A Composite Study of Explosive Cyclogenesis in Different Sectors of the North Atlantic. Part I: Cyclone Structure and Evolution

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ABSTRACT

General characteristics of the dynamical and thermal structure and evolution of strong explosive cyclones in the northwestern Atlantic near North America (18 cases) and the extreme northeastern Atlantic near Iceland (19 cases) are compared and contrasted through a composite study. Twice-daily gridded analyses from the European Centre for Medium-Range Weather Forecasts at 2.5° resolution from January 1985 to March 1996 are used. In the process of case selection, it is found that the frequency of rapid cyclogenesis in the Greenland–Iceland region is higher than previously thought, and some of the events can be extremely violent.

Many dynamically consistent differences are found when composite cyclones in the two sectors of the North Atlantic are compared. The upper-level forcing that triggers the development in the northeast Atlantic (NEA) is no less intense at the onset of rapid deepening. The NEA cyclones are also associated with lower static stability and locally concentrated but shallower thermal gradient, with less overall environmental baroclinicity. These factors lead to rapid depletion of available potential energy and result in a faster evolution and a shorter life cycle. Therefore, low-level thermal gradient and upper-level forcing components all weaken immediately after rapid deepening. The low-level incipient low in the NEA composite is also stronger, with a distinct potential vorticity (PV) anomaly visible at least 24 h prior to most rapid deepening, and the development produces a more pronounced warm core seclusion. Explosive cyclones in the northwest Atlantic, on the other hand, tend to have a higher stability and a greater amount of environmental baroclinicity, with temperature gradients in a broader area and deeper layers. These factors correspond to slower evolution and a longer life cycle.

For cases in the NEA near Iceland, it appears that both upper-level forcing and initial system strengths affect the maximum deepening rate. The close proximity of this region to the high PV reservoir in the lower stratosphere is helpful in the generation of very strong forcing and a violent development under favorable synoptic conditions, when a “parent cyclone” with appreciable strength exists to the north/northeast of the incipient system.

1. Introduction

Rapid intensification of extratropical cyclones, or explosive cyclogenesis, has been an area of active meteorological research during the past two decades due to onshore and offshore hazards and the failure of early operational models to predict it (Silberberg and Bosart 1982; Sanders 1986b, 1987; Sanders and Auciello 1989). Sanders and Gyakum (1980) first studied explosive cyclones (or “bombs”) and defined them as low systems whose central sea level pressure falls at an averaged rate of at least 1 Bergeron (or 1 Ber, in hPa day⁻¹), and

\[ 1 \text{ Ber} = 24 \sin(\phi)/\sin(60^\circ), \]

where \( \phi \) is the latitude of the cyclone center. Since the criterion is latitudinally adjusted to reflect the increase in geostrophic wind speed, the required 24-h pressure drop for 1 Ber increases from 13.9 hPa at 30° to 27.7 hPa at the poles. The deepening rate in some extreme cases can well exceed 1 Ber, and two famous examples are the Queen Elizabeth II storm in 1978 (Gyakum 1983a; Kocin and Uccellini 1985) and the Presidents’ Day cyclone in 1979 (Bosart 1981).

Sanders and Gyakum (1980) produced the first explosive cyclogenesis climatology, updated by Roebber (1984), showing it to be primarily a cold season maritime phenomenon with highest frequencies over the northwestern Pacific and northeastern Atlantic along major baroclinic zones. Although they suspect that both the frequency and intensity are likely underestimated over regions of sparse data coverage, studies by them and several others (e.g., Gyakum et al. 1989; Chen et al. 1992) provide us with the best explosive cyclogenesis climatologies to date.

Studies during the past two decades indicate that ex-
Explosive cyclones are like ordinary ones, fundamentally driven by baroclinic instability (Sanders 1986a; Manobianco 1989b; Wash et al. 1992), with other factors also contributing significantly to the deepening rate. These include 1) strong upper-level forcing (Rogers and Bosart 1986; Macdonald and Reiter 1988; Hirschberg and Fritsch 1991; Lupo et al. 1992), 2) intrusion of stratospheric high potential vorticity (PV) air (Bosart and Lin 1984; Uccellini et al. 1985; Zehr and Keyser 1991; Reader and Moore 1995), 3) latent heat release (Gall 1976; Anthes et al. 1983; Gyakum 1983b; Emanuel et al. 1987; Mullen and Baumhefner 1988; Kuo et al. 1991b; Whitaker and Davis 1994), 4) surface energy fluxes from the ocean (Atlas 1987; Davis and Emanuel 1988; Fantini 1990; Nuss and Kamikawa 1990; Kuo et al. 1991a; Hedley and Yau 1991), and 5) enhanced local baroclinicity from differential diabatic heating (Bosart 1981; Rogers and Bosart 1991). Rapid development is the result of nonlinear interaction among all these processes, although all factors need not be present (Uccellini 1990; Gyakum et al. 1995) and their importance could vary considerably in individual cases (e.g., Reed et al. 1993). Nonetheless, since strong upper-level forcing exists both over land and ocean, it is quite clear that the effect of abundant moisture, supplied through marine boundary layer processes, in lowering the stability is what makes explosive cyclogenesis an almost exclusive maritime phenomenon. In more recent years, high-resolution datasets obtained through field experiments (e.g., Dirks et al. 1988; Hadlock and Kreitzberg 1988) allow analysis of detailed mesoscale structure and evolution (e.g., Neiman and Shapiro 1993; Neiman et al. 1993; Blier and Wakimoto 1995). More complete reviews can be found in Uccellini (1990), Hoskins (1990), and Shapiro and Keyser (1990).

Prior studies have focused on northwestern Atlantic events due to their socioeconomic impacts along the North American coast and better data coverage. The Greenland–Iceland region has received much less attention and is traditionally thought to have little cyclogenetic activity with the exception of polar lows (e.g., Sardie and Warner 1985; Businger and Reed 1989; Nordeng and Rasmussen 1992). More recently, Sarreze et al. (1993) and Sarreze (1995) have shown that cyclogenesis is common near Iceland during cold seasons, and Rogers (1997) found that when the storm track is active in extreme northeastern Atlantic, over 20% of the days experience 24-h pressure changes in excess of 20 hPa. Examples of rapid cyclogenesis in this region include the record-setting case on 14–15 December 1986 (Gadd et al. 1990), the Braer storm (McCullum and Grahame 1993), the one on 30 September 1995 (L., D. Keigwin 1996, personal communication), and those studied by Sinclair and Elsberry (1986), Reed et al. (1988), Gronås (1995), Kristjánsson and Thorsteinsson (1995), and Wernli (1997). These cases (and the present study) demonstrate that rapid cyclogenesis also occurs near Iceland and can be very violent, but a comprehensive study is yet to be done. The purpose of this paper is, therefore, to compare and contrast explosive cyclones in different sectors of the North Atlantic through a composite study using a large number of cases, with an emphasis on northeastern Atlantic.

The present paper (Part I) employs a more qualitative approach and focuses on cyclone structure and evolution. Part II (in preparation) presents results of a vorticity budget analysis from a quantitative perspective. Section 2 describes the data and methodology used to identify, select, categorize, and composite our cases. Composite results of key variables at various levels and on vertical cross sections are presented in sections 3 and 4, respectively. Section 5 provides further discussion, and section 6 summarizes the major findings of this paper.

2. Data and methodology

a. ECMWF dataset

In this study, we use the European Centre for Medium-Range Weather Forecasts (ECMWF) operational global analysis (Level III-b), archived at the National Center for Atmospheric Research from January 1985 to March 1996. Data are twice daily (0000 and 1200 UTC) on a regular 2.5° × 2.5° latitude–longitude grid, and variables include geopotential, temperature, and three-dimensional u, v, and ω (omega) wind components at 14 pressure levels (1000, 850, 700, 500, 400, 300, 250, 200, 150, 100, 70, 50, 30, and 10 hPa), as well as relative humidity from the 1000–300-hPa level. The 925-hPa level has been added since January 1992. Other surface variables include mean sea level pressure (MSLP), surface pressure and temperature, temperature and dew-point at 2-m height, and u and v winds at 10-m height (at 2 m since February 1994).

The ECMWF operational model is a primitive-equation-based global spectral model with T213 truncation (roughly 0.55° resolution) and 31 levels (L31) from surface to 10 hPa (Simmons et al. 1995). It was T106/L19 before September 1991 and T63/L16 before May 1985, with various improvements over the period (Simmons et al. 1989, 1995; Bengtsson 1991). The model employs a four-dimensional data assimilation (FDDA) scheme to analyze data with multivariate three-dimensional optimum interpolation (3D OI) and an injection period of 6 h (Lorenz 1981), and is initialized four times a day (0000, 0600, 1200, and 1800 UTC) using nonlinear normal mode initialization (Wergen 1989).

The ECMWF FDDA system uses the 6-h forecast from the previous model run as the first guess (FG) field, then incorporates all available observations to produce the analysis before model initialization. The 3D OI scheme processes many data points and variables simultaneously by using large analysis volumes, and applies statistical methods to perform data selection and quality control at the same time. All data are subject to
checks for internal consistency, and against climatology, the FG field, and nearby reports (Lorenc 1981; Lönngberg and Shaw 1984; Shaw et al. 1987). The system has been shown to be useful in monitoring the observational network (Hollingsworth et al. 1986). The analyses are generally of high quality in the Northern Hemisphere extratropics (Trenberth and Olson 1988, Lambert 1988), and have been used for diagnostic studies of explosive cyclones by Sinclair and Elsberry (1986), Wash et al. (1988, 1992), and Manobianco (1989a).

b. Potential case identification

Potential explosive cyclogenesis cases in the dataset were identified through an objective approach. Automatic detection and tracking algorithms have been developed individually by Serreze (1995) for MSLP fields of gridded analyses from National Centers for Environmental Prediction (formerly the National Meteorological Center), and by Sinclair (1994) for vorticity centers based on ECMWF analyses. Here two programs, one for storm detection and the other for tracking, are developed to obtain a life history of low pressure systems in the Northern Hemisphere. Since all 162 cases finally selected are manually checked and potential errors of the programs are inconsequential to our composite results, the two programs are only briefly described below.

The detection program takes the ECMWF MSLP array as input and identifies synoptic-scale low pressure systems between 15\(^\circ\)N and the North Pole. A low is identified when the MSLP at a grid point is the lowest within a 750-km radius, is at least 1 hPa lower than the average at surrounding eight points, and has a positive Laplacian. With the low’s extent defined by the outermost closed isobar at 0.5-hPa intervals, the program discards lows smaller than 40 000 km² and those smaller than 120 000 km² if the MSLP deficit between the low’s center and boundary is less than 2 hPa after scanning the array. Such a setting excludes small and weak lows and helps keep the tracking easier and more accurate, although on average there are still as many as 16.9 lows per system. The details of the procedure are rather complicated, but in short the program attempts to find a match for the larger, deeper, and stronger lows earlier, and gives higher priority to isolated pairs. For less obvious pairs, the program employs a “score approach” and matches the pair that yields the highest score, if it is higher than a threshold value. In general, a potential pair gives a higher score if the two lows are closer to each other and have similar system characteristics (such as central MSLP, pressure deficit, and size, etc.). After this, the program determines whether remaining lows at \(t_1\) dissipate or merge with other lows, and whether those at \(t_2\) are new systems or split from existing ones.

Once the life history of each low is obtained, Sanders and Gyakum’s criterion of an averaged deepening rate of 1 Ber for 24 h is used to identify possible explosive cyclogenesis cases north of 30\(^\circ\)N. A total of 1369 events, 800 in the North Pacific and 569 in the North Atlantic, are identified in the dataset. The smoothed geographical distribution of cyclone positions at the midpoint of 24-h periods (Fig. 1) shows three distinct frequency maxima in the North Atlantic, including a main center off the coast near Nova Scotia and two other centers south of Greenland (near 50\(^\circ\)N, 45\(^\circ\)W) and slightly south of the Denmark Strait. When compared with Sanders and Gyakum (1980, their Fig. 3) and Roebber’s (1984, their Fig. 7c), our figure shows a higher degree of spatial variability in the North Atlantic, as well as a higher explosive cyclogenesis frequency in the Greenland–Iceland region.

We evaluated the performance of detection and tracking programs using manual tracking over two 3-month periods from December to February of 1988/89 and 1991/92. A total of 100 errors are found among matches for 6120 identified lows, and thus 98.4% of them were correctly matched. However, 24 of the errors involve bombs in the North Atlantic, including 18 occasions of misses, 3 of false generation, and 3 of underestimation. Manual detection identified 79 cases while the programs identified 64; so, overall the frequency in the North Atlantic is underestimated by nearly 20%. This deficiency arises mainly because explosive cyclones in the North Atlantic often travel fast and exhibit rapid changes in system characteristics, producing a lower score, sometimes below the threshold. In the North Pacific, the programs perform significantly better during the eval-
vation periods, as 82 cases are identified with only six errors. Thus, the North Atlantic distribution in Fig. 1 somewhat underestimates the climatology that would be obtained using manual tracking on ECMWF data, although the programs are no doubt useful in providing a large sample of potential cases.

Since the program's discrepancy is a tendency to miss rather than falsely generate bomb events, the North Atlantic maxima in Fig. 1 are quite real in ECMWF data. The higher frequency near Iceland is in broad agreement with finding of Serreze et al. (1997) that there has been an increase in cyclone frequency in this region since the early 1980s, which might be linked to the recent highly positive phase of the North Atlantic oscillation (e.g., Hurrell 1995). Moreover, later we will show that explosive cyclones in the Greenland–Iceland region exhibit a shorter life cycle and were more likely to be missed by the sparse and somewhat random ship observations (and therefore in manual analysis) compared to events at lower latitudes. This provides another explanation for their frequencies to be underestimated in the past.

c. Composite categories

Following Fig. 1 we divide the North Atlantic basin into three sectors such that each frequency maximum falls into either the northwest Atlantic (NWA), north-central Atlantic, (NCA), or the extreme northeast Atlantic (NEA) sector (Fig. 2). While all cyclones in Fig. 1 satisfy the Bergeron criterion for at least one 24-h period, their maximum 12-h deepening rate is used to group them into three intensity classes since it corresponds to the highest temporal resolution of the data. The classes are strong (ST, \( \geq 1.80 \) Ber), moderate (MO, 1.40–1.79 Ber), and weak (WE, 1.00–1.39 Ber), so chosen to yield roughly comparable numbers of cases in each category. Thus, all potential North Atlantic cases are grouped into a total of nine categories, three sectors times three intensity classes.

d. Final case selection and storm track correction

Several rules are applied to further select only the most ideal cases for the composites, with a goal to choose between 15 and 20 cases in each category. An ideal case 1) has only one dominant deepening period, whose maximum 12-h deepening rate is at least 0.5 Ber greater than that of any other deepening period it may have; 2) deepens primarily in its corresponding sector for at least 50% of its rapid development period, and 3) occurs between September and April. After all three rules were applied, cases farther away from their regional maximum in Fig. 1 were also dropped for categories that still had too many cases (>20). A final total of 162 cases were selected, but in this paper only the results of the ST NWA (18 cases) and ST NEA (19 cases) categories will be presented due to limited space (see Table 1 for a list of cases). For each case, data during a 7-day period from \( t_{0} \) to \( t_{72} \) for the Northern Hemisphere are extracted with \( t_{0} \) defined as the end point of the maximum 12-h deepening period, and used to produce composites.

During the final case selection process, MSLP maps
were used to manually correct any errors made by the
detection and tracking programs and to extend the storm
histories backward to early stages. Since composites are
made using a moving coordinate (quasi-Lagrangian ap-
proach) with the sea level low center being the common
reference point, a cautious treatment of storm tracks is
needed. Therefore, PV maps for a chosen potential tem-
perature ($\theta$) surface (usually between 300 and 309 K)
were also used in conjunction with MSLP maps to make
final surface storm track corrections for the cases se-
lected. Here, (Ertel’s) PV is defined as

$$PV = -g \left( \frac{\partial \theta}{\partial p} \right) (\zeta + f)_u,$$  \hspace{1cm} (2)

where $g$ is gravitational acceleration, and $f$ is Coriolis
parameter and must be computed on isentropic surfaces.
This is done by first finding the pressure level of chosen
$\theta$ at each grid point then interpolating $u$ and $v$ winds
onto the $\theta$ surface to obtain isentropic absolute vorticity,
while the vertical $\theta$ gradient is computed as described
in section 2e. This procedure only made a few minor
changes at early stages (mostly before $t_{-36}$) when the
PV value near the system center is approximately con-
erved.

e. Composite techniques and variables

Composites for a category are made for each variable
at each level by averaging values from all cases in that
category. Variables from $t_{-44}$ to $t_{-72}$ are first extracted
at all levels inside a finite domain constantly moving
with the sea level system for each case. The domain is
defined in latitude–longitude space with a fixed size of
75° longitude $\times$ 50° latitude ($31 \times 21$ grid points),
and the grid point (19, 11) [with (1, 1) at the northwest cor-
nor] is always placed at the sea level low center. Thus,
values for grid points at the same latitude–longitude
position relative to the sea level low center, and at the
same stage relative to $t_o$, are averaged to make the com-
posites. This differs from Manobianco’s (1989a) method
in which the 500-hPa vorticity center, rather than sea
level storm center, is the reference point. Data near
northern (southern) domain edges are occasionally ex-
cluded from the composite near the end (beginning) of
the 7-day period because the storm center is north of
65°N (south of 25°N).

Variables not provided by the ECMWF dataset are
calculated subsequently. For most of them (e.g., layer
thickness), the result is not influenced by whether the
variable is computed before or after compositing. How-
ever, this is not the case for nonlinear quantities (e.g.,
advection), and these variables are always computed
first at all grid points in the Northern Hemisphere before
finite-domain data extraction and compositing. This ap-
proach also ensures better approximation by finite dif-
fferences along domain boundaries, and is used for all
variables unless specified otherwise in the text. When
taking derivatives in the zonal direction, we skip an
increasing number of grid points between 62.5° and
87.5°N to avoid problems due to convergence of me-
ridians near the North Pole. Here, only enough points
are skipped to give a spacing at least that at 60°N ($\sim$139
km), and this same method is used for all calculations
for consistency. Vertical derivatives at a level are com-
puted by finite differences using interpolated values at
5 hPa below and 5 hPa above. The two levels used are
each bounded by 1000 and 10 hPa, such that layers of
5-hPa thickness are used for these two ends while those
of 10-hPa thickness are used for all intermediate levels.
Using such thin layers of equal thickness helps estimate

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Table 1. Cases used in the strong northwest Atlantic (ST NWA) and strong northeast Atlantic (ST NEA) composites. Variables (left to right)
are time, longitude ($^\circ$E), latitude ($^\circ$N), and central MSLP (hPa) at $t_o$. Time and date of each case are listed.

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<th>Lat</th>
<th>MSLP</th>
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gradients more accurately, since vertical data spacing is uneven and rather large in middle troposphere. For the cross-section analysis in section 4, composite values are interpolated onto levels at 50-hPa intervals for plotting.

3. Composite results

a. Mean storm track and evolution

Figure 3 presents mean storm tracks of explosive cyclones for the ST NWA and ST NEA categories. The ST NWA mean storm track (solid) starts near the Rockies at $t_{-36}$ and extends eastward just to the south of Nova Scotia at $t_0$, then northeastward across Newfoundland toward the southern tip of Greenland. The ST NEA cyclones have a mean track (dashed) farther east, with the storm center at $t_0$ to the southwest of Iceland near 58°N and extending toward Scandinavia near $t_{+24}$. These mean $t_0$ cyclone positions correspond well with the frequency maxima in NWA and NEA sectors in Fig. 1.

Table 2 shows that the MSLP of the mean ST NWA cyclone drops more rapidly from $t_{-36}$ to $t_0$, and slower afterward to gradually reach a minimum value of 967.3 hPa at $t_{+24}$. The mean NEA storm has more dramatic pressure drops between $t_{-24}$ and $t_0$, and obtains the lowest MSLP of 949.6 hPa at $t_{+12}$. This difference in pressure falls is due mainly to the latitudinal adjustment of the Bergeron criterion. In terms of deepening rates, more significant deepening (positive values) for ST NEA cyclones occurs from $t_{-36}$ to after $t_{+12}$, but extends longer to near $t_{+36}$ for NWA ones (Table 2). The averaged maximum deepening rate is 2.42 Ber for ST NEA cyclones and 2.17 Ber for ST NWA ones, suggesting that NWA cases are on average slightly more violent than NWA ones. After explosive development, the ST NEA mean deepening rate drops rapidly to below zero values by $t_{+24}$, indicating faster filling than NWA storms.

b. Mean sea level pressure and clouds

Figures 4 and 5 show the MSLP (hPa, solid) and approximated cloud regions (shaded) from $t_{-36}$ to $t_{+24}$ for the ST NWA and NEA composites. The cylindrical projection with the latitude to longitude ratio at 45°N, and a plotting domain of 45° longitude × 30° latitude, are used for these and all remaining figures in section 3. The ST NWA low (Fig. 4) deepens from 1008.6 to 999.5, then to 980.8 hPa from $t_{+24}$ to $t_0$, by 9.1 and 18.7 hPa, respectively, and afterward its central MSLP continues to decrease to 967.7 hPa at $t_{+24}$ (also Table 2). Throughout the period the system is accompanied by increasing clouds, which display a distinct "comma" shape from $t_0$ onward (Figs. 4d–f). In the NEA sector (Fig. 5), the projection-induced distortion is more severe toward later stages as the mean storm moves farther north (cf. Fig. 3). The ST NEA cyclone’s central MSLP values are consistently lower than that in NWA, with larger 12-h pressure drops from $t_{+24}$ to $t_0$ (by 14.7 and 27.4 hPa). Its central MSLP minimum of 949.6 hPa is reached at $t_{+12}$, then the system starts to fill (Figs. 5e and 5f).

The NEA incipient low develops from a pressure trough, with a “parent cyclone” to the northeast. The two lows move closer and the parent cyclone (989.7 hPa) is well inside the plotting domain at $t_{+12}$ (Fig. 5c). They rotate around each other cyclonically, and merge sometime between $t_{+12}$ and $t_0$ as the most rapid 12-h deepening occurs (Fig. 5d). It is worthwhile to note that since the composite origin is the incipient low center, the averaged strength of parent cyclones in reality is stronger than it appears in Fig. 5. On the other hand, little indication exists for either the parent cyclone or pressure trough in NWA (Fig. 4).

Other differences are also noticeable when the ST NWA and NEA cases are compared. The ST NEA cyclone appears to be stronger with a tighter MSLP gradient, especially immediately to its south and southeast before and during explosive deepening (but the increase
in near-surface wind should be comparable based on the geostrophic relationship). The NEA mature storm tends to weaken more rapidly after \( t_{0} \) as its central MSLP rises by \( t_{24} \), while the NWA low has not started to fill (also Table 2). Cloud area appears larger in NEA, but a greater portion of the difference after \( t_{0} \) may be artificially introduced by the distortion.

c. The 850-hPa height and temperature

The 850-hPa-level geopotential height (m, solid), temperature (K, dash), and positions of absolute vorticity (\( \eta \)) maxima are shown in Figs. 6 and 7. For ST NWA, the disturbance at \( t_{36} \) and \( t_{24} \) appears in the height field as a short-wave trough with a well-defined vorticity maximum (Figs. 6a and 6b). Significant development starts by \( t_{12} \), and the 850-hPa low deepens by 160 to 1180 m at \( t_{24} \), with the strongest height gradient immediately to the southeast of the center (Figs. 6c–f). The temperature disturbance has a rather small initial amplitude but lags the height wave by about \( \frac{1}{4} \) wavelength with the strongest gradient near the surface low, ideal for baroclinic development (Sanders 1971; Warrenfeltz and Elsberry 1989). By \( t_{12} \) the self-development process (Petterssen 1956a, b) has started to further amplify both temperature and height waves while promoting frontogenesis (Fig. 6c). The thermal wave becomes severely distorted by \( t_{0} \) (Fig. 6d), with a ridge extending toward the cyclone center from the southeast, and the thermal gradient starts to weaken by \( t_{24} \) (Fig. 6f).

In NEA (Fig. 7), the basic evolution is similar but the 850-hPa response is again greater than in NWA. The central height falls by 226 m from \( t_{12} \) to \( t_{0} \) and reaches an impressive 892 m at \( t_{24} \) (lowest among all categories, Figs. 7c–e). The height gradient and \( \eta \) maximum (Table 3) associated with the short wave are stronger in NEA and weaker in NWA. The temperature gradient at \( t_{36} \) and \( t_{24} \) in NEA is stronger and concentrated in a narrow zone across the surface low, and its thermal field distortion is greater during rapid development (Figs. 7a–d). After \( t_{0} \) a warm core seclusion forms and the temperature gradient deteriorates dramatically in NEA (Figs. 7e and 7f). The warm core seclusion, most often observed in rapid marine cyclogenesis, forms when warmer air is wrapped around the low center by colder air (e.g., Neiman and Shapiro 1993; Reed et al. 1994). The evolution of ST NEA cyclones at 850 hPa is quite sim-
ilar to the model proposed by Shapiro and Keyser (1990). In contrast, the NWA temperature gradient remains relatively strong (Fig. 6), and a warm core seclusion is not apparent in NWA composites, which is perhaps related to data resolution. In NEA, the parent cyclone is again visible at 850 hPa at $t_{12}$, and merges with the incipient disturbance (Figs. 7c and 7d).

d. The 500-hPa height and vorticity advection

Figures 8 and 9 depict composite 500-hPa geopotential height (m, thick), absolute vorticity advection $(10^{-9} \text{ s}^{-2}$, thin), and positions of vorticity maxima. At $t_{36}$ and $t_{24}$ the incipient trough in NWA lies about $7^\circ$–$10^\circ$ west of the surface low, with cyclonic vorticity advection (CVA) and divergence near the latter (Figs. 8a and 8b). The 500-hPa wave amplifies significantly at $t_{12}$ (Fig. 8c), with an enhanced height gradient at the trough base and near the surface low, and concurrent development occurs at lower levels (Figs. 4 and 6). This enhances the vorticity maximum (Table 3) and CVA, and helps to initiate the self-development process. The 500-hPa CVA grows from $2.8 \times 10^{-9} \text{ s}^{-2}$ at $t_{12}$ to $4.0 \times 10^{-9} \text{ s}^{-2}$ at $t_0$ (so does the vorticity maximum), and a strong positive and negative vorticity advection dipole develops at the vorticity center (Fig. 8d). In response to the increasingly distorted thermal field in the lower troposphere, the trough turns from positive (NE–SW) to negative (NW–SE) tilt. Further wave amplification after $t_0$ eventually causes the low to cut off by $t_{24}$ (Fig. 8f). Due to the self-limiting process, the vorticity advection dipole weakens even though the height gradient southeast of the low remains strong. The evolution is in very good agreement with sea level (Fig. 4), as MSLP continues to fall through about $t_{24}$ when CVA directly above the low becomes too small.

Several differences are evident when Figs. 8 and 9 are compared. First, at $t_{36}$ and $t_{24}$ the height gradient across the region of development is stronger in NEA, consistent with a stronger 850-hPa thermal gradient. This produces stronger CVA and reinforces the stronger low-level incipient disturbance discussed earlier. Second, the 500-hPa CVA forcing is also stronger during rapid development in NEA, where the CVA maximum exceeds $4.0 \times 10^{-9} \text{ s}^{-2}$ at $t_{12}$ (Fig. 9c) and reaches $4.9 \times 10^{-9} \text{ s}^{-2}$ at $t_0$, with higher values (shaded) over regions of strong diffluence and along the narrow area just downwind from the trough (Fig. 9d). The dipole structure is also stronger and the growth in vorticity larger during this stage in NEA (Table 3). Third, after rapid development the CVA in NEA weakens dramatically and moves farther away from the surface center (Figs. 9e and 9f), while that in NWA remains strong and close. This is tied to the formation of a cutoff low,
which occurs shortly after $t_0$ in NEA but much later in NWA, and in NEA is consistent with the diminishing thermal gradient. Finally, the parent cyclone is present in NWA even at 500 hPa, suggesting its deep vertical extent, and shows interaction with the developing low. As the parent cyclone approaches the incipient low from the northwest, the strong forcing causes the latter to intensify and become the more dominant feature of the two, then they merge (Figs. 9c±e). The role of the parent cyclone in the NEA sector will be further discussed in section 5.

e. The 250-hPa jet stream and divergence

Figures 10 and 11 show the jet stream forcing and evolution at 250 hPa with winds (streamlines and isotachs) and divergence (10$^{-5}$ s$^{-1}$, thin) plotted. In NWA, the jet shows an overall southward movement relative to the surface storm, and intensifies from 45 to 55 m s$^{-1}$ and becomes narrower between $t_{-36}$ and $t_0$ as simultaneous development occurs at all levels (Figs. 10a–d). The 250-hPa trough inferred from the streamline pattern evolves similarly to the 500-hPa trough, except that it is slightly farther west and no closed circulation forms up to $t_{-24}$. A region of divergence exists ahead of the trough but initially its magnitude is relatively small. As the jet core moves to the trough base, placing the surface low under its left-front quadrant, the approaching trough amplifies significantly. This combined CVA–jet streak forcing dramatically raises the maximum divergence from $1.4 \times 10^{-5}$ s$^{-1}$ at $t_{-24}$ to $3.3 \times 10^{-5}$ s$^{-1}$ at $t_{-12}$, and further to $4.1 \times 10^{-5}$ s$^{-1}$ at $t_0$ (Figs. 10b–d). The divergent center gradually moves toward the northeast after rapid deepening, as the surface low approaches its peak intensity (Figs. 10e and 10f).

During the incipient stage the jet in NEA also intensifies from 49 to 53 m s$^{-1}$ but its center is northeast of the surface low up to $t_{-12}$ (Figs. 11a–c). The divergence reaches a peak of $3.8 \times 10^{-5}$ s$^{-1}$ early at $t_{-12}$ but remains comparable through $t_0$, when a new wind maximum of 50 m s$^{-1}$ forms south of the storm center (Fig. 11d). As the cyclone migrates farther north, the separation between the jet and the surface low increases after $t_0$, and streamlines indicate formation of closed circulation at 250 hPa by $t_{+24}$ (Fig. 11f).

Many differences are also found by comparing Figs. 10 and 11. The stronger jet in NWA indicates that the baroclinicity here is of greater depth, while that in NEA, although stronger in a narrow zone at lower levels, is shallower. The primary jet center is southwest of the surface low in NWA during development, suggesting
the role of jet core dynamics. In contrast, the strongest NEA winds appear near the ridge before \( t_0 \), to the north-east of the surface low center (Figs. 11a–c). In NWA, the divergence is relatively weak before \( t_{24} \), but after the rapid deepening it remains comparable for a lengthy period, as the maximum decreases by merely \( 0.2 \times 10^{-5} \, s^{-1} \) between \( t_0 \) and \( t_{24} \) (Figs. 10). The NEA divergence, on the other hand, is considerably larger than in NWA up to \( t_{12} \), but drops rapidly after \( t_0 \) to \( 1.4 \times 10^{-5} \, s^{-1} \) at \( t_{24} \), consistent with the evolution in the MSLP field (Figs. 7 and 11). The jet and divergent region in NEA also move farther away from the surface low center after explosive deepening, consistent with lower levels and the formation of closed circulation at 250 hPa, which does not occur in NWA.

4. Vertical cross-section analysis

Composite vertical cross sections are also constructed along the lines AB in Figs. 4 and 5, along a direction roughly perpendicular to the upper-level wind above the surface storm center. The orientation of the AB line rotates gradually and changes from northwest–southeast initially to nearly west–east at \( t_{24} \). The sections pass through the sea level cyclone center at midpoint (marked as “0”), while the horizontal axis distance at each time is computed using the mean storm latitude in Fig. 3.

a. Northwest Atlantic composites

Figure 12 shows AB cross sections from \( t_{-36} \) to \( t_{24} \) for ST NWA. While the composites are unable to resolve features such as tropospheric fronts and tropopause folds at their true intensity, they are useful for qualitative comparison. The approach of an upper-level PV [defined in Eq. (2)] anomaly, associated with a lower tropopause (approximated by the 2 PVU line), before \( t_0 \) is evident (Fig. 12). By \( t_{12} \), a tropopause undulation has clearly formed above the surface low in the composite (Figs. 12e and 12f). Depicted by winds normal to the plane

<table>
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<th>Time (h)</th>
<th>( t_{-36} )</th>
<th>( t_{-24} )</th>
<th>( t_{-12} )</th>
<th>( t_0 )</th>
<th>( t_{12} )</th>
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<td>16.2</td>
<td>26.0</td>
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<td>28.9</td>
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<td></td>
<td></td>
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<td></td>
</tr>
<tr>
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<td>16.9</td>
<td>20.4</td>
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<td>18.9</td>
<td>26.7</td>
<td>26.3</td>
<td>25.0</td>
</tr>
</tbody>
</table>
b. Northeast Atlantic composites

Cross sections for the ST NEA, corresponding to Fig. 12, are presented in Fig. 13. The PV evolution in the AB cross sections is similar to that in NWA; however, the near-surface maximum (0.7 PVU) appears by \( t_{-36} \) at 920 hPa, then grows to 1.0 PVU and rises slightly at \( t_{-12} \) (Figs. 13a–c). The increase in the jet’s horizontal wind shear is quite dramatic between \( t_{-12} \) and \( t_0 \) near the PV center, along with a rapid upward extension of winds left of the storm center (Fig. 13c and 13d). At sea level the cyclone circulation reaches a peak of \( \sim 23 \) m s\(^{-1}\) at \( t_{-12} \) (Fig. 13e), but the associated PV anomaly has started to weaken considerably from 1.2 PVU. The \( \theta \) cross section (not shown) indicates downward bending of \( \theta \) surfaces over regions near the storm center by \( t_0 \), suggesting warm core development and a zone of weak stability, in good agreement with the 850-hPa results (Fig. 7). Such a downward bending of \( \theta \) surfaces is imposed by a warm anomaly below (Bretherton 1966; Hoskins et al. 1985) and is commonly found in oceanic explosive cyclones (e.g., Gyakum 1991).

c. Comparison between NWA and NEA

When NWA and NEA cross sections (Figs. 12 and 13) are compared, differences consistent with those found in section 3 can be seen. The upper-level PV values during development are higher in NEA than in NWA at the same pressure level (with a lower tropopause), especially at \( t_{-12} \) and \( t_0 \). At \( t_0 \), more high-PV air has been advected over the cyclone center in NEA,
while afterward the anomaly continues to grow more in NWA.

Regarding the upper-level jet, strong winds extend downward from its center in NEA during the incipient stage, best depicted by the 20 m s⁻¹ isotach (Figs. 13a and 13b). This extension in NWA is weaker and isotachs have a broader shape, suggesting a larger area with appreciable thermal wind and thus baroclinicity on the plane (Figs. 12a and 12b). By the same reasoning, the thermal gradient in early NEA stages must be strongest where vertical wind shear is large and minimized where isotachs are nearly vertical. The change in jet strength during development differs greatly, as there is virtually no reduction from $t_{-12}$ to $t_{-24}$ in NWA but substantial reduction (from 44 to 28 m s⁻¹) in NEA (Figs. 12 and 13).

The low-level PV anomaly, when it forms, is clearly separated from the upper-level anomaly in both sectors, but the appearance of a distinct center occurs much earlier in NEA (before $t_{-36}$) than in NWA (after $t_{-12}$). Its strength is also greater in NEA partly due to stronger shearing vorticity linked to the downward extension of strong winds. The NEA anomaly reaches maximum at $t_0$ then weakens rapidly, but that in NWA continues to grow through $t_{-12}$ and remains comparable at $t_{-24}$. The upward extension of cyclone circulation left of center between $t_{-12}$ and $t_{+12}$ is quicker in NEA. The vertical wind shear in NEA diminishes after development, indicative of a more equivalent barotropic structure, but remains relatively strong in NWA (Figs. 12 and 13). In the $\theta$ cross sections (not shown), the warm core structure (downward bending of $\theta$ surfaces) by $t_0$ is more visible in NEA, and the $\theta$ surfaces near the storm center are separated by greater distances. The latter indicates a lower static stability, which will be further discussed in the following section.

5. Discussion

Many features seen in sections 3 and 4 are consistent between levels and among variables in a dynamical sense and, therefore, did not arise randomly from compositing. These structural and evolutionary characteristics of ST NWA and ST NEA composite explosive cyclones are further discussed below in different aspects, with an emphasis on the northeast Atlantic sector.

a. Cyclone evolution and life cycle

For the ST NEA cyclones, results in sections 3 and 4 indicate a faster evolution and shorter life cycle, compared to ST NWA ones. This is manifested in the relatively early peak of upper-level forcing during the development and, after the rapid deepening, in 1) fast
weakening of sea level cyclone and forcing components, 2) deterioration of thermal and height gradients, and 3) the rapid occlusion and establishment of a more equivalent barotropic structure. In his pioneer work, Charney (1947) showed that for waves with the same wavelength, baroclinic instability increases with vertical wind shear, lapse rate (i.e., decreases with stability), and latitude. Hence, further works were carried out to examine the evolution in stability and thermal wind during the life cycle of composite cyclones.

Figure 14 shows the 1000–500-hPa (pressure weighted) stability factor computed from composite $\theta$ and averaged over a $5 \times 5$ box ($\sim 1250 \times 1250$ km$^2$) centered at the cyclone center from $t_{-36}$ to $t_{+12}$ for the ST NWA and ST NEA. The stability factor is $-g(\partial \theta/\partial p)$ in Eq. (2), and for the ST NWA (solid) its value is about 5.5 × 10^{-3} m$^2$ K kg$^{-1}$ at early stages, decreases slightly at the onset of rapid development, rises from $t_{-15}$ to $t_{+12}$, then further decreases again. On the other hand, the ST NEA stability factor (dashed), though comparable to ST NWA values initially, decreases significantly through rapid deepening stage to reach about $3.4 \times 10^{-3}$ m$^2$ K kg$^{-1}$ near $t_{+48}$. Such a condition facilitates the development of vertical motion, which plays a key role during the conversion process from available potential energy (APE) to kinetic energy as the baroclinic instability is released (see, e.g., section 8.3 of Holton 1992). This enhanced destabilization in NEA is apparently caused by both the faster approach of an upper-level cold low (or trough) and low-level warm core development. For instance, the 500-hPa height at cyclone center falls by 460 m from $t_{-36}$ to $t_{0}$ in the ST NEA, but only by 270 m in the ST NWA (Figs. 8 and 9).

Time series of the composite 1000–500-hPa thermal wind speed averaged over an $11 \times 11$ box ($\sim 2750 \times 2750$ km$^2$) is plotted to represent the evolution of overall environmental baroclinicity (Fig. 15). The ST NWA thermal wind (solid) increases during early stages and reaches a maximum of $\sim 24$ m s$^{-1}$ at $t_{0}$. In contrast, the peak thermal wind speed in the ST NEA (dashed) is only 20 m s$^{-1}$ at $t_{-48}$, and the implied baroclinicity decreases through the development from $t_{-48}$ to $t_{+36}$, to a value less than 10 m s$^{-1}$, as the release of baroclinic instability is clearly visible. Moreover, our results also suggest that the temperature gradient near $t_{-34}$ in ST NEA is strong only in a narrow zone near the incipient system in the lower and middle troposphere, while the gradient in NWA is less concentrated but deeper. This reduced overall environmental baroclinicity in NEA, combined with the weaker stability (Fig. 14), is likely
to cause a rapid depletion of APE as the thermal wind vanishes [see Eq. (3.38) of Holton 1992], in accordance with the evolution in Fig. 15 and the short life cycle of NEA cyclones seen in earlier sections. Thus, the ST NEA cyclones weaken immediately after $t_0$, while the NWA ones continue to develop for a longer period and/or to weaken more slowly.

The rapid evolution of explosive cyclones in the Greenland–Iceland area offers yet one more reason for the underestimation of their frequency in the past besides inadequate data coverage. That is, since NEA cyclones have faster life cycles, rapid development was more likely to be missed (compared to NWA ones) in manual analysis because the time span for a key ship observation to capture the phenomenon is shorter. Perhaps partly related to this underestimation, past studies mainly focused on rapid cyclogenesis closer to the North American coast (e.g., Bosart 1981; Gyakum 1983a, b; Anthes 1990; Uccellini 1990; Kuo et al. 1991a, b). More recently, with the aid of satellite data, operational models with dense enough grids can better capture these events at higher latitudes through internal dynamics, and some cases have started to appear in the literature (e.g., Gadd et al. 1990; McCallum and Grahame 1993; Grønås 1995; Kristjánsson and Thorsteinsson 1995).

b. Initial low-level system strength

Gyakum et al. (1992) show that for explosive maritime cyclones the frictionless vorticity equation at a surface low center can be reduced to

$$\frac{\partial \eta}{\partial t} = -\eta \nabla \cdot \mathbf{V},$$

where $\mathbf{V}$ is the horizontal wind vector and the cyclone vorticity grows roughly exponentially under constant low-level convergence (presumably induced by divergence aloft). Therefore, an antecedent surface vorticity development can be a useful indicator for further, more rapid, development. Compared to NWA, the ST NEA composite incipient cyclone is considerably stronger at sea level and 850 hPa, but weakens more rapidly after $t_0$ (Figs. 4–7). The distinct near-surface PV maxima in the ST NEA appears much earlier, before $t_{12}$ (section 4c). The stronger initial NEA disturbances are at least partially maintained by the forcing aloft (section 3), but could also be linked to the fact that they have moved over warmer water well before $t_0$, much longer than have NWA systems (Fig. 3). The stronger initial system strength at low levels in NEA suggests that it might play a greater role in determining...
the subsequent maximum deepening rate, and applying the antecedent development concept could have more success here, at least for stronger cases. Indeed, since explosive cyclones in NEA evolve faster and only a limited time is available to achieve a large deepening rate before environmental APE is depleted, the initial system strength must become important for a more violent development.

c. Role of parent cyclone

From earlier sections, the upper-level forcing we examined at the commencement of the explosive development is stronger in NEA, in terms of 500-hPa CVA, 250-hPa divergence, or cross-section PV anomaly. The produced net divergence during rapid deepening is also greater in NEA, as confirmed by examining the evolution of vertically integrated column mass fluxes at the cyclone center (not shown). These results are in agreement with the Bergeron definition, since a larger sea level pressure drop is needed to achieve the same deepening rate at higher latitudes. The interesting aspect, however, is that the forcing, strong enough to trigger some rather extreme cases listed in Table 1 (such as cases 3 and 12), still exists near Iceland at latitudes as high as 60°N. In this respect, the proximity of the NEA sector to the stratospheric high-PV reservoir certainly favors the creation of very strong IPV forcing, but upper-level northerly winds obviously must also exist to advect high-PV air southward to interact with the developing low, as discussed by Hoskins et al. (1985).

On the 315-K isentropic surfaces of our ST NEA category (not shown), the composite PV anomaly is also created through advection by northwesterly winds. Table 4 shows that the association between the incipient low and its parent cyclone is closer in stronger cases in the NEA sector. Near the onset of most rapid deepening \( t_{12} \), the composite parent cyclone in the ST NEA has deep vertical extent and is just to the north of the incipient low (section 3). At upper levels, the northerly branch of its circulation enhances the southward advection of high-PV air southward to interact with the developing low. At low levels, the confluence along the leading edge of cold air advection associated with the parent cyclone also enhances local baroclinicity (and promotes moisture convergence and surface energy transfer). Thus, a strong
parent cyclone for NEA cases tends to create favorable conditions at both upper and lower levels in a timely fashion, and help to create further rapid development. Sequential cyclogenesis, similar to the process just discussed, is perhaps the origin of the "cyclone family" concept developed at the Bergen School in the 1920s.

d. Other aspects

Although their scale and associated pressure perturbation are vastly different, a number of characteristics of explosive cyclones in the extreme northeast Atlantic obtained here are to some degree similar to those of polar lows, at least in a qualitative sense. These include a short life cycle, strong low-level baroclinicity, and weak stability in the vertical (Businger and Reed 1989), and their developments are both initiated by an upper-level CVA (or PV anomaly) under northerly winds (Rasmussen and Lystad 1987; Rasmussen et al. 1992). Emanuel and Rotunno (1989) used a theoretical hurricane model to simulate polar low development, and suggest that the mechanism of wind-induced surface heat exchange (WISHE; Emanuel 1986a; Rotunno and

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**Fig. 13.** Same as Fig. 12 but along AB lines in Fig. 5 for the ST NEA.

**Fig. 14.** The 1000–500-hPa pressure-weighted stability factor computed from composite $\theta$ and averaged over a $5 \times 5$ grid point box centered at the cyclone center from $t_{-36}$ to $t_{+32}$ for the ST NWA and ST NEA cases.

**Fig. 15.** The 1000–500-hPa thermal wind speed (m s$^{-1}$) of composites averaged over an $11 \times 11$ gridpoint box centered at the cyclone center from $t_{-36}$ to $t_{+32}$ for the ST NWA and ST NEA cases.
Emanuel 1987) for tropical cyclogenesis, although not spontaneous, is capable of sustaining a minimum sea level pressure of about 922 hPa (their “warm and moist” experiment using estimated sounding in a polar environment), a value comparable to those in some extreme NEA cases in Table 1. Emanuel (1986b) also proposed that the relative importance of WISHE to baroclinic instability for different types of lows during their most rapid deepening stage is, in decreasing order, as follows: tropical cyclones, polar lows (Bear Island type and comma cloud type), and ordinary cyclones. Based on our findings that the low-level PV anomaly during the incipient stage of ST NEA cyclones is important to further development, we suspect that a mechanism similar to WISHE could also contribute more in strong explosive cyclogenesis in the northeast Atlantic than in lower latitudes. This question will be further investigated in Part II of this paper, which presents results of a vorticity budget analysis from a quantitative perspective using all 162 cases selected.

6. Conclusions and summary

The present study employs twice-daily, gridded ECMWF operational analysis at 2.5° latitude–longitude resolution from January 1985 to March 1996, and carries out a composite analysis on explosive cyclones in the northwest (18 cases) and extreme northeast Atlantic (19 cases). To select ideal cases for the composites, the geographical distribution of explosive cyclogenesis derived from ECMWF data is plotted, and the results suggest a higher explosive cyclogenesis frequency in the Greenland–Iceland region, and therefore the frequency was likely underestimated in the past.

Characteristics of the dynamical and thermal structure and evolution of strong composite explosive cyclones (>1.8 Ber) in the northwest and extreme northeast Atlantic are compared and contrasted. The differences observed between cyclones in the two regions are dynamically consistent, and major findings are summarized below.

1) Although the Bergeron criterion requires a larger sea level pressure drop at higher latitudes for the same deepening rate, a fair number of strong explosive developments still occur in the extreme NEA sector, and some can be very violent. The composite upper-level forcing, as indicated by 500-hPa CVA, 250-hPa divergence, and cross-section PV anomaly, is no less intense at the onset of the explosive deepening in NEA than in the NWA sector off the North American coast.

2) Strong NEA explosive cyclones are associated with lower static stability and locally concentrated but shallower thermal gradient with less overall environmental baroclinicity. These factors lead to a rapid depletion of APE and, thus, a faster evolution and a shorter life cycle, with thermal gradient and forcing components diminishing immediately after rapid development. In contrast, the NWA cases, being closer to the major land–sea boundary, tend to have greater overall environmental baroclinicity, higher stability, and evolve slower.

3) The initial low-level disturbances in NEA also tend to be stronger with earlier formation of a distinct PV anomaly near the surface. This initial system strength appears to play a greater role in determining the subsequent maximum deepening rate in NEA cyclones, consistent with their shorter life cycle. The warm core structure produced during rapid deepening is also more pronounced in the NEA events.

4) For NEA cyclones, the close proximity of the Greenland–Iceland region to the stratospheric high-PV reservoir is one favorable condition to generate a strong upper-level forcing. In addition, a parent cyclone with appreciable strength through a deep layer to the north or northeast of the incipient low appears essential in producing a rather extreme explosive cyclogenesis. The suitable synoptic conditions are created as the parent cyclone’s circulation advects high-PV air southward at upper levels and enhances local baroclinicity at low levels simultaneously.

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