Hurricane Bonnie (1998) : maintaining intensity during high vertical wind shear and an eyewall replacement cycle

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HURRICANE BONNIE (1998): MAINTAINING INTENSITY DURING HIGH VERTICAL WIND SHEAR AND AN EYEWALL REPLACEMENT CYCLE

by

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ABSTRACT

Hurricane Bonnie (1998) was an unusually resilient hurricane that maintained intensity in 12–16 ms\(^{-1}\) vertical wind shear and during an eyewall replacement cycle from 23 – 25 August. This remarkable behavior was examined using observations from flight-level data, microwave imagery, radar, and dropsondes over the two-day period. The symmetric and asymmetric aspects of Bonnie’s eyewall replacement cycle were documented and compared to eyewall replacement cycles in other hurricanes. Similar to other observed eyewall replacement cycles, Bonnie exhibited the development, strengthening, and dominance of a secondary eyewall while a primary eyewall decayed. However, Bonnie’s structure was highly asymmetric due to strong 12–16 ms\(^{-1}\) vertical wind shear, in contrast to the more symmetric structures observed in other hurricanes undergoing eyewall replacement cycles. It is hypothesized that strong shear preferentially forced convection downshear, which was able to extend upshear and form a secondary eyewall through enhanced surface fluxes upshear. The larger radius of maximum winds after the eyewall replacement cycle completed might have aided Bonnie’s resiliency in shear by increasing the likelihood that diabatic heating would fall inside the radius of maximum winds. These observations of Hurricane Bonnie’s ability to maintain intensity in spite of high shear and an eyewall replacement cycle, both of which usually result in weakening, provide a new direction from which to view the intensity change issue in hurricanes.
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1. Introduction

Tropical cyclone intensity change is a critical topic to address in order to better warn communities of the potentially devastating impacts from these systems. Yet, understanding tropical cyclone intensity change is a lingering issue in the research and forecasting communities, as evidenced by little improvement to tropical cyclone intensity forecasts since the 1990s (Rappaport et al. 2009). The myriad of factors affecting tropical cyclone intensity complicates this issue, as both internal dynamics and external factors affect storm intensity. Two particularly well-cited influences on tropical cyclone intensity change include vertical wind shear (Simpson and Riehl 1958; DeMaria 1996; Frank and Ritchie 2001; Riemer et al. 2010; Tang and Emanuel 2010) and eyewall replacement cycles (ERCs; Willoughby et al. 1982; Houze et al. 2007; Sitkowski et al. 2011; Kossin and DeMaria 2016).

The prevailing understanding of vertical wind shear is that it weakens tropical cyclones, though studies differ in the mechanism by which this occurs. DeMaria (1996) found that tropical cyclone intensity change was negatively correlated with shear magnitude, strongest at a lag of 24 h. The weakening observed by DeMaria (1996) was attributed to the development of a mid-level warm anomaly—a consequence of the vortex maintaining thermal wind balance under a shear-induced tilt. This warm-anomaly stabilized the storm, decreased convection, and ultimately weakened the simulated tropical cyclone (DeMaria 1996). Frank and Ritchie (2001) arrived at different results than DeMaria (1996), asserting that the weakening of a tropical cyclone in shear is a top-down process, due to the erosion of the upper-level warm core. They hypothesized that a shear induced asymmetric structure eventually became strong enough to flux high potential vorticity and equivalent potential temperature outwards at upper levels. However, Riemer et al. (2010) pointed to the importance of the boundary layer in weakening a sheared tropical cyclone.
They found that shear induces an asymmetric convective structure reminiscent of a stationary band complex (Willoughby et al. 1984) that produces downdrafts and flushes low equivalent potential temperature air into the tropical cyclone core via inflowing boundary layer air (Riemer et al. 2010). By this mechanism, low equivalent potential temperature air reaches the eyewall and reduces the efficiency of the tropical cyclone’s “Carnot cycle” (Emanuel 1986). In addition to this low-level pathway proposed by Riemer et al. (2010), Tang and Emanuel (2010) suggested that ventilation of the tropical cyclone core with midlevel dry air by eddies is an important pathway by which shear weakens a tropical cyclone.

Shear also has an asymmetric effect on tropical cyclone structure. Using lightning as a proxy for deep convection, Corbosiero and Molinari (2002) found that lightning preferentially occurred downshear in tropical cyclones. This relationship was stronger in higher shear and for more intense storms (but only in the inner core). Corbosiero and Molinari (2002) used Frank and Ritchie’s (2001) findings to explain their results, discussing how asymmetric deep convection maintains a vortex’s vertical alignment in the face of shear-induced tilt. However, shear-induced asymmetric convection might prevent the storm from deepening, due to asymmetric heating outside the inertially strong core and based on the mechanisms noted in the previous paragraph.

Even so, tropical cyclones do not always weaken in high shear. A number of case studies have observed storms that intensified while embedded in strong shear. In Hurricane Irene (1999), shear-induced asymmetric convection was strong enough that it increased the azimuthal average diabatic heating within the radius of maximum winds and contributed to intensification (Nguyen and Molinari 2012). Similarly, Hurricane Claudette (2003; Shelton and Molinari 2009) and Tropical Storm Gabrielle (2001; Molinari and Vollaro 2010) intensified despite at least 10 m s⁻¹ shear as intense convection developed within the radius of maximum winds. These authors
suggested that shear helped to generate the intense, asymmetric convection that allowed the storms to intensify.

Studies have also hypothesized that tropical cyclones may resist shear due to vortex tilt evolution. Modeling the evolution of a barotropic vortex, Jones (1995) found that the initial response of a vortex in sheared flow is to tilt, and for the upper and lower vortex centers to rotate cyclonically about the mid-level center. The precession of these upper and lower level vortex centers opposed environmental wind shear by either inducing sheared flow of opposite sign or reducing tilt of the vortex by vortex realignment (Jones 1995). Reasor et al. (2004) proposed that the tilt of a vortex may additionally be reduced by a vortex Rossby wave (VRW) damping mechanism. A small tilt magnitude and stable left of shear (minimizes the net vertical shear) tilt configuration was seen in Hurricane Guillermo (1997; Reasor et al. 2012), and supports the ability of the above mechanisms in a tropical cyclone’s ability to resist shear.

ERCs also produce substantial intensity changes in tropical cyclones. In their seminal study, Willoughby et al. (1982) observed tropical cyclone weakening as an outer convective ring contracted and intensified around a decaying inner eyewall. Once the inner eyewall vanished and the outer eyewall replaced it, the hurricane reintensified (Willoughby et al. 1982). Sitkowski et al. (2011; hereafter S11) further detailed the stages of ERCs in their climatological study of North Atlantic hurricanes. The three expected stages of intensity changes associated with ERCs are: 1) intensification, during which an inner wind maximum reaches its peak intensity while an outer wind maximum associated with rainbands appears, 2) weakening, during which the inner wind maximum weakens while the outer wind maximum contracts and intensifies, forming a concentric eyewall, and 3) reintensification, during which the outer wind maximum’s intensity exceeds that of the inner, as the inner one decays and vanishes (S11).
Observational and modeling studies of ERCs support the series of events described by Willoughby et al. (1982) and S11. A commonly studied example of an ERC is that observed in Hurricane Rita (2005). Rita had a textbook case of a secondary eyewall, displaying a symmetric ring of convection around the inner eyewall, separated by a clear moat. At the start of its ERC, Hurricane Rita was at its maximum intensity (Didlake and Houze 2011), weakened during its ERC (Houze et al. 2007; Judt and Chen 2010; Didlake and Houze. 2011), and briefly stopped weakening at the end of its ERC (Didlake and Houze 2011). Aside from weakening due to shear after its ERC, Rita’s intensity fluctuations are consistent with Willoughby et al. (1982) and S11. Hurricane Edouard (2014) and Hurricane Gonzalo (2014) exhibited similar intensity fluctuations as Rita (2005), intensifying at the start of their ERCs and weakening during their ERCs (Abarca et al. 2016; Didlake et al. 2016). Modeling studies have also reproduced the weakening that occurs during ERCs as rainbands intensify and contract to form a secondary eyewall (Zhu and Zhu 2014; Zhang et al. 2017).

Despite the common occurrence of ERCs and secondary eyewall formation (SEF) in hurricanes, their cause is still relatively unknown. Nong and Emanuel (2003) suggested that a sufficiently strong external forcing, such as an upper-level trough, is necessary for SEF. Many other studies point to the importance of internal processes in producing a secondary eyewall. Vortex Rossby waves are one such mechanism that can accelerate the mean tangential wind at some radius from the center of the vortex through wave-mean flow interactions (Montgomery and Kallenbach 1997). This occurs through the outward propagation of waves on a negative radial vorticity gradient and eventual stagnation at a particular radius. The importance of the axisymmetric vorticity structure in SEF was highlighted in TerweyJv and Montgomery’s (2008) beta-skirt hypothesis, in which convectively generated vorticity occurring in a region of a gentle
negative radial vorticity gradient (beta-skirt) can transfer energy into the mean flow. Additional hypotheses suggest that unbalanced boundary layer dynamics are sufficient for SEF (Huang et al. 2012; Abarca and Montgomery 2013, 2015). It is feasible that both external conditions and internal storm dynamics are important to SEF, as suggested by Kossin and DeMaria (2016) and Kossin and Sitkowski (2009).

Findings from Kossin and Sitkowski (2009) show that hurricane SEF generally occurs in lower shear, which studies have recently begun to reexamine in more depth. In a case study of Hurricane Gonzalo, Didlake et al. (2016) observed asymmetric eyewalls with respect to moderate (4–7 m s\(^{-1}\)) shear. The differing asymmetries between the primary and secondary eyewalls in Gonzalo were hypothesized to be the result of tilt and interaction with a stationary band complex. However, Didlake et al. (2016) did not speculate about the role of shear in the development Gonzalo’s ERCs. This question was instead addressed in ERC simulations performed by Dai et al. (2017) and Zhang et al. (2017). In Dai et al.’s (2017) simulation, eddy flux convergence from a midlatitude jet produced an asymmetric stratiform cloud from which a secondary eyewall formed. Zhang et al.’s (2017) study similarly suggested that moderate environmental wind shear provided forcing for outer rainbands to develop and a secondary eyewall to form from the rainbands. Results from Dai et al. (2017) and Zhang et al. (2017) generally support Nong and Emanuel’s (2003) idea that a strong external forcing can trigger a secondary eyewall. These results suggest the link between the environmental influences and SEF is an important one.

Though these previous studies examined various aspects of shear and ERCs, none have conducted a detailed case study of an ERC in a highly sheared, steady-state storm, as was the case in Hurricane Bonnie (1998). This is the focus of this study and will be accomplished.
through use of radar, flight-level data, microwave imagery, and dropsondes in Bonnie from 23–25 August 1998. The objective is to document the evolution of symmetric and asymmetric aspects of Bonnie’s ERC and to compare these results with other ERCs. Details of Bonnie’s storm history are discussed in Chapter 2, followed by data and methodology in Chapter 3. The evolution and structure of Bonnie’s ERC is shown in Chapter 4. Comparison of Bonnie’s ERC with other ERCs, as well as possible reasons for Bonnie’s SEF and resiliency in high shear are discussed in Chapter 5. Conclusions are made in Chapter 6.

2. Bonnie’s Storm History

At 1200 UTC 19 August 1998, Bonnie formed as a tropical depression and intensified into a tropical storm a day later as it tracked west/northwestward around a Bermuda high (Fig. 2.1; Rogers et al. 2003; Zhu et al. 2004). By 1200 UTC 21 August, Bonnie was located northeast of Puerto Rico and entered into a region of low environmental wind shear and warm ocean waters (Molinari and Vollaro 2008). These favorable environmental conditions allowed Bonnie to rapidly intensify late on 21 August to obtain hurricane status by 0600 UTC 22 August (Fig. 2.1). Bonnie continued its rapid intensification throughout the rest of the day until 1200 UTC 23 August, even as northwesterly shear increased from 5 to 12 m s⁻¹ due to proximity to an upper-level trough over the Southeastern United States (Braun et al. 2006). After 1200 UTC 23 August, Bonnie transitioned into a steady-state hurricane, maintaining a constant 51 m s⁻¹ windspeed (Fig. 2.2). The lowest minimum pressure of 954 mb was recorded at 0000 UTC 24 August when vertical wind shear magnitude was 15 m s⁻¹. Despite small pressure fluctuations, this steady-state-high-shear regime persisted in Bonnie until midday on 25 August, at which point shear dropped from 13 m s⁻¹ at 0600 UTC 25 August to 7 m s⁻¹ at 1800 UTC 25 August. Bonnie maintained its 51 m s⁻¹ intensity all the way until landfall in North Carolina on 27 August.
Prior studies on Bonnie suggested that increasingly warm sea surface temperatures (SSTs) may have contributed to Bonnie’s intensification and resiliency in high shear (Heymsfield et al. 2001; Molinari and Vollaro 2008). SSTs were observed to rise from 28°C on 20 August to 30.5°C on 25 August by Heymsfield et al. (2001). These values are similar to Reynolds SST values, which are obtained a week prior at a resolution of 100 km (Wentz et al. 2000). In contrast, the Tropical Rainfall Measuring Mission (TRMM) microwave imager (TMI) observed a significant cold wake in Bonnie from 24–26 August, with SSTs ranging from 25–27°C (Wentz et al. 2000). At a finer spatial (50 km) and temporal (daily) resolution, the TMI SSTs are a more accurate representation of Bonnie’s local SSTs (Wentz et al. 2000).

Two observational studies and four modeling studies of Bonnie have examined shear-related impacts on Bonnie’s intensity and structural evolution (see Molinari and Vollaro 2008 for a complete review of these studies). Heymsfield et al. (2001) noted the presence of vigorous convective cells (convective bursts) in Bonnie’s eyewall during the intensification period from 1800 UTC 21 August through 23 August. These convective bursts repeatedly occurred downshear (Heymsfield et al. 2001) and were noted to have supercellular characteristics (Molinari and Vollaro 2008). After this intensification period, Zhu et al. (2004) observed the development of a secondary eyewall at 0000 UTC 24 August during a period of high vertical wind shear. However, due to a shear-induced asymmetric structure, Zhu et al. (2004) chose to focus on a more axisymmetric ERC that occurred at approximately 0000 UTC 26 August.

While the SEF at 0000 UTC 24 August noted by Zhu et al. (2004) was skimmed over in their paper, this SEF and ERC is of particular interest in the present study. In addition to maintaining resiliency in the face of high vertical wind shear, the occurrence of an ERC in Bonnie further deviates from the expected behavior of a sheared tropical cyclone. Many
questions arise from Bonnie’s surprising behavior, such as how did Bonnie maintain intensity during its ERC in 12–16 m s⁻¹ shear? Did shear play a role in Bonnie’s ERC? And could the occurrence of an ERC have helped Bonnie maintain resiliency in high shear? These questions motivate this study.
Figure 2.1. Hurricane Bonnie’s (1998) track from 19–31 August (courtesy of Dave Vollaro).

Figure 2.2. Bonnie’s maximum windspeed (m s\(^{-1}\); black line and left y-axis) and 850–200 hPa shear magnitude (m s\(^{-1}\); red line and right y-axis) from 0000 UTC August 22 – 0000 UTC August 26. Windspeed and shear magnitude were obtained from the Best Track dataset. The three grey rectangles denote three separate U.S. Air Force flights analyzed in this study, where A1 is 0813–1733 UTC 23 August, A2 is 1956 UTC 23 August–0718 UTC 24 August, and A3 is 0757–1848 UTC 24 August.
3. Data and Methods

3.1. Flight-level data

Data obtained from eight different flights from 23–25 August were considered for the purposes of this study (Table 3.1): A1 (~ 1200 UTC 23 August), A2 (~0000 UTC 24 August), P1 (~0000 UTC 24 August), P2 (~0000 UTC 24 August), A3 (~ 1200 UTC 24 August), A4 (~0000 UTC 25 August), P3 (~0000 UTC 25 August), and P4 (~0000 UTC 25 August). Half of the flights included aircraft reconnaissance missions by the United States Air Force (USAF) WC-130, which collected nearly continuous 1-Hz frequency measurements of temperature (T), dewpoint temperature (T_d), and wind from midday on 23 August to early on 25 August at 700 mb. The National Oceanic and Atmospheric Administration (NOAA) WP-3D N42RF and N34RF (P-3) aircraft flying at 600 mb obtained the same measurements as the USAF flights late on 23 August and 24 August, with additional measurements including vertical velocity (w) and radar data.

Aircraft data were transformed by personnel at NOAA’s Hurricane Research Division (HRD) into storm-relative coordinates following Willoughby and Chemlow’s (1982) center-position determination method, and then interpolated into 0.5 km bins along a 150-km radial leg. Four to ten of these 150 km radial legs (Fig. 3.1) comprised each of the eight flights and provided ample azimuthal coverage of Bonnie. The post-processing of aircraft data allowed for the calculation of storm-relative radial (u) and tangential wind (v) from the wind measurements and equivalent potential temperature (Θ_e) following Bolton (1980). Instrument wetting events in hurricanes described by Eastin et al. (2002) can result in spurious values of Θ_e, but application of the Zipser et al. (1981) correction to flight-level T and T_d during post-processing mitigated this issue. This correction reduced wetting errors by approximately 30–50% (Eastin et al. 2002) and
was shown to be of adequate use when compared with Eastin et al.’s (2002) corrections (Corbosiero et al. 2005). These findings prove satisfactory to proceed with using flight-level derived $\Theta_e$.

To characterize the shape of flight-level $v$ radial profiles, an alpha parameter ($\alpha$) was calculated:

$$\alpha = \frac{\ln(\bar{v}_2)}{\ln(\bar{v}_1)}$$

(1)

where $r_1$ ($r_2$) is the radius of maximum wind (three times the radius of maximum wind) and $\bar{v}_1$ ($\bar{v}_2$) are the axisymmetric $v$ values at the respective radii (Eqn. 3 in Mallen et al. 2005).

Bonnie’s $\alpha$ was calculated for flights A1 and A2 (see Table 3.1), but not flight A3 due to lack of observations at three times the radius of maximum winds. The median $\alpha$ value in 72 Atlantic and eastern Pacific tropical cyclones (TCs) is 0.37, with a range of 0.04–0.67, where smaller values indicate a more gradual decrease in $\bar{v}$ and thus a broader vortex (Table 2 from Mallen et al. 2005).

Additional quantities calculated from the flight-level data include the symmetric component of relative vorticity ($\zeta$) and absolute angular momentum ($m_a$), given by:

$$\zeta = \frac{\partial v}{\partial r} + \frac{v}{r}$$

(2)

$$m_a = rv + \frac{fr^2}{2}$$

(3)

A centered finite difference approximation was used to calculate $\zeta$ in Eqn. 2.
To isolate major structural changes and minimize local fluctuations, a 10\textsuperscript{th} order polynomial was fit to \(v\), \(u\), \(\zeta\), and \(m_a\). This 10\textsuperscript{th} order polynomial compares well to the raw fields, as shown by an example in Fig. 3.2 of \(v\). The major changes in \(v\) are adequately captured by the polynomial fit, while small-scale fluctuations in raw \(v\) are smoothed out. Therefore, all radial plots and Hovmöllers of flight-level variables are shown in terms of this 10\textsuperscript{th} order polynomial fit.

3.2. Radar data

The lower fuselage (LF) and tail (TA) radars aboard the NOAA P3 aircraft collected data during flights P1, P2, P3, and P4 (Table 3.1). The LF is a C-band (5 cm) horizontal polarization radar that scans azimuthally from its location at the base of the plane. The TA radar is an X-band (3 cm) vertical polarization radar located in the plane’s tail (Jorgensen 1984a). Given its vertical configuration, the TA radar can be operated in two scanning modes – normal and fore/aft scanning technique (F/AST). When operated in its normal mode, the TA radar scans in 360° circles perpendicular to the long axis of the plane, while in F/AST, the TA radar scans in cones tilted 20° forward and aft normal to the axis of rotation (Fig. 3.3; Gamache and Marks 1995). Both TA radar scanning modes were in operation during flight P1, while flights P2 and P3 operated in normal mode, and flight P4 operated exclusively in F/AST mode (Table 3.1).

While the LF radar solely provides flight-level reflectivity data, additional information was gleaned through processing the TA radar reflectivity and Doppler velocity data using Gamache’s (1997) automated variational algorithm. Solving the radar projection equations and continuity equation, solutions from Gamache’s (1997) algorithm produced a three-dimensional gridded (swath) analyses with a 2 x 2 km\(^2\) horizontal and 0.5 km vertical resolution over a 400 x 400 km domain size (Reasor et al. 2009; Rogers et al. 2015). These swath analyses were then
transformed into cylindrical coordinates with a 1° azimuthal and 2 km radial resolution, which allowed for creation of radius vs. height graphs extending out to 150 km and up to 15 km. In both the swath analyses and radius-height cross sections, data included vertical and horizontal winds and reflectivity (Rogers et al. 2012).

Additional information on Bonnie’s vertical structure was gained from use of vertical incidence (VI) data, which refers to when the TA radar beam was pointing directly up or down during a normal scanning mode. This configuration collects reflectivity and Doppler velocities, with the Doppler velocities measuring the up and down motion of precipitation particles relative to the aircraft motion. Vertical air velocity ($w$) can then be calculated following Marks and Houze (1987):

$$w = W - v_t$$

(4)

where $W$ is Doppler velocity (m s$^{-1}$) and $v_t$ (m s$^{-1}$) is the precipitation particle fallspeed. $v_t$ was determined using an empirical relationship among reflectivity and altitude relative to the estimated melting level (Marks and Houze 1987). Determining $v_t$ depends on precipitation type and storm region, with magnitudes of uncertainties in $v_t$ ranging from 1.0–1.5 m s$^{-1}$ in ice and 1–3 m s$^{-1}$ for graupel in convective regions near the melting layer and 3–10 m s$^{-1}$ in rain (Marks and Houze 1987). Therefore, a temporal (spatial) average of 30 s (5 km) was necessary to obtain meaningful estimates of $w$. Subsequent to smoothing and post-processing, VI data has a 2–4 s temporal, 1.6 km horizontal, and 0.3 km vertical resolution.

Slight differences exist among the various radar and flight-level measurements. Since $w$ is computed via integration of divergence using data points 3 km apart from the TA radar, spatial averaging over an area of 9 km$^2$ results in $w$ values and locations that might differ from both VI
data and flight-level data (Marks et al. 1992; Reasor et al. 2009; Rogers et al. 2012). In comparing w distribution composites from numerous storms, Rogers et al. (2012) found decent agreement between VI and TA w, with only slight differences due to the VI data’s finer spatial resolution. Reasor et al. (2009) and Marks et al. (1992) similarly discovered small differences between TA and flight-level w, with differences on the order of 0.4–5 m s⁻¹. However, according to Reasor et al. (2009), TA w exhibited a positive bias (i.e., stronger updrafts), but Marks et al. (1992) found a negative bias (stronger downdrafts). This discrepancy was attributed to differences in navigation correction techniques, scanning technique, or synthesis technique (Reasor et al. 2009).

3.3. Dropsondes

From 0800 UTC 23 August–0200 UTC 25 August, the USAF C-130, NOAA P3, and NASA DC-8 aircrafts dropped 71 dropsondes within 150 km of the center in Hurricane Bonnie. Missing data reduced the number of usable sondes to 65. Dropsondes collected T, T_d, u, v, w, and pressure measurements from flight level (600 or 700 mb, depending on the aircraft) to the sea surface and were interpolated 100 m in the vertical. Using Willoughby and Chemlow’s (1982) storm center detection method, storm motion and sonde drift were removed so dropsonde data is storm relative. P-3 and USAF dropsonde data was post-processed and quality-controlled using the Editsonde software developed by HRD, while DC-8 dropsondes were post-processed using the National Center for Atmospheric Research’s (NCAR’s) Atmospheric Sounding Processing Environment (ASPEN) software. As mentioned by Powell (2006), work conducted by Aberson and Black suggested the superiority in using Editsonde for research purposes, but Barnes and Fuentes (2010) found little difference between the processing from Editsonde and ASPEN. Based on the latter results, all dropsonde data were combined.
Of the 65 dropsondes examined for this study, a majority sampled late on 23 and 24 August when all four aircraft were flying, with sparser dropsonde observations earlier on both days (Table 3.1). Based on the azimuthal and radial dropsonde distribution (Fig. 3.4), dropsonde composites were created for the secondary eyewall region. The dropsonde data provided essential information about how the thermodynamic and kinematic structure of Bonnie evolved during its ERC.

3.4. Additional data

Vertical wind shear data was obtained from the Statistical Hurricane Intensity Prediction Scheme (SHIPS). Shear is calculated in SHIPS as the magnitude difference between 850 and 200 hPa wind vectors, averaged over a 600 km radius from the vortex center. 850–200 hPa shear magnitude and direction is available every six hours during Bonnie’s lifecycle (DeMaria and Kaplan 1994).

Additional tropical cyclone environmental and intensity data was obtained from variables included in Kossin and Sitkowski’s (2009) ERC prediction model. Eighty-two different features taken from the SHIPS dataset were fed into the ERC prediction model, which covered a time period from 1997–2006 and included 45 SEF events (Kossin and Sitkowski 2009) among North Atlantic Hurricanes. Statistics of these features for SEF events and non-events were compared with those of Hurricane Bonnie to better place Bonnie’s ERC in context with other TCs.

3.5. Hovmöllers

Radius versus time Hovmöllers were created using USAF (700 mb) flight data in order to understand the evolution of axisymmetric quantities (v, \( \zeta \), u, and \( m_a \)). Flight legs from four different flights (i.e., flights A1–A4) were placed into nine six-hourly time bins. This resulted in
six to nine flight legs in five of the time bins and one to two legs in four of the time bins (Table 3.2). Flight legs were averaged together in each time bin to provide ample azimuthal coverage, except for time bins consisting of only one to two legs, for which linear interpolations of the previous and subsequent six hour time bins were taken. Hovmöller plots display the midpoint of each time bin (i.e., at hour 3).

Azimuth vs. time Hovmöllers were also created using LF reflectivity data obtained from flights P1 and P2 (600 mb). Six LF images were collected during this time frame and placed into one hour bins between 2015 UTC 23 August–0021 UTC 24 August (Table 3.3). This resulted in one to two radar sweeps in each bin, except for the time bin at 2215 UTC that contained no radar sweeps and was linearly interpolated from neighboring time bins to mitigate the lack of data. Separate azimuth vs. time Hovmöllers were created for the primary eyewall and secondary eyewall by averaging over their respective radii (i.e., 71–115 km for the secondary eyewall and 20–51 km for the primary eyewall) to understand how their azimuthal coverage changed with time.
Table 3.1: Description of flight level data used in this study of Hurricane Bonnie (1998) from 23–25 August. 

#Pass refers to the number of center passes in each flight, # drops refers to the number of dropsondes dropped, LF (TA) stands for lower-fuselage (tail) radar, Dual Dop stands for Dual-Doppler, and VI stands for vertical incidence. See text for details.

<table>
<thead>
<tr>
<th>Flight</th>
<th>#Pass</th>
<th>Pressure (mb)</th>
<th>Start Time</th>
<th>End Time</th>
<th>Middle Time</th>
<th># Drops</th>
<th>LF</th>
<th>TA</th>
<th>Dual Dop</th>
<th>VI</th>
</tr>
</thead>
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<tr>
<td>A1</td>
<td>6</td>
<td>700</td>
<td>0813 UTC</td>
<td>1733 UTC</td>
<td>0823 UTC</td>
<td>12</td>
<td>No</td>
<td>No</td>
<td>No</td>
<td>No</td>
</tr>
<tr>
<td>A2</td>
<td>10</td>
<td>700</td>
<td>0823 UTC</td>
<td>0718 UTC</td>
<td>0824 UTC</td>
<td>7</td>
<td>No</td>
<td>No</td>
<td>No</td>
<td>No</td>
</tr>
<tr>
<td>P1</td>
<td>8</td>
<td>600</td>
<td>1736 UTC</td>
<td>0230 UTC</td>
<td>0824 UTC</td>
<td>24</td>
<td>Yes</td>
<td></td>
<td>Yes</td>
<td>No</td>
</tr>
<tr>
<td>P2</td>
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<td>600</td>
<td>1726 UTC</td>
<td>0240 UTC</td>
<td>0824 UTC</td>
<td>-</td>
<td>Yes</td>
<td>Normal</td>
<td>No</td>
<td>Yes</td>
</tr>
<tr>
<td>A3</td>
<td>8</td>
<td>700</td>
<td>0757 UTC</td>
<td>1848 UTC</td>
<td>0824 UTC</td>
<td>7</td>
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<td>No</td>
<td>No</td>
<td>No</td>
</tr>
<tr>
<td>A4</td>
<td>10</td>
<td>700</td>
<td>1951 UTC</td>
<td>0643 UTC</td>
<td>0825 UTC</td>
<td>6</td>
<td>No</td>
<td>No</td>
<td>No</td>
<td>No</td>
</tr>
<tr>
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<td>600</td>
<td>1834 UTC</td>
<td>0200 UTC</td>
<td>0825 UTC</td>
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<td>Normal</td>
<td>Yes</td>
<td>Yes</td>
</tr>
<tr>
<td>P4</td>
<td>10</td>
<td>850</td>
<td>1825 UTC</td>
<td>0310 UTC</td>
<td>0825 UTC</td>
<td>-</td>
<td>Yes</td>
<td>FAST</td>
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<td>No</td>
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</tbody>
</table>
Figure 3.1. 2300 UTC 23 August infrared satellite image overlaid with National Oceanic and Atmospheric Administration’s P3 (blue and white), U.S. Air Force (pink), and DC8 (black) flight tracks +/- 3 hours of the satellite imagery. The blue asterisk denotes Hurricane Bonnie’s (1998) center position.
Figure 3.2. a) Raw axisymmetric tangential wind (m s\(^{-1}\)) and b) axisymmetric tangential wind fit with a 10th order polynomial for three separate U.S. Air Force (700 mb) flights: 0813–1733 UTC 23 August (Flight A1, red), 1956 UTC 23 August–0718 UTC 24 August (flight A2, black), and 0757–1848 UTC 24 August (flight A3, blue). Six to ten radial legs were averaged for each flight to obtain an axisymmetric quantity.
Figure 3.3. Schematic of the tail-band Doppler radar displaying its two scanning modes: normal, when the radar scans in 360° cones perpendicular to the long axis of the plane, and forward/aft scanning technique (F/AST), in which these conical scans are tilted 20° fore or aft normal to the axis of rotation (Gamache and Marks 1995, their Fig.1. ©American Meteorological Society. Used with permission.).

Figure 3.4. Distribution of 65 dropsondes relative to Bonnie’s center between 1200 UTC 23 August–05 UTC 25 August, with range rings every 50 km. Blue dots denote dropsondes in the primary eyewall (20–50 km; 1200 UTC 23 August–1400 UTC 24 August), red dots are dropsondes in the secondary eyewall (70–115 km; 1200 UTC 23 August–1400 UTC 24 August), and white dots are at all other times.
Table 3.2. Distribution of U.S. Air Force flight legs in the six hourly time bins used to create radius versus time Hovmöllers of tangential wind, relative vorticity, angular momentum, and radial wind. Shaded rows denote time bins in which a linear interpolation was performed using the previous and subsequent time bins.

<table>
<thead>
<tr>
<th>Day</th>
<th>Time bin (UTC)</th>
<th># flight legs</th>
</tr>
</thead>
<tbody>
<tr>
<td>8/23</td>
<td>1200–1800</td>
<td>6</td>
</tr>
<tr>
<td>8/23 – 8/24</td>
<td>1800–0000</td>
<td>1</td>
</tr>
<tr>
<td>8/24</td>
<td>0600–1200</td>
<td>1</td>
</tr>
<tr>
<td>8/24</td>
<td>1200–1800</td>
<td>7</td>
</tr>
<tr>
<td>8/24 – 8/25</td>
<td>1800–0000</td>
<td>2</td>
</tr>
<tr>
<td>8/25</td>
<td>0000–0600</td>
<td>8</td>
</tr>
<tr>
<td>8/25</td>
<td>0600–1200</td>
<td>1</td>
</tr>
<tr>
<td>8/25</td>
<td>1200–1800</td>
<td>7</td>
</tr>
</tbody>
</table>

Table 3.3. As in Table 3.2, except for one hourly time bins of lower fuselage radar sweeps from NOAA P-3 aircraft used to create azimuth vs. radius Hovmöllers of reflectivity.

<table>
<thead>
<tr>
<th>Day</th>
<th>Time bin (UTC)</th>
<th># flight legs</th>
</tr>
</thead>
<tbody>
<tr>
<td>8/23</td>
<td>2000–2100</td>
<td>2</td>
</tr>
<tr>
<td>8/23</td>
<td>2100–2200</td>
<td>2</td>
</tr>
<tr>
<td>8/23</td>
<td>2200–2300</td>
<td>0</td>
</tr>
<tr>
<td>8/23</td>
<td>2300–0000</td>
<td>1</td>
</tr>
<tr>
<td>8/24</td>
<td>0000–0100</td>
<td>1</td>
</tr>
</tbody>
</table>
4. Results

4.1. Symmetric ERC evolution

Radial profiles of $\bar{v}$ and $\bar{\zeta}$ obtained from three consecutive USAF flights from 23–24 August capture three distinct stages in Bonnie’s ERC. During flight A1 (Fig. 4.1), a single $\bar{v}_{\text{max}}$ of 45 m\,s\(^{-1}\) was located at a radius of 35 km. Winds gradually decreased beyond this radius and flattened out to ~35 m\,s\(^{-1}\) beyond 90 km. Though there was a distinct, single $\bar{v}_{\text{max}}$, the flat wind profile outside of this is unlike the more rapid $\bar{v}$ decrease that characterizes intense hurricanes (Mallen et al. 2005) and was suggestive of the forthcoming increase in winds at outer radii. This increase was apparent during the next flight, flight A2, in which a secondary $\bar{v}_{\text{max}}$ of 42 m\,s\(^{-1}\) formed and intensified at 90 km. Concurrently, the primary $\bar{v}_{\text{max}}$ weakened to 35 m\,s\(^{-1}\). In flight A3, the primary $\bar{v}_{\text{max}}$ disappeared and the secondary $\bar{v}_{\text{max}}$ became the only $\bar{v}_{\text{max}}$.

The series of events shown in Fig. 4.1 is typical for ERC events (Willoughby et al. 1982; S11; Didlake et al. 2016), except for the lack of secondary $\bar{v}_{\text{max}}$ contraction. However, no formal definition of what constitutes an ERC exists in the literature. Bonnie’s ERC commenced with the detection of a noticeable secondary $\bar{v}_{\text{max}}$ in one quadrant of the storm (see Chapter 4.3) when the primary $\bar{v}_{\text{max}}$ was distinct (flight A1; Fig. 4.1). SEF occurred in the middle of Bonnie’s ERC when the secondary $\bar{v}_{\text{max}}$ projected onto the axisymmetric mean at 90 km (flight A2; Fig. 4.1). The conclusion of Bonnie’s ERC was signified by the lack of the primary $\bar{v}_{\text{max}}$ at 35 km (flight A3; Fig. 4.1). This ERC definition used in Bonnie is fairly consistent with S11.

Changes in Bonnie’s $\bar{v}$ structure during its ERC are characterized by $\alpha$ (Table 4.1). Prior to flight A1, Mallen et al. (2005) calculated Bonnie’s $\alpha = 0.35$ from 0300–0600 UTC 23 August. This value is fairly close to the median for all tropical cyclones, but lower than that of major
hurricanes (α = 0.48; Mallen et al. 2005). For the radial profile during flight A1, α = 0.26 and during flight A2, α = -0.12 (since the outer wind maximum exceeded the inner). The decreased α value in Bonnie from 0.35 to 0.26 between 0300 and 0600 UTC 23 August and Flight A1 indicates a broadened $\bar{v}$ profile. Bonnie’s α values during flight A1 and A2 were lower than the median α value of 0.37 for Atlantic and Eastern Pacific tropical cyclones (Mallen et al. 2005), indicating a broader than average vortex.

A broadening of $\bar{v}$ prior to SEF has been noted in several studies (Huang et al. 2012; Abarca and Montgomery 2013; Zhu and Zhu 2014; Zhang et al. 2017). As $\bar{v}$ broadened in Bonnie, shear increased from 9 to 13 m s$^{-1}$. Whether the vortex broadened in response to the increased shear or independently of the shear is unable to be addressed due to lack of observations between Mallen et al.’s (2005) 0300–0600 UTC time frame and flight A1, but would be an interesting question to address in future research.

A radius vs. time Hovmöller displays the changes in $\bar{v}$ from a different perspective. The weakening of the primary $\bar{v}_{\text{max}}$ at 40 km radius is observed from 1500 UTC August 23–1500 UTC 24 August, over the course of the entire ERC (Fig. 4.2). The secondary $\bar{v}_{\text{max}}$ became evident at approximately 0000 UTC 24 August at a radius of 90 km. Upon its appearance, the secondary $\bar{v}_{\text{max}}$ was the same 45 m s$^{-1}$ intensity of the primary $\bar{v}_{\text{max}}$, which deviates from prior studies in which the secondary $\bar{v}_{\text{max}}$ during formation is usually weaker than the primary $\bar{v}_{\text{max}}$ (S11; Zhu and Zhu 2014; Didlake et al. 2016). Throughout the rest of the ERC (i.e., until 1400 UTC 24 August), the secondary $\bar{v}_{\text{max}}$ maintained its initial 45 m s$^{-1}$ intensity. There was also little change in the location of the secondary $\bar{v}_{\text{max}}$, as it remained centered around 80–90 km until the end of the ERC at 1400 UTC 24 August.
Accompanying changes in $\nabla$ during Bonnie’s ERC were changes in axisymmetric relative vorticity ($\bar{\zeta}$). Prior to SEF during flight A1 (Fig. 4.3), maximum $\bar{\zeta}$ (24*10^{-4} s^{-1}) was located at the 20 km radius, inwards of the primary $\bar{v}_{\text{max}}$, and sharply declined from this maximum out to 55 km. From r (radius) = 55–100 km, $\bar{\zeta}$ more gradually decreased and displayed small variations beyond 100 km. This negative radial gradient in $\bar{\zeta}$, or beta skirt, is consistent with the vorticity structures seen in other mature hurricanes (Mallen et al. 2005). During SEF in flight A2, $\bar{\zeta}$ noticeably increased to approximately 7*10^{-4} s^{-1} from r = 75–90 km, corresponding to the strengthening of the secondary $\bar{v}_{\text{max}}$. The decreased inner $\bar{\zeta}$ maximum (15*10^{-4} s^{-1}) reflects the weakened inner $\bar{v}_{\text{max}}$. By flight A3, $\bar{\zeta}$ exhibited a broad decrease with radius, consistent with a broader circulation post-ERC. Notice that throughout Bonnie’s ERC, the structure of $\bar{\zeta}$ fundamentally changed.

Variations in $\bar{\zeta}$ are also seen in a radius vs. time Hovmöller (Fig. 4.4). An initial $\bar{\zeta}$ maximum corresponding to the primary $\bar{v}_{\text{max}}$ weakened throughout the entire ERC. This weakening was most rapid prior to the development of the secondary $\bar{v}_{\text{max}}$ (Fig. 4.2) at 0000 UTC 24 August. While the inner $\bar{\zeta}$ maximum weakened during the ERC, an outer $\bar{\zeta}$ maximum developed from 60–80 km at 0000 UTC 24 August, just inside of where the secondary $\bar{v}_{\text{max}}$ formed (Fig. 4.2). This outer $\bar{\zeta}$ maximum remained elevated throughout Bonnie’s ERC.

Possible contributions to the development of the secondary $\bar{v}_{\text{max}}$ were axisymmetric $\bar{m}_a$ and $\bar{u}$, shown in Fig. 4.5 and 4.6, respectively. Prior to and during SEF, contours of $\bar{m}_a$ bowed inwards (Fig. 4.5), suggesting an inward advection of $\bar{m}_a$. Given the tight radial gradient in $\bar{m}_a$ from 60–100 km in radius, this advection was likely substantial. This result is consistent with $\bar{m}_a$ advection contributing to the development of a secondary $\bar{v}_{\text{max}}$. The tight radial gradient and
inward bowing $\bar{m}_a$ contours existed even after the conclusion of the ERC, through 1500 UTC 25 August.

Moderate (2–4 m s$^{-1}$) inflow outside of 100 km in radius (Fig. 4.6) likely assisted the inward advection of $\bar{m}_a$ and formation of secondary $\bar{v}_{\max}$. This inflow persisted through the entire ERC. Inside of the inflow maximum was weak 0–1 m s$^{-1}$ outflow in the vicinity of the primary $\bar{v}_{\max}$ (less than 50 km in radius) through 0700 UTC 24 August, which became weak 0–1 m s$^{-1}$ inflow through the end of the ERC. Notice that between the primary and secondary $\bar{v}_{\max}$ there was moderate 2–4 m s$^{-1}$ outflow centered at 70 km that maximized from 0100–0900 UTC 24 August, just after the development of the secondary $\bar{v}_{\max}$. Interestingly, post-ERC, inflow strengthened to 3 m s$^{-1}$ inside 70 km and to 5 m s$^{-1}$ beyond 110 km from 1500 UTC 24 August–0300 UTC 25 August. This may have fueled intense convective cells noted during that time (Molinari and Vollaro 2008).

4.2. Comparison with other ERCs

The three distinct stages in Bonnie’s ERC (based on S11) are denoted in a time series of shear magnitude and $\bar{v}_{\max}$ (Fig. 4.7a). Bonnie’s ERC started with the detection of a secondary $v_{\max}$ downshear at 1400 UTC 23 August just following the rapid intensification phase and increase in shear to 12 m s$^{-1}$. SEF occurred just 10 h later at 0000 UTC 24 August at which time shear was 14 m s$^{-1}$. Fourteen hours later at 1400 UTC 24 August, during Bonnie’s steady-state-high-shear phase, the inner $\bar{v}_{\max}$ disappeared to conclude Bonnie’s ERC.

Several differences exist (Fig. 4.7) between intensity changes associated with Bonnie’s ERC and S11’s conceptual model (Fig. 4.7b, their Fig. 4.2). In S11, an asymmetric secondary $v_{\max}$ is first detected during a rapid intensification phase, while in Bonnie, this detection occurred
subsequent to intensification. However, S11 notes that out of twenty-four ERCs events, seven did not have an intensification phase because the primary $\bar{v}_{\text{max}}$ had already attained peak intensity, which was the case in Bonnie. During the phase in which Bonnie’s SEF occurs, there was no change in intensity, yet in S11, this same stage is associated with weakening due to the decay of the primary $\bar{v}_{\text{max}}$. To conclude the ERC, S11 noted a typical reintensification phase during which the secondary (now symmetric) $\bar{v}_{\text{max}}$ exceeds the primary in terms of intensity and contracts inward (S11; Fig. 4.7b). Yet in Bonnie, no reintensification occurred, as Bonnie maintained 51 m s$^{-1}$ intensity as the primary $\bar{v}_{\text{max}}$ vanished.

Bonnie’s unusual ERC is further highlighted when compared to variables from Kossin and Sitkowski’s (2009) ERC predictor model in ERCs in 45 North Atlantic TCs (Table 4.2). While most of the mean values for these variables are similar between TCs with ERCs and Bonnie (at 0000 UTC 24 August), those related to shear (200 mb zonal wind, 850–200 mb shear magnitude, and standard deviation of GOES infrared brightness temperature) clearly differ. 200 mb zonal wind was 17 kt in Bonnie and only 11 kt for TCs with ERCs, shear magnitude was 29 kt in Bonnie and 11 kt in other ERCs, and standard deviation of GOES infrared brightness temperature was 16.7°C in Bonnie and 12.6°C in other TCs. The 0–600 km average symmetric tangential wind was also stronger in Bonnie at 17.3 m s$^{-1}$, whereas it was 10.9 m s$^{-1}$ in other TCs with ERCs, suggestive of Bonnie’s broader vortex (see Chapter 4.1).

Bonnie’s anomalous values are apparent in probability density functions (PDFs; Fig. 4.8) of these variables for TCs with ERCs, without ERCs, and Bonnie’s value at a single time. For both 200 mb zonal wind (Fig. 4.8a) and shear magnitude (Fig. 4.8b), Bonnie falls outside the highest range usually encountered in TCs with ERCs. Bonnie’s standard deviation of GOES infrared brightness temperature (Fig. 4.8c) and 0–600 km average symmetric tangential wind
(Fig. 4.8d) are on the upper-end of the distribution. Thus Bonnie, unlike most TCs, experienced higher shear, was more asymmetric, and broader during its ERC.

4.3. Asymmetric evolution

To visualize Bonnie’s asymmetric evolution, SSMI 85-GHz microwave images overlaid with shear magnitude and direction were examined (Fig. 4.9; Naval Research Laboratory: https://www.nrlmry.navy.mil/tc-bin/tc_home2.cgi ). The 85-GHz frequency detects ice, and thus, deep convective clouds, but cannot always detect low-level clouds at this frequency (Cecil et al. 2002). Data was retrieved from the F14 satellite for all SSMI images, except for 1200 UTC 22 August, which was collected by the F11 satellite (Wentz et al. 2015).

At ~0100 UTC August 22 (Fig. 4.9a), a large area of convection emanating from the north beyond 250 km wrapped radially inwards to approximately 100 km in the southwest. The beginning of an eyewall was apparent at a radius of 50 km to the south. Eleven hours later while in low (3 m s\(^{-1}\)) shear (Fig. 4.9b), the eyewall convection organized and intensified, while multiple rainbands existed beyond 150 km, with the most notable located to the north of center. Shear increased from 3 to 9 m s\(^{-1}\) from the west from 1200 UTC 22 August–0100 UTC 23 August (Fig. 4.9c) and rainbands were displaced downshear. At this time, the eyewall formed near the 50 km radius, with the most intense convection located downshear. A rainband that spiraled inwards from 150 km southwest to 100 km northeast also exhibited a convective maximum downshear. As shear increased an additional 3 m s\(^{-1}\) to reach a magnitude of 12 m s\(^{-1}\) by 1330 UTC 23 August (Fig. 4.9d), Bonnie took on a highly asymmetric structure, with both inner core convection and an outer stratiform (brightness temperatures \(> 220\) K in green, suggested by Cecil et al. 2002) area located almost exclusively downshear, as it might be expected in high shear. It is interesting that this broad stratiform area with embedded convection
at the 150 km radius to the southeast appeared to develop from the outer spiral rainbands observed 12 hours prior. By 0029 UTC 24 August (Fig. 4.9e), the convection in the broad stratiform area intensified downshear and wrapped radially inwards to ~80–120 km upshear. This quasi-circular banded feature (brightness temperatures > 240 K) was the nascent secondary eyewall forming and overtaking the downshear primary eyewall which was located at 30–50 km. Just 13 h later (Fig. 4.9f), the primary eyewall was no longer evident while the secondary eyewall (now the only eyewall) remained intense, indicating the conclusion of Bonnie’s ERC.

LF radar images (Fig. 4.10) provide a detailed evolution covering a four-hour period as the secondary eyewall developed. At 2012 UTC 23 August (Fig. 4.10a), an intense and highly asymmetric primary eyewall was located downshear-left at approximately 35 km. A strong shear signature is seen by the lack of convection upshear in the core. The primary eyewall remained intense and anchored downshear-left throughout the four-hour period. Surrounding the primary eyewall at 2012 UTC was a rainband that wrapped around from downshear-right to upshear-left from 80–120 km (Fig. 4.10a). This rainband intensified and organized from 2115 UTC (Fig. 4.10c)–2350 UTC 23 August (Fig. 4.10e), growing in size with > 30 dBZ reflectivity from 75–150 km. The most intense rainband region to the north/northeast (left-of-shear) was centered at the 100 km radius at 2350 UTC 23 August. By 0021 UTC 24 August (Fig. 4.10f), the rainband transitioned to a secondary eyewall, as evidenced by the nearly concentric ring of strong reflectivity extending from downshear-right at 100 km to upshear-left at 90 km.

To highlight azimuthal variations in the primary eyewall and secondary eyewall, azimuth vs. time Hovmöllers of reflectivity were created for both the primary eyewall and secondary eyewall (Fig. 4.11). The narrow azimuthal confinement of the primary eyewall is evident during SEF with the maximum 30–40 dBZ reflectivity consistently centered to the east (downshear; Fig.
4.11a). The primary eyewall also weakened from 40 dBZ to 35 dBZ between 2000–2200 UTC 23 August and reintensified to 40 dBZ at approximately 0000 UTC 24 August. In contrast, the secondary eyewall covered a greater azimuthal extent (Fig. 4.11b) with 25–35 dBZ covering from south (right-of-shear) to north (left-of-shear). The area of maximum intensity (35 dBZ) to the northeast at 2100 UTC 23 August expanded over time to cover the region from south to north by 0000 UTC 24 August.

Downshear and upshear dropsonde composites of u, v, windspeed, and \( \Theta_e \) in the SEF region (70 – 100 km) during SEF (2000 UTC 23 August–0020 UTC 24 August) are seen in Fig. 4.12. Downshear was characterized by a maximum of 17 m s\(^{-1}\) inflow at 975 mb that decreased to zero at 775 mb (Fig. 4.12a). Upshear displayed a 15 m s\(^{-1}\) inflow maximum at the surface, falling to zero at 900 mb. Tangential velocity (Fig. 4.12b) also exhibited shear-related asymmetries, with downshear having a deeper layer of considerable 40–45 ms\(^{-1}\) magnitude from 950–725 mb, whereas upshear displayed these high velocities over a shallower layer from ~950–850 mb. Both inflow and tangential velocity were lower downshear than upshear at the surface, which contributed to a sharply lower surface windspeed downshear (25 m s\(^{-1}\)) than upshear (38 m s\(^{-1}\); Fig. 4.12c). Unlike the velocity profiles, shear related differences in \( \Theta_e \) were less discernable (Fig. 4.12d). \( \Theta_e \) was nearly the same at the surface and at 600 mb, and downshear was only a couple of degrees higher than upshear.

Radial profiles of v and \( \zeta \) are examined as in Chapter 4.1, except stratified by shear-relative quadrant (Fig. 4.13). During flight A1 (Fig. 4.13a), an inner \( v_{\text{max}} \) associated with the primary eyewall was located between the 25–35 km radii in all quadrants. A secondary \( v_{\text{max}} \) was located at 120 km but most prominent only in the downshear-left quadrant. Though a distinct secondary \( v_{\text{max}} \) did not appear downshear-right, a flat v profile from 70–100 km was suggestive of a
forthcoming secondary $v_{\text{max}}$. Profiles of $\zeta$ confirm a developing downshear secondary $v_{\text{max}}$, with a secondary $\zeta_{\text{max}}$ located at 80 km (110 km) downshear-right (downshear-left; Fig. 4.13b). Notice that in addition to the lack of a clear secondary $\zeta_{\text{max}}$, upshear profiles displayed a primary $\zeta_{\text{max}}$ 10 km radially outwards of those downshear. This difference, in addition to the secondary $v$ and $\zeta$ downshear maxima, was likely due to asymmetric forcing by shear.

By flight A2 (Fig. 4.13c), secondary $v$ and $\zeta$ maxima appeared in all quadrants of the storm. This secondary $v_{\text{max}}$ was focused at 90–95 km upshear and 80 km downshear. Secondary $\zeta_{\text{max}}$ (Fig. 4.13d) were located radially inwards of the secondary $v_{\text{max}}$ at 80 km upshear and 65–70 km downshear. The appearance of secondary $v$ and $\zeta$ maxima in all quadrants of the storm during flight A2 shows that the initial downshear maxima were able to extend upshear to form a secondary eyewall between the two flights. Although this upshear extension was able to occur, shear still had an asymmetric effect on the $v$ and $\zeta$ profiles during Flight A2, as values for these kinematic quantities were slightly higher downshear than upshear. This result may be expected, given Bonnie’s asymmetric precipitation structure at this time (Fig. 4.10).

4.4. Concentric eyewall structure- downshear

Vertical profiles of the primary and secondary eyewall are examined downshear using cross-sections from the VI data and TA radar. During SEF from 2141–2152 UTC 23 August (Fig. 4.14a), a deep intense primary eyewall was characterized by high reflectivity ($> 30$ dBZ) up to 12 km in height and 35–45 km in radius. Strong updrafts ($> 4$ m s$^{-1}$; Fig. 4.14b) centered at 35 km fed the primary eyewall up to the melting level ($\sim 5$ km), and maximized at 8 m s$^{-1}$ above 6 km, likely due to latent heat of fusion (Molinari et al. 2012). These vigorous updrafts were flanked by downdrafts on either side that maximized above, and near, the melting level as
hydrometeors fell out of the intense updrafts. Outside of the primary eyewall from 50–60 km was the moat region containing an area of depressed reflectivity (Fig. 4.14a), where most values fell between 15–20 dBZ. Downdrafts (0–2 m s\(^{-1}\)) existed above 6 km in the moat region (Fig. 4.14b), with weak 0–2 m s\(^{-1}\) updrafts below this.

Two hours later from 2338–2359 UTC 23 August (Fig. 4.14c and d), marked changes in reflectivity and w occurred in association with SEF. The primary eyewall weakened, as seen by a decreased area of >30 dBZ reflectivity that was confined to 2–6 km in height at the 40 km radius (Fig. 4.14c). While updrafts intensified at upper levels to > 10 m s\(^{-1}\) (Fig. 4.14d), low-level updrafts weakened considerably to 0–2 m s\(^{-1}\). This change in w is significant because decreased low level updrafts could not continue to support convection in the primary eyewall. Compared to the previous time, precipitation in the moat increased (Fig. 4.14c), particularly below 2 km, where reflectivity values reached 38 dBZ. Most notably, the moat took on a stratiform signature (Marks and Houze 1987; Fig. 4.14d), with weak (0–2 m s\(^{-1}\)) updrafts occurring above the melting level and strong downdrafts (2–6 m s\(^{-1}\)) below this.

Another way in which to view Bonnie’s vertical structure during SEF is using TA radar cross-sections overlaid with u and w wind vectors (i.e., the secondary circulation; Fig. 4.15). The slightly different position of the eyewall reflectivity maxima and a greater radial range compared to the VI data was due to differences in scanning technique and resolution (Chapter 3.2). At 2152 UTC 23 August (Fig. 4.15a), the deep, intense primary eyewall was observed as in Fig. 4.14a, and a forming secondary eyewall was characterized by a > 30 dBZ shallow reflectivity tower at 100 km. While some u and w data was missing during this time, outflow from r = 40–50 km above the 6 km level was seen in the primary eyewall and weak, descending inflow was observed from 50–60 km below 6 km. Outside of the primary eyewall to approximately 70 km,
inflow occurred at lower and upper levels. In the vicinity of the forming secondary eyewall, from 80–90 km (no u and w wind data exists beyond 90 km), a weak secondary circulation was developing – inflow with a slight upward component appeared below 2 km and upward, and outward, turning air occurred above 3 km.

Two hours later at 2348 UTC 23 August (Fig. 4.15b), a weakened primary eyewall and strengthened secondary eyewall is apparent in the reflectivity and secondary circulations. While strong updrafts remained in the primary eyewall, with inner-edge updrafts below the melting level transitioning to strong outflow above, inflow weakened outside of the primary eyewall. This occurred coincident with a dissipating primary eyewall reflectivity tower, suggesting that the nascent secondary eyewall choked off the primary eyewall from necessary inflow. The recently formed secondary eyewall intensified and contracted over the two hours, as a > 36 dBZ reflectivity tower from the surface to 4 km was centered at 85 km. Associated with the intensified secondary eyewall was a stronger secondary circulation, seen by the in-up-out pattern in the u and w winds. Particularly strong outflow from this secondary circulation occurred above 4 km, consistent with a more mature secondary eyewall.

4.5. Concentric eyewall structure–3D Doppler analyses

Three-dimensional Doppler analyses from the TA radar centered at approximately 0000 UTC 24 August provide further insight into the double eyewall structure. As was observed from the LF radar, the TA radar depicts an intense primary eyewall to the east and a secondary eyewall wrapping around cyclonically from west to north, with maximum reflectivity to the north (Fig. 4.16). The primary eyewall was deep (Fig. 4.16d), with reflectivities greater than 20 dBZ extending up to 11 km. The secondary eyewall’s reflectivity structure was also coherent up to 11 km in its northern section, as seen by the greater than 12 dBZ band (Fig. 4.16d). Broad
windspeed maxima over 40 m s\(^{-1}\) were located on the eastern and northern side of Bonnie at 2 km in altitude (Fig. 4.17a), consistent with the location of primary and secondary eyewall reflectivity maxima. Above 2 km, windspeed maxima were more discrete (Fig. 4.17b–d), but still predominantly seen on the eastern and northern side of Bonnie (except at 11 km, where a speed maximum is located to the west).

North–south cross sections (Fig. 4.18) display the vertical double eyewall structure, in which shear (and tilt) induced asymmetries are apparent. To the north (left-of-shear), the inner eyewall is seen by a narrow reflectivity tower at approximately 30 km in radius extending up to 16 km (Fig. 4.18a). A broader area of high reflectivity from 70–125 km in radius characterized the secondary eyewall. Within the secondary eyewall, there existed a maximum (>35 dBZ), outwardly tilted reflectivity tower with slightly lower (30 dBZ), but still elevated, reflectivity radially outside. To the south (right-of-shear), reflectivity was generally more suppressed, especially in the primary eyewall region, where a coherent structure was lacking. However, a reflectivity maximum was located in the secondary eyewall region, at approximately 85–100 km to the south. Associated with these double eyewall reflectivity maxima were double wind maxima (Fig. 4.18a), most distinct around 3 km in height. These wind maxima were collocated with eyewall reflectivity maxima and ranged in magnitude from 40–50 ms\(^{-1}\) (40–45 ms\(^{-1}\)) to the north (south). The northern secondary eyewall, which exhibited the strongest convection, also contained the deepest wind maximum (up to 4 km in height).

A vertical velocity cross section (Fig. 4.18b) reveals a clear wavenumber-1 asymmetry between the shear-relative quadrants, particularly at the radius of the primary eyewall. Left-of-shear was characterized predominantly by downward motion, while right-of-shear contained mostly upward motion. Downdrafts were strongest at approximately 25–30 km in radius.
throughout most of the troposphere in the northern region, where values ranged from 2–5 m s\(^{-1}\). Updrafts maximized at 45–50 km in radius to the south through much of the troposphere, reaching magnitudes of 2–5 m s\(^{-1}\). In the vicinity of the secondary eyewall, around 75 km, both the north and south contained upward motion. However, to the south, these updrafts covered a broader region from approximately 60–110 km and only were 1–2 m s\(^{-1}\) at their maximum. To the north, a band of 0.2–5 m s\(^{-1}\) updrafts started at the surface 75 km in radius and ascended in height radially outwards to approximately 125 km in radius. Directly beneath this band of updrafts from 5–8 km in height were 0.5 to 2 m s\(^{-1}\) downdrafts from 85–100 km in radius. Downdrafts of this extent near the secondary eyewall were not observed to the south.

The vertical vorticity cross section (Fig. 4.18c) is consistent with the wavenumber one asymmetry observed in the previous plots. To the south, where there is predominantly rising motion, vorticity appeared enhanced compared to the north at approximately 20 km in radius. Vorticity was generally maximized within 50 km both north and south near the primary eyewall, consistent with observations of mature hurricanes (Mallen et. al 2005). Another region of positive vorticity existed outside of the primary eyewall, at 75 km in radius to the north. This positive vorticity band was nearly collocated with the ascending updrafts, windspeed maximum, and reflectivity maximum that characterized the secondary eyewall.
Figure 4.1. As in Fig. 3.2b, except where flight A2 (1956 UTC 23 August–0718 UTC 24 August) is labeled “SEF” and flight A3 (0757–1848 UTC 24 August) is labeled “ERC”. These labels denote times when secondary eyewall formation occurred and the eyewall replacement cycle completed, respectively.

Table 4.1. Alpha parameter ($\alpha$) for various times prior and during Bonnie’s secondary eyewall formation. Calculations follow Mallen et al. (2005; their Eqn. 3), with the * indicating a value that was calculated in Mallen et al. (2005).

<table>
<thead>
<tr>
<th>Time</th>
<th>$\alpha$</th>
</tr>
</thead>
<tbody>
<tr>
<td>0300–0600 UTC 23 Aug.</td>
<td>0.35*</td>
</tr>
<tr>
<td>0813–1733 UTC 23 Aug.</td>
<td>0.26</td>
</tr>
<tr>
<td>0757 UTC 23 Aug. – 1848 UTC 24 Aug.</td>
<td>-0.12</td>
</tr>
</tbody>
</table>
Figure 4.2. Radius vs. time Hovmöller of flight-level (700 mb) tangential wind (m s$^{-1}$) from 1500 UTC 23 August–1500 UTC 25 August.
Figure 4.3. As in Figure 4.1, but for relative vorticity (\(*10^{-4}\ \text{s}^{-1}\)). The gentle negative radial gradient in relative vorticity during flight A1 is denoted as the beta-skirt.
Figure 4.4. As in Fig. 4.2, but for relative vorticity ($10^{-4}$ s$^{-1}$).
Figure 4.5. As in Fig. 4.2, but for absolute angular momentum ($10^6 \text{ m}^2 \text{ s}^{-1}$).
Figure 4.6. As in Fig. 4.2, but for radial velocity (m s\(^{-1}\)).
Figure 4.7. A comparison of intensity changes during eyewall replacement stages in a) Hurricane Bonnie (from Fig. 2.2) and b) an adaption of Sitkowski et al.’s (2011) conceptual model (their Fig. 8. ©American Meteorological Society. Used with permission.). Open dots denoted the time when the outer wind maximum is first detected, grey dots denote the time of secondary eyewall formation (appearance of concentric rings on microwave imagery) in Hurricane Bonnie (Sitkowski et al. 2011), and black dots denote the time the inner wind maximum is gone. Dashed lines are used to separate these three stages by expected intensity change.
Table 4.2. Statistical Hurricane Intensity Prediction Scheme (SHIPS) features used in Kossin and Sitkowski’s (2009) Eyewall Replacement Cycle Predictor (pERC) model. The preference for an eyewall replacement cycle (ERC) for each feature is given, along with mean values for hurricanes undergoing and not undergoing an ERC (indicated by “ERC yes” and “ERC no”, respectively.) The values for these features in Bonnie at the time of secondary eyewall formation (0000 UTC August 24) is also shown for comparison. Data provided by Dr. James Kossin.

<table>
<thead>
<tr>
<th>SHIPS Feature</th>
<th>ERC yes (mean)</th>
<th>ERC no (mean)</th>
<th>Bonnie (0000 UTC 24 Aug.)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Current intensity</td>
<td>110.41 kt</td>
<td>86.17 kt</td>
<td>100 kt</td>
</tr>
<tr>
<td>Latitude</td>
<td>21.67°</td>
<td>27.04°</td>
<td>24.80°</td>
</tr>
<tr>
<td>Climatological depth of 26°C isotherm</td>
<td>95.43 m</td>
<td>66.48 m</td>
<td>54 m</td>
</tr>
<tr>
<td>200 mb zonal wind (200-800 km from center)</td>
<td>11.04 kt</td>
<td>8.62 kt</td>
<td>16.9 kt</td>
</tr>
<tr>
<td>500–300 mb relative humidity</td>
<td>50.21%</td>
<td>45.14%</td>
<td>49%</td>
</tr>
<tr>
<td>0–600-km average symmetric tangential wind at 850 mb from NCEP analysis</td>
<td>10.88 m/s</td>
<td>8.84 m/s</td>
<td>17.3 m/s</td>
</tr>
<tr>
<td>Azimuthally averaged surface pressure at outer edge of vortex</td>
<td>1013.9 mb</td>
<td>1015.5 mb</td>
<td>1012 mb</td>
</tr>
<tr>
<td>850–200 mb shear magnitude</td>
<td>11.13 kt</td>
<td>18.15 kt</td>
<td>29.1 kt</td>
</tr>
<tr>
<td>Maximum potential intensity</td>
<td>136.35 kt</td>
<td>118.73 kt</td>
<td>145 kt</td>
</tr>
<tr>
<td>St. dev (from axisymmetry) of GOES IR brightness temp between 100 and 300 km</td>
<td>12.59° C</td>
<td>16.61° C</td>
<td>16.7° C</td>
</tr>
<tr>
<td>Average GOES IR brightness temp between 20 and 120 km</td>
<td>-63.09° C</td>
<td>-54.74° C</td>
<td>-68.°8 C</td>
</tr>
</tbody>
</table>
Figure 4.8. Probability density function (PDF) of a) 200 mb zonal wind (kt), b) shear magnitude (SHRD; kt) c) standard deviation of bright temperature (°C) and d) 0–600-km average symmetric tangential wind (m s⁻¹) for TCs undergoing (“Yes ERC”) and not undergoing an ERC (“No ERC”) from Kossin and Sitkowski’s (2009) database of 45 North Atlantic hurricanes. The dashed black line indicates Bonnie’s value at the time of its secondary eyewall formation at 0000 UTC 24 August. Data and plotting routine for PDFs courtesy of Dr. James Kossin.
Figure 4.9. 85-GHz SSMI images of Hurricane Bonnie at a) 0054 UTC 22 August, b) 1209 UTC 22 August, c) 0041 UTC 23 August, d) 1326 UTC 23 August, e) 0029 UTC August 24, and f) 1314 UTC 24 August. Images are obtained from the Naval Research Laboratory (https://www.nrlmry.navy.mil/tc-bin/tc_home2.cgi). Shear direction and magnitude obtained from the Statistical Hurricane Intensity Prediction Scheme (SHIPS) are indicated by the white arrow and text and range rings are every 50 km, out to 250 km.
Figure 4.9. (Continued)
Figure 4.10. Lower fuselage radar reflectivity images during Bonnie’s secondary eyewall formation at a) 2012 UTC 23 August, b) 2047 UTC 23 August, c) 2115 UTC 23 August, d) 2152 UTC 23 August, e) 2350 UTC 23 August, and f) 0021 UTC 24 August. Range rings are every 50 km and shear direction is denoted by the black arrow. The black horizontal lines in d) and e) denote the location of vertical cross sections shown in Fig. 4.15 and 4.16.
Figure 4.10. (Continued)
Figure 4.11. Azimuth vs. time Hovmöllers of reflectivity in the a) primary eyewall (20–50 km) and b) secondary eyewall (70–115 km). Reflectivity was obtained from the lower fuselage radars onboard the NOAA P3 aircraft from 2000 UTC 23 August–0030 UTC 24 August.
Figure 4.12. Shear-relative (DS = downshear, US = upshear) dropsonde composites of a) radial wind, b) tangential wind, c) windspeed, and d) equivalent potential temperature in the secondary eyewall region (70–100 km). DS (US) composites include five (seven) dropsondes from the NOAA P-3 flights.
Figure 4.13. Shear relative profiles of a), c) tangential wind and b), d) relative vorticity from a), b) 0813–1733 UTC 23 August (flight A1) and c), d) 1956 UTC 23 August–0718 UTC 24 August (flight A2). A single flight leg was used for each radial profile. DSR = downshear right (green), DSL = downshear left (blue), USL = upshear left (red), USR = upshear right (black).
Figure 4.14. Vertical incident cross-sections east of Bonnie’s center at a), b) 2141–2152 UTC 23 August and c), d) 2338–2359 UTC 23 August of a), c) reflectivity and b), d) vertical velocity. Locations of cross sections from 2141–2152 UTC August (2338–2359 UTC 23 August) are indicated by the black horizontal line in Fig. 4.10d (e). Images were created with assistance from Nancy Griffin at the Hurricane Research Division.
Figure 4.15. Tail radar cross-sections east of center of reflectivity and secondary circulation (u and w wind vectors; magnitudes are indicated by the length of the arrows) at a) 2152 UTC 23 August and b) 2348 UTC 23 August. Cross sections are in the same location as Fig. 4.14.
Figure 4.16. Tail Doppler reflectivity at a) 2 km, b) 5 km, c) 8 km, and d) 11 km. These reflectivity analyses are centered at approximately 0000 UTC 24 August. Black north-south vertical lines denote the cross-section location used in Fig. 4.18. Images created by Dr. Robert Rogers.
Figure 4.17. As in Fig. 4.16, but for wind speed (color) and direction (arrows).
Figure 4.18. Tail Doppler cross-sections indicated by the black lines in Figs. 4.16 and 4.17, from south (S; left side of x-axis; right-of-shear) to north (N; side of x-axis; left-of-shear) of a) reflectivity (shaded; dBZ) and wind speed (contours; every 5 m s$^{-1}$), b) vertical velocity (m s$^{-1}$), and c) relative vorticity (10$^{-4}$ s$^{-1}$). Cross sections are centered at approximately 0000 UTC 24 August. Images created by Dr. Robert Rogers.
5. Discussion

5.1. Differences from typical ERCs

The strong shear-induced asymmetries observed in Hurricane Bonnie differ from the typical view of an ERC being largely symmetric. These differences are highlighted by Bonnie’s stronger 200 mb zonal wind, larger 850–200 mb shear magnitude, greater standard deviation of GOES-R brightness temperature, and stronger 0–600 km average symmetric tangential wind (Table 4.2) when compared to other hurricanes that underwent ERCs. Thus, Bonnie was anomalously broader, more asymmetric, and encountered higher shear during its ERC than other storms.

The effect of strong shear on Bonnie’s ERC was clearly seen during SEF, in which the secondary eyewall appeared to develop from localized convection within a highly asymmetric stratiform area (Fig. 4.9d–e). Dai et al. (2017) similarly observed this in their simulation of a vortex interacting with mid-latitude jet. The secondary eyewall was unable to form a completely concentric ring (Fig. 4.10f), with suppressed convection upshear right. This result is consistent with Reasor et al. (2013), who found that upshear right contained minimum precipitation among highly sheared storms. Upshear also contained lower \( v \) and \( \zeta \) in the secondary as well as primary eyewall (Fig. 4.13), as would be expected in a high-shear regime (Reasor et al. 2013).

Though Bonnie clearly displayed a wavenumber one asymmetry during its ERC, the extent and location of this asymmetry differed between the primary and secondary eyewall. Windspeed was higher left-of-shear compared to right-of-shear in both eyewalls (Fig. 4.18a), but the above surface windspeed maximum was deeper in the secondary eyewall than the primary. This result was likely due to the deeper layer of higher reflectivity in the secondary eyewall at
approximately 00 UTC 24 Aug. (Fig. 4.18a), which could have helped generate the deeper wind maximum. The left-of-shear reflectivity maximum seen in the secondary eyewall (Fig. 4.10, 4.16, and 4.18) was shifted downwind of the downshear primary eyewall reflectivity maximum (Fig. 4.10 and 4.16). Didlake et al. (2016) similarly saw a downwind shift in maximum reflectivity between the primary and secondary eyewall in a sheared storm, which was hypothesized to be the result of shear induced tilt and interactions with a rainband complex.

It is plausible that Didlake et al.’s (2016) hypothesis occurred in Bonnie’s double eyewall, but it does not explain why Bonnie’s primary eyewall was more asymmetric than the secondary. A similar structure was observed in Hence and Houze (2012) and is highlighted in Fig. 4.11, in which the secondary eyewall covered a greater azimuthal extent than the narrowly confined primary eyewall. Shear influenced the primary eyewall more than the secondary eyewall, as seen by the strong wavenumber one asymmetry in w near the primary eyewall region (Fig. 4.18b). Strong updrafts (downdrafts) at approximately 40 km (30 km) in radius characterized right-of-shear (left-of-shear) consistent with shear/tilt induced asymmetries (Jones 1995). However, in the vicinity of the secondary eyewall (~70–100 km in radius) this relationship was less discernable, given the presence of similar strength updrafts both right and left-of-shear.

Potential vorticity arguments might be able to explain the primary eyewall’s greater susceptibility to shear. Upshear subsidence is driven by negative potential vorticity advection by cross-storm flow (Bender 1997), and for a given cross-storm flow, advection is greatest where there is the largest potential vorticity radial gradient. Initially, during Bonnie’s ERC, the largest potential vorticity radial gradient likely existed in the primary eyewall, which resulted in greater subsidence. However, after the secondary eyewall replaced the primary, it would be expected
that the secondary eyewall was more susceptible to shear, due to its stronger potential vorticity gradient. Didlake et al. (2016) found evidence that near or after the secondary eyewall had replaced the primary, it exhibited shear-related asymmetries more similar to the initial primary eyewall. Thus, this potential vorticity explanation seems feasible, but further research would be needed to see if this was the case.

In addition to shear-related asymmetries, Bonnie further deviated from other ERCs in that it did not contract or change intensity. Once the secondary eyewall emerged, it remained relatively stationary at approximately 85 km in radius (Fig. 4.1 and 4.2). It is possible that strong shear prevented the secondary eyewall from contracting, but the mechanism by which this occurs is uncertain. The lack of intensity change in Bonnie (Fig. 4.7) was also anomalous, based on current literature of ERC behavior, and is addressed below in Chapter 5.4.

5.2. Similarities to typical ERCs

Despite the results from the previous section, Hurricane Bonnie was similar in many regards to “classic” ERCs, such as those observed by Willoughby et al. (1982), Houze et al. 2007, and S11. A robust asymmetric secondary $v_{\text{max}}$ was first detected in one area of the storm (downdraft) during flight A1 (Fig. 4.13a) to commence the ERC, as in S11. By flight A2, this asymmetric $v_{\text{max}}$ projected onto the azimuthal mean (i.e., secondary $\bar{v}_{\text{max}}$; Fig. 4.1 and 4.2). The subsequent development, strengthening, and dominance of the secondary $\bar{v}_{\text{max}}$ that occurred while the primary $\bar{v}_{\text{max}}$ decayed in Bonnie (Fig. 4.1) is consistent with the progression of wind maxima described by S11 and Willoughby et al. (1982).

Associated with these changes in $\bar{v}$ were changes in $\bar{\zeta}$, in which the primary (secondary) $\bar{\zeta}$ maximum decayed (strengthened) during the ERC (Fig. 4.3 and 4.4). The decay of the primary
maximum could have been indicative of mixing processes (Kossin and Eastin 2001) or just a reflection of the weakening primary v$_{\text{max}}$. Changes in the overall radial $\zeta$ structure, particularly in regards to the development of the secondary $\zeta$ maximum, were similarly observed by Abarca and Corbosiero (2011) in hurricanes Katrina and Rita during SEF.

Bonnie’s double eyewall and moat vertical structure downshear (Fig. 4.14 and 4.15) also had similarities to other observed ERCs. The deep inner reflectivity tower fed by strong updrafts (Fig. 4.14 and 4.15) and shallower outer reflectivity tower (Fig. 4.15) in Bonnie have been observed in Hurricane Rita (Houze et al. 2007; Didlake and Houze 2011), Hurricane Gonzalo (Didlake et al. 2016), and in 37 concentric eyewall cases (Hence and Houze 2012). Bonnie’s moat at 2141–2152 UTC 23 August (Fig. 4.14b) displayed descent above 6 km and weak ascent below. Descent maximized above 6 km in the moat was similarly observed in Rita (2005), though Rita also featured descent in lower levels (Houze et al. 2007; Didlake and Houze 2011), unlike Bonnie. In contrast to Rita’s w structure, Bonnie’s moat at 2338–2359 UTC 23 August (Fig. 4.14d) had a stratiform signature. Though the moat is usually characterized by a relatively precipitation-free region, as in Hurricane Rita (Houze et al. 2007), light stratiform precipitation could exist if strong enough downward motion is not present to suppress precipitation. Hence and Houze (2012) and Marks and Houze (1987) similarly discovered a stratiform region outside the primary eyewall, which they attributed to the outward transport and melting of hydrometeors produced by the primary (or single) eyewall. Given Bonnie’s strong updrafts in the primary eyewall, this idea is a plausible explanation for the stratiform moat region.

5.3. Bonnie’s SEF

Though many factors likely contributed to Bonnie’s SEF, results suggest that shear played an important role in this process. Based on Nong and Emanuel’s (2003) findings, strong
shear in Bonnie could have provided a sufficient external forcing needed to initiate a secondary eyewall. After shear increased to 10 m s\(^{-1}\) at 1300 UTC 23 August, Bonnie displayed a highly asymmetric stratiform cloud pattern with localized convection embedded downshear (Fig. 4.9d), similar to Dai et al.’s (2017) results. Secondary \(v\) and \(\zeta\) anomalies appeared first downshear (Fig. 4.13a and b), likely due to shear-induced tilt dynamics (Reasor et al. 2013). Thus, shear created a favorable environment for enhanced upward motion, winds, and convection downshear, as would be expected (Frank and Ritchie 2001; Corbosiero and Molinari 2002).

The process by which the downshear rainband convection became a secondary eyewall could involve a couple of mechanisms. It is feasible that persistent shear forcing allowed for sustained latent heating downshear outside of the primary eyewall that became strong enough to project onto the azimuthal mean and then form a secondary eyewall (Rozoff et al. 2012; Zhu and Zhu 2014). From Figs. 4.9d and e and Fig. 4.10, it is clear that rainband activity increased prior to SEF, which likely contributed to the spin-up of a secondary wind maximum and SEF, consistent with Rozoff et al. (2012) and Zhu and Zhu (2014).

An additional mechanism by which the secondary eyewall could have formed was through the upshear extension of downshear convection by enhanced surface fluxes. Dropsonde profiles (Figs. 4.12c and d) reveal that similar \(\theta_e\) existed at the surface both downshear and upshear, while surface windspeeds were nearly 35% higher upshear during SEF. Higher windspeeds upshear continued even after SEF (Fig. 5.1b), as the eyewall extended around further upshear. This suggests surface fluxes were greater upshear, which could have allowed downshear convection to extend upshear and form a secondary eyewall. This explanation for SEF is favored over the axisymmetrization argument (Kuo et al. 2004), since axisymmetrization alone cannot produce the observed \(\bar{\zeta}\) increase during SEF, as a local \(\zeta\) maximum would be strained out by the
mean vortex flow. However, convectively generated $\zeta$ possibly aided by enhanced surface fluxes upshear could act to increase $\bar{\zeta}$ in the secondary eyewall region.

Numerous other SEF theories exist in the literature, some of which might have played a role in Bonnie’s SEF. The beta-skirt region has been known to be favorable for SEF, as it is conducive to the generation and outward propagation of vortex Rossby waves (VRWs; Corbosiero et al. 2006, Abarca and Corbosiero 2011) and can transfer convectively generated vorticity perturbations into the azimuthal mean flow (Terwey and Montgomery 2008). Bonnie displayed a beta-skirt in its radial $\bar{\zeta}$ structure during flight A1 from 55–100 km (Fig. 4.3), which encompassed the region in which the secondary $\tilde{v}_{\text{max}}$ developed at 85 km (Fig. 4.1). The appearance of the secondary $\tilde{v}_{\text{max}}$ at slightly less than three times the primary $\tilde{v}_{\text{max}}$ might indicate the presence of VRWs (Montgomery and Kallenbach 1997; Corbosiero et al. 2006). However, with limited observational data, particularly radar data, the presence of VRWs cannot be determined. Regardless of the mechanism, the increased $\bar{\zeta}$ associated with SEF that occurred in the beta-skirt region supports Terwey and Montgomery’s (2008) hypothesis.

The role of unbalanced boundary layer dynamics has also been cited as an important process by which the secondary eyewall forms. From this viewpoint, strong boundary layer inflow and inward advection of $\bar{m}_a$ outside the primary eyewall follows a broadening of the tangential wind field (Huang et al. 2012; Abarca and Montgomery 2013). Supergradient winds develop in response to strong inflow, which is associated with convergence in and above the boundary layer where the secondary eyewall forms (Huang et al. 2012; Abarca and Montgomery 2013; 2014). The inward bowing of $\bar{m}_a$ (Fig. 4.5) contours from 60–90 km and strong inflow outside 100 km (Fig. 4.6) during Bonnie’s ERC is consistent with these previous studies and SEF. However, these results in Bonnie are at flight-level, not in the boundary layer, which
deviates from the aforementioned studies. Given limited boundary layer observations, it is
difficult to assess the presence of supergradient winds. Yet, strong inflow that switches to
outflow at 775 hPa (900 hPa) downshear (upshear; Fig.4.12a) is consistent with the presence of
supergradient winds in Rita’s secondary eyewall (Didlake and Houze 2011). Bonnie thus
exhibited processes observed in other SEF cases, though shear appeared to be one of the
fundamental influences.

5.4. Bonnie’s constant intensity

Previous studies of Bonnie suggested that warm sea surface temperatures played a role in
intensification (Heymsfield et al. 2001; Molinari and Vollaro 2008). Presumably, these warm
SSTs (over 28°C) would have contributed to Bonnie’s constant intensity during the high shear
period from 23–25 August. However, the TRMM TMI observed significantly cooler SSTs (25°–
27°C) in Bonnie from 24–26 August (Wentz et al. 2000). Airborne Expendable
BathyThermograph (AXBT) measurements confirmed Wentz et al.’s (2000) findings, with SSTs
of approximately 27°C occurring at 2100 UTC 23 Aug (Table 5.1). These results suggest that
SSTs did not play a major role in Bonnie’s resiliency during high shear.

It is instead speculated that Bonnie’s larger radius of maximum winds during and post-
ERC contributed to Bonnie’s resiliency in high shear. The radius of maximum wind between
1400 UTC 23 August and 1400 UTC 24 August expanded from 35 km to 85 km, which indicates
Bonnie became a larger storm. Jones (1995) and DeMaria (1996) both found that larger storms
are more resistant to the effects of vertical wind shear, which possibly was the case in Bonnie.
Additionally, an increased radius of maximum wind means diabatic heating within the radius of
maximum winds was more likely, which would increase the efficiency of intensification (Nolan
et al. 2007; Pendergrass and Willoughby 2009; Vigh and Schubert 2009). Diabatic heating within Bonnie’s larger radius of maximum winds was seen by the intense primary eyewall situated inside the secondary eyewall at 0000 UTC 24 August (Fig. 4.10f). Additionally, radar and satellite observations post-ERC indicate that intense convection occurred within 50–100 km (not shown), supporting this hypothesis. However, due to the detrimental effects of strong shear and cool SSTs, Bonnie was limited in the extent of intensification and could likely only maintain intensity. It still remains a question, however, as to how Bonnie maintained intensity at the beginning of its ERC, before the radius of maximum winds expanded.
Figure 5.1. As in Fig. 4.12c., from (a) 2000 UTC 23 August–0020 UTC 24 August and b) 1200 UTC 23 August–0000 UTC 25 August. DS (US) composites include a) five (seven) dropsondes and b) eight (14) dropsondes from NOAA P-3 flights.

Table 5.1. Airborne Expendable BathyThermograph (AXBT) sea surface temperature (SST) measurements at various locations in Bonnie between 2047–2124 UTC 23 August.

<table>
<thead>
<tr>
<th>Time (UTC)</th>
<th>SST (°C)</th>
<th>Distance from center (km)</th>
<th>Location</th>
</tr>
</thead>
<tbody>
<tr>
<td>2047</td>
<td>27</td>
<td>10</td>
<td>eye</td>
</tr>
<tr>
<td>2116</td>
<td>27.5</td>
<td>26</td>
<td>eyewall</td>
</tr>
<tr>
<td>2124</td>
<td>26.5</td>
<td>50</td>
<td>eyewall</td>
</tr>
</tbody>
</table>
6. Conclusions

This study examined an ERC during a highly sheared, steady-state period in Hurricane Bonnie (1998) from 1200 UTC 23 August–1500 UTC 25 August using flight level, radar, and dropsonde observations. Bonnie exhibited many of the hallmarks of a classic ERC, including the full progression of the development, strengthening, and dominance of a secondary eyewall over the primary eyewall. However, Bonnie’s structure was highly asymmetric due to strong shear, which was a departure from the classic view of SEF and ERCs being symmetric in nature. It is hypothesized that shear may have acted as external forcing to preferentially generate intense convection downshear radially outside of the primary eyewall. This convection was likely able to extend upshear to form a nearly concentric secondary eyewall due to enhanced surface fluxes.

While it is unclear as to why Bonnie did not experience intensity changes at the beginning of its ERC, it seems feasible that later during the ERC and after it concluded, the ERC aided Bonnie’s resiliency in high shear. Later and post-ERC, the radius of maximum winds expanded, which implied that Bonnie could have more effectively resisted shear due to being a larger storm. A larger radius of maximum winds also created a greater area in which convection could fall, which is favorable for intensification (or maintaining intensity).

Many questions are raised by this study and would be avenues for future research. It is still unclear why SEF occurred in Bonnie during adverse environmental conditions and how shear played a role in this process. If shear did help to initiate the SEF, then why do more sheared storms not form secondary eyewalls or undergo ERCs? Furthermore, what is the role of external versus internal dynamics in initiating ERCs? Is one mechanism more dominant than the other? These sorts of questions are unable to be addressed in the present study, but would be interesting to address from a modeling framework.
REFERENCES


