-div (e) $F_s$-def (f) $F_s$-vort.
Figure 6.8: Same as Figure 6.5 (a) and (b) 10/20 0800z; (c) (d) 10/20 0711z
NOAA-14.
Figure 6.9: Same as Figure 6.7 except at 10/20 0800z.
Figure 6.10: (a) and (b) 10/20 1532z NOAA-14.

Figure 6.11: Same as Figure 6.7 except for 10/20 2100z and (a) F_n-total (b) F_s-total (c) F_n-div.
Figure 6.12: Same as Figure 6.5 (a) and (b) 10/20 2100z; (c) and (d) 10/20 1915z NOAA-15.
Figure 6.13: Same as Figure 6.5 (a) and (b) 10/21 0700z; (c) Ch.4 10/21 0525z NOAA-14 (d) and (e) 10/21 1031z NOAA-15.

Figure 6.14: Same as Figure 6.7 except for 10/21 0700z and (a) $F_n$-total (b) $F_s$-total.
Figure 6.15: Same as Figure 6.5 (a) and (b) 10/21 2000z; (c) (d) 10/21 1853z NOAA-15.

Figure 6.16: Same as Figure 6.7 except for 10/21 2100z and (a) $F_n$-total (b) $F_s$-total.
Figure 6.17: AVHRR Ch.4 NOAA-17 (a) 10/05 1443z (b) 10/05 2032z (c) 10/06 1421z (d) 10/07 0140z.
Figure 6.18: (a) QuikSCAT swath 0.25° (contour, ms$^{-1}$) with wind direction (unit vector) at 10/06 2201z and (b) 10/07 0922z (c) Ch.4 AVHRR 10/06 2008z NOAA-12 (d) 10/07 0620z NOAA-16.
Figure 6.19: Same as Figure 6.18 for (a) 10/07 1440z (b) 10/07 2133z (c) 10/08 0023z and (d) 10/07 1354z NOAA-17 (e) 10/07 2208 NOAA-17.
Figure 6.20: Same as Figure 6.7. except for (a) $F_n$-total and (b) $F_s$-total at 10/07 2200z and (c) $F_n$-total and (d) $F_s$-total at 10/08 0000z.

Figure 6.21: Same as Figure 6.5 (a) 10/08 0700z (b) 10/08 0602z NOAA-17.
Figure 6.22: Vector frontogenesis functions with QuikScat derived winds 0.5° and ECMWF derived potential temperature from CEOF (Chapter 2.6) for 10/08 0700z: (a) $F_n$ (scalar frontogenesis) (b) $F_s$ (rotational frontogenesis) with components (c) $F_n$-def (d) $F_n$-div (e) $F_s$-def (f) $F_s$-vort.
Figure 6.23: Same as Figure 6.5 at (a) 10/08 1400z (b) 10/08 2100z (c) 10/09 0700z (d) 10/08 1327z NOAA-17 (e) 10/08 2145z NOAA-17 (f) 10/09 0826z NOAA-17.
Figure 6.24: Same as Figure 6.21 except for (a) $F_n$-total and (b) $F_s$-total at 10/08 1400z and (c) $F_n$-total and (d) $F_s$-total at 10/08 2100z.
Figure 6.25: Same as Figure 6.22 except for (a) $F_n$-total and (b) $F_s$-total at 10/09 0000z and (c) $F_n$-total and (d) $F_s$-total at 10/09 0700z and (e) $F_n$-total and (f) $F_s$-total at 10/09 1300z.
Figure 6.26: Top row QuikScat 2.5 km resolution wind speed (ms$^{-1}$ contoured according to scale) (courtesy of Dr. David Long BYU) (a) 09/06 0924z (b) 09/07 0135z (c) 09/07 0856z and AVHRR temperature (d) 09/06 1135z NOAA-17 (e) 09/07 0120z NOAA-17 (f) 09/07 1112z NOAA-15.
Figure 6.27: QuikSCAT 0.25° swath wind speed (ms$^{-1}$ contoured according to scale) (a) 09/07 2110z (b) 09/08 0113z (c) 09/08 0835z and AVHRR temperature; bottom row: (d) 09/07 2326z NOAA-17 (e) 09/08 1049z NOAA-15.
Figure 6.28: (a) ~0.25° QuikSCAT swath wind speed (m s\(^{-1}\)) at 09/081325z (b) AVHRR Ch. 4 09/08 1312z NOAA-15 (c) 0.5°QuikScat gridded wind speed (m s\(^{-1}\)) 09/08 2100z (d) 09/09 0700z (e) 09/08 2132z NOAA-17 (f) 09/09 0833z NOAA-15.
Figure 6.29: Same as Figure 6.7 except for (a) $F_n$-total and (b) $F_s$-total at 09/08 0800z and (c) $F_n$-total and (d) $F_s$-total at 09/08 1400z and (e) $F_n$-total and (f) $F_s$-total at 09/08 2100z (g) $F_n$-total and (h) $F_s$-total at 09/09 0700z.

Figure 6.30: 0.5° QuikScat gridded wind speed (ms$^{-1}$) (a) 09/09 1300z and (b) 09/09 2100z.
7. CONCLUSION AND IDEAS

The evolution of post-extratropical transition tropical cyclones is not adequately explained by one cyclone paradigm, either observational or theoretical. For the two hurricanes studied that underwent warm seclusion in the Shapiro-Keyser framework, the upper-level flow pattern associated with Irene and Kate were markedly different. Irene interacted head-on with a zonal confluent jet while Kate paralleled a high-amplitude, meridionally oriented jet. Rapid cyclogenesis ensued as Irene superimposed its deformation pattern associated with the strong TC vortex upon the baroclinic zone. A frontal fracture and intrusion of high-PV stratospheric air superimposed its cyclonic circulation upon the diabatically generated low-level PV found in the warm frontogenesis region. The resulting warm core seclusion occurred rapidly with very strong winds well in excess of hurricane force on the southwestern flank of the bent-back front along tightened thickness gradients as seen by QuikSCAT. Kate, on the other hand, translated much further north and only interacted with the midlatitude trough to the west after its deformation pattern sufficiently phase locked in terms of rotational frontogenesis. The frontal fracture was much weaker than in the case of Irene, yet a warm seclusion did develop along with the concomitant strong winds. From the theoretical cyclone paradigm viewpoint, Irene developed in a strongly cyclonic shear environment indicative of an LC2-type wave while Kate experienced more of an LC1 type cyclone with characteristic trough thinning. It is suggested that categorizing extratropical storms of tropical origin, especially according to one specific paradigm fails to capture the preconditioning of midlatitude environment by the TC. Fabian failed to accelerate the intervening jet streak and did not enter regions favorable for cyclogenesis.

In order to understand the processes responsible for frontal fracture and warm seclusion, a dynamical model with the temporal resolution and spatial scale is necessary. Questions continue about the exact mechanism for the very strong winds along the bent-back front, which deal with much more than simple pressure gradients and translation of the storm. Many warm seclusions affect western Europe and pound the continent with heavy winds and precipitation. Thus, it is helpful to look at the mechanisms behind the surge of winds especially in terms of the stratospheric high-PV intrusion above the low-level diabatic PV anomaly. Vertical momentum budgets are currently being studied with respect to contraction and expansion of the winds during extratropical transition. QuikSCAT winds at high resolution (~2.5 km) at different stages of cyclone evolution would be invaluable in assessing the fine scale changes in the system as well as strengthening a hypothesis about the evolution of the strong winds around the bent-back front.
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Hi-resolution QuikSCAT derived wind fields courtesy of Dr. David Long, BYU.

BIOGRAPHICAL SKETCH

Ryan Nicholas Maue was born on January 3, 1981 in Manistee, Michigan along the eastern shore of Lake Michigan. The usually chilly waters affected Manistee’s weather during the summer with cool breezes and during the winter with heavy lake effect snow. Ryan marveled at the panoramic view the lakeshore provided and enjoyed watching thunderstorms over the lighthouses at night. After graduating from high school as valedictorian, he attended the University of Michigan with the intention of majoring in mathematics, a subject he excelled in throughout his high school years. Later in his sophomore year, he dual enrolled in the College of Engineering as a student in the Atmospheric Oceanic and Space Sciences department and the College of Literature, Science and Arts Honors Program with a dual major in History. Achieving a balance between the two disciplines proved challenging yet very rewarding. Historically speaking, Ryan was fascinated with communism in particular the rise and fall of Stalin and the situation on the Korean Peninsula. He authored a senior thesis using original research at the Gerald R. Ford Presidential Library and met the former President.

Ryan graduated in April 2002 with degrees in history (A.B., high distinction) and atmospheric science (B.S., summa cum laude). His academic achievement earned him induction into the distinguished Phi Beta Kappa honor society. He reached a crossroads early in his final semester at Michigan concerning the decision to attend graduate school in meteorology or go to law school. The ultimate decision did not preclude the other option, only delayed it. Thus, he enrolled at the Florida State University for graduate study at the Center for Atmosphere and Ocean Prediction Studies (COAPS) in August 2002 to pursue a Masters of Science degree in meteorology. Both tropical and extratropical meteorology applications of scatterometer wind data interested Ryan throughout his experience at FSU. He plans on finishing his PhD. by early 2007 and attending law school soon afterwards to become a high-powered, yet conservative, environmental lawyer.