Observations of the Early Evolution of an Explosive Oceanic Cyclone during ERICA IOP 5. Part II: Airborne Doppler Analysis of the Mesoscale Circulation and Frontal Structure

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ABSTRACT

Using airborne Doppler radar data collected onboard the NOAA P-3 aircraft during ERICA IOP 5, the three-dimensional wind field of the circulation center of an explosive extratropical cyclone is shown. The cyclone was entering its rapid intensification stage at the time of the analysis. The circulation formed along a frontal boundary and was characterized by mean vertical velocity, vorticity, and divergence values comparable to those derived for mature hurricanes. In addition, the circulation at this early stage of cyclone development was relatively shallow (<2.5 km AGL). Intense convection occurred within and surrounding the center of the circulation suggesting that diabatic effects played a role in its development. It is believed that the mesoscale circulation is the preexisting, shallow low that has been shown to form in association with explosive cyclogenesis. The wind synthesis also revealed the structure of the bent-back warm front, which had undergone a scale contraction. The intensity of this warm front, based on the kinematic depiction of the strength of the secondary circulation and the small spatial scale of the frontal discontinuity, may have been greater than any other case documented in the literature and was comparable to examples of intense cold fronts.

1. Introduction

One of the most remarkable achievements in synoptic meteorology was the development of the polar front theory of cyclones advanced by Bjerknes (1919) and Bjerknes and Solberg (1922). In these landmark papers, the formation of a cyclone as a result of an instability on the polar front (a surface of discontinuity separating subtropical and polar air masses) is described. Initially, the extratropical cyclone begins as a wave along the front. As the cyclone evolves, the warm sector diminishes in size and finally becomes occluded as the low-level cold air completely encircles the low center.

Although numerous features of the polar front theory still exist today in meteorological textbooks, several modifications have been proposed based on new theoretical developments and the introduction of sophisticated observing systems. Charney (1947) and Eady (1949) proposed the theory of baroclinic instability in the westerly current to explain cyclogenesis. This theory, coupled with the inability of earlier work to obtain satisfactory solutions of frontal instability (e.g., Bjerknes and Godske 1936), led to the acceptance of baroclinic instability as the primary cause of extratropical cyclogenesis.

Petterssen et al. (1955), however, suggested that cyclogenesis occurs when an upper-level baroclinic wave becomes superimposed upon a low-level frontal system (often referred to as type B development). Palmén and Newton (1969) discuss how the most rapid cycloic development can occur when upper-level divergence is superimposed on a preexisting circulation along a surface front. In the study of the Presidents' Day storm, Bosart (1981) documented the importance of an upper-level short-wave trough that interacted with a preexisting, shallow cyclone [also shown by Gyakum (1983); Bosart and Lin (1984); Rogers and Bosart (1986); Takayabu (1991); and Gyakum et al. (1992)]. It should be noted that even before the work of Chárney and Eady, Bjerknes and Holmboe (1944) recognized that perhaps a synergistic mechanism could exist between an initial frontal instability and an upper-level baroclinic disturbance in cyclogenesis.

More recently, there has been considerable interest in a certain class of extratropical disturbances called "explosive" or rapidly deepening cyclones (Hadlock and Kreitzberg 1988). These cyclones are characterized by a minimum deepening rate of 24 mb (24 h)$^{-1}$ (normalized to 60°N) (Sanders and Gyakum 1980) and have been shown to attain an intensity comparable to that of hurricanes. Kuo et al. (1992) caution that this does not mean that the dynamics of these systems are
similar to tropical cyclones. Synoptic studies of explosive cyclogenesis by Bosart (1981), Gyakum (1983), Reed and Albright (1986), among others, have shown that the rapid intensification of these cyclones can occur over a short time interval and lead to a low central pressure. Neiman and Shapiro (1993) and Neiman et al. (1993) have shown that this rapid intensification can also occur over an extended period (~18 h). Data-based diagnostic studies (e.g., Tracton 1973; Robertson and Smith 1983; Reed et al. 1988) and numerical simulations (e.g., Kuo and Reed 1988; Mailhot and Chouinard 1989; Manobianco 1989; Kuo and Low-Nam 1990) have suggested that latent heat release plays an important role in the deepening of explosive cyclones. Observational studies are still lacking, however, in determining the exact nature and organization of the latent heating associated with this type of cyclogenesis. Neiman et al. (1993) have provided diagnostics of diabatic heating along the warm front for an intense oceanic cyclone.

Further refinements of the polar front model occurred when it was discovered that cyclogenesis could lead to intense frontogenesis, rather than being a consequence of the latter (e.g., Hoskins and West 1979). This finding did not change the observational fact that preexisting fronts are favored areas of cyclogenesis (Reed 1990). More recently, Shapiro and Keyser (1990) proposed several modifications to the Norwegian cyclone model based on in situ observations with an aircraft and high resolution numerical simulations of baroclinic disturbances. In particular, they highlight three modifications:

(i) loss of cold-frontal baroclinity (frontolysis) near the triple point during the early phases of cyclogenesis (referred to as the "frontal fracture");
(ii) the cyclone-relative westward migration of the warm frontal structure back into the northerly flow west of the intensifying cyclone center (referred to as the "bent-back" warm front); and
(iii) the exclusion of intermediate temperature air behind the cold front in the fully developed cyclone. The term "T-bone" is used to refer to the frontal orientation that results when the advancing cold front becomes oriented perpendicular to the "bent-back" warm front.

One of the primary objectives of the present paper is to delineate the kinematic structure of the warm front. Surprisingly, there have been relatively few studies that have examined the detailed wind flow patterns of warm fronts (Browning and Harrold 1969; Heymsfield 1979; Hertzman et al. 1988; and Crochet et al. 1990).

Since the inception of the Norwegian cyclone model, the evolution of the extratropical cyclone and its attendant fronts has been the focus of considerable interest. However, to the authors' knowledge there has never been a detailed kinematic study of the internal structure of an extratropical cyclone. Studies such as Bosart (1981), Gyakum (1983), and Reed and Albright (1986) have come close by combining comprehensive surface analyses with much coarser upper-level data. Shapiro and Keyser (1990), Neiman et al. (1993), and Wakimoto et al. (1992) have presented some finescale depictions of cyclones based on in situ and single-Doppler velocity data recorded by a research aircraft. In addition, there have been a few polar low studies in the literature that have reconstructed reasonably detailed analyses of the cyclone structure (Shapiro et al. 1987; Douglas et al. 1991; Bond and Shapiro 1991).

This is the second in a series of papers that examines the extratropical cyclone of 19 January 1989 observed during intensive observing period (IOP) 5 of ERICA (Experiment on Rapidly Intensifying Cyclones over the Atlantic) (Hadlock and Kreitzberg 1988). In addition to routinely available synoptic data during the cyclone event, observations were taken by deployed ocean buoys and from research aircraft. The aircraft released dropwindsondes and collected high-resolution radar reflectivity and Doppler velocity information with onboard radars. In Part I (Bier and Wakimoto 1995), the large-scale synoptic conditions during the developing stage of the cyclone were presented along with mesoscale analyses combining surface observations, dropwindsondes, in situ measurements, and radar reflectivity data collected by research aircraft. In this paper, the Doppler velocity data recorded during several of the flight legs flown by the aircraft are highlighted. These airborne Doppler radar data are unique for several reasons:

1) The radar was able to collect radial velocity data from a complete synthesis of the circulation center of the IOP 5 cyclone. The only other cyclone studies with comparable spatial resolutions are those attempted on hurricanes (Marks and Houze 1984, 1987; Marks et al. 1992). The reconstructed flow field provides an unprecedented view of the kinematic structure of the circulation center of the low during its developing stage as an upper-level short-wave trough approached the area.

2) A three-dimensional reconstruction of the bent-back warm front was obtained. The circulation center was found to be situated along an intense front resembling those presented in the literature that have undergone a scale contraction (e.g., Neiman and Shapiro 1993; Neiman et al. 1993). There have been relatively few studies of warm fronts in the literature.

3) The organization and nature of the convection near and within the circulation center was resolved by the radar reflectivity measurements. Characterizing the convective field is important given the recent interest on the effects of diabatic processes in rapid deepening of extratropical cyclones. Past observational studies have shown the existence of convection based primarily on satellite imagery (e.g., Bosart 1981; Gyakum 1983; Rogers and Bosart 1986; Gyakum and Barker...
1988). Neiman et al. (1993) did show a greater than 40-DBZ convective echo for a mesocirculation during ERICA IOP 4 cyclone (see their Fig. 5).

Section 2 discusses the data and methodology. The single- and pseudo-dual-Doppler analyses of the circulation center are presented in section 3, and section 4 provides a summary and discussion.

2. Data and methodology

The two NOAA (National Oceanic and Atmospheric Administration) P-3 research aircraft were the primary data platforms during ERICA and were deployed from the Naval Air Station in Brunswick, Maine. A total of four missions were flown during IOP 5:

- flight 1: 0445 UTC 19 January–1355 UTC 19 January
- flight 2: 1110 UTC 19 January–2000 UTC 19 January
- flight 3: 1757 UTC 19 January–0331 UTC 20 January
- flight 4: 0041 UTC 20 January–0958 UTC 20 January

Owing to its endurance and design, the P-3 aircraft is ideally suited for collecting data during long excursions over the ocean. The aircraft typically has a range of 4300 km and can remain aloft for 8–10 h. For more information on the characteristics of the aircraft, the reader is referred to Jorgensen et al. (1983). This paper only uses data collected during flight 1.

The P-3 aircraft are equipped with 2 radars: a C-band (5.59 cm) lower fuselage radar and an X-band (3.22 cm) tail Doppler radar. The lower fuselage radar records surveillance scans of radar reflectivity and has horizontal and vertical beamwidths of 1.1° and 4.1°, respectively. The radar was nominally set to scan at 2 rev min⁻¹, with a pulse repetition frequency (PRF) of 200 s⁻¹. The tail radar was designed to allow 360° scanning at 8 rev min⁻¹ in a plane normal to the aircraft’s ground track. This radar transmits at a PRF of 1600 s⁻¹, resulting in an unambiguous velocity interval of 12.88 m s⁻¹. The beamwidth is 1.35° across the sweep angle and 1.90° along the sweep angle. The aircraft were also equipped to record in situ measurements of standard thermodynamic and kinematic parameters.

The unique aspect of the current case study was the track flown during flight 1 that completely encircled the center of the extratropical cyclone at a height of 300 m AGL (see Fig. 1). The single-Doppler velocity data collected during legs 1, 2, 3, and 4 were used to synthesize the three-dimensional kinematic wind field of the circulation center associated with the low. The analysis area encompassed by Fig. 1 was shown as a small black box in Fig. 13 in Part I.

Radar data collected during legs 1 and 2 were able to reconstruct the wind field associated with the bent-back warm front extending northeast from the circulation center while legs 3 and 4 were able to collect the crucial data necessary to resolve the vortex. These two separate syntheses were merged to produce the final wind field. The radar data were time-space adjusted using a velocity of 18.5 m s⁻¹ from 240°, corresponding to the propagation of the circulation center. Unlike Doppler analyses using ground-based radars, the target of interest is often assumed to be quasi-stationary for a longer period of time while the aircraft flies the legs that can be used to reconstruct the wind field. Marks et al. (1992) refer to this as a "pseudo"-dual-Doppler analysis because it requires the use of data from the same radar at two different times. Accordingly, the synthesis of the first two legs must be viewed with some caution, owing to the time separation of approximately 35 min as indicated in Fig. 1. The intent of the analysis shown in this sector of the cyclone is to highlight the flow structure of the front and not to resolve the individual convective features developing in the region. The synthesis of the circulation center is not seriously affected by the time stationarity assumption since the total elapsed time to traverse legs 3 and 4 was only 6 min 45 s.

3. The flight track around the circulation center

a. Overview

As previously discussed in section 2, Fig. 1 depicts the path taken by flight 1 around the circulation center between 1005:58 and 1057:00 UTC. The cyclone is early in its rapid intensification stage based on the results presented in Part I. Plotted along the track at 1-min intervals are the flight level data of pressure, air and dewpoint temperature, and wind speed and direction. Unfortunately, no thermodynamic data were collected after approximately 1051 UTC. The black lines along the track represent intervals when high resolution single-Doppler velocity data were used to perform the pseudo-dual-Doppler analysis.

Within the smaller box on the figure is the reconstructed wind field based on the Doppler velocity information at 300 m AGL (the details of the wind synthesis are discussed in section 3c and the appendix). The streamlines (drawn in gray) are an estimate of the flow field based on the flight-level data and the Doppler analysis.

The locations where the P-3 penetrated the bent-back warm and secondary cold fronts east and south of the circulation center, respectively, are shown. The flight-level data for both penetrations were shown in Part I. In particular, the penetration of the secondary cold front corresponds to the location of the fine line shown in Figs. 11a and 14 in Part I. The interpolation of the frontal locations inside the circuit flown by the P-3 were based on the convergence zones resolved by the pseudo-dual-Doppler synthesis. Note that the circula-
tion center was situated along the frontal boundary. A mean circulation of about $3.1 \times 10^6$ m² s⁻¹ was calculated for the flight track shown in the figure.

b. Single-Doppler analysis

An example of a lower fuselage radar scan recorded at 1049:19 UTC is shown in Fig. 2 (this image is similar to those shown in Figs. 11a,b in Part I). The cyclone has acquired a T-bone structure (Shapiro and Keyser 1990) as the cold front perpendicularly intersects the warm front to the east of the low center. The radar scan in Fig. 2 suggests, as highlighted by the convective activity, that the warm-frontal structure extends continuously to the west of this intersection point and then toward the south as it approaches the circulation center.

The arrow in Fig. 2 denotes the location of the dry slot discussed in Part I. This dry, southerly flow produces a small discontinuity west of the T-intersection with the bent-back warm front displaced farther toward the north (also suggested in the color image at 1055:30 UTC shown in Part I, see Fig. 11b). The areal distribution of precipitation shown in Fig. 2 is in reasonable
agreement with Special Sensor Microwave/Imager (SSM/I) satellite observations shown by Reed et al. (1993a). The black dot on the figure represents the location of the circulation center based on the pseudodual-Doppler analysis presented in section 3c. Note that the convective activity only extends 10 km to the south of this point.

Also labeled on Fig. 2 are the aircraft positions and orientations of two vertical cross sections of radar reflectivity and Doppler velocity at 1038:14 and 1043:00 UTC shown in Figs. 11e and 11f in Part I, respectively. The cross section in Fig. 11e illustrates the vertical structure of the bent-back warm front as the aircraft was flying toward the northeast, approximately parallel to and north of the front. The top panel in the figure depicts the convective echoes near the frontal region in excess of 50 dBZ. A radar bright band produced by melting precipitation (Battan 1973) is apparent at distances less than 20 km from the aircraft at 1.5 km AGL. A limitation of single-Doppler velocities is that they only reveal the radial component of the flow toward and away from the radar; therefore, inferences concerning the flow field must be viewed with caution. However, combined with the synthesized three-dimensional wind field, such data still reveal many important features.

Two distinct flow regimes are apparent in the bottom panel of Fig. 11e (Part I). At levels below 1–2 km, the flow is from the north (positive Doppler velocities) with the indication of a possible convergence zone at the leading edge of the front at approximately 32 km from the aircraft position. Unfortunately, this convergence zone is not distinct, owing to the attenuation of the X-band tail radar by the intervening precipitation (Hildebrand et al. 1981). A second area of northerly flow can be seen at \( \approx 4 - 5 \) km AGL and more than 30 km from the radar. This is probably a result of the thunderstorm outflow at upper levels from the strong cells seen in the radar reflectivity image. Above 2 km the flow reverses with a southeasterly component (negative Doppler velocities) measured in excess of 14 m s\(^{-1}\) at about 3 km AGL. Similar to the results presented by Wakimoto et al. (1992), Neiman and Shapiro (1993), and Neiman et al. (1993) the front is very intense with horizontal discontinuities on very small scales. Other researchers have noted analogous discontinuities associated with cold-frontal collapse and its apparent similarity with density currents such as those at the leading edge of outflow boundaries from thunderstorms (e.g., Carbone 1982; Hobbs and Persson 1982; Shapiro 1984; Shapiro et al. 1985). Although it is not clear that the flow depicted in Fig. 11e is thermodynamically similar to a density current, the elevated head and stronger velocities away from the aircraft near the sea surface between 24 and 30 km are strikingly similar to these types of laboratory flows (e.g., Simpson 1969). Examples of warm-frontal collapse have been shown in numerical simulations (Kuo and Reed 1988; Hedley and Yau 1991; Kuo et al. 1991; Reed et al. 1993b).

The vertical structure (labeled EF in Fig. 2) approximately parallel to and slightly north of the bent-back warm front as the aircraft was flying toward the southeast is shown in Fig. 11f (Part I). A pronounced bright band is evident in the top panel at about 2 km AGL. The bottom panel in Fig. 11f reveals the location of the frontal boundary based on the kinematics of the flow. A northeasterly component of flow (\( \approx 15 \) m s\(^{-1}\)) is observed below the bent-back warm front (estimated to be near the zero isopleth of Doppler velocities at about 1.4 km AGL) and rapidly changes direction to a southeasterly component above the boundary.

Horizontal plots of interpolated radar reflectivity and single-Doppler velocities when the P-3 was flying along leg 1 are shown in Fig. 3. The black dot on the figure represents the location of the circulation center. Prominent at 0.3 km AGL is the small rotational couplet associated with the cyclonic circulation with a diameter of about 20 km. The axis connecting the centers of the positive and negative maxima in the couplet is not parallel to the aircraft track but is rotated clockwise by about 25°–30°. Doviak and Zrnić (1984) have shown that a clockwise rotation of the axis occurs when a circulation has a significant inflowing component. The zero isopleth of Doppler velocity approximately delineates the frontal convergence boundary, which is well correlated with the area of intense convection. The couplet has weakened and appears to show a more prominent rotational signature at 1.3 km. The couplet becomes even less distinct at 2.3 km, consistent with the pseudo-dual-Doppler analyses to be shown in sec-

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**Fig. 2.** Radar reflectivity surveillance scan at low elevation angle from the lower fuselage radar on board the NOAA P-3 at 1049:19 UTC 19 January 1989. Radar reflectivities are represented by a gray scale with high values in white. Convection along the warm, cold, and bent-back warm front are evident. Range rings are in kilometers from the aircraft position. The aircraft position, time, and orientation of two vertical cross sections labeled CD and EF are shown (these vertical cross sections are shown in Figs. 11e and 11f in Part I). The black dot represents the location of the circulation center and the black arrow denotes the approximate location of the dry slot.
tion 3c. Also apparent at this height is a dramatic reversal in the single-Doppler velocity field compared with the 0.3-km analysis with the zero isopleth of Doppler velocity now identifying an axis of strong divergence. Maximum radar reflectivities at all three levels exceed 45 dBZ. Actual divergence values based on pseudo-dual-Doppler analyses are shown in the next section.

The results presented in Figs. 2 and 3 illustrate that the developing cyclone during IOP 5 was imbedded in an area that was convectively active. The role of convection in cyclogenesis has been a subject of debate in the literature. Studies by Tracton (1973), Bosart (1981), Gyakum (1983), Smith et al. (1984), Rogers and Bosart (1986), and Gyakum and Barker (1988) suggest that the influence of latent heat release associated with cumulus convection near the cyclone center plays an active role in rapid cyclogenesis. Indeed, sensitivity tests by Reed et al. (1993b) have shown that latent heat release had a large impact on the storm intensity during IOP 5. Without moist processes, only a minor storm developed.

An example of the nature of the convection located slightly northeast of the circulation center is shown by the vertical cross section (segment $AB$ in Fig. 3) in Fig. 11d (Part I). The top panel shows the overshooting tops of the thunderstorm at a range of approximately 26 km from the radar. The strong horizontal convergence at low levels ($<2.5$ km AGL) into the circulation at a range of about 25 km from the radar is apparent in the bottom panel of single-Doppler velocities.

c. Pseudo-dual-Doppler analysis

A unique aspect of this study is the pseudo-dual-Doppler analysis of the circulation center of the ERICA IOP 5 cyclone. Previous case studies (e.g., Carbone 1982, 1983) have examined small-scale circulations during a tornadic event that were nearly one order of magnitude smaller than the one shown here. The editing and synthesis procedures using the tail Doppler radar data are explained in the appendix.

Horizontal cross sections of pseudo-dual-Doppler winds (in a ground-relative frame of reference) of the circulation center and attendant front combined with radar reflectivity, vertical vorticity $\zeta$, divergence, and vertical wind speed are presented in Fig. 4. The mesoscale circulation is prominent at low levels in Fig. 4a and is centered within a band of convection as shown in Fig. 3. In a modeling study of a rapidly intensifying marine cyclone, Kuo et al. (1991) showed that the cyclone exhibits a tighter inner core when diabatic processes, owing to latent heat release, were included in the model run compared to the adiabatic baroclinic simulations.

There appear to be two primary flow regimes in the wind field near the circulation. Southeast and northwest of the center the flow is predominantly south-southwesterly and north-northeasterly, respectively. The streamlines at this level (shown in Fig. 5) suggest that both flows curve cyclonically and subsequently converge into the center of the circulation, as is evident in the $\zeta$ and divergence fields. In this regard, it was decided for this paper to define the center of the circulation as the maximum in convergence (which exceeds $-8 \times 10^{-3}$ s$^{-1}$) in Fig. 4a. Note that this center is near but not collocated with the $\zeta$ maximum. Although the cyclonic vorticity and convergence values along the frontal boundary northeast of the circulation center in Fig. 4a are significantly larger than those documented by Browning and Harrold (1969), Heymsfield (1979), and Hertz-
FIG 4. Horizontal cross sections of pseudo-dual-Doppler winds in a ground-relative frame of reference combined with four fields at (a) 0.3, (b) 1.3, (c) 2.3, and (d) 3.3 km AGL. The winds are superimposed on radar reflectivity (top left), vertical vorticity (top right), divergence (bottom left), and vertical motion (bottom right). Reflectivity, vorticity, divergence, and vertical motion values are contoured with gray lines with values greater than 40 dBZ, $6 \times 10^{-3}$ s$^{-1}$, $-6 \times 10^{-3}$ s$^{-1}$, and 6 m s$^{-1}$ shaded gray, respectively. Locations of two vertical cross sections through the bent-back warm front and the circulation center are labeled BB' and AA', respectively.

man et al. (1988) along warm fronts, they are less then the values presented by Carbone (1983) for a strong cold front that spawned a tornado. The wind field also depicts an area of surface divergence located west of the circulation center. On the basis of the circulation calculated in section 3a, a mean vor-
ticity within the area encircled by the P-3 was determined to be about $7.02 \times 10^{-4} \text{ s}^{-1}$. Accordingly, the mean vorticity over the P-3 circuit was approximately one order of magnitude less than the peak values shown in Fig. 4.

At 1.3 km AGL (Figs. 4b and 5), the flow field has become asymmetric with a pronounced southwesterly flow and a weaker northeasterly flow located southeast and northwest of the circulation center, respectively. The magnitude and areal extent of $\zeta$ have increased, and a maximum in convergence at the location of the circulation center is still evident. The appearance of the maximum in vertical vorticity at this level rather than at the surface can be attributed to a portion of the radar...
beam striking the surface. This effect, which is pronounced at low levels, would reduce the magnitude of the measured Doppler velocities and, hence, the convergence values. The strong updrafts along the front and within the circulation center (>8 m s⁻¹) are now apparent.

At 2.3 km AGL (Figs. 4c and 5), no distinct rotation center is seen in the predominantly diffuent wind field. This observation and the results shown in Fig. 3 suggest that the circulation center is a relatively shallow phenomenon. The vertical velocities along the front are, again, much larger than any observed values in the lit-
temperature except for those reported by Carbone (1982) and Shapiro et al. (1985). Recall that both of the latter cases were for intense cold fronts that exhibited a scale contraction. In particular, published values on warm fronts by Heymsfield (1979) and Hertzman et al. (1988) do not exceed 0.6 m s\(^{-1}\). A numerical simulation of an intense warm front that undergoes a scale contraction by Kuo et al. (1991), however, was associated with updrafts greater than 1.5 m s\(^{-1}\), even at a grid resolution of 45 km. In addition, these Doppler-derived vertical velocities shown in Fig. 4 are also comparable to those described by Marks et al. (1992) in the eyewall of Hurricane Norbert. The diffluent wind pattern is still evident at 3.3 km (Figs. 4d and 5).
reflectivities associated with the convection are less pronounced since this cross section is closer to topping the storms. The updrafts are also weaker at this level. The shallowness of the convection in Fig. 4, particularly given the high reflectivities and vertical velocities, is noteworthy.

A vertical cross section perpendicular to the bent-back warm front along $BB'$ is shown in Fig. 6a. The warm front is characterized by a cyclonic shift of the horizontal winds across the front and the sloping axis of $\zeta$ with values greater than $4 \times 10^{-3}$ s$^{-1}$. The strong convergence and sloping updraft in excess of 4 m s$^{-1}$ are also apparent. This compares well with the vertical incidence scans of single-Doppler velocities (not shown), which recorded 5 m s$^{-1}$ updrafts (after removing the effect of terminal fall speeds) as the aircraft penetrated the front. The slope of the front is approximately 1:4 and appears to be typical for contracted warm fronts over the ocean (Neiman et al. 1993; Wakimoto et al. 1992). The intensities of the updraft and the sharpness of the frontal boundary shown in Fig. 6a are related to the diabatic effects of convection. Strong latent heat release by condensation within frontal upglide (Williams et al. 1981; Thorpe and Emanuel 1985; Kuo et al. 1991) and evaporation of falling precipitation (Huang and Emanuel 1991) provides for a positive interaction with adiabatic frontogenesis during the frontal-scale collapse via two processes: 1) the direct heating (on the warm side) and cooling (on the cold side) owing to the changes of state, and 2) the low-level, cross-frontal convergence associated with the diabatically induced upward (on the warm side) and downward vertical motion (on the cold side) as part of the ageostrophic circulation, reinforcing the low-level
frontogenesis (Kuo et al. 1991). Thorpe and Emanuel (1985) illustrate that latent heat release can collapse the frontal updraft into a narrow plume of strong ascent.

Consistent with the vertical cross section by the tail Doppler radar shown in Fig. 11d (Part I), strong convection is collocated with the circulation center in Fig. 6b. Convergence extends up to 2–2.5 km AGL with divergent flow aloft. The maximum in vertical motion is centered at 2 km AGL. The slight displacement between the ζ maximum at the surface and the center of the circulation as determined by the convergence is apparent. An examination of the shift in the direction of the total horizontal wind vector with height in the figure is consistent with the results from Fig. 4, suggesting that the circulation does not extend above 2 km AGL. Other cross sections (not shown) are consistent with this scenario.

Vertical profiles of ζ, divergence, and vertical motion horizontally averaged over the analysis region (∼20 km × 30 km) resolved by legs 3 and 4 are plotted in Fig. 7. These smoothed values are comparable to the low-level estimates derived by Frank (1984) based on a composite analysis of a mature hurricane. The shallowness of the convergence compared to the strong divergence aloft is exemplary of a growing and deepening cyclone (Petterssen et al. 1955). Note that the updraft attains a maximum above the level of nondivergence owing to the density weighting in the anelastic continuity equation. The maximum in vorticity, located at the lowest levels, is consistent with the shallow circulation center but is cyclonic throughout the depth of the analysis region.

In addition to the synthesis shown in Fig. 4, another wind field was generated using legs 1 and 4 (Fig. 8). This was not considered to be the primary analysis since a time stationarity assumption would be required for about 51 min (i.e., the beginning of leg 1 until the end of leg 4). However, there are two important observations that can be made:

1) There is excellent agreement between the results shown in Figs. 4 and 8 suggesting that the time stationarity assumption used in the primary pseudo-dual-Doppler synthesis was a good one. In addition, since
the results are replicated with a different synthesis, it reinforces the accuracy of the analysis shown in Fig. 4.

2) The wind field shown in Fig. 8 has the advantage of revealing the flow pattern in the northwest to north-northwest quadrant of the cyclone that was not available in Fig. 4. Of particular interest is the suggestion of a trough in the wind field at 4.3 and 5.3 km AGL. Based on the analysis shown in Part I (Fig. 6), it is known that the upper-level, short-wave trough at 700 mb was achieving a proper phasing (westward tilt) with respect to the surface cyclone. There is a suggestion in Fig. 8 that the trough axis is slightly ahead (toward the east) of the circulation center; however, its precise location is not known given the limited amount of data in the northwest quadrant of the synthesis.

4. Summary and discussion

This paper has presented airborne radar observations of the center of an extratropical cyclone and its attendant fronts during ERICA IOP 5. As shown in Part I,

the cyclone was entering the rapid deepening stage during the analysis time. By synthesizing Doppler velocity data, the detailed kinematic flow pattern associated with the circulation center was resolved for the first time. This circulation may be the preexisting, shallow low that has been documented in the past in association with explosive cyclogenesis (e.g., Gyakum 1983; Bosart and Lin 1984; Rogers and Bosart 1986; Takayabu 1991; and Gyakum et al. 1992). The maximum values

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1 Recall that the first synthesis shown in Fig. 4 relied on legs 1 and 2 in this area. The latter leg did not provide the necessary data in the northwest quadrant.
of vertical velocity ($\approx 0.9$ m s$^{-1}$), vorticity ($\approx 15 \times 10^{-4}$ s$^{-1}$), and divergence ($\approx 11 \times 10^{-4}$ s$^{-1}$) averaged over the cyclone region were comparable to those based on composite analyses of hurricanes. In addition, the local maxima were approximately one order of magnitude greater than these averages.

The mesoscale circulation developed on the frontal boundary and appeared to be achieving a proper phasing (westward tilt) with an upper-level, short-wave trough. Recently, there has been a renewed interest in the area of frontal waves and cyclones of horizontal length scales of 1000 km or less (e.g., Moore and Pelletier 1987; Joly and Thorpe 1990; Schär and Davies 1990; and Thorncroft and Hoskins 1990). In particular, Joly and Thorpe (1990) present a theoretical study that may be relevant to the current analysis. In the presence of active moist processes (clearly shown in the present study), they document the development of high potential vorticity anomalies leading to the formation of a frontal wave of only 800 km and a growth rate of about 1 day$^{-1}$. The suggestion that the size of the disturbance is reduced and growth rates increase is consistent with the IOP 5 storm; however, the scale of their derived frontal wave may still be too large to be applicable.

The exact location of the circulation center associated with an extratropical cyclone has never been clearly defined in the literature. A recent example by Kuo et al. (1992) suggests that the low center is west of the intersection of the cold and warm fronts in the early stage of development (see their Fig. 8). The present study places the circulation center on the front.

The present paper documents the three-dimensional flow pattern associated with the cyclone’s warm front with maximum updraft velocities approximately three times greater than the numerical simulations by Kuo et al. (1991). This is considered an excellent comparison since the grid resolution in their model run was 45 km. Moreover, the single-Doppler cross section of radar reflectivity and single-Doppler velocity in Fig. 11e (Part I) suggests a similarity between the kinematic structures of the bent-back warm front and the density current. Features such as a well-defined head with accompanying maximum horizontal velocities were evident in the figure.
Several of the frontal features proposed by Shapiro and Keyser (1990), which challenge the classical Norwegian model of extratropical cyclone development, appear to be confirmed in Parts I and II. A T-bone structure was apparent in the orientation of the warm and cold fronts. In addition, the front to the west of this point appeared to represent a continuous extension of the warm front (bent-back warm front). It is also possible that this front may have developed as a frontogenetical feature associated with the strong deformation and convergence fields of the cyclone as documented by Kuo et al. (1992) rather than as a westward migration of the warm front. However, the present analysis appears to support the Shapiro and Keyser model owing to the rather long extension of the frontal boundary south of the low discussed in Part I (analyzed as a secondary cold front as shown in Figs. 13a in Part I). The existence of this frontal feature is not addressed in the numerical simulations of Kuo et al. (1992). An important question for future research is to determine the percentage of the strong kinematic signature (i.e., vorticity, divergence, and vertical motions) that can be directly attributed to the circulation of the marine cyclone and its attendant fronts and the fraction that can be attributed to the deep convection in the vicinity of the circulation center.

The ERICA database has provided a unique opportunity to study the mesoscale structure of marine cyclones with finescale observations collected by in situ measurements and Doppler radar. However, more research is necessary in order to determine the relationship between these small-scale features, such as the intense mesoscale circulation documented in this study, and the large-scale, three-dimensional structure of the extratropical cyclone. This study has concentrated on the kinematics of the mesoscale circulation; future work will include an attempt to retrieve the perturbation pressure field using the techniques first proposed by Gal-Chen (1978).

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APPENDIX

Doppler Radar Analysis

a. Editing

Each radar scan was carefully edited using the NCAR (National Center for Atmospheric Research) Research Data Support System (RDSS) computer software (Oye and Carbone 1981) to remove sea clutter and sidelobe echoes and to unfold the Doppler velocity field. Range corrections were made by comparing the altimeter readings with the range to the ocean surface recorded by the tail radar. Antenna-pointing errors result from situations when the tail antenna is scanning in a mode that is not precisely perpendicular to the ground track. This error can usually be identified by noting fixed ground velocities that are nonzero or pronounced asymmetries in a scan of Doppler velocities in the directions immediately above and below and also to the left and right of the antenna. This error was ± 1 \( \text{m s}^{-1} \). The velocity data were also corrected for the error induced by the aircraft vertical velocity. The magnitude of this correction was ± 1.0 \( \text{m s}^{-1} \).

b. Synthesis technique

The scan rate of the tail radar was approximately 48 deg s\(^{-1}\), or 7.5 s per scan, and the average aircraft speed within the analysis time period was about 115 m s\(^{-1}\). Accordingly, the horizontal separation between consecutive scans was about 920 m. In the vertical direction, the beamwidth was a maximum of 1160 m wide in the analysis domain (1.9° beamwidth at 35 km), however, the effective sampling was 550 m (or half the beamwidth) based on the PRF of 1600 s\(^{-1}\) and the number of samples (32) per pixel. On the basis of the above calculations, it was decided to interpolate the reflectivity and Doppler velocity data onto a Cartesian grid with horizontal and vertical grid spacing of 1.0 and 0.5 km, respectively.

A point in space must be sampled by the radar along at least two different flight legs that should, optimally, be oriented 90° relative to each other so as to minimize the errors produced in the synthesis. This type of geometry is similar to dual-Doppler techniques using ground-based radars. As shown in Fig. 1, the flight legs labeled 1, 2, 3, and 4 were nearly ideal for performing a wind synthesis. For more information on airborne radar techniques, the reader is referred to Ray and Jorgensen (1988).

The radar data was time—space adjusted to a reference time of 1045 UTC using a velocity of 18.5 m s\(^{-1}\) from 240°, corresponding to the propagation speed to the circulation center. A Cressman (Cressman 1959) filter was used in the interpolation process with a radius of influence of 1.1 km. The data were then synthesized within CEDRIC (custom editing and display of reduced information in Cartesian space; Mohr et al. 1986). The hydrometeor terminal fall speeds based on reflectivity used in the synthesis procedure was estimated from the relationship established by Joss and Walldvogel (1970) with a correction for the effects of air density (Foote and du Toit 1969). Since the elevation angles used in the present synthesis are restricted to elevation angles less than 10°, the radial velocities are primarily sampling the horizontal wind component with a very small contribution from the vertical motion. Accordingly, the wind vectors shown in Figs. 4 and 8 are considered to be highly accurate.
Based on the criteria established by Carbone et al. (1985), motions having wavelengths of approximately 5.5 km or greater are resolvable in this case. Accordingly, a two-step Leise filter (Leise 1982) was implemented, which considerably damps wavelengths up to 5.5 km and eliminates those less than 4.0 km. Vertical velocities were obtained from the anelastic continuity equation by integrating the horizontal convergence fields using variational adjustment (O’Brien 1970) and assuming $w = 0 \text{ m s}^{-1}$ as the upper and lower boundary conditions. A fractional boundary condition (e.g., Kingsmill and Wakimoto 1991) is used at the lowest grid point at 0.3 km since it is not coincident with the ocean surface. The upper boundary was chosen to be 7 km where the tropopause was situated based on the mean observed location of the thunderstorm anvils. Errors in the horizontal wind, given the uncertainty in the radial velocity, were estimated to be $2 - 3 \text{ m s}^{-1}$. This is consistent with the range discussed by Marks et al. (1992). For the vertical velocity, error estimates were approximately $3 \text{ m s}^{-1}$.

REFERENCES


