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RAPID DEVELOPMENT OF CYCLONES AT HIGH LATITUDES OVER THE NORTH ATLANTIC

DISSERTATION

Presented in Partial Fulfilment of the Requirements for the Degree Doctor of Philosophy in the Graduate School of The Ohio State University

By

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ABSTRACT

General characteristics of North Atlantic explosive cyclogenesis are compared and contrasted through a composite study and a vorticity budget analysis. Gridded surface and upper level operational analyses from the European Centre for Medium-Range Weather Forecasts (ECMWF) at 2.5° latitude-longitude resolution from January 1985 to March 1996 are used. A total of 162 explosive cyclogenesis cases are selected and classified into nine different categories, comprising three geographical sectors (the Northwest, the North-central and the extreme Northeast Atlantic) and three intensity classes (strong, moderate and weak), based on their 12-h maximum deepening position and deepening rate.

Results confirm the conclusions of earlier studies that explosive cyclogenesis is fundamentally driven by baroclinic instability, regardless of the intensity of development or geographic sector. At a smaller scale near the cyclone center, however, the total thermal contribution toward changing the layer thickness, largely from diabatic heating, is at least comparable to the contribution from upper level geostrophic forcing during the explosive deepening period. The diabatic heating arises mainly from (1) convergence of latent heat released and sensible heat transferred from the ocean by low level ageostrophic winds during the spin-up process, and (2) an increase in latent heat release at resolvable scales.
The low level convergence of heat promotes efficient warm core development, in a manner similar to tropical cyclone development.

Differences that are consistent in a dynamical sense are found in the evolutionary and structural characteristics of composite explosive cyclones in different sectors of the North Atlantic. Composite cyclones of the Northwest Atlantic sector are associated with weaker upper level forcing just before and during the most rapid deepening. These systems have weaker initial strengths at low levels, higher static stability, and stronger overall baroclinicity through a greater tropospheric depth, as they are closer to land-sea boundaries. Their development shows slower spin-up and spin-down processes, a longer life cycle, and a less pronounced warm core structure. Resolvable scale latent heating contributes relatively more toward the total diabatic effect near the storm center, consistent with more abundant moisture and a lower Bowen ratio. The self-development process in stronger Northwest Atlantic events appears similar to that of weaker events, but is simply more intense. The maximum deepening rate appears to be determined by the strength of upper level forcing, which is related to the storm environment baroclinicity.

Composite Northeast Atlantic cyclones near Iceland are accompanied by stronger upper level forcing, since a larger pressure drop is required to reach the same deepening rate. The close proximity of this region to the stratospheric high potential vorticity (PV) reservoir is helpful in creating stronger upper level PV anomalies under suitable synoptic conditions (such as significant cold air advection). Low level incipient systems also tend to be stronger, with distinct low level PV anomalies visible at least 24 h prior to the onset of the most rapid 12-h deepening. Their static stability is considerably lower and the
overall baroclinicity is weaker, with a shallow but concentrated temperature gradient near the system center before the rapid deepening. These factors and a larger Coriolis parameter result in rapid spin-up and spin-down processes and a shorter life cycle with rapid depletion in available potential energy immediately after the explosive deepening stage.

The maximum deepening rate of explosive cyclones in the extreme Northeast Atlantic tends to be determined by the strength of both the upper level forcing and low level incipient system, consistent with their short life cycle. Apart from the explosive deepening stage, the thermal contribution is important in maintaining incipient system strength and providing early destabilization, leading to significantly lowered stability that facilitates a more violent development. The warm core structure tends to be more pronounced in this sector and a larger portion of the diabatic heating is from low level convergence of added heat from condensation and surface fluxes while the rest is from latent heat release at resolvable scales, in agreement with a higher Bowen ratio.

It appears that a rather extreme explosive cyclogenetic event in the Northeast Atlantic requires that both strong upper level forcing and a strong incipient disturbance with lowered stability come together in a timely fashion. The existence of a “parent cyclone” of appreciable strength to the north or northeast of the incipient low tends to create such a favorable condition, by advecting high PV air farther southward toward the incipient low in upper levels, as well as by providing confluence to enhance the local temperature gradient at low levels.
Dedicated to my mother
I wish to thank my advisor, Dr. Jeffrey C. Rogers, for his continuous guidance and financial support throughout the course of this work, and for his patience in correcting both my scientific and stylistic errors. I am grateful to have the opportunity to work with Dr. Rogers, and without him this work would not have been possible. Thanks also go to my dissertation committee members, Dr. Jay S. Hobgood, Dr. Ellen Mosley-Thompson, and Dr. A. John Arnfield, as well as to Dr. John N. Rayner for their insightful suggestions, valuable comments and encouragement during this study.

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CHAPTER 1

INTRODUCTION

Explosive cyclogenesis is the rapid intensification of extratropical cyclones, first studied and defined by Sanders and Gyakum (1980). It is primarily a cold season maritime event and has been an area of active meteorological research during the past two decades. Attention arose because of the safety concerns associated with these intense winter cyclones, and the systematic failure of early operational models to predict their rapid development. A related scientific issue is whether these cyclones are dynamically different from ordinary cyclones.

Due to intensive research on this topic and rapid development of computer and numerical weather prediction (NWP) technologies, the operational models' skills in forecasting explosive cyclogenesis have been greatly improved, as has our knowledge regarding the deepening process and other aspects of these storms (Chapter 2). It is recognized that explosive cyclones are like ordinary ones, fundamentally driven by baroclinic instability, but latent heat release and surface energy fluxes also significantly contribute to the rapid deepening process. Thus, additional deepening rate is provided through abundant moisture supply in the marine boundary layer (MBL), and this energy
source makes explosive cyclogenesis a maritime phenomenon. Uccellini (1990) points out that the trend during the past 20 years toward emphasizing latent heat release and making analogies between explosive cyclones and hurricanes has nearly brought us full circle to the position of the “thermal theory of cyclones” put forward about a century ago (Fig. 1.1). Of course, studies during the first half of the twentieth century demonstrated the vital role of upper level dynamical processes in the development of extratropical cyclones, a concept not included in the “thermal theory.”

Past climatological studies suggest that explosive cyclogenesis occurs most frequently in mid-latitudes over the northwestern Pacific and northwestern Atlantic, along major baroclinic zones near land-sea boundaries (Section 2.1). This research, however, demonstrates that explosive cyclogenesis also occurs in the extreme Northeast Atlantic near Iceland, as well as in the North-central Atlantic to the south of Greenland with appreciable frequencies (Chapter 3). Since the Greenland-Iceland region is traditionally viewed as a region with little cyclogenetic activity, studies addressing explosive cyclogenesis there have been few and did not appear in the literature until recent years. All of these investigations have focused on individual events (Section 2.5) and a comprehensive study is needed. Therefore, the purpose of this dissertation is to compare and contrast explosive cyclones in different sectors of the North Atlantic, and to investigate whether there are discernible differences in the general characteristics of their development and evolution. Such differences are expected since the subpolar environment in high latitudes differs from the environment of mid-latitudes in many ways. An additional focus of this study is to emphasize the role of low level convergence of heat
Figure 1.1: Summary of major developments in the study of cyclogenesis from the 1800s to 1990 (from Uccellini 1990).
generated through diabatic processes (latent heat release and surface energy fluxes) besides the latent heating increase in the rapid deepening process of explosive cyclones, which is a point that tends to be overlooked in the literature.

Gridded operational analyses from the European Centre for Medium-Range Weather Forecasts (ECMWF) at 2.5° latitude-longitude resolution for both surface and upper levels are employed to conduct a composite study and a diagnostic vorticity budget analysis. The data set is generally of high quality in the Northern Hemisphere extratropics (Trenberth and Olsen 1988; Lambert 1988), and has been used by several investigators to study explosive cyclones (e.g., Wash et al. 1988; Manobianco 1989a; Wash et al. 1992). The vorticity budget analysis method developed in this dissertation, however, has not been seen in previous studies, and provides quantitative evidence and additional insight into the causes of the observed phenomena. During cold months (September to April) from January 1985 to March 1996, a total of 162 explosive cyclogenesis cases are selected, and classified into nine different categories (three geographical sectors of the North Atlantic times three intensity classes according to their maximum deepening rate). Composites are produced and the mean cyclones in different categories are compared and contrasted. It should be noted that these results represent the general characteristics of cyclones in different sectors and intensity classes, and variations among individual cases are expected, as shown by the modeling studies and sensitivity experiments discussed in Section 2.5.

Chapter 2 provides a literature review of explosive cyclogenesis covering aspects related to its climatology, the performance of operational models in detecting it, baroclinic instability theory, effect of moisture, the isentropic potential vorticity (IPV) viewpoint,
modeling and diagnostic studies, and the detailed mesoscale structure of explosive cyclones. Also reviewed in Section 2.7 are studies of polar lows, a group of mesoscale cyclones that exhibit similarities to explosive cyclones. Chapter 3 describes the ECMWF data set and the methodology employed to (1) identify potential cases of explosive cyclogenesis in the North Atlantic basin, (2) select the cases included in the study, (3) classify selected events into different categories, and (4) produce composites and the computation of different variables. Chapter 4 presents composite analysis results for key variables at various atmospheric levels and along cross sections. Results are discussed both from the traditional Quasi-Geostrophic (QG) perspective and from the IPV viewpoint. Vorticity budget analysis results are presented and discussed in Chapter 5, which also includes the derivation of the equation set based on the Petterssen-Sutcliffe development equation and the computational aspects used in this diagnostic study. The vorticity budget analysis is performed for different size areas near the cyclone center, as well as for different quadrants relative to the center, in an effort to investigate variations in vorticity budget variables toward the storm center and in different directions away from the center. Chapter 6 summarizes the major findings of the composite and vorticity budget analyses.
Chapter 2

Literature Review

This chapter reviews previous studies of explosive cyclogenesis, and is divided into seven sections. Section 2.1 reviews climatological aspects of this phenomenon and provides examples of rapid cyclogenesis in the Greenland-Iceland region. Section 2.2 summarizes the historical performance of operational models in predicting marine explosive cyclones. Sections 2.3 and 2.4 review theoretical aspects of baroclinic development in a moist atmosphere and the isentropic potential viewpoint, respectively. Section 2.5 discusses past modeling and diagnostic studies focused specifically on explosive cyclogenesis. Section 2.6 reviews some recent studies on the detailed mesoscale structure near the center of explosive cyclones and along their associated fronts. Due to their similarities to synoptic scale explosive cyclones, polar lows in the North Atlantic are also reviewed in Section 2.7. Additional general reviews are available in Hoskins (1990), Reed (1990), Shapiro and Keyser (1990), and Uccellini (1990).
2.1 Climatology of Explosive Cyclogenesis in the North Atlantic

Explosive cyclogenesis over the extratropical oceans in winter has been an area of active meteorological research during the past two decades. Rapidly intensifying cyclones often bring hurricane force winds and dangerous high seas that threaten the safety of sailing, fishing, oil-drilling and other local activities, as well as large amounts of precipitation along the Atlantic coast of North America. Attention also arose because early operational models systematically failed to predict the rapid intensification of these storms (Sanders 1986b). The phenomenon was first studied by Sanders and Gyakum (1980) who referred to these cyclones as atmospheric “bombs.” They defined the criterion for explosive cyclogenesis as a central sea level pressure (SLP) fall at an average rate of at least 1 Bergeron (or 1 B, in hPa h⁻¹) for 24 hours, and

\[ 1 \, B = \frac{\sin(\phi)}{\sin(60^\circ)}, \quad \text{Eq. (2.1)} \]

where \( \phi \) is the latitude of the cyclone center. The required 24-h pressure drop for a bomb varies from about 13.9 hPa at 30°N, to 24.0 hPa at 60°N, and to 27.7 hPa at the North Pole. It is defined like this because cyclones at different latitudes do not produce the same geostrophic wind speed for the same pressure gradient (c.f., Eqs. (5.1) and (5.2)), and the range of pressure fall needs to be adjusted geostrophically to reflect the actual corresponding wind speed increase. Therefore, the Bergeron definition is related more to changes in wind than pressure, since the wind and induced high seas are of major concern in explosive cyclones.
In some explosive cyclogenesis cases, the deepening rate can well exceed 1 B. Two famous examples are the *Queen Elizabeth II* (or *QE II*) storm in 1978 and the Presidents' Day cyclone in 1979, with average maximum deepening rate of about 2.9 B for 24 h and 3.9 B for 6 h, respectively (Bosart 1981; Gyakum 1983a; Kocin and Uccellini 1985). An analogy sometimes is made between these extreme events and tropical cyclones (Bergeron 1954; Riehl 1980), as the extraordinarily large pressure gradient near the storm center produces a comparable radial wind profile, and an eye-like cloud structure may be present in satellite images.

Sanders and Gyakum (1980) also present the first climatology of explosive cyclones. Using twice-daily maps of the 1978-79 cold season and a criterion of a 1 B deepening rate for 12 h, their seasonal distribution of explosive cyclones is shown in Fig. 2.1. Among a total of 150 cases, 61 occurred in the North Atlantic and 86 in the North Pacific, while 3 events were in neither basin. It is clear that explosive cyclogenesis occurs mainly in winter, as bomb frequency is quite low before November and after April. December and January have the highest frequency, when almost one cyclone reaches the criterion every day (Fig. 2.1a). The Atlantic basin peak occurs in February and the frequency drops rapidly thereafter (Fig. 2.1b), while the Pacific experiences an early peak in December and frequencies remain relatively high through much of the winter (Fig. 2.1c).

On the frequency distribution of deepening rates, the analysis by Roebber (1984) is the most intriguing. By using a one year sample of all Northern Hemispheric cyclones (*n* = 1130) and a reference latitude of 42.5°N (therefore 1 B = 19 hPa per 24 h), his resultant
Figure 2.1: Monthly frequency of explosive cyclogenesis for the 1978-79 cold season. Data used are twice-daily surface weather maps prepared by National Meteorological Center (NMC) from 1 September 1978 to 31 August 1979, and the criterion is 1 B deepening rate for 12 hours (from Sanders and Gyakum 1980).
distribution shows a range of adjusted 24-h pressure change between -43 and +20 hPa with a general trend of fewer cases toward greater pressure changes (Fig. 2.2a). The distribution also contains a tail at the rapid deepening side that causes the mean to be greater than the mode. Moreover, the chi-square goodness-of-fit test indicates that this distribution fails to be “normal” at 5% significance level, but can be fitted well by the sum of two separate normal curves that have different mean and standard deviation. One pair of such normal curves is shown in Fig. 2.2b, suggesting that the distribution can be produced by 975 ordinary cyclones with a mean of -3.75 hPa plus 155 bombs with a mean of -22.30 hPa. Roebber’s (1984) analysis therefore provides statistical evidence that explosive cyclones develop differently from ordinary ones in some meaningful way, and some mechanism (or mechanisms) other than the large scale baroclinic instability may significantly contribute to the deepening process.

The geographic distribution of explosive cyclogenesis events in the Northern Hemisphere during three cold seasons (Sanders and Gyakum 1980) is shown in Fig. 2.3, and the distribution of maximum deepening positions during a 7-year period (Roebber 1984) is shown in Fig. 2.4. From these figures, it is clear that explosive cyclogenesis is primarily a maritime event with appreciable overland frequency only in the eastern United States. The phenomenon occurs most frequently over the northwestern Pacific and northwestern Atlantic along major baroclinic zones where the temperature contrast between cold continents and warm oceans is large during winters. These locations also correlate well with regions of large sea surface temperature (SST) gradient along the edge of the warm Kuroshio current and Gulf Stream. In the North Pacific, other regions with
Figure 2.2: Distribution of adjusted 24-h deepening of all cyclones in the Northern Hemisphere (from 130°E eastward to 10°E) between February 1980 and January 1981. The adjusted deepening is defined as $\Delta p = \Delta p \sin(42.5^\circ)/\sin\phi$, where $\phi$ is the latitude of cyclone center. (a) The distribution is fitted with one normal curve, and the deepening criterion of bomb is indicated. (b) The same distribution is fitted with two normal curves with different mean and standard deviation (from Roebber 1984).
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Figure 2.4: Geographic distribution of 24-h maximum deepening positions of explosive cyclones during a 7-year period from 1976 to 1982. Data are once-daily (12 Z) and the same smoothing method as in Fig. 2.3 is applied (from Roebber 1984).
maximum activities are found to the south of the Bering Strait and Gulf of Alaska, while in the North Atlantic, few events occur north of 50°N near the Greenland-Iceland area. Although both the frequency and intensity of events are likely subject to underestimation, especially over regions with sparse data coverage (Sanders and Gyakum 1980), these and other studies (e.g., Gyakum et al. 1989; Chen et al. 1992) provide the best explosive cyclogenesis climatology to date.

The Greenland-Iceland region has traditionally been viewed as a resting place of migratory storms with little synoptic scale cyclogenetic activity. However, Serreze et al. (1993) and Serreze (1995) have recently shown that cyclogenesis is common in this region during the cold season, and Rogers (1997) finds that over 20% of the days experience 24-h pressure changes in excess of 20 hPa when the storm track is active in the extreme northeastern Atlantic. In Chapter 3, it will be shown that the frequency distribution derived from ECMWF operational analyses shows a considerably higher spatial variability in the North Atlantic, and past studies might have underestimated the frequency of explosive cyclogenesis in the high latitude regions of North-central and extreme Northeast Atlantic.

Figures 2.5 and 2.6 are 12-h National Meteorological Center (NMC, now National Centers for Environmental Prediction, or NCEP) final analysis of two rather extreme examples of explosive cyclogenesis in the Greenland-Iceland region. The first (Fig. 2.5) occurred on December 14-15 of 1986 and has been described by Gadd et al. (1990). At 00 Z, December 14 (Fig. 2.5a), the incipient low (968 hPa) is near 52°N, 42°W, to the south of a much stronger synoptic scale low (936 hPa). During the next 24 h, it deepens
Figure 2.5: A series of four NMC hemispheric final sea level pressure analysis at 12-h intervals for the North Atlantic region from (a) 14 December 00 Z to (d) 15 December 12 Z of 1986.
Figure 2.6: A series of four NMC hemispheric final sea level pressure analysis at 12-h intervals for the North Atlantic region from (a) 09 January 12 Z to (d) 11 January 00 Z of 1993.
explosively to 928 (Fig. 2.5b) and to nearly 900 hPa (Fig. 2.5c), setting a new all-time record of lowest pressure achieved by extratropical cyclones in the North Atlantic. For the latter time, Sanders (1987) reported an estimate by Bosart of 908 hPa, with the analysis certain to within a few hPa. After the rapid development, its central SLP rises at an equally impressive rate to 932 hPa by 12 Z of December 15 (Fig. 2.5d).

The second example took place from January 9 to 11 of 1993. The incipient low (992 hPa) is located near 47°N, 37°W at 12 Z on January 9 (Fig. 2.6a), again to the south of a synoptic scale low (958 hPa). It deepened rapidly to 960 hPa at 00 Z, January 10 (Fig. 2.6b) and further to 919 hPa twelve hours later (Fig. 2.6c), before the pressure gradient surrounding the center weakens at 00 Z on January 11 (Fig. 2.6d). Other examples of rapid deepeners include the storm in late September 1995 that damaged the research vessel *JOIDES Resolution* (Keigwin 1996; personal communication) and the one studied by Kristjánsson and Thorsteinsson (1995). These cases demonstrate that explosive cyclogenesis does occur in the Greenland-Iceland area and can be very violent, and they are all included in the sample of strong events in the extreme Northeast Atlantic in the present study (Chapter 3).

### 2.2 Prediction of Maritime Explosive Cyclogenesis

Early operational models often failed to predict explosive development of marine cyclones. Sanders and Gyakum (1980) evaluated the NMC 6-layer and 7-layer Primitive-Equation (PE) models during two cold seasons from 1977 to 1979. The two models had
essentially the same formulation, except that the 381 km horizontal resolution of the 6-layer model was doubled in the 7-layer model and the latter had one additional layer in the stratosphere. The 6-layer model predicted an average 12-h pressure fall of only 4.7 hPa for 46 western Atlantic storms, while the observed deepening was 16.3 hPa. The 7-layer model predicted 5.4 hPa of the observed 16.5 hPa mean deepening for another sample of 67 cases. Based on the minor improvement of the 7-layer model over the 6-layer one, Sanders and Gyakum (1980) concluded that further improvement in forecasts was unlikely to be achieved simply by increasing the model horizontal resolution, and some ingredients such as the bulk effect of cumulus convection, planetary boundary layer (PBL) processes, and adequate vertical resolution, must have been missing in the models.

Silberberg and Bosart (1982) evaluated the NMC's Limited-Area Fine Mesh Model (LFM) during the 1978-79 cold season, and found that the LFM had a tendency to underforecast oceanic cyclogenesis and overforecast continental events to the lee of major mountain barriers. Sanders (1986b) also evaluated the LFM's performance using 51 North Atlantic explosive cyclogenesis cases west of 53° W between 1981 and 1984. The averaged pressure fall between 12-h and 24-h model forecasts was about 9 hPa, 58 % of the mean observed deepening of 15.5 hPa. The correlation coefficient between observed and predicted deepening was 0.55, explaining about 30 % of the variance in the analyzed deepening data. Mean position errors indicated that the LFM tends to predict a storm track slightly too fast and slightly too far to its right. Although discrepancies existed, the LFM showed distinct improvement over the earlier 6-layer and 7-layer PE models.
The Nested-Grid Model (NGM) became the NMC’s operational regional model in March 1985. The model has two levels of nesting and three different grid sizes, a 240 km A-grid, a 170 km B-grid, and a 85 km C-grid that covers North America, the northwestern Atlantic and northeastern Pacific (Hoke et al. 1989). Sanders (1987) evaluated the NGM performance during the 1986-87 cold season. The averaged prediction of 12-h deepening for Northwest Atlantic cyclones was 9.9 hPa for the C-grid, accounting for 90% of the observed 11 hPa value. Compared with the LFM, the NGM clearly performs better in predicting Northwest Atlantic explosive cyclogenesis. Sanders (1987) attributed the improvement to better model resolution, improved analyses and model treatment of PBL processes, and suggested that the nature of the explosive cyclogenesis research problem has altered from one of discovering missing ingredients to one of improving model performance and extending the range of predictability.

The C-grid domain of the NGM was enlarged after 1987. Sanders and Auciello (1989) evaluated the performance of the NGM and the “aviation run” of the global spectral model (AVN) during the winter of 1987-88. They found that the forecast skill had improved since the preceding year, as the probability of detection of an event in a specified 24-h period was improved and the false alarm rate was reduced. The NGM mean predicted 12 to 24-h deepening was 9.4 hPa, on average 2 hPa less than the observed values, with a correlation coefficient of 0.55. The AVN model had similar values. The correlation for the short range forecast was limited mainly by the inability of the models to fully capture the initial onset of rapid deepening, although they tend to catch up later.
During the Experiment on Rapidly Intensifying Cyclones over the Atlantic (ERICA, Section 2.6), Sanders (1992) evaluated the performance of the Medium-Range Forecast (MRF) model. For 40 ordinary and explosive cyclones, the correlation coefficients between predicted and analyzed 24-h pressure changes for forecast days 0-1, 1-2, 2-3, 3-4, and 4-5 were 0.91, 0.82, 0.82, 0.68, and 0.57, respectively. This indicates that the ability of operational models in forecasting explosive cyclogenesis has been improved significantly, and is mainly a result of improved model resolution, initial conditions, and better treatment of convection, PBL processes and other subgrid scale processes (Kuo 1991).

Today’s operational models are clearly capable of successfully forecasting explosive cyclogenesis events. However, initial conditions are still limited by the sparsity of oceanic data (Sanders 1987), and model forecast skills often exhibit considerable run-to-run variability. Roebber (1990) showed that the NGM predicted an average deepening of 10.8 hPa for 52 cases during 1988-89 season, much weaker than the observed 17.0 hPa, and suggested that underforecasting is more serious for stronger events. Reed et al. (1988) presented three explosive cyclogenesis cases in the North Atlantic during a one-week period in January 1986. The earlier two events were well predicted by the ECMWF T106 operational model but the third one was not, although the model usually has high skill predicting intense marine cyclogenesis (Bengtsson 1990). Kuo et al. (1995) showed an example of a rapid mesoscale cyclogenesis event in eastern United State that was poorly predicted by NMC’s LFM, NGM and Eta models. In the latter two studies, the
authors suggest that quality of initial conditions appears to be the primary cause for the operational forecast failures.

2.3 Baroclinic Development in a Moist Atmosphere

It has long been recognized that development of extratropical cyclones in the mid-latitudes is driven by baroclinic instability. Charney (1947) explained the growth of baroclinic disturbances in terms of net conversion of available potential energy (APE) in the zonal current to kinetic energy (KE). As shown in Fig. 2.7, meridional air motion represented by solid arrows is thermally direct and baroclinically unstable with APE being converted into KE. Motion depicted by dashed arrows is baroclinically stable and the displaced air parcel will return to its original position (Holton 1992). Since the relative magnitude of the vertical to horizontal displacement (i.e., the slope of arrows) depends on the wavelength, only waves with intermediate wavelengths are baroclinically unstable and thus will grow for a given meridional temperature gradient (Fig. 2.8) and there is a shortwave cutoff beyond which waves will not grow (Wiin-Nielsen 1978). As the horizontal temperature gradient increases, so do the slopes of the isentropes in Fig. 2.7, and an air parcel moving along the same solid arrows would experience a greater temperature difference to its ambient environment and thus accelerates faster. This produces a more efficient APE to KE conversion process, and a larger growth rate and a wider range of unstable waves, together with a decrease in the wavelength of the shortwave cutoff (Fig. 2.8).
Figure 2.7: Slopes of parcel trajectories relative to the zonal mean potential temperature surfaces (or isentropes) for a baroclinically unstable disturbance (solid arrows) and for a baroclinically stable disturbance (dashed arrows) (from Holton 1992).
Figure 2.8: Theoretical limits of baroclinic instability as a function of wavelength ($L_c$ in $10^3$ km) and thermal wind ($U_T$) of the basic current. Along the vertical axis, four units of $U_T$ correspond to a horizontal temperature gradient of $2.8 \times 10^{-3}$ K km$^{-1}$, or a vertical wind shear of approximately 1.1 m s$^{-1}$ km$^{-1}$. The horizontal dashed line represents the typical value of $U_T$ at mid-latitudes, and $L_c$ (about 3000 km) is referred to as the shortwave cutoff and all waves shorter than it are stable regardless the strength of horizontal temperature gradient (from Wiin-Nielsen 1978).
Since explosive cyclogenesis occurs mainly in strong baroclinic zones (Figs. 2.3 and 2.4), the strength of upper level forcing initiating the release of instability and the baroclinicity of the environment are expected to be important. The region about one quarter the distance from a mobile upper level trough to its downstream ridge is most favorable for explosive cyclogenesis based on observational studies (Colucci 1985; Hirschberg and Fritsch 1991), composite analyses (Sanders and Gyakum 1980) and model simulations (Weng and Barcilon 1987; Warrenfeltz and Elsberry 1989). Sanders (1986a) found that stronger bombs of the northwestern Atlantic tend to be associated with faster-moving and stronger 500 hPa vorticity maxima. The phenomenon of explosive marine cyclogenesis appears fundamentally driven by the same mechanism as continental cyclogenesis, but the self-development process described by Petterssen (1956) is more intense and the low level response to upper level forcing is dramatically larger (Bosart and Lin 1984; Rogers and Bosart 1986; Sanders 1987; Lupo et al. 1992). This is consistent with findings that factors helpful to ordinary cyclogenesis, such as short wave trough, low level convergence, upper level divergence and vertical motion, are also important in more rapid deepeners (Macdonald and Reiter 1988; Rogers and Bosart 1991; Wash et al. 1992).

Since explosive cyclones tend to occur near regions of a strong SST gradient, the differential diabatic heating across the SST gradient inside the PBL might play a role in enhancing the local baroclinicity (Sanders and Gyakum 1980). Sanders (1986a) suggests that stronger Atlantic bombs move a longer distance along the strong SST gradient at the edge of Gulf Stream than do less intense cases. Bosart (1981) examined the evolution of the Presidents’ Day cyclone and found that coastal frontogenesis was initiated along a
zone of strong SST gradient and was enhanced by the cold air damming along the eastern side of Appalachians. The onset of explosive deepening was triggered by an intense mid-level short wave trough. Since bombs also occur in the northeastern Pacific and Atlantic where the sea surface temperature is nearly isothermal, strong SST gradients seem contributory but may not be crucial in all cases (Sanders 1986a).

Near the surface cyclone center where the three dimensional vorticity advection and tilting effect are small, the tendency of absolute vorticity ($\eta$) for adiabatic and frictionless motion (Gyakum et al. 1992) can be roughly approximated by

$$\frac{\partial \eta}{\partial t} = -\eta \nabla \cdot \mathbf{V}.$$  \hspace{1cm} \text{Eq. (2.2)}

Based on Eq. (2.2), the rate of cyclone vorticity generation is determined by the strengths of surface convergence (induced by upper level divergence) and of the system itself. Even forced by a constant convergence, a cyclone will not have a linear response and its absolute vorticity will grow exponentially with time. Therefore, rapid vorticity generation during the explosive deepening stage should be preceded by development at a slower rate, a concept that may be useful in forecasting explosive cyclogenesis. Gyakum et al. (1992) found that strong cyclogenesis cases tend to have a strong circulation at the onset of their most rapid intensification, as well as a longer period (more than 12 h) of antecedent development. Consistent with this theoretical aspect, Sanders and Gyakum (1980) found that most of the deepest lows qualify as bombs, while Roebber (1984) suggests that the mean deepening period for explosive cyclogenesis is much longer than that for ordinary cyclones.
Although explosive cyclogenesis is primarily a manifestation of the release of baroclinic instability (Sanders 1986a; Manobianco 1989b; Warrenfeltz and Elsberry 1989; Wash et al. 1992), some other mechanism or mechanisms must also contribute since baroclinic instability alone cannot fully account for the observed large deepening rates (e.g., Gyakum 1983a; Mullen 1983). One process related to upper level forcing is the intrusion of stratospheric high potential vorticity (PV) air during intense upper level frontogenesis and tropopause fold, a phenomenon first discussed by Reed and Sanders (1953) and Reed (1955). Keyser and Shapiro (1986) reviewed the structure and dynamics of upper level frontal zones by applying the two-dimensional secondary circulation model developed by Saywer (1956) and Eliassen (1962). In response to surface frontogenesis, a thermally direct secondary circulation is induced along the cross-frontal section plane to help maintain thermal wind balance, with warmer air rising to the south and colder air sinking to the north of the surface frontal zone (Fig. 2.9a, Keyser and Pecnick 1985). At upper levels, in the presence of cold air advection (CAA), the confluence brought about by the upper branch of the secondary circulation and the CAA (helped by the tilting effect associated with a local thermally indirect circulation just north of the jet core) can result in frontogenesis and intensification of the upper level jet (Fig. 2.9b). There is a marked difference in tropopause height across the jet and the air in the sinking branch of the transverse circulation originating in the lower stratosphere can penetrate the troposphere down to about 700 hPa level in 48 hours (Fig. 2.9c). The air of stratospheric origin is characterized by high PV values (due to strong stability) and high concentration of ozone and other radioactive substances. As it sinks and vertically stretches absolute vorticity is
Figure 2.9: Cross frontal sections of potential temperature (dashed) and (a) transverse ageostrophic circulation \( (v_{\text{ag}}, w) \) after 24 h integration, (b) along-front wind component \( (u) \) after 48 h integration, and (c) air trajectories originating at 0 h near the tropopause and ending at 48 h at \( x = 0 \) km. Integration is performed with a two-dimensional PE model of frontogenesis for case of confluence in the presence of upper level cold advection (from Keyser and Pecnick 1985).
generated rapidly in order to conserve PV (also Hoskins 1972; Shapiro 1976). This process may help to produce explosive cyclogenesis near the surface.

Uccellini et al. (1985) provide observational evidence that a tropopause fold formed and stratospheric air descended into the troposphere prior to the explosive deepening of the Presidents’ Day cyclone. The high PV air was nearly collocated with the surface cyclone center when rapid development occurred. In this particular case and others like it, the dramatic increase of low level absolute vorticity is also related to the intrusion of stratospheric air (Bosart and Lin 1984; Reader and Moore 1995). However, since significant intrusion of stratospheric air is not reported in most cases of explosive cyclogenesis, this process may only play an important role in some events.

The baroclinicity of the atmosphere and strength of upper level forcing are only part of the story, since they cannot explain why explosive cyclogenesis occurs almost exclusively over oceans. A steep slope of isentropic surfaces (Fig. 2.7), and therefore a more efficient APE to KE conversion, can also occur due to the same horizontal temperature gradient but with a lower static stability. In such a situation, some initially stable shorter waves become unstable, and wavelengths of both the shortwave cutoff and the most unstable waves shift toward the left in Fig. 2.8. Since the compensation by vertical motions is reduced when the stability is low, the APE to KE conversion is further facilitated and a larger growth rate can be achieved.

The release of latent heat with condensation of water vapor in the atmosphere is one process by which stability can be substantially reduced over the course of extratropical cyclone development. Since observations reflect the combined results of all atmospheric
physical processes and the effects from one process cannot be easily isolated, most of past studies on this topic have been done through numerical simulations and sensitivity tests. Gall (1976) used a 9 sigma-layer, finite-difference PE nonlinear general circulation model (GCM) with a channel domain to investigate the effect of latent heat release on growing baroclinic waves. With idealized initial conditions, his results show that growth rates of all wavelengths are significantly increased when moisture is present in the model, although the most unstable wavelengths remain unchanged. He suspects this to be caused mainly by the poor model resolution of shorter waves. Emanuel et al. (1987) employed a two-dimensional, 2-layer semi-geostrophic (SG) model (Hoskins and Bretherton 1972) to obtain the analytical relationship between growth rate, wavelength and stability. Their solutions confirm that maximum growth rate increases significantly as the stability approaches conditional symmetric neutrality (e.g., Bennetts and Hoskins 1979; Emanuel 1979), and the most unstable wavelength shortens to about 60% that of the dry mode, together with a shortening of the shortwave cutoff. In this work, Emanuel et al. (1987) also performed simulations using a multilevel version of the model, which produced similar results.

Gutowski et al. (1992) examined the interaction between moisture and baroclinic eddies using a 10 sigma-layer version of the NMC PE global spectral model (GSM) which includes both stable precipitation and subgrid scale moist convection. They concluded that the greatest effect of latent heating on the life cycle of baroclinic waves is a stronger conversion of eddy APE to eddy KE, associated with an increased ratio of vertical to meridional heat flux (cf. Fig. 2.7). This enhanced conversion exhausts the APE reservoir
more rapidly and thus induces a faster life cycle in moist experiments despite the additional APE generated by latent heating. In a more recent study, Whitaker and Davis (1994) used a two-dimensional linear QG Eady model and a nonlinear three-dimensional hydrostatic PE model to simulate the effect of condensation associated with stable precipitation by forced uplifting alone, assuming that all rising (sinking) air is saturated (unsaturated) and there is no potential instability anywhere in the models. The most unstable mode in their linear model moist run has a slightly shorter wavelength and slightly faster growth rate. The surface cyclone grows significantly faster however, in their nonlinear model when the moisture is included. The fronts, especially the warm front, are much sharper in the moist simulation. They conclude that this rapid surface cyclone growth, which cannot be anticipated in the linear model, results from the superposition of low level vorticity, associated with the PV anomaly due to latent heating, upon the circulation associated with the advective PV anomalies in the troposphere (Whitaker and Davis 1994).

The effect of latent heating associated with cumulus convection near the storm center and conditional instability of second kind (CISK) has been suggested as a possible mechanism responsible for the additional deepening rate observed, since many cases of explosive cyclogenesis have developed a warm core structure in the lower troposphere before the onset of their rapid deepening and have exhibited cloud features reminiscent of those around tropical cyclones (Bosart 1981; Gyakum 1983a, b; Anthes et al. 1983; Uccillini 1986; Gyakum 1991). In the investigation of the Presidents’ Day cyclone (Bosart 1981) the environment was apparently able to sustain a super-adiabatic lapse rate near the ocean surface to support the convection. Bosart suggested that PBL processes and
convective scale forcing may be fundamental in providing differential diabatic heating, so the possible role of CISK to the continuing cyclone development cannot be ruled out. Sanders and Gyakum (1980) and Gyakum (1983b) also suspected that insufficient representation of the bulk effect of cumulus convection may be one of the reasons for the failure of early operational models in forecasting explosive cyclogenesis. The importance of latent heating associated with condensation and precipitation, and the effect of PBL processes and other factors have been investigated mostly through numerical simulations, and will be reviewed and discussed in Section 2.5.

2.4 Isentropic Potential Vorticity Viewpoint

Hoskins et al. (1985) developed an IPV viewpoint and discussed baroclinic development as the result of interaction between an advancing upper level IPV anomaly and a low level warm anomaly. In isentropic coordinates following an adiabatic frictionless motion, the conserved quantity is Ertel's potential vorticity, defined as

$$P = -g \left( \frac{\partial \theta}{\partial p} \right) (\zeta_\theta + f),$$

Eq. (2.3)

where $g$, $\theta$, $\zeta_\theta$, and $f$ are gravitational acceleration, potential temperature, isentropic relative vorticity, and Coriolis parameter (or planetary vorticity), respectively. $P$ has units of $\text{m}^3 \text{s}^{-1} \text{K} \text{kg}^{-1}$, and is simply the product of isentropic absolute vorticity ($\zeta_\theta + f$) and a stability factor $\sigma^{-1}$, where $\sigma$ is the mass density in $(x, y, \theta)$ space defined as
\[ \sigma = -\frac{1}{g} \left( \frac{\partial P}{\partial \theta} \right). \]  

Eq. (2.4)

As shown by Hoskins et al. (1985) and Thorpe (1985), an isolated positive IPV anomaly (relative to the global mean of \( P \) for that isentrope) is associated with cyclonic circulation which extends vertically, and the opposite is true for a negative IPV anomaly (Fig. 2.10). Like an isobaric vorticity maximum, the IPV anomalies can be advected and have similar dynamical significance in the development of synoptic scale weather systems. When an upper level IPV anomaly, similar to the one in Fig. 2.10a, is approaching a particular location, the air column in that location must stretch vertically to increase \( \zeta_o \) in order to conserve \( P \) in response to the lowered stability without changing the local Coriolis parameter \( f \). Thus a cyclonic circulation is induced throughout the depth of the troposphere. It has furthermore been shown that a surface warm anomaly is effectively equivalent to a positive IPV anomaly without such a variation in temperature, and similarly a surface cold anomaly is equivalent to a negative IPV anomaly (Bretherton 1966). An example of the circulation induced by a pair of warm and cold anomalies is given in Fig. 2.11 showing that the relationship between temperature and IPV anomalies is established through the effect of temperature anomalies in changing static stability.

Hoskins et al. (1985) proposed an ideal model of cyclogenesis induced by the arrival of an upper level positive IPV anomaly over a low level baroclinic zone (Fig. 2.12). The thermal advection by the induced low level circulation creates a warm anomaly ahead of the upper level IPV anomaly at the surface. This induces its own circulation and reinforces the anomaly aloft by advecting high PV air equatorward west of the IPV
Figure 2.10: Circularly symmetric flows induced by simple, isolated, positive (a) and negative (b) IPV anomalies near the tropopause. Two sets of thin contour lines are isentropes and transverse velocity at 5 K and 3 m s\(^{-1}\) intervals, respectively. The tropopause is depicted by the thick line, and the location of IPV anomalies is stippled. The basic stability is uniform in the troposphere and six times larger in the stratosphere. The domain size is 5000 km in the horizontal and 16.67 km in the vertical (from Hoskins et al. 1985).
Figure 2.11: Circularly symmetric flows induced by simple, isolated, positive (a) and negative (b) temperature anomalies at the lower boundary. The variables, contour method, and domain size are the same as in Fig. 2.10 (from Hoskins et al. 1985).
Figure 2.12: A schematic picture of cyclogenesis associated with the arrival of an upper air IPV anomaly over a low level baroclinic zone. In (a) the upper level cyclonic IPV anomaly (solid plus sign) has just arrived over a region of significant low level baroclinic activity. The circulation induced by the anomaly (solid arrows) leads to a warm temperature anomaly somewhat ahead of the upper IPV anomaly (open plus sign) in (b). The warm anomaly then induces its own cyclonic circulation (open arrows). If the equatorward motion at upper levels advects high PV, lower stratospheric air, and the poleward motion advects low PV, upper tropospheric air, then the upper level circulation induced by the surface warm anomaly will, in effect, reinforce the upper air IPV anomaly and slow down its eastward progression (from Hoskins et al. 1985).
maximum (Fig. 2.12b), subsequently reducing the transient speed of the upper level IPV. Since the circulation of the IPV anomaly also slows down the low level wave by shifting the temperature advection maxima slightly to the west than where they otherwise would be, in this process both anomalies not only tend to amplify each other but also tend to "phase-lock" through the interactions between them. Here, static stability still plays a vital role because it determines how easily the circulations associated with the IPV and temperature anomalies can penetrate vertically and thus how effectively they can interact (Hoskins et al. 1985).

The IPV viewpoint can be particularly useful in the case of explosive cyclogenesis for several reasons. First, cyclogenesis is viewed as the result of interactions between upper and lower level IPV anomalies, so the upper level dynamical forcing and the low level thermodynamic component (such as temperature advection and latent heating) are no longer conceptually treated as two independent processes. This is more consistent with the nonlinear nature of the deepening process. Second, the effect of intrusion of stratospheric air into the middle and even lower troposphere (Section 2.3) is implicitly included in the model through an enhanced upper level IPV anomaly and the downward extension of the induced circulation. Third, the low level warm anomaly does not have to be produced by temperature advection, and any other physical process that can produce or enhance the anomaly (such as latent heating, other diabatic effect, and PBL fluxes) is contributory to the development. In addition, the warm anomaly can be a pre-existing one, not necessarily induced by the approaching upper level PV anomaly. Given the dynamical significance of IPV, there has been an increasing application of the "IPV
viewpoint" in more recent literature, such as Reed et al. (1992, 1993b, 1994), Gyakum et al. (1995), Davis et al. (1996), Kuo et al. (1996), and Rivals et al. (1998).

2.5 Modeling and Diagnostic Studies of Explosive Cyclogenesis

As mentioned in Section 2.3, it is difficult to study effects of individual processes using observational data directly, and this is particularly true for explosive cyclogenesis since routine observations over the ocean are usually of inadequate spatial and temporal resolution for detailed diagnosis of the deepening process. Therefore, except for climatological studies (Section 2.1) and a few individual events (Section 2.3), most studies are performed either through numerical modeling and sensitivity tests or by utilizing model output for diagnosis (as the approach employed in this dissertation). Without rapid and continuous development of NWP techniques during recent decades, such an approach could never be successful (Anthes 1990). Major factors investigated include effects of latent heat release and surface energy fluxes, model resolution, initial conditions and forecast range.

Anthes et al. (1983) used the Pennsylvania State University/National Center for Atmospheric Research (PSU/NCAR) mesoscale model and carried out a series of simulations with different initial conditions, horizontal resolution, and model treatment of condensation and PBL processes to investigate their relative importance on the QE II storm development. Their results suggest that baroclinic instability was the primary mechanism for the early intensification, but latent heating became progressively more
important toward the later stage when the storm and the associated vertical motion grew stronger. The major effect of latent heating is in the storm intensity and storm track. Similar results are obtained by Sardie and Warner (1983) and Fosdick and Smith (1991).

Kuo et al. (1991a) further performed two experiments on the QE II storm using a newer version of the PSU/NCAR model (MM4), having a horizontal resolution of 45 km and 16 sigma-layers. The simulation that included latent heating effects predicted the storm track significantly better, and produced a much stronger storm after 24 h of integration starting from the same initial field. Although the adiabatic simulation can predict the synoptic scale deepening, the model cannot capture the intense mesoscale inner core and its large pressure gradients without latent heating. Kuo et al. (1991a) suggest that latent heating in the QE II storm may enhance the secondary circulation associated with the frontal system, establishing a positive feedback process similar to that in CISK, such that explosive cyclogenesis should be viewed as a moist baroclinic instability process with nonlinear interactions between the baroclinic dynamics and the latent heating process. Rogers and Bosart (1991) similarly suggested that latent heating associated with deep convection plays a crucial role in the storm deepening and tight pressure gradient near the cyclone center.

Reed et al. (1992) used MM4 model output to study the PV structure from a successful simulation of a rapid marine cyclogenesis event. They indicate that a low level positive PV anomaly in and near the frontal cloud band formed rapidly through condensational heating as air from the PBL ascended along the warm frontal surface, while sensible heat flux played little role in its formation. In an earlier study using
ECMWF forecasts, Reed *et al.* (1988) also suggested the effect of latent heating in two explosive cyclones near Iceland contributed about 40 to 50% of the rapid deepening.

Reed *et al.* (1993b) performed successful simulation and sensitivity tests on the ERICA Intensive Observation Period (IOP) 5 storm using the MM4 model, showing that this case had high sensitivity to latent heating and moderate sensitivity to surface energy fluxes, Gulf stream position, and model resolution. This particular case was not well predicted by the NGM model, and about 60% of the 13 hPa error in forecast storm central pressure is attributable to the finer grid size and the Grell convective parameterization scheme (*Grell et al.* 1994) of the MM4, while the remaining 40% is due to other factors. The result of their diagnostic analysis indicates that diabatically produced low level PV also appeared in the warm frontal cloud mass, and the storm environment was neutral or even unstable to vertical and slantwise convection.

Sensible and latent heat fluxes in the PBL have also been suggested as important processes that help the rapid development of marine cyclones (e.g., Bosart 1981; Atlas 1987; Nuss and Anthes 1987; Sanders and Davis 1988; Davis and Emanuel 1988; Hedley and Yau 1991). In regions with strong oceanic SST gradients, sensible heat flux can modify the atmosphere and enhance the low level baroclinicity through differential heating. The fluxes from the warmer ocean surface can simultaneously heat and moisten the lower levels and thus destabilize the atmosphere to produce conditional or convective instability. The addition of water vapor through evaporation supplies the moisture available for latent heat release in the updrafts when condensation occurs. In experiments by Anthes *et al.* (1983), replacing the bulk PBL parameterization by an explicit PBL model with three
additional layers produced a considerably deeper and stronger model cyclone. Atlas (1987) used the Goddard Laboratory for Atmospheres (GLA) model to simulate the development of the Presidents' Day cyclone and found that the model correctly predicted the intense coastal cyclogenesis in the control run, but failed to predict any cyclogenesis when surface fluxes are excluded from the model since about 36 h before the most rapid deepening. In the PE model simulations by Fantini (1990), including surface fluxes and the slantwise convective adjustment, the model baroclinic waves exhibit deepening in two phases, an early exponential growth phase followed by an explosive growth phase, with the latter corresponding to hurricane-like kinetic and thermal structures.

Mullen and Baumhefner (1988) used the NCAR 9-layer spectral R31 global Community Climate Model (CCM) to carry out a sensitivity study on 11 eastern Pacific explosive cyclones. Their results suggest that baroclinic dynamics account for roughly one-half of the cyclones' deepening, while total diabatic heating accounts for the remaining half. For the portion from diabatic heating, latent heat release contributes nearly one-half of the central SLP drop, but the storms cannot reach their maximum intensity without PBL processes that keep supplying a large amount of sensible heat and moisture. By comparing the control experiments with simulations having only dry dynamics, Mullen and Baumhefner (1988) also report a much warmer lower troposphere (by about 12 K) and an apparent slower traveling speed in the later stage of control run cyclones. The modeling and diagnostic study by Manobianco (1989b) using the Florida State University Global Spectral Model (FSUGSM) suggests a smaller contribution to surface pressure tendency from the fluxes, less than 10 %, while latent heat release
accounts for no more than 40% with the remainder due to dynamical adiabatic forcing in three cases.

Rogers and Bosart’s (1991) diagnostic study of two intense cyclones along the east coast of North America suggests the importance of oceanic heat and moisture fluxes in enhancing low level baroclinicity and destabilizing the air mass along the cyclone’s path, but no widespread area of convection and convective instability was observed during the explosive deepening. Kuo et al. (1991b) conducted a series of eight experiments on seven explosive cyclones using the PSU/NCAR MM4 model to investigate effects of surface energy fluxes both during and preceding the rapid deepening. They found that fluxes occurring during the 24-h period prior to the onset of most rapid deepening are important to storm development, but the concurrent fluxes during the explosive deepening stage had very little effect. They suggested that the major effect of energy fluxes is to precondition the marine environment to one favorable for explosive development, and to supply moisture as an energy source of latent heating when storms intensify rapidly. Once this potential is realized, rapid deepening can occur without concurrent surface fluxes (Kuo 1991).

The performance of numerical models, operational or not, is also affected by other factors, including the horizontal and vertical resolution of the model, the quality of the initial field starting the simulation, as well as the forecast range. The improvement in model resolution, initial conditions, and improved model treatment for convection, PBL processes and other subgrid scale processes, together with the increase in computing power, are the primary reasons for success in forecasting explosive cyclogenesis by
operational models in recent years (Sanders 1987; Kuo 1991). Anthes et al. (1983) estimated that at least 4 vertical layers are needed below 700 hPa for successful simulation of explosive deepening, while doubling the horizontal resolution from 90 to 45 km resulted in only modest improvement (also Kuo and Reed 1988). Kuo and Low-Nam (1990) found that a reduction in grid size from 80 to 40 km produced an average additional pressure fall of 3 hPa during the rapid deepening stage for 9 cases, while the results showed a slight damping effect when the number of model layers increased from 15 to 23. They suggested that a model with grid size around 80 km should have an optimal vertical resolution near 15 layers.

Hedley and Yau (1991) consider a high horizontal resolution approaching 10 km very important in successfully simulating the structure of the intense warm front and the rapid deepening of the cyclone. Balasubramanian and Yau (1994) also produced a realistic explosive cyclone by using a two layer PE model with slantwise convection parameterization, and suggested that the formation of the bent-back warm front due to intense convection near the storm center is closely related to the onset of the explosive cyclogenesis phase in some events. It therefore appears that a fine grid size may not produce a much deeper cyclone, but is important to resolve the detailed mesoscale structure of pressure, temperature, wind and precipitation near the center of explosive cyclones and along their associated fronts. Further discussion regarding this topic is presented in Section 2.6.

The effect of initial conditions on the QE II storm simulation was also investigated by Anthes et al. (1983). They showed that a successful forecast can be achieved by
adding a subjectively prepared, supplementary data set to the initial field, by using an explicit treatment for condensation, and by coupling the model with a multilayer PBL model. However, without the supplementary data set, the simulation showed little improvement over forecasts by operational models. Kuo and Reed (1988) also found that a poorer initial state can have two negative impacts; a weaker storm and a delay in the onset of rapid deepening. Similar results are also obtained by Atlas (1987). Mullen and Baumhefner (1989) and Kuo and Low-Nam (1991) both showed that the error growth rate in the explosive cyclogenesis environment is significantly higher than that in an averaged mid-latitude flow, while the latter study found that as the forecast range increases the forecast skill shows serious degradation in terms of the predicted pressure fall, the onset of rapid deepening, and the position of the storm center.

Manobianco et al. (1992) also investigated the impact of data assimilation on the simulation of the O E II storm using the Goddard Mesoscale Atmospheric Simulation System (GMASS). Experiments that were initialized using data assimilation and with supplementary data both showed improved result in terms of cyclone intensity and position during the initial phase of rapid development. These runs captured the developing cyclone, while the experiment without supplementary data or assimilation delayed the onset of deepening by 6 to 9 hours. Quality of initial conditions is also crucial in successfully simulating a continental cyclogenesis event, as the diabatic processes played a vital role in the development of this case (Kuo et al. 1995).

Keyser (1991) notes that explosive cyclogenesis research is showing signs of maturity. Extensive research during the past decade has recognized that the strength of
upper level forcing, latent heating, PBL processes, as well as effective static stability are all important in producing explosive marine cyclogenesis. For successful simulations of these systems, model treatment in subgrid scale processes (particularly cumulus convection and surface energy fluxes), model resolution, and initial conditions are also important. Uccellini (1990) stated that the individual dynamic and diabatic processes should be viewed as necessary for rapid cyclogenesis, but not sufficient to produce large deepening rates when acting alone. Therefore, the rapid development phase of extratropical cyclones depends not on the individual contribution of these processes, but on nonlinear interactions among them.

The importance of individual processes to development, however, can vary to a large degree from one explosive cyclone to the next, as shown in Table 2.1 for effects of condensational heating and surface energy fluxes based on several modeling studies (Reed et al. 1993b). The inclusion of latent heat release in the model has positive impact on the development of all cases, but the ratio between simulated 24-h pressure falls in control runs and experiments without latent heating (column 7) varies from 1.11 (slightly positive) to 3.36 (extremely important). The impact of surface energy fluxes on model cyclones (column 9) does not vary to a nearly as large degree, and can be slightly negative (0.90) to moderately positive (1.27). The difference is partly due to the different models used in the studies, as well as their differing parameterization schemes for subgrid scale convection and PBL processes. From studies reviewed in this section, it is nonetheless clear that for certain explosive cyclogenesis cases the model performance depends critically upon successful reproduction of moist processes (Reed et al. 1993b).
Table 2.1: Effects of condensational heating and surface energy fluxes on deepening rate during the 24-h period of most rapid deepening (from Reed et al. 1993b).
2.6 Mesoscale Structure of Explosive Cyclones

Two field experiments, the Genesis of Atlantic Lows Experiment (GALE, Dirks et al. 1988) and ERICA (Hadlock and Kreitzberg 1988) were launched in 1986 and during the winter of 1988-89, respectively. Their goals were to obtain observational data with adequate horizontal and temporal resolution for explosive cyclogenesis research and to validate model simulation, eventually leading to improved knowledge and forecasts. Data collected especially by buoy network and aircraft allow much more detailed study of the evolution and mesoscale structure of pressure, temperature, winds, cloud and precipitation patterns near the center of explosive cyclones and along their associated fronts during the rapid deepening stage.

Neiman and Shapiro (1993) and Neiman et al. (1993) documented the evolution of the ERICA IOP 4 storm that deepened 60 hPa in only 24 hours. As shown in Figs. 2.13 and 2.14, observed low level features include intense warm and cold fronts, a bent-back front that extends from the warm front westward relative to the storm into the polar airstream, a secondary cold front that develops out of the bent-back front to the south of the low center, as well as a warm core seclusion that forms as air behind the cold front wraps around a pocket of relatively warm air near the storm center. The cyclone center shows a displacement relative to the triple point to the southwest. These features are also reported by Chang et al. (1996), and observed in the ERICA IOP 5 storm by Reed et al. (1993a) and Blier and Wakimoto (1995). During the mature stage, the warm front and particularly the bent-back front are very intense, with a temperature gradient on the order
Figure 2.13: Sea level pressure (hPa) analysis at (a) 06 Z, (b) 12 Z, (c) 18 Z of 4 January 1989, and (d) 00 Z of 5 January 1989. Wind vector flags are 25 m s$^{-1}$; full barbs are 5 m s$^{-1}$; half-barbs are 2.5 m s$^{-1}$ (from Neiman and Shapiro 1993).
Figure 2.14: 850 hPa temperature (°C) analysis at (a) 06 Z, (b) 12 Z, (c) 18 Z of 4 January 1989, and (d) 00 Z of January 1989. Wind flags and barbs are the same as in Fig. 2.13 (from Neiman and Shapiro 1989).
of 0.2 K km\(^{-1}\) (Reed et al. 1993a; Wakimoto et al. 1995), and are associated with strong ascent and precipitation. The warm core seclusion is associated with a pool of large vorticity (Reed et al. 1994). On a finer scale, the wind, pressure, temperature and precipitation fields near the cyclone center also show considerable complexity, with multiple circulation and pressure centers (Fig. 2.15, Neiman et al. 1993). The estimated sensible and latent heat fluxes in the MBL show strong spatial variability and are extremely large (approaching a combined 3000 W m\(^{-2}\)) along the high wind speed, ring-like region encircling the low center (Fig. 2.14d, Neiman and Shapiro 1993). They suggest that the evolution of the ERICA IOP 4 storm matches the conceptual marine cyclogenesis model (Fig. 2.16) proposed by Shapiro and Keyser (1990).

The latest numerical models can simulate the observed mesoscale structure near the center of marine explosive cyclones with considerable degree of realism. Reed et al. (1993a) used the PSU/NCAR MM4 model with full physics and a nested grid (30 km and 90 km) to reproduce the extreme thermal gradients associated with the fronts and rapid deepening of the ERICA IOP 5 storm, in good agreement with observations. Chang et al. (1996) performed a simulation of the ERICA IOP 4 storm using the Naval Research Laboratory (NRL) limited-area model (about 30 km resolution with 16 sigma-layers). Features such as the comma-shaped precipitation pattern, the bent-back warm front, secondary cold front, and warm core seclusion are reproduced, with model surface energy fluxes magnitudes over the Gulf Stream that are comparable to those estimated by Neiman and Shapiro (1993). Their sensitivity tests indicate that both surface energy fluxes and latent heating contribute to rapid deepening, but dynamical processes are still the primary
Figure 2.15: Fine scale analysis of (a) 350-m above ground level (AGL) streamlines, (b) 350-m AGL streamlines and NOAA WP-3D PPI Doppler radar reflectivity (dBZ), (c) 350-m AGL equivalent potential temperature (K), and (d) sea level pressure (hPa) at 06 Z of 4 January 1989. The analysis domain is a small area (about 3° longitude by 1.5° latitude) near the low pressure center and triple point in Fig. 2.12a (from Neiman et al. 1993).
Figure 2.16: An alternative model of frontal-cyclone evolution: incipient broad-baroclinic phase (I), frontal fracture (II), bent-back front and frontal T-bone (III), and warm-core frontal seclusion (IV). Upper: sea level pressure (solid), fronts (bold), and cloud signature (shaded). Lower: temperature (solid), and cold and warm air currents (solid and dashed arrows, respectively) (from Shapiro and Keyser 1990).
mechanism for development, as in other ERICA cyclones (Reed et al. 1993a; Blier and Wakimoto 1995; Kuo et al. 1996).

Kuo et al. (1996) used the non-hydrostatic version of the PSU/NCAR model (MM5, Dudhia 1993) to simulate the ERICA IOP 5 storm and to assess the model’s sensitivity to various subgrid scale cumulus parameterization and explicit moisture schemes at 20 and 60 km horizontal resolutions. The simulated mesoscale lows at low levels are characterized by large diabatically produced PV maxima and by nearly coincident rainfall maxima when some schemes are used, but such characteristics are not seen with other cumulus parameterization schemes. The authors suggest that distribution and intensity of precipitation, the atmospheric thermodynamic structure, and the evolution of mesoscale lows in the model are very sensitive to the choice of cumulus parameterization scheme.

2.7 Polar Lows in the North Atlantic

Over high latitude regions of the North Atlantic (and North Pacific), polar lows are another type of cyclone that threatens the safety of shipping and other local activities during winters (e.g., Wilhelmsen 1986). Compared with synoptic scale cyclones, polar lows are shallower (below 500 hPa) and smaller (with diameter of a few hundred to more than 1000 km), and develop within the cold air mass on the poleward side of the jet stream, often along the Arctic front or sea-ice boundaries (Sardie and Warner 1983; Rasmussen et al. 1992). In the North Atlantic, regions prone to polar low development
are Greenland, the Norwegian and Barents Seas, Denmark Strait and areas to the south and east of Iceland (Sardie and Warner 1985), as well as the Labrador Sea and Davis Strait to the west of Greenland (Boyer 1993).

According to Businger and Reed’s (1989) classification, at least three types of polar lows exist in the North Atlantic. The short-wave/jet-streak type polar lows are characterized by comma-shaped cloud patterns and are related to PVA forcing associated with a secondary vorticity maximum aloft (Duncan 1977; Reed 1979). They are usually larger but less frequent in the Atlantic (Carleton 1985). The Arctic-front type polar lows are associated with shallow, but strong, baroclinicity created through differential diabatic heating from the underlying surface in the PBL, and are triggered by an upper level PVA or IPV anomaly (Forbes and Lottes 1985; Rasmussen et al. 1992). The cold-low type systems develop farther back within the inner core of deep cut-off lows and are characterized by weak baroclinicity, strong surface fluxes and deep convection (Rasmussen 1979, 1985; Rasmussen and Lystad 1987; Nordeng and Rasmussen 1992). They are triggered by an upper level IPV anomaly, and sometimes display a spiral-shaped cloud pattern reminiscent of a hurricane at the mature stage (Rasmussen et al. 1993). In all three types, destabilization from below during arctic outbreaks is another common favorable condition (Bratseth 1985; Shapiro et al. 1987), and the small sizes of polar lows are related to a larger Coriolis parameter, low static stability (Rasmussen and Lystad 1987) and the effect of latent heat release on baroclinic wave development (Mullen, 1979).

Different mechanisms have been investigated for polar low development. Baroclinic instability is the major process for short-wave/jet-stream type polar lows (e.g.,
For Arctic-front and cold-low type systems, the atmospheric baroclinicity is either shallow or weak, and the baroclinic instability is still a necessary mechanism but tends to only trigger development (Reed and Duncan 1987; Nordeng and Rasmussen 1992). Conditional instability of the second kind (CISK) or air-sea interaction instability (ASII), both related to development of tropical cyclones, is subsequently responsible for more rapid development (Økland 1977, 1987; Emanuel and Rotunno 1989). Numerical modeling studies suggest that latent heating from deep convection is necessary to account for the observed traveling speed and growth rate of polar lows (Sardie and Warner 1983, 1985; Aakjær 1992). The pressure perturbation produced by this heating is comparable to the magnitudes observed in the atmosphere, on the order of 10 hPa (Rasmussen et al. 1992). The rapid deepening of polar lows occurs coincident with the deep convection outbreaks (Businger 1985; Montgomery and Farrel 1991), while the often observed inversion cap in polar environments may play an important role in accumulating convective available potential energy (CAPE) before the convection and in regulating the horizontal distribution of latent heating during deep convection outbreaks (Rasmussen et al. 1992, 1993).

The ASII mechanism, first proposed by Emanuel (1986a) and Rotunno and Emanuel (1987) to replace CISK (Charney and Eliassen 1964; Ooyama 1964) as the primary mechanism for tropical cyclone formation, is now widely accepted (e.g., Holton 1992). In ASII theory, the positive feedback between vortex scale and cumulus scale still exists as in CISK, but the loop is established through the anomalously large sensible heat...
and moisture fluxes above the tropical ocean and the low level convergence that transports this energy into the storm center. An initial weak depression is required to provide local convergence and stronger surface wind to raise the energy fluxes and establish the feedback. This is consistent with the observation that tropical cyclones evolve from pre-existing disturbances such as easterly waves and do not form spontaneously. In ASII theory tropical cyclones therefore form due to a gradual increase of energy input from the ocean which is the result of the growing vortex and local convergence, rather than due to the release of CAPE stored in the atmosphere as suggested in CISK. For the cold-low type polar lows, Emanuel and Rotunno (1989) also suggest that the ASII mechanism plays a role in producing further deepening once they are triggered by the upper level IPV anomaly and is capable of sustaining a much larger pressure drop than is observed in a typical polar environment. Table 2.2 shows the conceptual model suggested by Emanuel (1986b) of the relative importance of baroclinic instability versus ASII in the development of various types of cyclones. Beside baroclinic instability, CISK and ASII, barotropic instability is one other possible mechanism that cannot be ruled out, but it is highly unlikely to be the sole mechanism for polar low development, and represents only a minor contribution (Mullen 1979; Businger and Reed 1989).

2.8 Relationship between Current and Previous Research

Previous climatological studies suggest that explosive cyclogenesis occurs most frequently in mid-latitudes over the northwestern Pacific and northwestern Atlantic, along
<table>
<thead>
<tr>
<th>Baroclinic instability</th>
<th>Air-sea interaction instability</th>
</tr>
</thead>
<tbody>
<tr>
<td>Ordinary cyclones</td>
<td>Bombs</td>
</tr>
<tr>
<td></td>
<td>Polar lows (comma clouds)</td>
</tr>
<tr>
<td></td>
<td>Polar lows (Bear Island type)</td>
</tr>
<tr>
<td></td>
<td>Tropical cyclones</td>
</tr>
</tbody>
</table>

Table 2.2: Proposed scale of relative importance of baroclinic instability versus air-sea interaction instability (ASII) in the development of different types of lows by Emanuel (1986b). Baroclinic instability dominates the development of ordinary extratropical cyclones, while ASII dominates that of tropical cyclones. The “comma-cloud type” polar lows correspond to the short-wave/jet-streak type lows in the text, while the “Bear Island type” polar lows are those “cold-low type” lows that occur near the Bear Island region between Svalbard and Scandinavia and exhibit spiral-shaped cloud signature.
major baroclinic zones near land-sea boundaries, while few events occur in the Greenland-Iceland region (Section 2.1). However, several more recent studies indicate that cold season cyclogenetic activities are also common near Iceland, and the two extreme examples in Section 2.1 demonstrate that explosive cyclogenesis does occur in this region and can be very violent. The frequency distribution derived from ECMWF operational analysis in this study (Chapter 3) exhibits considerably higher spatial variability in the North Atlantic, with distinct local maxima immediately off the North American coast near Nova Scotia, to the south of Greenland near 50°N, and to the south of Denmark Strait. Thus, explosive cyclone frequencies in the North-central and extreme Northeast Atlantic were likely underestimated during the past. This difference may arise due to the lack of key observations in the past and the use of ECMWF analysis instead of conventional weather maps, and may also be linked to the low frequency interdecadal atmospheric circulation variability in the North Atlantic, as the cold season mean Icelandic Low was stronger from the early 1980s to mid-1990s and was accompanied by a general increase in regional cyclone activity (Serreze et al. 1997).

Since the Greenland-Iceland region is traditionally viewed as a region with little cyclogenetic activity, previous studies addressing explosive cyclogenesis there were few and did not appear in the literature until recently (Section 2.5). All these investigations focused on individual events, and a comprehensive study investigating explosive cyclogenesis in different sectors has yet to be done. The purpose of this dissertation is, therefore, to compare and contrast explosive cyclones in different sectors of the North Atlantic, and to investigate whether there are discernible differences in the general
characteristics of their development and evolution. Such differences are expected since the subpolar environment in high latitudes differs from that of mid-latitudes in many ways.

Research during the past two decades has shown the importance of both stronger upper level dynamical forcing, and the effect of moisture in producing explosive cyclogenesis through the lowering of the static stability (Sections 2.3 and 2.5). Explosive cyclones are like ordinary ones, fundamentally driven by baroclinic instability, but latent heat release and boundary layer energy fluxes also contribute significantly to the rapid deepening process. Thus, the abundant moisture supply in the MBL provides the energy for the additional deepening rate and makes explosive cyclogenesis a maritime phenomenon, since strong upper level forcing exists both over land and oceans. However, previous studies did not adequately address details of how moist processes and latent heating are important during the explosive deepening stage. Through a vorticity budget analysis, this study also attempts to shed some light on this question, and to investigate how various dynamical and physical processes might contribute differently in explosive cyclones in different sectors of the North Atlantic.

The IPV viewpoint developed by Hoskins et al. (1985) reviewed in Section 2.4 is used in the discussion of isentropic and cross section analysis results (Sections 4.7 and 4.8). Due to the improvement in model resolution, initial conditions, and representation of subgrid scale convection and PBL processes, today’s operational models have much higher skills in predicting explosive cyclogenesis (Section 2.2), and high resolution mesoscale models are capable of simulating the fine structure near the center and the fronts of explosive cyclones more realistically than previously possible (Section 2.6).
CHAPTER 3

DATA AND METHODOLOGY

Data used and the methodology employed in this study are described in this chapter. The methodology includes the approaches used to identify potential cases in the North Atlantic basin and to further select those to be included in the composite, the criteria used to classify selected cases into different categories, and the methods to produce composites and to compute variables.

3.1 ECMWF Data

The data set used is the European Centre for Medium-Range Weather Forecasts (ECMWF) Tropical Oceans Global Atmosphere (TOGA) Global Surface and Upper Air Analysis (Level III-b), archived and managed by the National Center for Atmospheric Research (NCAR), for January 1985 to March 1996. Operationally produced, uninitialized data are provided twice daily (00 and 12 Z) on a regular 2.5° × 2.5° latitude-longitude grid (144 × 73 for the globe). Upper level variables include geopotential, temperature, $u$ and $v$ wind components, and vertical velocity at 14 standard pressure levels.
(1000, 850, 700, 500, 400, 300, 250, 200, 150, 100, 70, 50, 30, and 10 hPa). Relative humidity is also available at levels below 300 hPa. Since January 1992, the 925 hPa level has been added to the data set. Surface variables include surface pressure and temperature, mean sea level pressure (MSLP), temperature and dew point temperature at 2 m height, and $u$ and $v$ winds at 10 m height (2 m height since February 1994). The temperature at 2 m height is the only variable not used in this study.

The ECMWF operational model is a primitive-equations (PE) based, global spectral model with a triangular truncation of T213 (roughly 0.55° latitude/longitude resolution) and 31 vertical levels from surface to 10 hPa (L31) as described by Simmons (1991). It was T106/1.125° and L19 before September 1991, and T63/1.875° and L16 before May 1985. Various other improvements were also made throughout the period (Trenberth and Olson 1988; Bengtsson 1991; Simmons et al. 1995). The model employs a Four-Dimensional Data Assimilation (FDDA) scheme to analyze the data with the multivariate three-dimensional optimum interpolation (3D OI) described in detail by Lorenc (1981) and an injection period of six hours ($\pm 3$ h within the time of initialization), and is initialized four times a day (00, 06, 12, and 18 Z) with the nonlinear normal mode initialization (Wergen 1989).

The ECMWF assimilation system uses the 6-h forecast from the previous model run as the first guess (FG) field, then incorporates all available observations (including those from land-based surface stations, ships, radiosonde network, drifting buoys, aircraft, and satellites) to produce the analysis before the final initialization process. The 3D OI scheme processes many nearby data points and variables simultaneously by using large
analysis volumes, and applies statistical methods to perform simultaneous data selection and quality control. All data are subject to checks for internal consistency and climatological reasonableness, for checks against the FG field, and checks against neighboring reports (Lorenc 1981; Shaw et al. 1987). The assimilation system has been shown to be useful for monitoring the performance of the observational network by 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), Manobianco (1989a), and Wash et al. (1992).

3.2 Potential Case Identification

To select cases included in the study, possible explosive cyclogenesis events in the data set must be first identified. For practical reasons, this needs to be done through an objective, automatic approach, rather than a subjective, manual one. Automatic detection and tracking algorithms have been developed by Serreze (1995) for low pressure systems using gridded NMC analyses, and Sinclair (1994; 1995) for vorticity centers using ECMWF data. Here two programs, the first one for storm detection and the second one for system tracking, are developed to obtain the history of individual low pressure systems in the Northern Hemisphere during the data period.

Sanders and Gyakum's (1980) criterion for “bombs” is used to identify possible explosive cyclogenesis cases. Their geographic distribution in the North Atlantic and
intensity of the development form the basis of the classification (discussed in Section 3.3). Finally, other rules are applied to filter out less desired cases and obtain the final list of cases used for the composites (described in Section 3.4).

3.2.1 Storm Detection Program

The storm detection program uses the ECMWF MSLP data array and identifies synoptic scale low pressure systems in the Northern Hemisphere at each time. The storm tracking program follows these centers through time (described in Section 3.2.2). In the detection program, a low pressure center is defined at a grid point when it meets all three following conditions: (1) the MSLP is lower than, or equal to any other point within a 750 km radius (great circle distance), (2) the MSLP is at least 1 hPa lower than the averaged MSLP at the surrounding 8 points, and (3) the MSLP has a positive Laplacian. After scanning through the array and picking out potential low pressure centers between 15°N and the North Pole, the program then determines the boundary and calculates the size and associated average MSLP Laplacian of each system. The boundary of a low pressure system is defined as the outermost closed isobar from the center using 0.5 hPa increments, while the size is the areal sum of all 2.5° × 2.5° boxes (which vary from about 3,400 km² at 87.5°N to 77,300 km² at the equator) enclosed by the boundary. The MSLP Laplacian is computed in local Cartesian coordinates, and an increasing number of grid points in the zonal direction (ranging from 1 point at 30° N to 23 points at 87.5° N) are skipped to avoid problems due to the convergence of meridians toward the pole. The program then
discards all low pressure systems with an area smaller than 40,000 km$^2$, as well as those smaller than 120,000 km$^2$ if the MSLP deficit between the center and the boundary is less than 2 hPa. Finally, the program re-calculates the boundary, size, and average MSLP Laplacian for each retained low pressure system and the results are written into the output file. Such a parameter setup in the storm detection program generally excludes small and weak systems and is helpful in making the storm tracking easier and more accurate.

Output variables of the storm detection program for low pressure systems include the date, time, central MSLP, latitude and longitude, column and row numbers in the MSLP array, zonally averaged MSLP at the latitude of the center, MSLP along the boundary, system size, central MSLP Laplacian, averaged system MSLP Laplacian, and a system identification number for tracking purposes (here the rank of lows in terms of MSLP is used). The program also detects high pressure centers with a different parameter setup, but this is beyond the focus of this study. A two-dimensional character map (144 × 33, 10° N to the North Pole) showing the position and coverage of identified systems is produced for each time step. All these above-mentioned variables (except for the zonally averaged MSLP) together with the character map, are used in the storm tracking program to help match systems in consecutive time steps.

3.2.2 Storm Tracking Program

The storm tracking program uses the output from the storm detection program, and tracks individual low pressure systems through time by matching those at a particular
time (designated $t_1$) with those 12 hours later (designated $t_2$). At the beginning of each run, every low pressure is assigned a "system number" starting from 1, and a low at $t_2$ is assigned the same number as the one at $t_1$ once they are "paired" (i.e., considered the same system by the program). After matching all possible pairs, the program then determines whether each un-paired low at $t_1$ is a cyclolysis event or a merger with another system at $t_2$, as well as whether each un-paired low at $t_2$ is a cyclogenesis event or a system that splits from another system at $t_1$. The next available system number is given for each un-paired low at $t_2$ that is identified as a newly formed system. Because explosive cyclogenesis is rare during the summer, the tracking program is run once for every 12-month period from July 1 to June 30 of the following year such that the tracking is continuous throughout each cold season. The only two exceptions are the runs from January to June of 1985 and from July of 1995 to March of 1996.

The details of the matching procedure are rather complicated because the program requires increasing detail to match remaining lows once initial steps in the procedure have matched the more "obvious" pairs. Nevertheless, the procedure can be broken down into seven steps: (1) to match large, isolated and strong lows, (2) to match large, isolated but moderate lows, (3) to match deep but small lows (which are usually multiple centers inside a broad low pressure region), (4) to match weaker but slow-moving lows, (5) to match the remaining lows, (6) to determine if un-paired lows at $t_1$ dissipate or merge with other systems, and (7) to determine if un-paired lows at $t_2$ are new form systems or split from existing lows. In step 5 the program employs a "score approach" and matches the pair that yields the highest total score. In general, a potential pair gives a higher total score if
the low travels a shorter distance in a more desirable direction (toward the east-northeast),
and poses less change in its characteristics such as the central MSLP, system size, and
Laplacian, etc. There is also a threshold of the score below which two lows under
consideration cannot be paired. In addition, no system can move toward the south for
more than 1000 km over a 12-hour period. This restriction is necessary to avoid incorrect
matching of a system at \( t_1 \) with another \( t_2 \) system which moves rapidly toward the north.
The output variables of the storm tracking program are the same as those of the detection
program, except that the system number and special codes indicating the status of the low
(such as "cyclogenesis", "dissipation", "merge", or "split") are added and the two-
dimensional character map is removed.

### 3.2.3 Possible Explosive Cyclogenesis Cases

With output files of the storm tracking program, the low pressure systems are re-
sorted by their system number, instead of by time, and the history of all cyclones is
obtained. Explosive deepening events are identified using Sanders and Gyakum's (1980)
criterion: a decrease in central sea level pressure at an averaged rate of at least 1 B during
a 24-h period (Eq.(2.1)). All events south of 30°N are discarded to exclude rapidly
deepening warm season tropical storms. The position of an explosive cyclone is taken to
be that at the mid-point of the 24-h period. A total of 1369 explosive deepening events
are identified between January 1985 and March 1996, 800 in the North Pacific and 569 in
the North Atlantic. These events comprise 846 individual cyclones, 484 North Pacific and
362 North Atlantic systems, since many of them have more than one 24-h period of explosive development. These explosive cyclogenesis cases in the North Atlantic form the population from which cases included in the composite study will later be drawn (described in Section 3.4).

Mesoscale polar lows (Section 2.7) tend to be under-represented in ECMWF fields (Forbes and Lottes 1985; Rasmussen et al. 1992) and their usual pressure perturbation on the order of 12 hPa (e.g., Sardie and Warner 1985; Nordeng and Rasmussen 1992) is considerably smaller than the 24-h bomb criterion. It is still possible however for our sample to include some polar lows that deepen explosively and eventually become a dominant synoptic-scale feature, since no special treatment was employed to ensure their exclusion. This type of cyclone originating in polar air streams is similar to those studied by Sinclair and Elsberry (1986).

3.2.4 Evaluation of Detection and Tracking Programs

The performance of the storm detection and tracking programs is evaluated by comparing their results to those obtained by manual tracking using data for two 3-month periods from December 1988 to February 1989 and December 1991 to February 1992. These periods are near the middle portion of the full data period, and correspond to more active seasons in the North Atlantic. Among a total of 6120 identified low centers for the Northern Hemisphere, 100 errors were found and thus the program tracks low systems correctly about 98.4 % of the time. Considering the fact that on there are on average as
many as 16.9 low pressure systems between 15° N and the North Pole at each time step, such an accuracy is quite good and the program performance would be satisfactory for studies of general climatology of all cyclones.

During each cold season, however, on average only about 120 explosive deepening events are identified, and incorrect tracking could have more serious effects. Table 3.1 qualitatively summarizes the performance of the tracking program, by comparing program results with manual tracking for the same six months. The North Atlantic basin is further divided into three sectors (Fig. 3.3, to be discussed in Section 3.3). Table 3.1 shows that the tracking program captured virtually all explosive deepening events in the North Pacific during that period, but missed 19% of those over the North Atlantic. The error (missing) rate is higher in the Northwest (17%) and North-central Atlantic (30%). A more detailed examination reveals that this deficiency arises mostly because explosive cyclones in the North Atlantic are often fast-moving and exhibit rapid changes in system characteristics over a short time. Such systems tend to produce a lower score in the matching procedure, sometimes lower than the required threshold value. Despite this discrepancy, the storm detection and tracking programs are useful for identifying possible explosive cyclogenesis cases, and provide a large number of potential candidates for the composite study.

3.3 Classification and Composite Categories

The spatial distribution of raw counts of explosive deepening events (as obtained with the approach described in Section 3.2), at full 2.5° resolution, exhibits considerable
<table>
<thead>
<tr>
<th></th>
<th>Northern Hemisphere</th>
<th>North Pacific</th>
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<tr>
<td></td>
<td>Total</td>
<td>Total</td>
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<td>NWA</td>
</tr>
<tr>
<td>True total</td>
<td>161</td>
<td>82</td>
<td>79</td>
<td>30</td>
</tr>
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<td>Identified</td>
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<td>1</td>
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<tr>
<td>Missing rate (%)</td>
<td>9.3</td>
<td>0.0</td>
<td>19.0</td>
<td>16.7</td>
</tr>
</tbody>
</table>

Table 3.1: Qualitative summary of the performance of storm tracking program as compared to manual tracking. Data during two 3-month periods (December 1988 to February 1989 and December 1991 to February 1992) are used for this evaluation. Numbers in the table are number of 24-h explosive deepening events, except for the missing rate where they are in percentage. Missing rate is the ratio between events that are not identified and the true total. Events in the North Atlantic are divided into three sectors (see Fig. 3.3), Northwest Atlantic (NWA), North-central Atlantic (NCA) and extreme Northeast Atlantic (NEA), in columns 4 to 6.
variability and further smoothing is needed. To do this, the counts are first aggregated into $5^\circ \times 5^\circ$ quadrilaterals by adding them together, then a smoothing method of 0.5 times the frequency at the central point plus 0.125 times the sum of frequencies at four adjacent points is applied. This produces sufficient smoothing while retaining considerable detail, and the resulting distributions of 24-h explosive cyclogenesis events ($n = 1369$) and their maximum deepening positions ($n = 846$) are shown in Figs. 3.1 and 3.2, respectively. In Fig. 3.2, every cyclone is counted only once when it reaches its maximum deepening rate, unless it has separate explosive deepening periods at least 30 hours apart.

The distribution of 24-hour explosive cyclogenesis in the North Atlantic (Fig. 3.1) shows three distinct frequency maxima along the primary storm track (Whittaker and Horn 1984). These maxima include a broad maximum off the coast of North America near Nova Scotia, a second center to the south of Greenland (near $45^\circ W, 50^\circ N$), and the third one slightly to the south of Denmark Strait, with frequencies of 18, 18, and nearly 12, respectively. These centers are also depicted clearly by the distribution of maximum deepening positions in Fig. 3.2, in which the maximum off the coast is displaced slightly to the southwest near $67^\circ W, 40^\circ N$. The patterns in Figs. 3.1 and 3.2 can be compared with those in Fig. 3 of Sanders and Gyakum (1980) and Fig. 7c of Roebber (1984), who both used the same smoothing method and whose figures are shown in Figs. 2.3 and 2.4, respectively. In the North Atlantic, Figs. 3.1 and 3.2 show a higher degree of variability in the spatial distribution and more frequent explosive cyclogenesis events toward the Northeast Atlantic with ECMWF MSLP analyses than are found in the earlier studies using conventional maps. Although deficiencies exist in the storm tracking program and
Figure 3.1: Geographic distribution of total 24-h explosive cyclogenesis events in the Northern Hemisphere derived from twice-daily ECMWF MSLP analysis for the period January 1985 to March 1996. Numbers of events are counted and smoothed in $5^\circ \times 5^\circ$ latitude-longitude quadrilaterals (see text).
Figure 3.2: Geographic distribution of maximum 24-h deepening positions of explosive cyclones in the Northern Hemisphere derived from twice-daily ECMWF MSLP analysis for the period January 1985 to March 1996. The smoothing method is the same as in Fig. 3.1.
the representativeness of such a partially model-based climatology remains to be
determined, these maxima are quite real in the ECMWF data set since the program tends
to miss, rather than falsely generate, explosive deepening events from incorrect tracking
(Table 3.1). In addition, this result is in broad agreement with the findings of Serreze et
al. (1997) indicating that since the early 1980s there has been an increase in synoptic scale
cyclone activities in the climatological Icelandic low region, corresponding to the recent
highly positive phase of the North Atlantic Oscillation (e.g., Hurrell 1995).

In the North Pacific, explosive cyclogenesis occurs most frequently near the
western margin of the basin, with the primary maximum centered near 150°E, 40°N to the
east of Japan (Figs. 3.1 and 3.2) and other minor local maxima extending eastward along
the primary storm track (Whittaker and Horn 1984). The frequency distributions for this
basin are generally in better agreement between Figs. 3.1 and 2.3, and between Figs. 3.2
and 2.4 in both the location and magnitude (after adjusting for data lengths and temporal
resolution between studies), especially for the maximum deepening positions (Figs. 3.1
and 3.2). Figure 3.2 is also in very good agreement with Fig. 12 of Gyakum et al. (1989).

The explosive cyclogenesis events in the North Atlantic shown in Fig. 3.1 are
classified into three different sectors according to Fig. 3.3, with each of the frequency
maxima falling into one sector. The Northwest Atlantic (NWA) includes the maximum
immediately off the coast of North America, the North-central Atlantic (NCA) includes
the one to the south of Greenland, and the extreme Northeast Atlantic (NEA) includes the
maximum near Denmark Strait. The abbreviations NWA, NCA, and NEA will be used
frequently in subsequent text and figures to denote each geographic group.
Figure 3.3: Three sectors in the North Atlantic used to classify explosive cyclogenesis events according to their geographic locations. The outlines of the sectors, Northwest Atlantic (NWA, Region I), North-central Atlantic (NCA, Region II), and extreme Northeast Atlantic (NEA, Region III), are drawn with thick solid lines. Names of various geographic regions are also plotted in the map.
Whereas all the explosive cyclone events in Fig. 3.2 deepened explosively for at least one 24-h period, their 12-h deepening rate (in terms of Bergeron) are calculated since this corresponds to the highest temporal resolution of the data set. These cases are also classified into three different intensity classes, weak (1.00 to 1.39 B), moderate (1.40 to 1.79 B), and strong (≥ 1.80 B), based on their maximum 12-h deepening rate. These criteria are chosen because they give roughly comparable numbers of cases in each intensity class. For the intensity classes, the abbreviations ST, MO, and WE will be used. Therefore, all the potential North Atlantic cases are classified into nine categories, three sectors times three intensity classes, and the number of cases in each category ranges from 29 to 63 with a total of 360, since two cases did not occur in any of the sectors.

3.4 Final Case Selection and Storm Track Correction

For practical reasons, other rules are employed to further select the most ideal cases for the composites, with a goal of having 15 to 20 cases in each category. These rules include (1) an ideal case should have only one dominant deepening period, whose maximum 12-h deepening rate should be at least 0.5 B greater than the maximum rate during its other deepening period, if there is one, (2) an ideal case should deepen primarily in its corresponding sector, at least 50% of the time during its explosive deepening period (or periods), and (3) an ideal case occurs between September and April. When all three rules were applied, about 25% of the cases are eliminated, and the number of cases in each category is now between 18 and 47. Finally, when there were more than enough
cases in a particular category, those farther away from the frequency maxima in Fig. 3.1 and those close to the boundaries between sectors (Fig. 3.3) were dropped.

During the case selection process, the MSLP fields over the life span of potential cases are plotted at 1 hPa intervals. These maps are used to manually correct any errors made by the storm detection and tracking programs and to extend storm history backward to an early development stage when the surface low is weak and not recognized by the detection program, sometimes showing up as a trough on the map without a closed center. Since 12-h deepening rates are used instead of 24-h rates, some cases identified earlier as being in one category were moved to another category, most often for an adjacent sector or intensity class. Using the procedure and rules described above, a final total of 162 explosive cyclogenesis cases are selected for the composite analysis (55 in NWA, 54 in NCA, and 53 in NEA). For each case, data during a 7-day period from \( t_{-72} \) to \( t_{-4} \) for the Northern Hemisphere are extracted from the ECMWF global data set, with \( t_0 \) defined as the end point of the maximum 12-h deepening period. These data files will be used as the source files for the composites. A list of these cases in each category is provided in Table 3.2, and their spatial distribution at \( t_0 \) is shown in Fig. 3.4. The total number of cases in each category ST NWA, MO NWA, WE NWA, ST NCA, MO NCA, WE NCA, ST NEA, MO NEA, and WE NEA, is 18, 19, 18, 20, 19, 15, 19, 18, and 16, respectively.

Because the composites were made using a moving coordinate system (Quasi-Lagrangian approach) with the sea level low pressure center being the common reference point (the origin), a cautious treatment of storm tracks is needed. Therefore, PV values along a chosen moist potential temperature \( (\theta_m) \) surface are computed and plotted for
Table 3.2: List of cases included in the composite analysis for each of the nine categories. Variables (from left to right) provided are time, longitude (°E), latitude (°N), and central MSLP (hPa) at \( t_0 \). For the time variable, its first, second, third, and last two digits indicate year, month, date, and time (Z), respectively. Total numbers of cases in each category and sector are also given in parentheses.
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Table 3.2.
Figure 3.4: Spatial distribution of explosive cyclogenesis cases included in the composite analysis at $t_0$. Numbers represent the frequency of cases at each grid point with 2.5° latitude/longitude resolution, and thick gray lines depict the boundaries of three North Atlantic sectors.
each of the cases selected, together with the MSLP maps, to make final corrections of the
sea level storm track between \( t_{36} \) and \( t_{72} \). The \( \theta_m \) level used is above the storm center
near, or slightly above the 500 hPa level at \( t_0 \), usually between 300 and 309 K. This
procedure made relatively few minor changes in low pressure positions in the early stage
(mostly before \( t_{36} \)) when the PV value near the storm should be approximately conserved.
After all these checks and corrections, the finalized storm tracks represent the best tracks
based on the author’s knowledge.

3.5 Compositing Techniques

Using the cases listed in Table 3.2, composites are made for each category of
explosive cyclones and further diagnostic and comparative study is performed to
investigate characteristics of systems in different sectors and with different intensities. In
short, the composites are made for each category using a moving latitude-longitude
coordinate system with its origin continuously placed at the defined sea level low pressure
center of each of the cases in that category throughout the 7-day period. More detailed
information on this compositing technique and its computational aspects are given below.

3.5.1 Composite Time and Domain

The composites are made for each variable for each category separately by using
all cases in that category with each having the same weight (i.e., arithmetic average). For
each case throughout the period from \(t_{-54}\) to \(t_{-72}\) (relative to \(t_0\) as given in Table 3.2),
values of the variable under consideration are extracted at all levels from 1000 to 10 hPa inside a finite domain constantly moving with the sea level center for each case. The domain is defined in latitude-longitude space with a size of 75° longitude \(\times\) 50° latitude (31 \(\times\) 21 grid points), and the grid point (19, 11) is always placed at the sea level low pressure center. The first coordinate index indicates the number of grid points in the zonal direction from west to east (positive x-direction), and the second index indicates the number of grid points in the meridional direction from north to south (negative y-direction). Figure 3.5 gives an example of the domain when a hypothetical low center moves from 60°W, 45°N, to 45°W, 50°N, then to 35°W, 55°N. The coordinate indexes at the low center and four corners are also indicated for the first domain.

The size of the domain is sufficient to include the entire region of interest for each case, but the 2.5° resolution is inadequate to resolve the mesoscale structure near the storm center, such as those described in Section 2.6. Once data in the finite domain are extracted for all cases in a category, the composites from \(t_{-54}\) to \(t_{-72}\) are made at each level simply by averaging the cases. Therefore, values for grid points having the same latitude-longitude position relative to the sea level low pressure center (but likely different actual great circle distance from the storm center) are averaged to make the composite. The composite method described here differs from the one used by Manobianco (1989a) in which composites are made relative to the 500 hPa vorticity maximum (at 24-h intervals) rather than the surface storm center. Toward the end of the 7-day period, a few cases (mostly those in the NEA sector) moved to the north of 65°N, and data are not available
Figure 3.5: An example of the moving finite domain with a size of 75° longitude by 50° latitude. The hypothetical low pressure system center moves from 60°W, 45°N, to 45°W, 50°N, then to 35°W, 55°N (dots and thick black lines), and the corresponding domains are plotted as thick gray lines. The coordinate indexes at the low center and four corners are also indicated for the first domain.
for the northernmost rows of the domain array. Missing data values are assigned for such portion, and are excluded in the composite calculation. Similarly, missing data in some other cases to the south of 25°N at cyclone inception are treated the same way. At this point, the composites at 925 hPa are based on only those cases after January of 1992.

3.5.2 Variables and Calculation

Variables that are not provided directly by the ECMWF data set are calculated subsequently. For some of them, the order of the calculations and creation of the composite does not influence the final result. One example is the thickness (computed from geopotential height), since the thickness of composite height is the same as the composite (average) of thickness of all cases. However, this is not the case for most variables, especially those involving partial derivatives. For those variables, values at all Northern Hemisphere grid points are always computed first for individual cases, then the finite domain data are extracted and composites are produced. This approach also ensures that values of variables involving derivatives are better approximated by the finite-difference technique along boundaries of the domain, and is used for virtually all variables unless it is otherwise specified in the text (when we are interested in certain variables computed from the composite field, instead of the composite of those variables). Again, in the x-direction, an increasing number of grid points need to be skipped when taking derivatives to avoid problems toward the North Pole. Here, for latitudes between 62.5 and 87.5°N, only enough grid points are skipped to give a distance at least that at 60°N.
(about 139 km), and this same method is used for all calculations for consistency.

There are also variables that involve derivatives in the vertical, such as the IPV. The vertical gradient of a variable at each standard pressure level is approximated by the finite difference technique using interpolated values at two levels, one 5 hPa below and the other 5 hPa above the level under consideration. The two levels used are each bounded by 1000 and 10 hPa, such that layers of 10 hPa thick are used to estimate the vertical gradient at all intermediate data levels (from 925 to 30 hPa) but those of 5 hPa thick are used for the lowest and highest levels (1000 and 10 hPa). Using such thin layers of equal thickness undoubtedly gives more accurate vertical gradient estimates, since data levels are unevenly spaced in the vertical and the spacing is rather large in the middle troposphere. The kinematic vertical velocity is another variable that requires calculation using thin layers, although it does not involve differentiation in the vertical. For this purpose the layer depth never exceeds 5 hPa, while the kinematic vertical velocity is adjusted such that the integrated total divergence of the air column is zero between sea level and 10 hPa.

For the cross sections analysis presented in Section 4.7, the composite variables are interpolated onto levels at 50 hPa intervals to produce evenly spaced data arrays for display. The values at 950 and 900 hPa levels, of course, are interpolated from those at 1000 and 850 hPa. However, during this procedure it seems disadvantageous to not make use of the composite value at 925 hPa from cases after 1992 to yield better estimates, since such cases constitute a fair portion, about 20 to 50 %, of total cases (Table 3.2). This is done by first obtaining a “best estimate” at 925 hPa that combines the composite value with the interpolated value following the method given in the Appendix, then by
incorporating this 925 hPa "best estimate" into the interpolation. The same method is also used for variables that are computed from the composites while a higher accuracy in the approximations is desired in the lower troposphere, such as the stability factor shown in Fig. 4.54.
CHAPTER 4

EVOLUTION AND STRUCTURE OF COMPOSITE CYCLONES

By using the composite technique described in the previous chapter, the dynamical and thermal structure, as well as the evolution of the mean cyclone from $t_{-24}$ to $t_{-72}$ for each category can be obtained. In this chapter, mean cyclones for different categories are discussed, compared and contrasted in terms of their characteristics and behavior. The results discussed here include the mean storm track and evolution at the surface, low level response and baroclinicity, jet stream and upper level forcing, IPV and cross-section analysis. Finally, further discussion and conclusion is provided based on the composite analysis in section 4.8. Due to limited space, discussion will focus primarily on strong and weak composite explosive cyclone cases for the three sectors and less on moderate cases.

4.1 Mean Storm Track and Evolution

The mean storm tracks of the explosive cyclone cases in each of the nine categories are shown in Fig. 4.1, with large open circles representing the cyclone position at $t_0$ in the NWA, NCA, and NEA cases. The mean tracks for NWA cases (solid lines)
Figure 4.1: Mean storm tracks of the explosive cyclone cases for each of the nine
categories. In each panel mean tracks of NWA, NCA, and NEA cases are drawn in thick
solid, long dashed, and short dashed lines, respectively, for (a) strong (ST), (b) moderate
(MO), and (c) weak (WE) categories. Mean cyclone center positions are drawn
interchangeably with solid (at $t_{44}$, $t_{60}$, and $t_{60}$) and open (at $t_{72}$, $t_{45}$, $t_{0}$, and $t_{72}$)
circles, with those at $t_{0}$ enlarged. Also shown are boundaries of the three sectors.
start near the Rockies at $t_{-48}$ and extend eastward with $t_0$ (ending time of maximum 12-h deepening) immediately to the south of Nova Scotia, then northeastward across Newfoundland toward the southern tip of Greenland. Mean NCA cases (long dash lines) take a path farther eastward and reach their maximum deepening period to the south of Greenland between 45 and 52°N, then move northeastward toward Iceland. NEA events (short dash lines) have mean tracks still farther east extending toward Scandinavia near $t_{-72}$, with mean storm position at $t_0$ to the south or southwest of Iceland near 59°N, well off the eastern coast of North America. The mean cyclone positions at $t_0$ correspond very well with the frequency maxima in Fig. 3.1, except for MO and WE categories in the NEA sector (Fig. 4.1c) where the positions at $t_0$ are slightly farther east because more cases closer to Iceland were selected (Fig. 3.4).

The corresponding evolutions in MSLP and deepening rate of the composite cyclones are depicted in Figs. 4.2 and 4.3, respectively. For all nine categories, the MSLP decreases more or less exponentially between $t_{-48}$ and $t_0$, after which the mean cyclone keeps deepening for an additional period of 12 to 36 hours before MSLP rises again (Fig. 4.2). The NEA storms have the largest pressure drop between $t_{-12}$ and $t_0$, and usually the lowest MSLP at $t_{-12}$ (950 hPa for ST, 964.1 hPa for MO, and 969.1 hPa for WE categories) among all sectors, except for the MO intensity class where the NCA cyclone reaches 961.3 hPa at $t_{-24}$. In contrast, the decreases in pressure for NWA cyclones are much less dramatic, and the minimum MSLP values reached are higher and occur later in the stage (967.3 hPa at $t_{-36}$, 971.0 hPa at $t_{-36}$, and 979.1 hPa at $t_{-24}$ for ST, MO, and WE classes, respectively). The difference in averaged MSLP drops among the sectors is
Figure 4.2: Composite mean sea level pressure (MSLP, in hPa) from $t_{-72}$ to $t_{+72}$ for (a) ST, (b) MO, and (c) WE categories. In each panel the MSLP curves for NWA, NCA, and NEA cases are drawn in thick solid, long dashed, and short dashed lines, respectively.
Figure 4.3: Composite deepening rate (in Bergeron) from $t_1$ to $t_2$ for (a) ST, (b) MO, and (c) WE categories. In each panel the curves for NWA, NCA, and NEA cases are drawn in thick solid, long dashed, and short dashed lines, respectively.
mainly due to the latitudinal adjustment of the Sanders and Gyakum's criterion, and is particularly significant for ST cases between $t_{-24}$ and $t_0$. Mean storms in the NCA tend to have MSLP curves between those of the other two sectors, but qualitatively are similar to the NEA curves (Figs. 4.2).

Positive deepening rates (Fig. 4.3) occur roughly from $t_{-48}$ to near $t_{-24}$ for NEA and NCA cyclones, and in general extend longer to about $t_{-36}$ for NWA storms. Due to the adjustment, the greater MSLP decreases in NEA and NCA cases produce very comparable maximum deepening rates for MO and WE intensity classes (between 1.56 and 1.60 B, and 1.24 and 1.29 B, respectively). For the ST categories (Fig. 4.3a), the maximum deepening rates suggest that the NEA storms (2.42 B) are indeed on average slightly more violent than the NCA ones (2.26 B), which in turn are slightly stronger than the NWA cyclones (2.17 B). After the explosive development, the deepening rates for ST NEA and ST NCA categories drop rapidly to negative values by $t_{-24}$ (Fig. 4.3a), suggesting a faster average filling rate than ST NWA storms (which is also evident in Fig. 4.2a).

4.2 Surface Mean Sea Level Pressure and Clouds

Figures 4.4 to 4.9 present the composite MSLP field (hPa) from $t_{-36}$ to $t_{-24}$ (a to f) for strong and weak classes in the three sectors. Areas with mean 1000-500 hPa relative humidity (RH) greater than 75 % are shaded to depict the cloudy regions. The latitude to longitude ratio at 45°N is used for all such figures to reduce the degree of distortion associated with the projection from the condition had a ratio of 1 been used. For the
Figure 4.4: Composite mean sea level pressure (MSLP, in hPa) from (a) t-36 to (f) t-24 at 12-h intervals for ST NWA category. Isobars are drawn every 4 hPa. Areas with mean 1000-500 hPa relative humidity (RH) greater than 75% are gray shaded to depict general regions of clouds. Labels along the left and bottom edges are latitude and longitude relative to the defined low pressure center in degrees, and the latitude to longitude ratio used corresponds to that at 45°N.
Figure 4.5: Same as Fig. 4.4 but for WE NWA category.
Figure 4.6: Same as Fig. 4.4 but for ST NCA category.
Figure 4.7: Same as Fig. 4.4 but for WE NCA category.
Figure 4.8: Same as Fig. 4.4 but for ST NEA category.
Figure 4.9: Same as Fig. 4.4 but for WE NEA category.
NWA sector, the ST composite low (Fig. 4.4) appears to be weaker at sea level than the WE one (Fig. 4.5) before $t_{-24}$, but deepens much faster between $t_{-12}$ and $t_0$ (a pressure fall of 18.7 versus 11.3 hPa for WE). After $t_0$ both composite cyclones continue to develop at a slower rate (Fig. 4.2), while the pressure gradient around the center of the ST cyclone remains considerably tighter with more closed isobars than the WE one. Both systems are accompanied by increasing clouds near the storm center through the composites, and display a distinct “comma” shape starting at $t_0$ (Figs. 4.4 and 4.5).

For the NCA sector, the ST composite (Fig. 4.6) exhibits lower central MSLP values than the WE storm (Fig. 4.7) by $t_{-12}$. The MSLP falls from $t_{-12}$ to $t_0$ by 23.4 hPa in ST and 12.9 hPa in WE, and the ST case remains about 15 hPa deeper than the WE one after $t_0$. The two cyclones stop deepening by $t_{-24}$, and maintain their strength. The “comma” cloud associated with the WE case (Fig. 4.7) is somewhat more distinct at both $t_0$ and $t_{-12}$, but the ST storm has a larger cloud area in general (Fig. 4.6).

For the NEA sector (Figs. 4.8 and 4.9), the distortion produced by the projection toward their later stages as storms move farther north is more severe (Fig. 4.1). The ST case has a stronger MSLP gradient and is stronger than the WE case throughout the period. An additional feature in the NEA composite MSLP fields is the “parent cyclone” which appears to the northeast of the developing low in early stage then merges with it sometime between $t_{-12}$ and $t_0$. For the ST composite, this “parent cyclone” (prior to $t_{-12}$) is also stronger and the merger produces the largest $t_{-12}$ to $t_0$ pressure drop (27.4 hPa) among all categories (Fig. 4.8), compared to 14.3 hPa in the WE case (Fig. 4.9). The difference between the intensity of ST and WE storms after $t_0$ is also dramatic, similar to
the NCA sector, and both systems reach their minimum central MSLP value at $t_{12}$. The cloud areas again display a "comma" shape by $t_0$, and is also in general larger for the ST category (Figs. 4.8 and 4.9).

When the evolution at sea level is compared among composite cyclones of the ST intensity class, several noticeable differences exist. The incipient ST NEA cyclone before $t_0$ is stronger with a tighter MSLP gradient surrounding its center especially to the southeast, while the mature storm after the rapid deepening tends to start weakening earlier than in the NWA or NCA (Figs. 4.4, 4.6, and 4.8). The NWA composite storm only weakens after $t_{24}$. Differences also exist in the strength of the "parent cyclone," its proximity to the developing storm, and the interaction between the two. In the NEA sector, the "parent cyclone" is stronger and closer to the incipient low to its northeast. The two lows rotate cyclonically around each other, and merge right before $t_0$ (Fig. 4.8). In the NCA sector, the "parent cyclone", if there is one, is weaker and much farther northeast, outside the plotting domain even at $t_{12}$, and there is no direct merger between the two (Fig. 4.6). In the NWA sector, the incipient low does not appear to develop from a pressure trough, and there is little evidence indicating the existence of a "parent cyclone" (Fig. 4.4). The cloud area tends to be greater toward the NEA sector, but after $t_0$ a greater portion of the difference may be introduced artificially by the more severe distortion in the projection. Similar differences among the three sectors are also quite apparent for the weak (Figs. 4.5, 4.7, and 4.9) and moderate classes (not shown).
4.3 Low Level Response and Baroclinicity

The 850 hPa level composite geopotential heights (m) and temperatures (K) corresponding to Figs. 4.4 to 4.9 are shown in Figs. 4.10 to 4.15. Also plotted in these figures are positions of maximum absolute vorticity ($\eta$) computed from the horizontal winds. For the ST NWA category, the 850 hPa disturbance associated with the cyclone at sea level in Fig. 4.4 appears in the height field (solid lines) as a short wave trough at $t_{.36}$ and $t_{.24}$ with a well defined vorticity maximum (Fig. 4.10a, b). Significant development at this level starts by $t_{.12}$, and the 850 hPa low center becomes enclosed by three height contours at $t_0$ (1180 m with a 12-h height fall of 160 m), continuing to deepen thereafter to a central height of 1034 m at $t_{.24}$ (Figs. 4.10c to f). During this rapid development period, the strongest height gradient is immediately southeast of the low center. In the temperature field (dashed lines) the disturbance has a rather small amplitude at $t_{.36}$ and $t_{.24}$, but lags the height wave by about 1/4 wavelength, an ideal condition for baroclinic development (Charney 1947), with the largest thermal gradient near (or immediately to the north of) the surface low center. At $t_{.12}$ the horizontal temperature advection increases to further distort the temperature wave (while promoting frontogenesis) and noticeably shortens the wavelength of the disturbance (Fig. 4.10c). After $t_0$, the distortion becomes increasingly severe, with the thermal ridge extending across the cyclone center from southeast to northwest and the thermal trough to the southwest (Figs. 4.10d to f). By $t_{.24}$ the wavelength of temperature disturbance is shortened to the order of 600 km, but the thermal gradient near the low has weakened somewhat.
Figure 4.10: Composite geopotential height (solid, m) and temperature (dashed, K) at 850 hPa level from (a) \( t_{-36} \) to (f) \( t_{-24} \) at 12-h intervals for ST NWA category. Contours are drawn every 40 m for height and every 3 K for temperature. The position of the maximum absolute vorticity center at 850 hPa is also plotted as an “X.” The plotting domain and meaning of labels are the same as in Fig. 4.4.
Figure 4.11: Same as Fig. 4.10 but for WE NWA category.
Figure 4.12: Same as Fig. 4.10 but for ST NCA category.
Figure 4.13: Same as Fig. 4.10 but for WE NCA category.
Figure 4.14: Same as Fig. 4.10 but for ST NEA category.
Figure 4.15: Same as Fig. 4.10 but for WE NEA category.
Comparing the evolution at 850 hPa for category WE NWA (Fig. 4.11) with that of Fig. 4.10, one notices that the short wave disturbance in ST NWA does not appear to be stronger before \( t_{-12} \). In fact, both WE NWA and MO NWA composites have greater local maximum \( \eta \) values than the strong case (Table 4.1). During this period, however, the ST NWA system is associated with stronger thermal gradient, which is helpful in further development in the classical QG model. At \( t_0 \) the difference in system strength between categories becomes apparent, with the ST event exhibiting a stronger thermal gradient near the low, greater distortion in temperature field, and better developed thermal (especially cold air) advection (Figs. 4.10d and 4.11d). These differences remain quite evident afterwards at \( t_{-12} \) and \( t_{-24} \).

Figures 4.12 and 4.13 show the 850 hPa condition for ST and WE composites in the NCA sector. In the height fields, the short wave (until \( t_{-24} \)) and closed low (since \( t_{-12} \)) disturbances of ST are associated throughout development with a considerably tighter height and thermal gradient than the WE case, but again are not decidedly more intense prior to \( t_{-12} \) (also Table 4.1). The ST case displays an impressive deepening between \( t_{-12} \) and \( t_0 \) with a height fall of 203 m, versus 113 m of the WE composite (Figs. 4.12 and 4.13, c and d). The thermal gradient near the center of the ST system during the incipient and most rapid deepening stages before \( t_{+12} \) is significantly stronger than the WE one. The explosive development in ST right before \( t_0 \) produces more distortion in the temperature field and leads to the seclusion of a pocket of warm air (272 K) coincident with the low center at \( t_{+12} \) (Fig. 4.12e). Similar to evolution at the sea level, the NCA storms cease further deepening by \( t_{-24} \) with the ST case exhibiting more weakening after \( t_{-12} \).
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<td>Weak</td>
<td>17.0</td>
</tr>
</tbody>
</table>

Table 4.1: Local absolute vorticity ($\eta$) maximum values (in $10^{-5}$ s$^{-1}$) associated with developing cyclones for each of the nine categories from $t_{-36}$ to $t_{+24}$ at 850 hPa (upper half) and at 500 hPa (lower half).
In the NEA sector, the differences between the ST and WE composites (Figs. 4.14 and 4.15) are similar to those for the NCA sector. The ST NEA disturbance is not easily discernibly stronger than WE before $t_{-12}$ (also Table 4.1), but is again associated with enhanced thermal and height gradients. At $t_{-12}$ the "parent cyclone" also appears at 850 hPa for both ST and WE, but its merger with the incipient low before $t_0$ is not as apparent as at sea level, especially for the WE composites. The height fall of 226 m from $t_{-12}$ to $t_0$ for ST is the largest among all categories, and is a reflection of the rapid spin-up process in higher latitudes (Figs. 4.14c, d). Both ST and WE NEA storms reach their minimum 850 hPa height (of 892 and 1057 m) at $t_{-12}$, and start to fill by $t_{-24}$. In the thermal field, the ST storm is associated with consistently stronger distortion than the WE case during the rapid deepening process, and the warm air seclusion occurs at both $t_{-12}$ and $t_{-24}$ for the ST storm (Figs. 4.14e, f), as well as at $t_{-12}$ for the WE one (Fig. 4.15e).

The sector development of the three strong 850 hPa composites can be compared (Figs. 4.10, 4.12 and 4.14). It is clear that during the incipient stage, the height gradient and local $\eta$ maxima associated with the developing short wave at 850 hPa are stronger in NEA than NCA, whose gradient and $\eta$ maxima are in turn stronger than NWA (also Table 4.1). In the thermal fields, the NEA temperature gradient is also stronger near the low center before $t_{-12}$ and the distortion during the rapid deepening stage is greater, while after $t_0$ the temperature gradient diminishes much faster despite a deeper storm at both the 850 hPa and sea level (Figs. 4.4, 4.6, and 4.8). The greater distortion in the NEA temperature field (and to a lesser degree the NCA field) is associated with the extension of the thermal ridge toward the low center and the warm core seclusion, a process unique in explosive
oceanic cyclogenesis and one that is caused by the relatively warm air being wrapped around the low center by colder air (Section 2.6).

During the incipient stage the 850 hPa disturbance in ST NEA also shows a closer association and interaction with the "parent cyclone" than those observed in the other two sectors. It is quite apparent that the incipient short wave in the NEA sector is embedded in the circulation of the parent cyclone to its northeast, while the two lows merge between \( t_{1.2} \) and \( t_0 \) (Fig. 4.14). The association between the NCA short wave and the parent cyclone (outside the domain) appears to be very limited with no mergers (Fig. 4.12), while there is little indication of the existence of a parent cyclone in NWA (Fig. 4.10). When the three weak cyclones at 850 hPa are compared (Figs. 4.11, 4.13 and 4.15), qualitatively similar differences to those between strong cases are found, except that the thermal gradient in the WE NEA case starts to weaken earlier than those in other sectors, just as it does in the ST case in that sector. Interestingly enough the MO systems (not shown) are the strongest among all intensity classes in their corresponding sectors during the incipient stages at both \( t_{1.2} \) and \( t_{1.2} \) (Table 4.1), but the thermal gradient is considerably weaker than the ST cases, only comparable to the WE cases. The differences between the three MO composites are otherwise similar to those occurring among the ST and WE cases.

4.4 Upper Level Forcing at 500 hPa

The composite results for the 500 hPa level are shown in Figs. 4.16 to 4.21 and include geopotential height, absolute vorticity (\( \eta \)) advection, and positions of local \( \eta \)
Figure 4.16: Composite geopotential height (thick, m) and absolute vorticity advection (thin, $10^{-9}$ s$^{-2}$) at 500 hPa level from (a) $t_{-36}$ to (f) $t_{+24}$ at 12-h intervals for ST NWA category. Contours are drawn every 60 m for height, and every $10^{-9}$ s$^{-2}$ for vorticity advection with areas greater than $2 \times 10^{-9}$ s$^{-2}$ shaded. Positions of maximum absolute vorticity centers are also plotted as "X." The plotting domain and meaning of labels are the same as in Fig. 4.4.
Figure 4.17: Same as Fig. 4.16 but for WE NWA category.
Figure 4.18: Same as Fig. 4.16 but for ST NCA category.
Figure 4.19: Same as Fig. 4.16 but for WE NCA category.
Figure 4.20: Same as Fig. 4.16 but for ST NEA category.
Figure 4.21: Same as Fig. 4.16 but for WE NEA category.
maxima. For the ST NWA category, the incipient 500 hPa trough lies about 7° to 10° west of the surface low with a well-defined local vorticity maximum. The divergence associated with the positive vorticity advection (PVA) helps maintain the surface low (Fig. 4.16a, b). At $t_{12}$ (Fig. 4.16c), with concurrent development at lower levels, the 500 hPa wave amplifies more rapidly with an enhanced height gradient at the base of the trough, directly above the intensifying surface system. This enhances both the vorticity maximum (Table 4.1) and resultant PVA and initiates the self-development process, in which the lower and upper level disturbances reinforce each other through a mutual interaction (cf., Fig. 4.10, Petterssen 1956). For strong explosive cyclogenesis, this process is depicted quite nicely even in the composites and is rather impressive. During the most rapid 12-h deepening period ending at $t_6$ (Fig. 4.16d), the 500 hPa $\eta$ and PVA maxima grow from $16.9 \times 20.4 \times 10^{-5} \text{s}^{-1}$ (Table 4.1) and from $2.8 \times 4.0 \times 10^{-9} \text{s}^{-2}$, respectively. A strong PVA/NVA (negative vorticity advection) dipole structure also develops, with the PVA and NVA maxima lying immediately downstream and upstream from the vorticity center. In response to the increasingly distorted temperature field below 500 hPa (e.g., Fig. 4.10), the tilting of the trough turns from positive (NE-SW) to negative (NW-SE) by $t_6$, and becomes increasingly negative afterwards. Associated with this is further amplification of the wave (at all levels), eventually leading to formation of the cut-off low before $t_{24}$ (Fig. 4.16f). When the upper and lower level lows are in-phase, the self-limiting process is in effect and mutual interaction between PVA and thermal advection ceases, and the PVA/NVA dipole weakens although the height gradient southeast of the low remains strong. The evolution of MSLP (Fig. 4.2) is in very good agreement with that of PVA.
directly above the surface center, and further deepening at sea level continues after \( t_0 \) until about \( t_{34} \) when the PVA directly above the low becomes too small.

Figure 4.17 shows conditions at 500 hPa for the WE NWA category. Compared with the ST NWA case, the weak composite shows very similar evolution although each of the components discussed above is weaker. Throughout the composites the height gradient at the base of the developing trough (Fig. 4.17) is consistently weaker, as are the magnitudes of the corresponding \( \eta \) maximum (Table 4.1), the PVA, and the PVA/NVA dipole structure. The difference between the strengths of various components in ST and WE cases is most evident at \( t_0 \) and \( t_{12} \).

The results for ST and WE NCA categories are presented in Figs. 4.18 and 4.19. During the incipient stage up to \( t_{12} \), the height gradient near the ST surface center is considerably stronger than in WE, and is accompanied by greater vorticity maximum (Table 4.1) and much stronger PVA directly above the surface low (Figs 4.18 and 4.19, a to c). At the beginning of the most rapid 12-h deepening period (\( t_{13} \)), the composite ST NCA 500 hPa trough has already become negatively tilted and develops very strong PVA ahead of its vorticity center. The maximum PVA value further increases from 3.7 to \( 4.3 \times 10^{-9} \) s\(^{-2} \) at \( t_0 \) with an associated diffluence region clearly visible (Fig. 4.18d). The PVA magnitudes and the height gradient near the trough base, as well as the strength of the PVA/NVA dipole in the WE NCA storm (Fig. 4.19d), are all considerably weaker than ST at \( t_0 \) although their general evolutions are quite similar. The cut-off lows in both cases form by \( t_{12} \), but it is likely that the ST composite forms earlier within that 12-h period (Figs. 4.18e and 4.19e). At the end of development, the ST cyclone reaches a lower
central height value (4968 m) at \( t_{-24} \) than the WE case (5106 m), corresponding to a lower MSLP value at the surface (Figs. 4.18f and 4.19f).

The differences between ST and WE evolutions at 500 hPa for the NEA sector (Figs. 4.20 and 4.21) are similar to those for the NCA sector in the early stage at \( t_{-24} \) and \( t_{-12} \). With a stronger height gradient, and thus geostrophic wind across the region of development, the PVA downstream from the ST trough is considerably stronger than for WE. At \( t_{-12} \), the ST disturbance develops a large area of significant PVA with a maximum exceeding \( 4.0 \times 10^{-9} \text{ s}^{-2} \) close to the deepening surface center (Fig. 4.20c). Associated with PVA development is the approaching of a closed 500 hPa upper level low (5021 m) from the north and further height gradient tightening for the NEA sector. This phenomenon is likely to play an important role in development of more extreme oceanic cyclogenesis near Iceland, and will be discussed further in section 4.8. Similar evolution can be seen between \( t_{-24} \) and \( t_{-12} \) in the WE NEA case, but is much less pronounced with the resultant PVA maximum reaching only \( 2.8 \times 10^{-9} \text{ s}^{-2} \) (Figs. 4.21b, c). At \( t_0 \), the larger shaded ST PVA values are located in regions of significant diffluence and along the narrow area immediately downwind from the vorticity ridge extending southward from the \( \eta \) maximum (Fig. 4.20d). The dipole structure at this time is very strong with a maximum PVA value of \( 4.9 \times 10^{-9} \text{ s}^{-2} \), the largest among all categories, and the closed low starts to form with strong concurrent development at all levels. It is clear that this 500 hPa low initially develops ahead of the approaching low, then deepens enough to replace the latter at \( t_{-12} \) as the cut-off low of the ST NEA category. A similar evolution cannot be seen in the WE class, for which the initially larger 500 hPa low to the north is far less apparent.
After the most rapid deepening stage and formation of the cut-off low (likely shortly after $t_0$), PVA associated with the ST cyclone weakens dramatically and moves away from the surface center at $t_{-12}$ and $t_{-24}$ (Fig. 4.20e, f). For the WE case, the 500 hPa cut-off low forms much later at almost $t_{-12}$ and significant PVA weakening does not start until then (Fig. 41e, f).

Comparing the three ST events (Figs. 4.16, 4.18, and 4.20), one notices that the 500 hPa height gradient across the region of development prior to $t_{-12}$ is stronger toward the northeast sector, consistent with the condition at 850 hPa, with much larger PVA values directly above the surface center in NCA and NEA sectors. During the most rapid deepening stage from $t_{-12}$ to $t_0$, the PVA and the dipole structure of the ST NEA storm are the strongest among the three sectors, while those of the ST NWA case are the weakest. The magnitudes of local $\eta$ maxima in these cases also agree well with the PVA (Table 4.1), but not all the difference in $\eta$ contributes to the PVA difference since a difference of nearly $3 \times 10^{-5}$ s$^{-1}$ in Coriolis parameter ($f$) occurs between storms in the NEA and NWA sectors at $t_0$ (Fig. 4.1). The increase in $\eta$ from $t_{-12}$ to $t_0$ for ST events is largest in the NEA sector (Table 4.1), and is mainly related to the rapid amplification of 500 hPa waves toward this region although the $\beta$ value ($\beta = \frac{df}{dy}$) also increases as the cyclones move farther north. The faster storm evolution toward higher latitudes is exemplified by the timing of the appearance of the 500 hPa cut-off low, which forms after $t_{-12}$ in NWA, a few hours after $t_0$ in NCA, and only shortly after $t_0$ in NEA. Closely related to the cut-off low formation is the PVA weakening and the movement of the PVA center away from the surface storm center, which is most visible in the NEA case at $t_{-12}$ and $t_{-24}$. Reflecting the
faster deterioration of thermal gradient in the lower troposphere, the condition of 500 hPa height gradient in the storm environment reverses after \( t_{-12} \), with a stronger gradient in NWA and weaker in NEA (Figs. 4.16, 4.18, and 4.20). The approaching upper level low during the NEA incipient stage (Fig. 4.20) is consistent with evolution at 850 hPa and sea level, suggesting a westward tilting with height and deep vertical extent of the “parent cyclone.” In the ST NCA case an approaching trough is noticeable, while such a feature is absent in the NWA sector (Figs. 4.16 and 4.18).

Qualitatively, the overall differences between the three WE cases at 500 hPa (Figs. 4.17, 4.19, and 4.21) are very similar to those between ST events. A few variations, however, are discernible. The initial PVA directly above the sea level low center at \( t_{-24} \) does not appear stronger toward the northeastern sector. For the WE NEA case, the approaching upper level low is considerably weaker, and the largest increase in maximum \( \eta \) value occurs between \( t_0 \) and \( t_{-12} \), instead of \( t_{-12} \) and \( t_0 \) (Table 4.1). This is associated with a rapid but comparatively late-stage amplification of the 500 hPa wave, and subsequently a later weakening of PVA after \( t_{-12} \) (Fig. 4.21).

4.5 Jet Stream Level Forcing

The 250 hPa level is chosen to show evolution of the jet stream and the associated forcing in the upper troposphere for ST and WE composites (Figs. 4.22 to 4.26). Variables plotted are winds and divergence produced by ageostrophic motion. For the ST NWA category, Fig. 4.22 shows that an upper level jet streak is present throughout the
Figure 4.22: Composite streamlines (thick, with arrows), wind speed (m s⁻¹), and divergence (thin, 10⁻¹ s⁻¹) at 250 hPa level from (a) t₋₃₆ to (f) t₋₁₂ at 12-h intervals for ST NWA category. Wind speeds are expressed in gray shades starting from 25 m s⁻¹, with darker colors indicating stronger winds. Isotachs are also drawn at 5 m s⁻¹ intervals between the shades as thick white lines. Contour intervals for divergence are 1 × 10⁻⁵ s⁻¹, and dashed lines indicate negative values (convergence). The plotting domain and meaning of labels are the same as in Fig. 4.4.
Figure 4.23: Same as Fig. 4.22 but for WE NWA category.
Figure 4.24: Same as Fig. 4.22 but for ST NCA category.
Figure 4.25: Same as Fig. 4.22 but for WE NCA category.
Figure 4.26: Same as Fig. 4.22 but for ST NEA category.
Figure 4.27: Same as Fig. 4.22 but for WE NEA category.
period in proximity to the surface system, and intensifies and becomes narrower between $t_{-36}$ and $t_0$. Due to the northward component of storm movement, smaller before and larger after $t_0$ (Fig. 4.1), the jet streak also appears to migrate southward with time in the figures. The 250 hPa trough, inferred from the streamline pattern, approaches the surface disturbance from the west and evolves similarly to the 500 hPa trough (Fig. 4.16) except that its position is slightly farther to the west and no closed circulation forms up to $t_{-24}$. Throughout the development, a region of divergence (thin lines) associated with streamline divergence exists ahead of the trough, in good agreement with the 500 hPa PVA but with a relatively small magnitude prior to $t_{-24}$ (Fig. 4.22b). Starting from $t_{-24}$, in response to amplification of the approaching trough and enhancement in thermal gradient (i.e., frontogenesis) at lower levels, the jet streak strengthens and moves to the trough base to the west-southwest of the surface cyclone, placing the latter under its left-front quadrant. Such a combination dramatically raises the PVA downwind from the trough and leads to an increase in divergence from $1.4 \times 3.3 \times 10^{-5}$ s$^{-1}$ at $t_{-12}$ (Fig. 4.22c), and further to $4.1 \times 10^{-5}$ s$^{-1}$ at $t_0$ (Fig. 4.22d). The positions of maximum divergence are closest to the surface center at $t_{-12}$ and $t_0$ (especially $t_{-12}$), producing the largest 12-h MSLP drop for the ST NWA composite cyclone (Fig. 4.2a). The divergence region gradually moves away from the surface center to its northeast at $t_{-12}$ and $t_{-24}$ as the cyclone approaches its maximum intensity (Fig. 4.22e, f).

During rapid development, ageostrophic winds also produce an enhancement of wind speed immediately downstream and a reduction immediately upstream from the region of maximum divergence. Between $t_{-12}$ and $t_{-12}$, the upper level ageostrophic winds
at 250 hPa for the ST NWA case flow out from the divergent center in nearly all
directions at a speed of 5 to 10 m s\(^{-1}\) (not shown), large enough to alter the structure of
the jet streak and possibly limit maximum forcing from the jet streak in the upper
troposphere. Such a phenomenon can be clearly seen at both \(t_{-12}\) and \(t_{-12}\) in Figs. 4.22c
and 4.22e for the present category, as well as in other figures shown later for remaining
categories.

Comparing results for the WE NWA category (Fig. 4.23) with Fig. 4.22, one
notices several differences. The wind speed of the WE jet streak is considerably smaller
for most of the time during development, especially at \(t_{-24}\) and \(t_{-12}\), consistent with its
weaker thermal and height gradients at lower levels. Despite this, however, the WE NWA
case has greater maximum divergence from \(t_{-36}\) to \(t_0\) than the ST composite (Figs. 4.22
and 4.23, a to d). This might seem surprising, but supports the condition at low levels that
initial strength of the ST disturbance is slightly weaker. A closer inspection reveals that
before \(t_0\) divergence of the ST event begins to exceed that of the WE (Figs. 4.22d and
4.23d), since regions with significant divergence in the ST case are larger as suggested by
the more widely spaced contour lines surrounding the maximum. The evolution of the
streamline patterns (and inferred height field) is very similar for the two categories.

Figures 4.24 and 4.25 are 250 hPa composites for ST and WE cases in the NCA
sector. The evolution of the upper level trough is, like that for the NWA sector, very
similar between the ST and WE categories, but in the ST case a closed circulation forms
by \(t_{-24}\) (Fig. 4.24f). The jet streak is well defined early in the ST NCA composites, and its
strength increases significantly from 48 to 58 m s\(^{-1}\) at \(t_{-24}\) (Fig. 4.24b), with the surface
cyclone also located in the left-front quadrant of the jet core through $t_0$ (Fig. 4.24d). The jet streak is in contrast much weaker for WE cases and is displaced farther west at $t_{12}$ and $t_0$, although a secondary wind speed maximum center (48 m s$^{-1}$) appears south of the storm center at $t_0$ (Figs. 4.25c, d). The magnitude of the divergence ahead of the trough for ST is only comparable to WE before $t_{24}$, but is considerably greater at $t_{12}$ and $t_0$, corresponding to a much greater pressure drop at the surface (Fig. 4.2a, c).

For the NEA sector, the streamline analysis again confirms formation of a closed 250 hPa circulation by $t_{24}$ for ST events (Fig. 4.26) but not so for the WE category (Fig. 4.27). In the ST NEA case, the jet streak before $t_{12}$ has its center to the northeast of the incipient surface cyclone, with a maximum speed of 49 m s$^{-1}$. The wind to the south and southwest of the storm center starts to strengthen by $t_{12}$ (Fig. 4.26c), forming a new jet maximum at $t_0$ with its speed reaching 50 m s$^{-1}$ (Fig. 4.26d). As the cyclone migrates farther north after $t_0$, the separation between the jet streak and the surface low increases. For the WE NEA event, the wind speed center to the west-southwest of the surface low appears as early as $t_{26}$, while the maximum downwind from the divergence also exists at $t_{26}$ and $t_{24}$ (Fig. 4.27a, b). The divergence associated with the ST case is significantly stronger than that of the WE case from $t_{24}$ to $t_{12}$. This is especially true at $t_{12}$ and $t_0$, when the ST divergence reaches 3.8 and $3.7 \times 10^{-5}$ s$^{-1}$, compared to smaller WE values of $2.6$ and $2.7 \times 10^{-5}$ s$^{-1}$ (Figs. 4.26 and 4.27, c and d).

The jet streaks from Figs. 4.22 to 4.27 exhibit a more complex behavior than the composite jet streak in Manobianco (1989a) for east coast explosive cyclogenesis. While it is possible that the more complex features seen in this study arise partly from different
composite techniques, their existence is physically reasonable. In baroclinic development, the upper level jet not only responds to changing conditions in the thermal field at lower levels in order to maintain thermal wind balance, but its structure must also be modified by the ageostrophic motion from at least three sources: (1) the forcing of the jet streak itself, (2) the QG PVA forcing ahead of the trough (enhanced by stronger winds), and (3) forcing from inertia, stronger when the amplitude of upper level waves is large and/or the wind from the trough to the downstream ridge is enhanced together with an increase in Rossby number (Bjerknes and Holmboe 1944). The ageostrophic motion is particularly strong in the case of explosive cyclogenesis, and as mentioned earlier, it enhances the wind speed immediately downstream and reduces it immediately upstream from the divergence center. Therefore, the divergence tends to promote the appearance of jet maximum downstream near the ridge and to delay the advance of the jet streak upstream near the trough, and appears to be the main cause of some features seen in Figs. 4.22 through 4.27.

A comparison between the three 250 hPa ST cases indicates that the streamline pattern evolves from a positive to a negative tilt faster, with a slightly greater amplitude, in the NCA and NEA cases than for NWA, in which no closed circulation forms up to $t_{124}$ (Figs. 4.22, 4.24, and 4.26). Overall, jet streak strength is greatest in the NCA case before $t_9$, and in the NWA afterwards. In the incipient stage, the divergence directly above the NCA and NEA surface centers is significantly stronger than for NWA, and helps maintain the initial low level disturbance strength, consistent with 500 hPa conditions. The faster cyclone evolution in the NEA sector, which shows an early peak in divergence at $t_{12}$ (Fig. 4.26c), is in good agreement with the lower levels. This is exemplified by comparing the
magnitude of maximum divergence among the three strong cases from $t_{-12}$ to $t_{-12}$, as the
NEA sector has the largest divergence early at $t_{-12}$, the NCA sector at $t_0$, and the NWA
sector later at $t_{+12}$ (Figs. 4.22, 4.24, and 4.26, c to e). Although this relationship does not
hold true for the WE or MO (not shown) cases, they also show a faster evolution in the
extreme Northeast Atlantic. The correspondence between the 250 hPa divergence during
the incipient stage and the intensity class is better toward the NEA sector, suggesting that
the initial disturbance strength might more strongly affect development in this region.

For the NEA sector, especially the ST case, the jet streak more frequently first
appears to the northeast of the surface center at $t_{-24}$ and $t_{-12}$ (Figs. 4.26 and 4.27, a and b).
Modified by the ageostrophic motion, these jet maxima are likely located to the south of
the parent cyclone in a region of strong 500 hPa height gradient (Fig. 4.20a, b). The
configuration between the jet maximum and the parent cyclone resembles the one between
the jet streak and the new explosive cyclone at around $t_{-12}$ and $t_{-24}$ (Fig. 4.20e, f), since
both are in a similar, but later, stage of their respective developments.

4.6 Isentropic Potential Vorticity Analysis

The results of isentropic potential vorticity (IPV) analysis are presented for ST and
WE categories (Figs. 4.28 to 4.33) for the 315 K moist potential temperature ($\theta_m$) surface.
For these analyses, the PV values must be computed on isentropic instead of isobaric
surfaces (Eq. (2.3)). This is done at each grid point at each time for each individual case,
by first finding the pressure level of the chosen $\theta_m$ value then interpolating the $u$ and $v$
Figure 4.28: Composite potential vorticity (in PVU, or $10^{-5}$ m$^2$ K s$^{-1}$ kg$^{-1}$) and wind vectors along the 315 K moist potential temperature ($\theta_m$) surface from (a) $t_{-36}$ to (f) $t_{+24}$ at 12-h intervals for ST NWA category. For potential vorticity, contour lines are drawn every 0.5 PVU in white, with the only exception in black at 2 PVU. Darker shades indicate higher PV values between 0 and 2 PVU, as well as from 2 PVU up. The length of wind vectors is proportional to the speed, and the distance between two adjacent ticks along the bottom edge equals 25 m s$^{-1}$. The plotting domain and meaning of labels are the same as in Fig. 4.4.
Figure 4.29: Same as Fig. 4.28 but for WE NWA category.
Figure 4.30: Same as Fig. 4.28 but for ST NCA category.
Figure 4.31: Same as Fig. 4.28 but for WE NCA category.
Figure 4.32: Same as Fig. 4.28 but for ST NEA category.
Figure 4.33: Same as Fig. 4.28 but for WE NEA category.
wind components onto the $\theta_m$ surface to obtain the isentropic relative vorticity ($\zeta_0$). The local change rate of $\theta_m$ in the vertical ($\partial \theta_m / \partial p$) is computed using thin layers no more than 10 hPa thick, as described in Chapter 3. For the ST NWA composite (Figs. 4.28a to f), the evolution in the PV field on the 315 K $\theta_m$ surface is characterized by the approach and subsequent arrival from the north and northwest of a high PV tongue over the surface cyclone. The 2.0 PVU contour line is drawn in black and serves as a rough approximation for the boundary between air of stratospheric and tropospheric origin. During the incipient stage from $t_{.36}$ to $t_{.12}$ (Fig. 4.28a to c), the isentropic PV and wind fields show patterns reminiscent of those at 500 and 250 hPa levels (Figs. 4.22 and 4.23), except that the PV field configuration (i.e., gradient) is opposite to height field. The resemblance is supported by the elevation of the 315 K $\theta_m$ surface, sloped downward from the upper troposphere ($\sim 280$ hPa) near the northern edge to middle troposphere ($\sim 620$ hPa) near the southern edge of the domain (cf., Fig. 4.34). The high PV tongue extends farther south and approaches the surface storm center from the northwest and west at $t_{.12}$ and $t_6$, and is mainly caused by the strong northwesterlies (with maximum speed $\sim 35$ m s$^{-1}$) to the west (Figs. 4.28c, d). At this time near the surface cyclone center, positive PV advection (PPVA) can be clearly seen, but the air on the 315 K $\theta_m$ surface is likely of tropospheric origin ($< 2$ PVU). The leading edge of the PV tongue develops a cyclonic "hook-like" shape at $t_{.12}$ as it encounters the strong southwesterlies immediately ahead, and the PV value directly above the storm center has increased to $\sim 2.6$ PVU (Fig. 4.28e). The value grows further to $\sim 3.2$ PVU at $t_{.24}$, at which time a pocket of relatively low PV air (2.4 PVU) is enclosed just to the west of the surface center (Fig. 4.28f).
For the WE NWA case (Fig. 4.29), the evolution of PV and wind fields along the 315 K $\theta_m$ surface is similar to that for ST NWA, while several differences are noticeable. Throughout development, although the PV tongue in the WE event also approaches and eventually overtakes the surface cyclone as the ST case, the strength of its positive PV anomaly is consistently weaker and stratospheric air is also farther from the storm center than the ST case at the same stage (except $t_{-24}$). This is most likely related to the larger amplitude of the upper level trough before $t_{-12}$ (Figs. 4.16, 4.17, 4.22 and 4.23, a to c) and therefore the stronger ST northerly wind component behind the PV tongue (Figs. 4.28 and 4.29, a to c). The PV gradient near the center of the developing storm is considerably larger in ST, so is its apparent PPVA at $t_{-12}$ and $t_0$. Toward the end of the deepening at $t_{-24}$, the “hook-like” structure also develops in the WE category as outlined by the 3.0 PVU contour (Fig. 4.29f).

Results of the IPV analysis for ST and WE categories in the NCA sector both show the cyclonic trajectories of the PV tongue movement relative to the sea level storm center as the former approaches the latter (Figs. 4.30 and 4.31). For the ST case, a pool of high PV air exceeding 5.0 PVU enters the domain from the north by $t_{-24}$ (Fig. 4.30b) and moves southward at $t_{-12}$ (Fig. 4.30c) along with a gradual tightening in the PV gradient and amplification of the circulation on the $\theta_m$ surface, as indicated by the wind fields. The stratospheric air overtakes the intensifying surface cyclone from the southwest shortly after $t_{-12}$, as the PV value directly above the storm center increases from about 1.5 to over 3.0 PVU at $t_0$ (Fig. 4.30d) and then to almost 4.0 PVU at $t_{-12}$ (Fig. 4.30e). At $t_{-24}$ the leading portion of the PV tongue has detached from its main body to form a pool of
high PV air (4.8 PVU), spinning coincident with the surface cyclone with evident closed circulation and strong isentropic relative vorticity in the wind field (Fig. 4.30f).

Although the evolution of the WE PV field is quite similar to its strong counterpart, the PV value on 315 K $\theta_a$ surface near the system center is constantly smaller throughout development (Fig. 4.31). The difference in PV value between the two intensity classes is enlarged after $t_0$, at which time stratospheric air reaches the WE storm center from the southwest but has overtaken the ST center for at least several hours (Figs. 4.30d and 4.31d).

The ST NEA composite shows a tongue of very high PV values initially extending southward to the northwest of the surface system center by about $t_{-12}$, then approaching the center along a cyclonic path from its west or west-southwest at $t_0$ (Fig. 4.32). The surface cyclone center is overtaken by stratospheric air as early as $t_{-24}$ (Fig. 4.32b) and has increasing PV values of about 2.9, 4.1, and 4.7 PVU afterward at $t_{-12}$, $t_0$, and $t_{-12}$, respectively (Fig. 4.32c to e). As with the ST NCA category, in the NEA sector at $t_{-24}$ a large pool of high PV air (5.3 PVU) has formed directly above the surface system and detached from the main body of the PV tongue, which is more than 20° to the west-northwest (Fig. 4.32f). Compared to the ST event, the WE NEA case shows a weaker PV anomaly approaching the developing low prior to $t_0$, as well as a weaker northwesterly flow to the west of the tongue before $t_{-12}$ (Figs. 4.33a to d). As a result, the PV values above system center are significantly smaller at $t_{-12}$ and $t_0$ (about 2.4 and 3.5 PVU, respectively). After $t_0$, on the other hand, the WE high PV air detaches earlier and has somewhat higher values than the strong case, but this is after the most rapid surface
deepening. The distribution pattern of the 315 K $\theta_m$ surface elevation for the ST NEA category shows a range from about 270 hPa to the north to 560 hPa to the south at $t_{36}$, and a much smaller range from about 250 to 380 hPa after development at $t_{24}$ (not shown) and therefore a more gentle isentrope slope, in agreement with the much diminished thermal gradient within the domain.

When the ST IPV analysis results are compared among the three sectors, it is clear that the tongue of high PV air in the NEA is significantly stronger, as is the leading edge of the tongue that approaches and overtakes the developing surface storm. To illustrate the difference in the PV tongue strength, the timing of the stratospheric air of at least 4.5 PVU to enter the domain, its areal extent, and proximity to the surface system are compared. In the NWA case, air $\geq 4.5$ PVU first enters the domain at $t_{-12}$, moves southward but becomes confined to the northwest after $t_{0}$ (Fig. 4.28). For the NCA sector (Fig. 4.30), air with the same PV value enters the domain before $t_{36}$, at least a full day earlier than in the ST NWA case, and expands and extends further southward to the latitude of the surface storm center at $t_{0}$, when its leading edge is about 700 km away from the center. A small pocket of air with PV $\geq 4.5$ PVU appears near the storm center at $t_{-24}$. For the NEA composite (Fig. 4.32), air with at least 4.5 PVU is also present before $t_{36}$, but it expands more rapidly with an areal extent constantly larger than in the NCA case throughout the period. Its leading edge moves as close to the surface center as $\sim 300$ km to its southwest at $t_{0}$. At $t_{-12}$ and $t_{-24}$, this 4.5 PVU air has reached the surface storm center and remains connected with the main body of the high PV tongue, with an areal extent much larger than its counterpart in the NCA. The PV values directly above the
cyclone center can also be compared for the significant development period from $t_{-12}$ to $t_{-12}$. In the NEA sector, these values are 2.9, 4.1, and 4.7 PVU (Fig. 4.32), considerably greater than the NCA values of 1.5, 3.1, and 4.0 PVU (Fig. 4.30), and those of 0.8, 1.7, and 2.6 PVU in the NWA sector (Fig. 4.28).

Similar differences in the high PV tongue strength and of PV values above the cyclone center on the 315 K $\theta_e$ surface are also evident for the three weak (Figs. 4.29, 4.31, and 4.33) and moderate composites (not shown). Such differences are mainly due to differences in the geographical locations of rapid deepening, and their proximity to the high PV reservoir in polar regions. The sector differences in proximity to the high PV air can be clearly illustrated by comparing the extent of the stratospheric air ($\geq$ 2 PVU) among the figures. The NEA composite events start development closer to the stratospheric air, and become embedded deeper within the high PV pool toward later stages as well (Figs. 4.28 to 4.33). Therefore, the explosive cyclogenesis events toward the NEA sector tend to be accompanied by a stronger PV anomaly (and thus PV forcing) on isentropic surfaces in the middle to upper troposphere since they typically occur in higher latitudes.

There is also a difference in PV tongue orientation among events in different sectors, especially during the incipient stage. The orientation agrees well with the upper level trough tilting at 500 and 250 hPa, and tends to be more positively tilted in the NWA sector. The winds behind the PV tongue of the NWA sector tend to have a stronger northerly component, while the winds ahead of the tongue in NEA tend to have a stronger southerly component (Figs. 4.28 to 4.33).
Throughout development, the gradual amplification in both the PV anomaly and its associated circulation is evident for all categories on the 315 K $\theta_m$ surface. The composite winds in these figures are “snap shots” at an instant in time, rather than trajectories. From the composite 315 K moist isentrope slopes in these cases (c.f., cross sections along lines AB to be discussed below in Section 4.7), it can be established that the northwesterlies behind the PV tongues are associated with adiabatic sinking motion, and therefore vertical compression of air column and development of anticyclonic vorticity, while the southwesterlies ahead are associated with rising motion, and vertical stretching and development of cyclonic vorticity. This means that an air parcel traveling across the PV tongue would follow a trajectory that has a greater amplitude than the streamlines, and the argument provides the basis for the amplification of the PV anomaly with time. However, the adiabatic argument can only explain the growth in the wave amplitude of existing PV anomalies through advection (mostly before $t_0$), not the generation or destruction of PV maxima or minima (as those seen near the storm center after $t_0$, more evident toward the NEA sector), which must arise from diabatic processes.

4.7 Vertical Cross Section Analysis

Two types of vertical cross sections are constructed along the thick solid lines of Figs. 4.28 through 4.33. The first is aligned from point A to point B in a direction approximately perpendicular to the wind on the 315 K $\theta_m$ surface at the system center, and runs initially from the northwest to the southeast of the storm and later from nearly west
to the east. The second cross section type is at a right angle to the first in the latitude-
longitude space roughly from the southwest to the northeast (point C to D). Both types
of cross sections pass through the surface system center at the mid point (marked as “0”),
while the distance along the horizontal axis corresponds to the latitude of composite storm
center positions shown in Fig. 4.1. The cross section orientations vary slightly among
categories for different sectors in the North Atlantic, but are kept the same for different
intensity classes in the same sector.

Figures 4.34 to 4.37 are the cross sections for the ST and WE NWA categories.
Sections along line AB span $t_{34}$ to $t_{24}$, and those along line CD are from $t_{12}$ to $t_{12}$.
While the composite cross sections are unable to resolve various features (such as
tropopause fold, and upper level and surface frontogenesis) at their true intensity in
individual cases, they are useful for qualitative comparison to determine basic similarities
and differences among categories. For the AB section of ST NWA (Fig. 4.34), the upper
troposphere and lower stratosphere PV field evolution (left column) is dominated by the
approach and arrival at storm center of a strong PV anomaly, which is accompanied by a
lower tropopause. Since the leading edge of the PV anomaly follows a cyclonic path
relative to the surface system (Fig. 4.28), high PV air actually moves out of the cross
section plane before $t_0$ (Figs. 4.34a, c and e) then turns around and moves into the plane
(from the west-southwest in $xy$ space) to overtake the surface cyclone center near and
after $t_0$ (Figs. 4.34g, i and k). Associated with the PV anomaly aloft is the jet maximum to
its southeast (or east) near the region of steepest tropopause slope, approximated by the 2
PVU contour (thick solid line). Reflecting the northward movement of the cyclone, the jet
Figure 4.34: Vertical cross section (along AB lines in Fig. 4.28) of left: composite potential vorticity (thick white lines and gray shades, in PVU) and horizontal wind speed normal to the plane (thin lines, in m s\(^{-1}\)), and right: composite moist potential temperature \((\theta_m)\), thick gray lines, in K) and horizontal wind and kinematic vertical motion parallel to the plane as vectors from \(t_{136}\) (a and b) to \(t_{224}\) (k and l) at 12-h intervals for ST NWA category. For the potential vorticity, contour intervals are 0.5 below 8 PVU, 1 from 8 to 12 PVU, and 2 afterwards, with an additional contour level at 0.75 PVU. The 2 PVU line is drawn in black and darker shades indicate higher PV values between 0 and 2 PVU, as well as from 2 PVU up. Contour intervals are 5 m s\(^{-1}\) for wind speed with solid and dashed lines representing positive (into the plane) and negative (out from the plane) values, and are 5 K for \(\theta_m\). The length of wind vectors is proportional to the speed, and a horizontal distance of 500 km represents 50 m s\(^{-1}\) and a vertical distance of 100 hPa represents 1 Pa s\(^{-1}\). The region at the lowest portion, if below sea level, is masked with medium gray.
Figure 4.34

Pressure (hPa) (a)  Pressure (hPa) (b)  Pressure (hPa) (c)

Distance (x 1000 km)

Pressure (hPa) (d)  Pressure (hPa) (e)  Pressure (hPa) (f)

Distance (x 1000 km) (to be continued)
Figure 4.34

Pressure (hPa)

Distance (x 1000 km)

(continued)
Figure 4.35: Vertical cross section (along CD lines in Fig. 4.28) of left: composite isentropic potential vorticity (in PVU) and horizontal wind speed normal to the plane (in m s⁻¹), and right: composite equivalent potential temperature (θₑ in K) and horizontal wind and kinematic vertical motion parallel to the plane as vectors from t₁₂ (a and b) to t₋₁₂ (e and f) at 12-h intervals for ST NWA category. For the potential vorticity, wind speed normal to the plane, and wind vectors parallel to the plane, the contour intervals, shading, and vector length are the same as in Fig. 4.34. The contour intervals for θₑ are also 5 K but contours are drawn as thick dashed gray lines, with potentially unstable region shaded with light gray. The region at the lowest portion, if below sea level, is masked with medium gray.
Figure 4.36: Same as Fig. 4.34 but along AB lines in Fig. 4.29 for WE NWA category.
Figure 4.37: Same as Fig. 4.35 but along CD lines in Fig. 4.29 for WE NWA category.
maximum shows an overall movement to the right. The upper level wind speed gradient tightens significantly (c.f., Fig. 4.22), together with an enhancement of the horizontal PV gradient across the tropopause, especially at $t_{12}$ and $t_0$ (Fig. 4.34e, g). The evolution in lower troposphere is characterized by PV generation (as the rapid cyclogenesis takes place) and appearance of a separate near-surface PV maximum prior to $t_0$, before the arrival of upper level PV anomaly. The interaction between the two PV anomalies is evident after $t_{12}$, as the high positive wind speeds to the right of the upper level PV anomaly extend downward from the jet maximum while the negative wind speeds to the left of the low level anomaly extend upward (Figs. 4.34e, g, i and k).

The cross sections of moist potential temperature in Fig. 4.34 (right column) confirm the 315 K $\theta_a$ surface elevation discussed in previous section, at least during incipient stages when the section has a more meridional orientation. The composite horizontal wind and kinematic vertical motion on the section plane also verify that the northwesterlies behind the PV tongue and the southeasterlies ahead of it (shown by winds normal to the plane) are associated with sinking and rising motion, respectively, and are more prominent from $t_{12}$ to $t_{-12}$. Low level PV generation during this period is also accompanied by upward motion, reaching $-0.70$ Pa s$^{-1}$ at 500 hPa at $t_0$ (Fig. 4.34h). The transverse circulation of the cyclone along the cross section is quite visible, with the upper level branch being stronger on the warm air side and the lower being stronger on the cold air side. In agreement with Fig. 4.22, the region of strongest upper level divergence moves away from the surface center after $t_0$, and so does the region with strongest upward motion (Figs. 4.34 j and l).
The same variables are also plotted in CD cross sections (t_{1.12} to t_{1.12}) in Fig. 4.35, except that \( \theta_a \) is replaced by the equivalent potential temperature (\( \theta_e \)) with potentially unstable regions lightly shaded. During this period, advection of high PV air by the strong northwesterly flow (from under the plane) produces a dramatic intensification of the PV anomaly with time on the plane and eventually a lower tropopause to the south (i.e., the development of a tropopause fold, Figs. 4.35a, c and e). Air with elevated PV values just to the left of the surface cyclone is then advected toward it by the southwesterlies on the section plane (Figs. 4.35b, d and f, also c.f., Fig. 4.28). In the lower troposphere, a separate PV anomaly again appears by \( t_{1.12} \), intensifies at \( t_6 \) (especially near the surface), and interacts with the upper level PV anomaly. A deep layer of potentially conditionally unstable air below about 700 hPa to the southwest of the surface system is present in the \( \theta_e \) field, while the stability is also very low near the surface cyclone center and the warm front, suggested by the steep slopes of \( \theta_e \) surfaces. The air movement along the plane (vectors) is more or less parallel to the \( \theta_e \) contours in most regions (especially at \( t_{1.12} \)) with stronger ascent along the warm front. The warm front significantly strengthens at \( t_6 \) to about 7 K per 100 km in the composite field due to strong low level convergence, while upward motion above it also nearly doubles to reach a maximum of -1.09 Pa s\(^{-1}\) at 500 hPa (Fig. 4.35d). At this time PV values are enhanced in the ascending region, and a separate center of 0.9 PVU appears near 700 hPa, suggesting that its formation is closely related to latent heating associated with strong upward motion (Fig. 4.35c).

Cross sections for WE NWA (Figs. 4.36 and 4.37) can be compared with those for the strong intensity class. For AB sections (Fig. 4.36) the upper level PV anomaly is
noticeably weaker throughout development, as is the associated maximum wind speed normal to the plane (by about 10 to 15% before $t_{12}$). The weaker jet center is consistent with conditions at 250 hPa (Fig. 4.23), and merely reflects the weaker tropospheric temperature and height gradients in the WE NEA composites. Near the surface at $t_{36}$ and $t_{24}$, however, the strength of the primary circulation associated with the incipient cyclone is somewhat stronger for the WE case (Figs. 4.34 and 4.36, a and c), in agreement with composites at sea level and 850 hPa. This is no longer the case before $t_6$ when interaction between upper and lower levels becomes much stronger in the ST event, although the low level PV anomaly in WE remains very comparable (Figs. 4.34g and 4.36g). The upward motion embedded in the ST transverse circulation exceeds that of the WE one convincingly at $t_6$, and confirms that greater upper level divergence exists, leading to subsequent faster low level PV generation. The northwesterly wind behind the ST PV tongue is indeed stronger throughout the entire troposphere, as is the associated sinking motion between $t_{12}$ and $t_{12}$.

Weaker northwesterlies in the left-half of the domain of CD sections for WE (Fig. 4.37) produce a weaker PV anomaly on the plane than ST (Fig. 4.35), especially at $t_6$. The evolution in the WE $\theta_e$ field is very similar to the ST case, with deep layers of potentially conditional instability in nearly the same region, but the warm front and low level convergence of wind vectors just to the northeast of the surface cyclone center are both somewhat weaker at $t_6$. The maximum vertical velocity produced above the frontal zone is $-0.93$ Pa s$^{-1}$ at $t_6$, also weaker than ST.
Figures 4.38 to 4.41 show the AB and CD cross sections for the ST and WE NCA categories. While the PV field evolution for ST NCA in Fig. 4.38 is similar to that for ST NWA, the low level PV maximum of 0.7 PVU (clearly separated from the upper level PV anomaly) appears earlier at $t_{24}$ near 850 hPa (Fig. 4.38c), and becomes well established (1.0 PVU) by $t_{12}$, before the arrival of high PV air aloft from the southwest (Fig. 4.38e, c.f., Fig. 4.30). The low level PV anomaly reaches its 1.5 PVU maximum at $t_0$ upon the mutual interaction between the upper and lower level PV anomalies (Fig. 4.38g). The primary circulation associated with the explosive cyclone also intensifies rapidly and it reaches the peak of about 20 m s\(^{-1}\) at the sea level near $t_{12}$ (Fig. 4.38i). By this time the low level PV center has started to weaken, and more interestingly, to ascend gradually upward apparently due to the export of local PV by vertical motion. The kinematic vertical motion field is consistent with the early appearance of the low level PV center, as the strongest middle-tropospheric upward motion ($-0.55$ Pa s\(^{-1}\)) actually occurs at $t_{24}$ but remains comparable through nearly $t_{12}$. Since the cyclone migrates farther north relative to the upper level jet after $t_0$ (Fig. 4.24), the ascending branch forced by the upper level dynamics gradually moves toward the right and becomes separated from the branch associated with the cyclone itself (Figs. 4.38j, l). In the $\theta_a$ field, development of a tropospheric warm core structure and zone of low static stability directly above the surface cyclone center is clearly visible starting from $t_0$, and is in good agreement with the 850 hPa warm air seclusion (Fig. 4.12). The downward bending of $\theta_a$ surfaces is imposed by a warm anomaly below (c.f., Fig. 2.11) and is commonly found in oceanic explosive cyclones (Chapter 2).
Figure 4.38: Same as Fig. 4.34 but along AB lines in Fig. 4.30 for ST NCA category.
Figure 4.39: Same as Fig. 4.35 but along CD lines in Fig. 4.30 for ST NCA category.
Figure 4.40. Same as Figure 4.34 but along AB lines in Figure 4.31 for WE NCA category.

(continued)
Figure 4.41: Same as Fig. 4.35 but along CD lines in Fig. 4.31 for WE NCA category.
The CD cross sections for the ST NCA category are shown in Fig. 4.39. Strong upper level northwesterlies exceeding 20 m s\(^{-1}\) again occur about 400 km to the left of center and advect high PV air from under the plane, producing a reversed tropopause height configuration. Evolution of the low level PV maximum is the same as Fig. 4.38, while its associated circulation also reaches maximum intensity at \(t_{-12}\), especially in the cold sector to the south-southwest of the system center with winds reaching 33 m s\(^{-1}\) (Fig. 4.39e). The potentially unstable region to the south extends from the surface to about 750 hPa at \(t_{-12}\) and later becomes shallower. The thermal gradient across the warm front at \(t_0\) is much weaker than in ST NWA, only about 3 K per 100 km. Ascent above the frontal zone in CD sections is strongest at \(t_0\) with a maximum of \(-0.96 \) Pa s\(^{-1}\) at the 500 hPa level, but weakens rapidly to almost half that value at \(t_{-12}\).

Comparing AB cross sections for the WE NCA category (Fig. 4.40) with those for ST NCA (Fig. 4.38), the WE upper level PV anomaly is consistently weaker with a higher tropopause, and the magnitude of the associated jet stream maximum normal to the plane is also smaller at \(t_{-24}\) and \(t_{-12}\) but not so at other times. The low level PV center first appears at \(t_{-36}\) (Fig. 4.40a) but is weaker than the ST case, especially at \(t_0\) as it reaches a 1.2 PVU maximum (Fig. 4.40g). This is consistent with the considerably stronger divergence and upward motion above the surface center noted in the ST composite between \(t_{-24}\) and \(t_0\) (also Figs. 4.24 and 4.25), with upward motion reaching a \(t_0\) peak of \(-0.43 \) Pa s\(^{-1}\) at 700 hPa for the WE category. The interaction in WE between the weaker upper and lower level PV anomalies is also likely less active, as suggested by the smaller vertical extent of both the positive and negative parts of the primary circulation since \(t_0\).
The circulation of the WE NCA explosive cyclone at sea level is maximized at $t_{-12}$ with a strongest wind speed of about 16 m s\(^{-1}\), 4 m s\(^{-1}\) weaker than the ST cyclone. In the $\theta_m$ field, the downward bending of the moist isentropes directly above the storm center does not become apparent until $t_{-24}$ (Fig. 4.401), in agreement with development of the warm core seclusion at 850 hPa. Also, the vertical separation between isentropes is smaller (most visible at $t_0$), suggesting somewhat more stable conditions near the storm center and supporting the less active interaction between upper and lower PV anomalies.

For cross sections along lines CD of the WE NCA composite (Fig. 4.41), the upper level PV anomaly is again weaker when compared with the ST case (Fig. 4.39), as a result of significantly weaker northwesterlies aloft at the left side of the domain. Evolution of the low level PV field is similar to that in Fig. 4.40, and the maximum does not start ascending after $t_{-12}$. The $\theta_e$ pattern of the WE case is very similar, with a very comparable potentially unstable region during the period, while the lower $\theta_e$ values to the left of storm center in ST at $t_0$ and $t_{-12}$ is obviously caused by much stronger cold air advection. The strongest $t_0$ upward motion for the WE NCA category is $-0.70$ Pa s\(^{-1}\), considerably weaker than the ST category but consistent with weaker divergence.

Cross sections for ST and WE NEA are presented in Figs. 4.42 to 4.45. In ST NEA sections along lines AB in Fig. 4.32 (Fig. 4.42), the approach and subsequent arrival of a strong upper level PV anomaly, together with the overall rightward movement of the jet stream near the tropopause are evident. The low level PV anomaly (0.7 PVU) appears very close to the surface at about 920 hPa even before $t_{-36}$ (Fig. 4.42a), then slowly rises to 850 hPa and grows to about 1.0 PVU at $t_{-12}$ (Fig. 4.42e) while remaining clearly
Figure 4.42: Same as Fig. 4.34 but along AB lines in Fig. 4.32 for ST NEN category.
(continued)

Figure 4.42.
Figure 4.43: Same as Fig. 4.35 but along CD lines in Fig. 4.32 for ST NEA category.
Figure 4.44: Same as Fig. 4.34 but along AB lines in Fig. 4.33 for WE NEA category.
Figure 4.45: Same as Fig. 4.35 but along CD lines in Fig. 4.33 for WE NEA category.
separated from the upper level anomaly. The narrowing of jet stream is quite dramatic between \( t_{12} \) and \( t_0 \), especially near the upper level PV center, along with a rapid upward extension of the negative part of the cyclone circulation (out from the plane) from the lower troposphere (Fig. 4.22e, g). The low level circulation continues to intensify after \( t_0 \) to a maximum at \( t_{12} \) (Fig. 4.22i), but its associated PV anomaly has already started weakening significantly from a peak value of 1.2 PVU. The right column of Fig. 4.42 shows that vertical motions become quite well established by \( t_{24} \) with maxima of \(-0.55\) and \(-0.66\) Pa s\(^{-1}\) at \( t_{12} \) and \( t_0 \) respectively. The separation between two branches of strong upward motion associated with the cyclone and the jet stream starts as early as \( t_0 \) (Fig. 4.42h), and the development of a warm core with relatively low stability is apparent after \( t_{12} \), similar to the ST NCA case.

The CD cross sections for ST NEA (Fig. 4.43) indicate a deep layer of northwesterly flow (in excess of 20 m s\(^{-1}\)) extending down to about 900 hPa and advecting air with higher PV values in the upper and middle troposphere (Fig. 4.43a). The growth of the upper level PV anomaly on the section plane between \( t_{12} \) and \( t_0 \) is very impressive, while the low level circulation reaches its maximum at \( t_{12} \) (Fig. 4.43e). The potentially unstable region to the southwest lies mostly below 800 hPa but another such region is also present to the northeast of the cyclone center below 850 hPa at \( t_{12} \) (Fig. 4.43d). The strongest vertical velocity above the relatively weak frontal zone increases from 0.53 to 0.86 Pa s\(^{-1}\) at \( t_{12} \), but only marginally to 0.91 Pa s\(^{-1}\) at \( t_0 \) before dropping rapidly to only 0.46 Pa s\(^{-1}\) afterwards. The overall reduction in the horizontal wind parallel to the section plane between \( t_{12} \) and \( t_{12} \) is quite substantial, but nonetheless
consistent with the earlier weakening of the thermal gradient at lower levels (e.g., Fig. 4.14).

The results along AB cross sections for the WE NEA category are shown in Fig. 4.44. When compared with Fig. 4.42, it becomes evident that the WE northwesterly flow is significantly weaker than in the ST case during the incipient stage between $t_{36}$ and $t_{12}$, leading to a weaker upper level PV anomaly accompanied by a higher tropopause. The WE jet stream is also weaker after $t_{24}$, particularly so at $t_0$ when the difference in maximum wind speed reaches 10 m s$^{-1}$ (Figs. 4.42g and 4.44g). Generation of low level PV is convincingly faster in ST, as its low level PV anomaly grows from 0.7 to 1.2 PVU from $t_{36}$ to $t_0$, while the one in the WE case only increases by 0.1 to 0.9 PVU during the same period. This is in good agreement with the weaker upward motion (with a maximum of $\sim 0.43$ Pa s$^{-1}$ at 700 hPa at $t_0$) in the WE event throughout the incipient stage. The interaction between upper and lower levels also appears to be stronger in the ST composites, as the portion of its cyclone circulation with negative wind speed extends to greater elevation after $t_{12}$. In the WE event, the splitting of the ascending branch into two occurs by $t_{12}$, one part associated with the upper level jet and the other associated with the cyclone. Simultaneously the warm core structure also becomes evident directly above the cyclone center (Fig. 4.44j). For the CD cross sections, the difference at $t_0$ in the strength of the upper level PV anomaly between the ST (Fig. 4.43) and WE composites (Fig. 4.45) is quite substantial, but not so at $t_{12}$ as if the timing of the arrival of the WE upper level PV anomaly is not optimal. Corresponding to a significantly weaker upper level divergence (Fig. 4.27), the WE composite also reaches a smaller upward motion.
maximum of $-0.62$ Pa s$^{-1}$ at $t_0$ in the middle troposphere, although the thermal gradient across the warm front is stronger at this time (Fig. 4.45d). After $t_0$, however, the strongest ascent remains comparable at $-0.58$ Pa s$^{-1}$ in the WE event (Fig. 4.45f). The potentially unstable regions in both cases are very similar.

When the cross sections along lines AB for the three strong cases in NWA, NCA, and NEA (Figs. 4.34, 4.38 and 4.42) are compared, several differences can be identified. These differences are consistent among variables, and can be linked to those differences found between composites in earlier sections of this chapter. Regarding evolution of the upper level PV anomaly, the 2 PVU contour line associated with the NEA anomaly tends to be lower, while the opposite is true in the NWA sector. The PV values therefore tend to be higher at similar pressure levels toward the NEA sector (and more so at the same height because of a colder troposphere), and such a tendency is more evident at $t_{12}$ and $t_0$. It is clear that in the ST NEA case more high PV air has arrived at regions directly above the surface cyclone at $t_0$ than occurs in the NWA case. After $t_0$, the upper level PV anomaly continues to grow through $t_{24}$ in the NWA sector (Fig. 4.34), while further enhancement stops near $t_{12}$ in both NCA and NEA (Figs. 4.38 and 4.42).

Regarding jet stream maxima identified in winds normal to the section plane, gradual right-ward movement is identified in all three cases from $t_{36}$ to $t_{24}$, but changes in jet strength during development differs greatly among sectors. Between $t_{12}$ and $t_{24}$ there is virtually no reduction in maximum jet intensity in NWA (Fig. 4.34), some reduction by about $6$ m s$^{-1}$ in the NCA (Fig. 4.38), and substantial reduction from $44$ to $28$ m s$^{-1}$ in the NEA case (Fig. 4.42). This agrees with the relatively strong thermal and height gradients.
to the southeast of the cyclone center in the lower and middle troposphere after $t_0$ in the ST NWA event (Figs. 4.10 and 4.16) and the faster weakening of these gradients in the ST NEA case at the same time (Figs. 4.14 and 4.20). During the incipient stage from $t_{36}$ to $t_{24}$, before the reduction of the maximum wind speed, there is a significant downward extension in NEA (and in NCA to a less degree) of the maximum winds from the jet to lower levels in areas southeast of the surface center, best depicted by the 20 m s$^{-1}$ isotach (Figs. 4.38 and 4.42, a and c). In contrast, such an extension in NWA (Figs. 4.34a, c) is much less significant and the overall shape of the isotach pattern is broader, suggesting a more even distribution of the thermal wind and thus a larger area with appreciable horizontal temperature gradient on the plane. By the same reasoning, the early stage NEA thermal gradient should be the strongest northwest (left) of the system center where the vertical wind shear is large, and should be minimized to the southeast (right) of the center where isotachs are nearly vertical.

In the lower troposphere, the earlier appearance of strong PV anomalies (separated from the upper level anomaly) toward the NEA sector is closely linked to the aforementioned downward extension of wind speed maxima, corresponding to stronger horizontal wind shear vorticity and circulation, also suggested by composites at the sea level and 850 hPa (Sections 4.2 and 4.3). The low level PV center of the ST NEA case reaches its maximum at $t_0$ then weakens rapidly afterwards, while the associated PBL wind peaks at $t_{-12}$ (Fig. 4.42). On the other hand, the distinct low level NWA PV anomaly does not appear until after $t_{-12}$ but continues to grow through $t_{+12}$ along with the PBL wind maximum (Fig. 4.34), before weakening slowly. The timing of the first appearance of the
NCA low level PV maximum is at $t_{24}$, between those for NWA and NEA, while its intensity reaches the largest value among all cases of 1.5 PVU at $t_0$ before quickly weakening (Fig. 4.38). The faster evolution toward NEA after the initiation of the rapid deepening is again quite evident, and can be illustrated by comparing the PBL wind strength from $t_{12}$ to $t_{24}$. The strongest PBL winds occur in the NEA sector at $t_{12}$, in the NCA sector at $t_0$ and $t_{12}$, and at $t_{24}$ in NWA. Among the three strong cases, the evolution of the primary circulation after the incipient stage exhibits a reduction in wind speed both downward from the jet stream in the positive portion and upward from the top of the PBL in the negative portion. This occurs much faster toward the NWA and slower toward the NEA sector, and represents a stronger overall NWA storm environment baroclinicity and a more equivalent barotropic atmosphere toward the NEA sector as cyclones become more deeply embedded into the polar air stream (Figs. 4.34, 4.38, and 4.42, g, i, and k).

The evolution of the $\theta_m$ field among the three sectors for ST cases is in good agreement with that of the wind field discussed earlier. During the incipient stage, in NWA (Figs. 4.34b, d) the tropospheric $\theta_m$ surfaces have more homogeneous slopes, while regions with steeper $\theta_m$ slopes in NEA and NCA are more confined in the lower to middle troposphere roughly from the surface cyclone center to 1000 km to the left (northwest), suggesting overall less available potential energy (APE) for baroclinic conversion (Figs. 4.38 and 4.42, b and d). Between $t_{36}$ and $t_0$ in the left-half of the domain, all three cases show a gradual increase in isentrope slopes, brought about by the confluence in winds along the section plane. At $t_{12}$ and $t_{24}$ it is also apparent that the slopes in the middle part
of the domain remain relatively strong in the NWA sector but are relatively weak in NCA and NEA. Development of the downward bending of the isentropes and warm core structure above the storm center since \( t_0 \) is visible only in NCA and NEA composites. Toward the NEA sector, the vertical separation between adjacent isentropes after explosive deepening is significantly greater than that in the NWA sector (Figs. 4.34, 4.38 and 4.42, j and l). This indicates a lower static stability, in agreement with the warm core in lower and middle levels and closed cut-off low in upper levels, and is consistent with the seemingly stronger interaction between the upper and lower PV anomalies. The stronger NEA and NCA sector ascent splits into two branches, one associated with the transverse circulation of the cyclone and the other forced by the jet stream dynamics (Figs. 4.24 and 4.26). A similar phenomenon, although more subtle, can be observed in the NWA case at \( t_{-12} \) and \( t_{+24} \), and the fact that the two branches are still intact as one at \( t_0 \) may be the primary reason for the NWA event to have strongest maximum upward vertical velocity \((-0.70 \text{ Pa s}^{-1})\) among all three cases (Fig. 4.22).

The CD cross sections can also be compared between the three strong composite cases. It is evident that the upper level PV anomaly in the NWA sector between \( t_{-12} \) and \( t_{-12} \) is accompanied by a higher tropopause with lower PV values (lighter gray shades), while anomalies in the NCA and NEA sectors have a lower tropopause and are significantly stronger (Figs. 4.35, 4.39 and 4.43). The difference in the upper level PV anomaly strength at \( t_0 \) is quite large as the surface cyclone undergoes its most rapid development, and agrees well with the advancing speed of the leading edge of the PV tongue along the 315 K \( \theta_m \) surface (Figs. 4.28, 4.30 and 4.32). In these cross sections,
the upper and lower level circulations are more discrete than in the AB sections, and their mutual intensification is also better shown. The low level circulation in all three cases reaches maximum strength at $t_{12}$, but is considerably stronger in the NCA and NEA sectors especially by $t_0$, consistent with the stronger initial strength of their low level PV anomaly. In the $\theta_e$ field, the warm front in the NWA case is much stronger than in the other two sectors, especially at $t_0$, while the potentially unstable layer to the southwest is also considerably deeper. The air motion along the cross section plane (wind vectors) shows in all cases that strong southwesterly flow enters the domain from the left, ascends along the warm front, and then exits to the right, resulting in an overall upward motion inside the domain. The strong ascent close to the storm center produces nearly neutral stability from the convective adjustment, as is shown by the very steep $\theta_e$ surfaces slopes. The strongest upward motion at $t_{12}$ between cases again occurs in the NEA sector ($-0.86 \text{ Pa s}^{-1}$) but afterwards switches to the NWA sector, supporting the notion of a faster evolution toward the NEA. The largest NWA $t_0$ vertical velocity of $-1.09 \text{ Pa s}^{-1}$ at 500 hPa (Fig. 4.35d) may be linked to the slower separation of the ascending branches noted earlier in this region. Due to the earlier separation of the jet stream relative to the cyclone in NEA, the southwesterly flow shows a large overall reduction in wind speed from $t_{12}$ to $t_{12}$, in correspondence to a much depleted baroclinicity in the storm environment (Figs. 4.39 and 4.43).

Many differences qualitatively similar to those found between cross sections for strong cases, also exist between weak (and moderate [not shown]) cases. For both types of cross sections, both the upper and lower PV anomalies tend to be stronger, and the
shape of isotachs tends to be narrower with a relatively confined region of significant baroclinicity toward the NEA sector during the incipient stage up to $t_0$ (Figs. 4.44 and 4.45). The faster-evolved development produces a warm core structure and a significant reduction in vertical wind shear of the southwesterly flow, together with a more equivalent barotropic storm environment and relatively low static stability. Toward the NWA sector (Figs. 4.36 and 4.37), both the upper and lower level PV anomalies tend to be weaker, and the low level anomaly appears later in the incipient stage. Prior to $t_{-12}$, the WE NWA isotachs have a shape broader than in the NCA and NEA cases, but not as broad as in the ST case of the same sector (Fig. 4.34), implying less APE in the environment. The development is slower-evolved and is associated with a smaller reduction in the southwesterly vertical wind shear, higher static stability, and little indication of a composite warm core structure.

4.8 Discussion and Conclusion

The purpose of this section is to synthesize results of the composite and cross section analysis presented in previous sections of this chapter, and to summarize some qualitative characteristics and differences in the dynamical and thermal structure as well as in the evolution of composite explosive cyclones in individual sectors. Many features observed in Figs. 4.4 through 4.45 and discussed in Sections 4.2 through 4.7 are reasonably consistent between levels and among variables in a dynamical sense, and therefore are not likely to be features that arise randomly from the compositing.
For the NWA sector, during the incipient stage from about \( t_{.36} \) to \( t_{.12} \) the low level system at the sea level and 850 hPa tends to be weaker than those of the NCA and NEA sectors. This tendency is also noted by comparing Figs. 4.46, 4.47 and 4.48, which show vertical profiles of composite geostrophic wind speeds averaged over a \( 5 \times 5 \) grid point box centered at the system center from \( t_{.48} \) to \( t_{.36} \) for the three sectors. The initial low level circulation strength does not correspond to the intensity class in this sector and thus does not appear to play an important role in determining the subsequent maximum deepening rate of development (Fig. 4.46, and Sections 4.2 and 4.3). Parameters more consistent with the intensity class during the incipient stage are those related to the horizontal temperature gradient, and thus the baroclinicity and amount of APE for development. Fig. 4.49 shows the mean 1000-500 hPa thermal wind speed of an \( 11 \times 11 \) grid point box, which covers nearly the entire cyclone system and curves represent environmental baroclinicity. For the NWA sector (Fig. 4.49a), the thermal wind magnitude increases during the early stage and reaches a maximum at \( t_{.12} \) or \( t_{.0} \), and corresponds to the intensity class well from \( t_{.48} \) to \( t_{.0} \), suggesting that the environmental baroclinicity of the more explosive events is indeed stronger. Stronger events also tend to have stronger thermal wind speed between \( t_{.44} \) and \( t_{.12} \) in the NEA sector (Fig. 4.49c), but similar correspondence is only marginally good between \( t_{.24} \) and \( t_{.12} \) for the NCA sector (Fig. 4.49b).

The largest differences between events in different intensity classes in the NWA sector are produced during the most rapid development between \( t_{.12} \) and \( t_{.0} \). Stronger PVA/divergence forcing in the upper troposphere, accompanied by a stronger height
Figure 4.46: Vertical profile of composite geostrophic wind speed (m s⁻¹) averaged over a 5 x 5 grid point box centered at the cyclone center from t = -48 to t = -12 for the NWA sector. The ST, MO, and WE cases are plotted as thick solid, dashed, and thin solid lines, respectively. The lower limit of each curve is 1000 hPa, or the averaged MSLP value of the box if it is less than 1000 hPa.
Figure 4.47: Same as Fig. 4.46 but for the NCA sector.
Figure 4.48: Same as Fig. 4.46 but for the NEA sector.
Figure 4.49: Composite 1000-500 hPa thermal wind speed (m s\(^{-1}\)) averaged over an 11 \times 11 grid point box centered at the cyclone center from \(t_4\) to \(t_{72}\) for (a) NWA, (b) NCA, and (c) NEA sectors. The ST, MO, and WE cases are plotted as thick solid, dashed, and thin solid lines, respectively.
gradient, $\eta$ maximum, PVA/NVA dipole and jet streak, is responsible for producing the larger deepening rate of the ST composite cyclone (Sections 4.4 and 4.5). Therefore, it appears that the self-development process in the ST case is fundamentally the same but more intense, with every one of its components being stronger owing to the stronger environmental baroclinicity (Fig. 4.49a). Figures 4.50 to 4.52 present the vertical profile of composite divergence averaged over the same $5 \times 5$ grid point box as in Fig. 4.46 for the three sectors, respectively, while Fig. 4.53 shows the vertically integrated mass flux computed from the mean divergence. For the NWA sector (Figs. 4.50 and 4.53a), the ST composite upper level divergence and low level convergence do not exceed those of MO and WE cases until $t_0$, as discussed in previous sections. The smaller total mass flux obtained for the ST case between $t_{12}$ and $t_0$, however, must be caused by the choice of a rather large grid point box, since directly above the cyclone center the ST case must have a greater outward total mass flux in order to have a larger MSLP drop (Fig. 4.53a). From an IPV point of view, the stronger upper level forcing in the ST case is translated into a stronger PV anomaly on the $315 \ K \ \Theta_m$ surface and on the cross sections, although the strength of these anomalies are weaker in the NWA sector than in the other two sectors. The low level PV anomaly appears about 12 hours prior to the arrival of the upper level anomaly in this sector, and the rapid cyclogenesis at the surface of the ST case, forced by the divergence associated with the upper level PV anomaly, is accompanied by significant PV generation in the lower troposphere (Section 4.7).

NWA composite explosive cyclones also show a slower evolution and life cycle, corresponding to a smaller Coriolis parameter and slower spin-up process. Systems
Figure 4.50: Vertical profile of composite divergence ($10^{-6}$ s$^{-1}$) averaged over a $5 \times 5$ grid point box centered at the cyclone center from $t_{-48}$ to $t_{+36}$ for the NWA sector. The ST, MO, and WE cases are plotted as thick solid, dashed, and thin solid lines, respectively. The lower limit of each curve is 1000 hPa, or the averaged MSLP value of the box if it is less than 1000 hPa.
Figure 4.51: Same as Fig. 4.50 but for the NCA sector.
Figure 4.52: Same as Fig. 4.50 but for the NEA sector.
Figure 4.53: Vertically integrated divergent component, convergent component, and total mass fluxes (kg s⁻¹) computed from the mean divergence shown in Figs. 4.50 to 4.52 for the column of a 5 x 5 grid point box centered at the cyclone center from $t_{24}$ to $t_{72}$ for (a) NWA, (b) NCA, and (c) NEA sectors. Divergent and convergent components are plotted as solid lines, and the total mass fluxes are plotted as dashed gray lines. Curves are thick for ST, intermediate for MO, and thin for WE cases, respectively.
continue to develop for a longer period of time and/or to weaken at considerably slower rates after $t_0$, a tendency clearly seen at various tropospheric levels (Sections 4.2 to 4.6). A comparison between Figs. 4.50 to 4.52 provides one more example, showing that the reduction in NWA upper level divergence after $t_0$ is much slower than in the NEA sector. Other characteristics in the structure and evolution of NWA composite explosive cyclones include a deeper potentially unstable layer to the south, stronger thermal contrast across the warm front at low levels, and stronger static stability during development. Figure 4.54 shows the mean stability factor (pressure-weighted) of the composite storms in the middle and lower troposphere between 1000 and 500 hPa for the same $5 \times 5$ grid point box surrounding the cyclone center. The stability factor ($\sigma^{-1}$) is the inverse of mass density defined in Eq. (2.4), except here $\theta_m$ is used instead of $\theta$. A lower $\sigma^{-1}$ value corresponds to a lower stability. As shown in Fig. 4.54a, $\sigma^{-1}$ values in NWA cases are between 4.9 and $5.7 \times 10^{-3}$ m$^2$ K kg$^{-1}$ during the incipient stage, then decrease slowly to about 4.5 to $5.0 \times 10^{-3}$ m$^2$ K kg$^{-1}$ toward $t_{c-2}$. In contrast, stability factors in NCA and NEA (Figs. 4.54b, c) start with comparable values early on, but decrease rapidly to below $4.0 \times 10^{-3}$ m$^2$ K kg$^{-1}$ during development (except for WE NCA). The slower destabilization of the atmosphere in NWA is consistent with the slower spin-up process and life cycle, corresponding to a slower conversion from APE to KE and therefore a slower reduction in vertical wind shear (Section 4.7 and also Fig. 4.49). In addition, a distinct warm core seclusion is not observed in NWA composite fields, and its absence is likely caused by the coarse resolution of the data set. However, results agree well with the stability evolution and suggest that this feature is less pronounced in the NWA sector than in NCA or NEA.
Figure 4.54: Pressure-weighted stability factor between 1000 and 500 hPa averaged over a 5 x 5 grid point box centered at the cyclone center from $t_{-72}$ to $t_{+72}$ for (a) NWA, (b) NCA, and (c) NEA sectors. The stability factor is defined in text, and is computed from the composite $\theta_e$ fields. The ST, MO, and WE cases are plotted as thick solid, dashed, and thin solid lines, respectively.
Since the NEA sector is farther away from land-sea boundaries, it is quite clear in Fig. 4.49 that the mean baroclinicity in the environment becomes weaker toward this region. The mean 1000-500 hPa thermal wind speed to reaches its peak toward the earlier stages in NEA, such that the implied amount of APE actually decreases before and during the development from \( t_{48} \) to \( t_{36} \) (Fig. 4.49c). For a smaller area in the vicinity of the cyclone center, a comparison between the vertical profiles of the geostrophic wind speed (Figs. 4.46 and 4.48) suggests that the horizontal temperature gradient in NEA during early stages (about \( t_{48} \) to \( t_{24} \)) is quite comparable to the NWA at lower levels (or even stronger in the ST case) but considerably weaker in the middle and upper troposphere. Thus, the baroclinicity in the NEA sector tends to be shallower and confined in a smaller region near the incipient cyclone, consistent with the results in Section 4.7.

The systems at lower levels in the NEA sector tend to be stronger than those in NWA, with a much earlier appearance of a distinct PV maximum near the surface before \( t_{36} \) mainly related to the fact that incipient lows have moved over warmer ocean well before \( t_0 \) (Fig. 4.1). The corresponding PBL geostrophic wind between \( t_{48} \) and \( t_{12} \) (Fig. 48) is on average about 3 m s\(^{-1}\) stronger than that in the NWA (Fig. 46), and hence further deepening at the same rate tends to produce a correspondingly stronger wind at \( t_0 \) and \( t_{12} \), especially for the ST case. After \( t_{12} \), low level winds weaken more rapidly in the NEA sector and become weakest among all sectors. The geostrophic wind above the PBL is also stronger up to the middle troposphere in the NEA during the incipient stage, in agreement with conditions at 500 hPa (Section 4.4). The low level PV anomalies in the NEA at \( t_{24} \) and \( t_{12} \), before onset of explosive deepening, are consistently stronger than
those of the NWA by an average of 0.3 to 0.4 PVU, especially in the ST and MO cases (Section 4.7). This suggests that the initial system strength may be important in producing explosive cyclogenesis in general for this sector, and applying the antecedent vorticity growth concept of Gyakum et al. (1992) could be more successful. The stronger incipient stage low level PV anomaly is at least partially maintained by stronger upper level PVA and divergence above the surface system (Sections 4.4 and 4.5). A comparison between Figs. 4.50 and Fig. 4.52 confirms that the NEA upper level divergence is indeed stronger before $t \cdot 12$ for the MO and WE cases, and more so before $t \cdot 12$ for the ST event. Fig. 4.53 also shows a tendency for the outward and total mass fluxes to be larger toward the NEA sector from $t \cdot 48$ to $t \cdot 12$. The smaller total mass flux of the MO case at $t_0$ when compared to the WE event in Fig. 4.53c is again related to the large size of the grid point box. In the NEA sector itself, the increase in upper level divergence during the incipient stage (Figs. 4.52 and 4.53c), upward motions, and the rate of low level PV generation (Section 4.7), correspond very well with the intensity class. Between $t \cdot 36$ and $t_0$, the growth in low level PV maxima is about 0.5, 0.3, and 0.1 PVU for the ST, MO, and WE composites, respectively (Figs. 4.42 and 4.44). The agreement between the intensity class and other components, such as temperature and height gradients, PVA magnitude above the center, and the strength of the PVA/NVA dipole structure during the incipient stage before $t \cdot 12$, is also better in the NEA sector than in NWA, although toward the lower levels the circulation of the ST case is only comparable to the MO case in strength at $t \cdot 36$ and $t \cdot 24$ (Table 4.1 and Section 4.3). From Figs. 4.48 and 4.49c it appears that the MO NEA
system has an initial strength at least comparable to the ST case, but the baroclinicity is considerably weaker and tends to limit the subsequent maximum deepening rate.

In NEA, the upper level PV anomaly on the cross sections, as well as the corresponding PV tongue and its leading edge on the 315 K $\theta_v$ surface, are significantly stronger than those in the NWA and are mainly due to closer proximity of the explosive cyclones to the stratospheric high PV reservoir to the north (Sections 4.6 and 4.7). The stronger upper level PV anomaly in the NEA sector is associated with a lower tropopause, and corresponds to the stronger upper level divergence discussed earlier. The strengths of the upper level PV anomaly and its associated circulation also correspond well to the intensity class. During the incipient stage, the northwesterly flow behind the ST PV anomaly is stronger throughout the entire depth of the troposphere, and helps advect high PV air southward to produce a stronger upper level PV anomaly (Section 4.7). Eventually, this stronger upper level PV anomaly overtakes the pre-existing low level anomaly and the interaction between the two leads to the more violent explosive cyclogenesis between $t_{12}$ and $t_6$. During the development of such a favorable condition, the “parent cyclone” observed in the NEA sector in previous sections is likely to play an important role. This is supported by Table 4.2, which clearly shows that the association between the incipient system and its “parent cyclone” becomes closer as the intensity of the development increases. Near onset of the most rapid deepening period ($t_{12}$), the composite “parent cyclone” in the ST NEA case has closed circulation in the lower and middle troposphere, and appears at 250 hPa level as a deep trough with large amplitude (Figs. 4.8, 4.14, 4.20 and 4.26). Since the composites are made relative to the center of
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<tr>
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<td>19</td>
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Table 4.2: The association between incipient and parent cyclones in the ST, MO, and WE categories of the NEA sector (Region III). From top to bottom: Total number of cases, number of cases with parent cyclone, average MSLP of parent cyclone at $t_{-12}$, number of cases with direct merging between incipient and parent cyclones, and number of cases with significant northerly wind at sea level to the west of incipient low at $t_{-12}$. 

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the incipient low, the parent cyclones tend to be considerably stronger in individual cases than they appear in the figures. Upon occlusion, the parent cyclone is located to the north of the upper level jet stream at $t_{.36}$ and $t_{.24}$ (Figs. 4.8 and 4.26), and the incipient low develops in the baroclinic zone underneath the 250 hPa jet. At upper levels, the northerly branch of the parent cyclone’s circulation enhances the southward advection of high PV air toward the incipient low, and the greater amplitude of the upper level trough promotes stronger divergence aloft and helps to maintain the incipient low, which shows up at 850 hPa as a short wave trough with a distinct vorticity center (Fig. 4.14). At low levels, the narrow baroclinic zone near the incipient low is enhanced by confluence and frontogenetic process along the leading edge of cold air to the rear of the parent cyclone, promoting local convergence of heat and moisture and formation of the PV anomaly near sea level.

Compared to systems in the NWA sector, the evolution of NEA composite cyclones is significantly faster and their life cycle is considerably shorter, corresponding to a greater Coriolis parameter in higher latitudes and a rapid spin-up (and spin-down) process. NEA cyclones reach their maximum strength near $t_{.12}$ at low levels (Sections 4.2 and 4.3, also Fig. 4.48), while various forcing components such as upper level PVA and divergence tend to weaken even earlier, immediately after $t_0$ (Sections 4.4 and 4.5), especially for stronger events. Such a tendency also appears in Figs. 4.52 and 4.53, showing that the upper level divergence weakens significantly after $t_0$ and nearly diminishes completely by $t_{.24}$, much more rapidly than in the NWA sector (Fig. 4.50). A faster life cycle (and therefore a faster APE to KE conversion), together with a more limited amount of APE in NEA (Fig. 4.49) implies a rapid depletion of APE and
associated temperature gradient, as shown at various levels and cross sections in previous sections. Figure 4.48 also shows significant vertical wind shear weakening in NEA even before $t_0$, and an equivalent barotropic condition is established by $t_{12}$, much earlier than in the NWA (Fig. 4.46). From the IPV viewpoint, this faster life cycle and rapid depletion of baroclinicity imply a limited time for the upper and lower level PV anomalies to interact and mutually intensify, and thus supports the earlier notion that the initial strength of the low level system/PV anomaly appears to be important in determining the maximum deepening rate toward the NEA sector.

Composite NEA cyclones are associated with a shallower layer of environmental moist potential instability to the south, a weaker warm front near the surface, and a stronger warm core structure and weaker static stability during development (Section 4.7). The development of the warm core structure between $t_0$ and $t_{24}$ also appears in Fig. 4.48, as the strength of the circulation decreases with increasing height. Similar structure in the vertical profiles for the NWA sector (Fig. 4.46) is not as evident, consistent with results from the cross section analysis. The mean stability factor near the storm center in NEA (Fig. 4.54c) decreases throughout the incipient stage, and is apparently produced both by the approach of the upper level cold trough and the low level warm core development. As an example, the height fall at 500 hPa directly above the system center is 460 m from $t_{36}$ to $t_0$ in the ST NEA case (Fig. 4.20), compared to 270 m in ST NWA (Fig. 4.16). The stronger low level convergence in the NEA sector is also consistent with the larger cloud area (Section 4.2). The stability in NEA eventually reaches a minimum between 3.4 and $3.8 \times 10^{-3} \text{ m}^2 \text{ K kg}^{-1}$ at about $t_{36}$. Again, the lower static stability near the system center
the system center is consistent with the faster life cycle, the rapid depletion of APE, and the apparently stronger interaction between the upper and lower level PV anomalies seen in the cross sections.

The characteristics in the structure and evolution of the NCA composite cyclones, related to the initial strength of the circulation, baroclinicity and upper level forcing, PV anomalies and their interaction, warm core seclusion, stability, and the length of life cycle, lie mostly between those for the NWA and the NEA sectors. Over a smaller area near the cyclone center, the ST NCA event is accompanied by large vertical wind shear near \( t_{24} \) (Fig. 4.47), corresponding to a narrow but very strong jet stream at 250 hPa (Fig. 4.24). The mean baroclinicity in the environment, however, is between the NWA and NEA sectors in both the magnitude and the timing of the maximum (Fig. 4.49). The evolution in the vertical profile of divergence (Fig. 4.50), mass fluxes (Fig. 4.53b), and stability (Fig. 4.54b) lie in general between the NWA and NEA sectors. Therefore, the composite cyclones in the NCA sector will not be discussed further.
CHAPTER 5

VORTICITY BUDGET ANALYSIS

The purpose of the vorticity budget analysis is to evaluate the contribution of various dynamical and physical processes in generating cyclone vorticity, and therefore to investigate characteristics of composite cyclones from a more quantitative perspective. The vorticity budget analysis method used here is based on the Petterssen-Sutcliffe development equation under the QG framework. The equation set is derived in Section 5.1 while computational aspects are discussed in Section 5.2. In summary, the method partitions the total vorticity change at 1000 hPa during each 12-h interval into geostrophic and ageostrophic components, and further breaks the geostrophic component into contributions from change in 500 hPa vorticity and from thermal effects that change the 1000-500 layer thickness. The thermal component is further partitioned into 12-h layer temperature changes due to horizontal advection, vertical transfer plus adiabatic effects, radiation, latent heat release, and sensible heating plus all other subgrid scale effects. Results of the vorticity budget analysis are presented and discussed in Section 5.3, while results near and at the cyclone center as well as for different quadrants of the cyclone are discussed in Section 5.4. Section 5.5 provides the conclusion of this chapter.
5.1 Equation Set Derivations

Under geostrophic balance and hydrostatic equilibrium in the $p$-coordinate system, the derivation of the Petterssen-Sutcliffe development equation (Sutcliffe 1947; Petterssen 1956) starts from the geostrophic wind equation

$$u_g = -\frac{1}{f} \frac{\partial \Phi}{\partial y} \quad \text{Eq. (5.1)}$$

and

$$v_g = \frac{1}{f} \frac{\partial \Phi}{\partial x} \quad \text{Eq. (5.2)}$$

where $u_g$ and $v_g$ are the zonal and meridional components of the geostrophic wind, $\Phi$ is the geopotential, and $f$ is the Coriolis parameter. The geostrophic relative vorticity ($\zeta_g$) is

$$\zeta_g = \frac{\partial v_g}{\partial x} - \frac{\partial u_g}{\partial y} = \frac{1}{f} \nabla^2 \Phi \quad \text{Eq. (5.3)}$$

and the geostrophic absolute vorticity ($\eta_g$) is

$$\eta_g = \frac{1}{f} \nabla^2 \Phi + f. \quad \text{Eq. (5.4)}$$

Here $\nabla^2$ is the horizontal Laplacian operator. The total absolute vorticity is simply

$$\eta = \eta_g + \eta_{ag}, \quad \text{Eq. (5.5)}$$

where $\eta_{ag}$ is the ageostrophic component of absolute vorticity, and therefore
\[
\frac{\partial \eta}{\partial t} = \frac{\partial \eta_g}{\partial t} + \frac{\partial \eta_{ag}}{\partial t}.
\]
Eq. (5.6)

For an air column between two arbitrary pressure levels, the difference in \( \eta_g \) at the top and the bottom is

\[
\eta_{gt} - \eta_{gb} = \frac{1}{f} \nabla^2 \Phi_t - \frac{1}{f} \nabla^2 \Phi_b = \frac{1}{f} \nabla^2 (\Phi_t - \Phi_b) = \frac{g}{f} \nabla^2 h,
\]
Eq. (5.7)

where subscripts \( t \) and \( b \) denote top and bottom levels, and \( h \) is the air column thickness.

By taking time derivatives of Eq. (5.7), the simplest form of the Petterssen-Sutcliffe development equation relating local change rates of \( \eta_g \) at the bottom and top levels is obtained as

\[
\frac{\partial \eta_{gb}}{\partial t} = \frac{\partial \eta_{gt}}{\partial t} - \frac{g}{f} \nabla^2 \left( \frac{\partial h}{\partial t} \right).
\]
Eq. (5.8)

By using Eqs. (5.6) and (5.8), the change rate of total vorticity at the bottom level can be broken down to the geostrophic and ageostrophic terms, while the former is attributed to geostrophic vorticity change at the top level and the thickness change.

The local change rate of thickness between two fixed pressure levels is related to the temperature change rate of the air column through the hypsometric equation as

\[
\frac{\partial h}{\partial t} = -\frac{R}{g} \int_b^t (\frac{\partial T}{\partial t}) d\ln \rho,
\]
Eq. (5.9)

and the local temperature change rate can be computed from the thermodynamic energy equation
where $\alpha$, $C_p$, and $\dot{Q}$ are the specific volume, specific heat of air at constant pressure, and diabatic heating rate, while $\alpha/C_p$ and $\partial T/\partial p$ are the dry adiabatic ($T_d$) and environmental ($\gamma$) temperature lapse rates in $p$-coordinate. The three right hand side terms of Eq. (5.10) represent effects of horizontal advection, stability (adiabatic heating and cooling plus transfer associated with vertical motion), and diabatic heating and cooling, respectively. The diabatic effect is computed as the residual of all other terms, and can be expressed as

$$
\frac{1}{C_p} \dot{Q} = \frac{1}{C_p} (\dot{Q}_R + \dot{Q}_L + \dot{Q}_S). 
$$

where $\dot{Q}_R$, $\dot{Q}_L$, and $\dot{Q}_S$ represent heating and cooling rates from radiative processes, latent heat release, and sensible heat flux that also accounts for all other subgrid scale effects and computational errors, respectively. The radiative effect $\dot{Q}_R$ in development of individual cyclones is usually limited, and its resultant temperature change can be estimated by assuming it is associated only with the diurnal cycle.

The temperature change from condensational heating in Eq. (5.11) is estimated by using the thermal dynamic energy equations for both $\theta_m$ and $\theta_e$ in the following manner. The moist potential temperature $\theta_m$ is computed using Poisson's equation as
\[ \theta_m = T \left( \frac{10^5}{p} \right)^{\kappa_m} \]  
Eq. (5.12)

where \( \kappa_m \) is the ratio between gas constant and specific heat at constant pressure of moist air. For adiabatic motion without phase change of water vapor, following the air movement \( \theta_m \) is conserved and the total change rate of \( \theta_m \) equals to zero.

\[ \frac{d\theta_m}{dt} = \frac{\partial \theta_m}{\partial t} + V \cdot \nabla \theta_m + \omega \frac{\partial \theta_m}{\partial p} = 0. \]  
Eq. (5.13)

Thus, for any (non-adiabatic) motion, the local change of \( \theta_m \) can be written as

\[ \frac{\partial \theta_m}{\partial t} = -V \cdot \nabla \theta_m - \omega \frac{\partial \theta_m}{\partial p} + RH_m + LH_m + SF_m + E_m \]  
Eq. (5.14)

and it is determined by the three-dimensional advection and all diabatic effects, which include radiative \( (RH_m) \) and latent \( (LH_m) \) heating and cooling, sensible heat flux \( (SF_m) \), and an error term \( (E_m) \). The corresponding equation for the equivalent potential temperature \( \theta_e \) is

\[ \frac{\partial \theta_e}{\partial t} = -V \cdot \nabla \theta_e - \omega \frac{\partial \theta_e}{\partial p} + RH_e + MF_e + SF_e + E_e \]  
Eq. (5.15)

and the equation does not contain a latent heating and cooling term since \( \theta_e \) is conserved for pseudo-adiabatic motion. Instead, it has a moisture flux term \( (MF_e) \). For the calculation of \( \theta_e \) itself, the formula derived by Bolton (1980) is used
\[ \theta_e = \theta_m \exp \left[ \left( \frac{3376}{T_{LCL}} - 2.54 \right) \cdot r (1 + 0.81r) \right], \quad \text{Eq. (5.16)} \]

where \( r \) is the mixing ratio, and \( T_{LCL} \) is the air parcel temperature at the lifting condensation level (LCL), given by

\[ T_{LCL} = \frac{1}{\frac{1}{T - 55} - \frac{\ln(U/100)}{2840} + 55}, \quad \text{Eq. (5.17)} \]

where \( U \) is the relative humidity in %. In Eqs. (5.14) and (5.15), the effects from subgrid scale radiation, sensible heat flux and other processes due to turbulent mixing are usually much smaller than the rest of the terms, and are likely similar for \( \theta_m \) and \( \theta_e \) at the same grid point in three-dimensional space. Therefore, it is assumed that

\[ RH_m + SF_m + E_m = RH_e + SF_e + E_e \quad \text{Eq. (5.18)} \]

in obtaining the following relationship

\[ LH_m = \frac{\partial \theta_m}{\partial t} + V \cdot \nabla \theta_m + \omega \frac{\partial \theta_m}{\partial p} - \frac{\partial \theta_e}{\partial t} - V \cdot \nabla \theta_e - \omega \frac{\partial \theta_e}{\partial p} + MF_e. \quad \text{Eq. (5.19)} \]

by subtracting Eqs. (5.15) from (5.14). The term \( MF_e \) is significant inside the PBL, and is the ultimate source of water vapor and condensational heating. It comes from the surface moisture flux \( (Q_e) \), which is positive for upward flux (from surface to air) and can be estimated using
\[ Q_E = \rho L_v C_E |V| (q_{sfc} - q_{air}) \]  

Eq. (5.20)

where \( \rho \), \( L_v \), \( C_E \) and \( |V| \) are air density, latent heat of vaporization, bulk transfer coefficient for water vapor and wind speed, while \( q_{sfc} \) and \( q_{air} \) are specific humidity of the surface and of the air, respectively. Here, the value used for \( C_E \) is \( 1.5 \times 10^{-3} \) (Blanc 1985), as used by Anthes et al. (1983) and Roebber (1993). Once the water vapor enters the lowest levels of atmosphere through evaporation and turbulent mixing, it raises \( r \) and thus \( \theta_e \) through Eq. (5.16). Further details will be given in Section 5.2 for this computation.

Therefore, the effect of the estimated surface moisture flux \( Q_E \) in changing \( \theta_e \) (the term \( MF_e \)) can be obtained, as can the effect of latent heating in changing \( \theta_m \) (the term \( LH_m \)) through Eq. (5.19). The latter is translated into the corresponding change in \( T \) due to condensational heating through Eq. (5.12). Having obtained the effect of \( \dot{Q}_L \) and \( \dot{Q}_R \) assumed in Eq. (5.11), the effect of \( \dot{Q}_S \) can be calculated as the residual. Thus, the diabatic term in Eq. (5.10) is further broken down into effects from these three different physical processes.

5.2 Computational Aspects and Composites

Most terms of the equations derived in Section 5.1 (e.g., the advection terms) can be readily obtained for each time from \( t_{-84} \) to \( t_{-72} \), using methods described in Section 3.5.2. However, given the coarse temporal resolution of the data set (12 h), it is not easy to accurately estimate the tendency terms (time derivatives) even when some special
techniques are utilized. One example of such techniques is the cubic spline fitting function used by Sinclair and Elsberry (1986). To lower the sensitivity of the vorticity budget analysis results to possible computational errors, the time-integrated form of the equation set in Section 5.1 is used instead, and thus Eqs. (5.6), (5.8), (5.10), (5.11) and (5.19) become

\[ \Delta \eta = \Delta \eta_g + \Delta \eta_{ag}, \quad \text{Eq. (5.21)} \]

\[ \Delta \eta_{gb} = \Delta \eta_{gb} - \frac{g}{f} \nabla^2 (\Delta h), \quad \text{Eq. (5.22)} \]

\[ \Delta T = \int_{t_1}^{t_2} (-V \cdot \nabla T) \, dt + \int_{t_1}^{t_2} \omega \left( \frac{\alpha}{C_p} \frac{\partial T}{\partial p} \right) \, dt + \Delta T_{dia}, \quad \text{Eq. (5.23)} \]

\[ \Delta T_{dia} = \Delta T_R + \Delta T_L + \Delta T_S, \quad \text{Eq. (5.24)} \]

\[ \Delta \theta_{mL} = \Delta \theta_m - \int_{t_1}^{t_2} (-V \cdot \nabla \theta_m) \, dt + \int_{t_1}^{t_2} \left( \omega \frac{\partial \theta_m}{\partial p} \right) \, dt - \Delta \theta_e \]

\[ + \int_{t_1}^{t_2} (-V \cdot \nabla \theta_e) \, dt - \int_{t_1}^{t_2} \left( \omega \frac{\partial \theta_e}{\partial p} \right) \, dt + \Delta \theta_{QE}, \quad \text{Eq. (5.25)} \]

The symbol \( \Delta \) represents change of variables during any 12-h interval from \( t_1 \) to \( t_2 \), and \( \Delta T_{dia}, \Delta T_R, \Delta T_L, \) and \( \Delta T_S \) are changes in \( T \) due to diabatic processes, radiation, latent heat release, and sensible heat flux, while \( \Delta \theta_{mL} \) and \( \Delta \theta_{QE} \) are changes in \( \theta_m \) from latent heating and in \( \theta_e \) from surface moisture flux \( Q_{ev} \), respectively. Thus, by using the time-integrated form of the equations, the need for evaluating tendencies of variables at specific instants of time is eliminated and the amount of change over each 12-h interval is computed instead.

The advection and stability terms in Eqs. (5.23) and (5.25), however, need to be integrated through time with values provided every 12 h. For this purpose, the cubic
spline fitting function (Shikin and Plis 1995) is used to obtain estimates at 1 h resolution from \( t_{-84} \) to \( t_{-72} \) (to ensure the continuity of the quantity throughout the entire period), then the integration is performed for each 12-h interval.

The layer for which the vorticity budget analysis is performed is chosen to be between the 500 (top) and 1000 (bottom) hPa levels, based partially on data availability. Using the method described above, all terms in Eqs. (5.21) and (5.22) can be obtained, while the diabatic heating term in Eq. (5.23) is calculated as the residual of all other terms. The 12-h temperature change from radiation (\( \Delta T_R \)) in Eq. (5.24) is assumed to be associated with the diurnal cycle only, and is computed for each cyclone case at each grid point in \((x, y, p)\) space as the averaged temperature change between 00 and 12 Z during the 7-day period. Then, a 1-4-6-4-1 weighting function is applied in the zonal direction to produce some spatial smoothing but the range of \( \Delta T_R \) in the meridional direction is preserved.

In Eq. (5.25) all terms on the right hand side are available except for \( \Delta \theta_{QE} \), the 12-h change in \( \theta_e \) due to surface moisture flux, which needs to be estimated. Here, surface evaporation is assumed to affect \( \theta_e \) values only at the 1000, 925, and 850 levels inside or near the top of PBL, since in the free atmosphere the subgrid scale moisture flux is usually negligible compared to the transport of \( \theta_e \) by vertical motion. This is especially true in the case of rapid cyclogenesis where most regions in the storm environment are dominated by either significant ascending or descending motion. The humidity variable at the lowest level provided by the ECMWF data set is dew point temperature (\( T_d \)) at 2 m height, so the 12-h net energy input due to addition of water vapor through surface
evaporation per unit area ($\Delta E_{QE}$) is estimated (also using spline fitting function) as

$$\Delta E_{QE} = 0.8 \int_{t_i}^{t_f} \left[ \rho L_v C_E |V_{2m}| (q^{* - q_{2m}}) \right] dt,$$

Eq. (5.26)

where $|V_{2m}|$ and $q_{2m}$ are wind speed and specific humidity of air at 2 m height, and $q^{*_{sf}}$ is computed as the saturation specific humidity at surface temperature ($T_{sf}$) and pressure ($p_{sf}$). The latter assumption is valid in the storm environment during most of the 7-day period when the underlying surface is ocean. Because $q$ and $|V|$ at 2 m height are used instead of surface values, a 0.8 factor is applied to adjust for the likely overestimation in the moisture flux $Q_E$. Before February 1994, near surface $u$ and $v$ winds are provided at 10 m height, so the wind speed is reduced to 2 m height using the logarithmic decay relationship

$$|V_{2m}| = |V_{10m}| \frac{\ln(z_2/z_0)}{\ln(z_{10}/z_0)},$$

Eq. (5.27)

where $z_2 = 2$ m, $z_{10} = 10$ m, and $z_0$ is the surface roughness length, assumed to be a constant $10^{-4}$ m suitable for open sea (Oke 1987). Although $z_0$ is also dependent on wind speed and stability, its value does not seriously affect $|V_{2m}|$ as long as it is much smaller than 1 m. The 12-h net energy input $\Delta E_{QE}$ is, of course, related to the change in water vapor content through

$$\Delta E_{QE} = L_v \Delta M_v = \frac{L_v}{g} \sum (\Delta q \cdot \delta p),$$

Eq. (5.28)

where $\Delta M_v$ is the net water vapor mass input per unit surface area, $\Delta q$ and $\delta p$ are the specific humidity change at each of the three PBL levels and the pressure depth.
represented by the levels. The saturation deficit can be estimated for each 12-h interval at each grid point. If this deficit is greater than the estimated net input of water vapor from evaporation, $q$ is raised by the same fraction toward saturation at the three PBL levels. Otherwise, all three levels are brought to saturation and additional water vapor is distributed among them such that $q$ increases by a ratio of 3:2:1 at 1000, 925, and 850 hPa levels. Here, $\Delta q$ is not limited by the saturation because even with saturated air, the moisture could be transported elsewhere later in the same 12-h interval, thereby allowing more room for additional water vapor. Having estimated the 12-h $\Delta q$ (and thus $\Delta r$) from evaporation, its effect in raising $\Theta_r$ (i.e., $\Delta \Theta_{rQE}$) can be obtained through Eqs. (5.16) and (5.17) by using the mean values of $T$, $\Theta_m$, $U$, and $r$ between $t_1$ and $t_2$. Substituting $\Delta \Theta_{rQE}$ back into Eq. (5.15) yields $\Delta \Theta_{ml}$, which then can be translated into $\Delta T_L$ using Poisson's equation with a mean value of $\kappa_m$ between $t_1$ and $t_2$. Thus, $\Delta T_S$ is finally obtained using Eq. (5.24) as the residual between $\Delta T_{daw}$, $\Delta T_R$, and $\Delta T_L$.

The vorticity budget analysis is also first performed for individual cases for each 12-h intervals at each grid point for the entire Northern Hemisphere, then the finite-domain data are extracted and composites are made for each category, in the same manner as described in Section 3.5, except that variables are 12-h changes and the common reference point (the origin) is the surface cyclone center at the end of each 12-h interval. To evaluate contributions from various dynamical and physical processes in generating cyclone vorticity near the storm center, the values are averaged over rectangular grid point boxes with differing sizes in producing areal means. For thermal variables, Eqs. (5.8) and (5.9) state that the geostrophic vorticity change is related to the Laplacian of the thickness
change, while the thickness change is proportional to the vertical integration of
temperature change throughout the layer with respect to $\ln p$. The temperature change
variables between 1000 and 500 hPa are summed in the vertical according to the pressure
depth that the levels represent ($p$-weighted) for simplicity, since the interest here is to
compare general characteristics of composite cyclones in different categories. The best
estimates at 925 hPa level (Appendix) are also used here for this calculation.

5.3 Results and Discussion

5.3.1 General Description and Comparison

Results of the vorticity budget analysis are presented for a $5 \times 5$ rectangular box
centered at the cyclone center for the ST, MO, and WE categories in NWA, NCA, and
NEA for the entire 7-day period from $t_{44}$ to $t_{72}$ in Figs. 5.1 to 5.9. Vorticity and thermal
components are plotted separately in panels (a) and (b), while contributions from thermal
diabatic processes appear in panel (c), with symbols placed at the middle of each 12-h
interval. The $5 \times 5$ grid point box is about $1250 \times 1250$ km$^2$, covering the region
surrounding the storm center of strong pressure or height gradient during development
(c.f., Figs. 4.4 to 4.15).

The time series of box-averaged 12-h changes in 1000 hPa total absolute vorticity
(TOT) are plotted in the top panels (a), and are broken down to geostrophic (GEO) and
ageostrophic (AGEO) components, as well as both 12-h changes in 500 hPa geostrophic
Figure 5.1: 12-h changes in absolute vorticity ($\Delta \eta$, $10^{-5}$ s$^{-1}$) and temperature ($\Delta T$, K) averaged over a 5 x 5 grid point box centered at the storm center for 1000-500 hPa layer from $t_{44}$ to $t_{72}$ for the ST NWA composite cyclone. Vorticity components (a): Total $\Delta \eta$ (TOT), and its geostrophic (GEO) and ageostrophic (AGEO) components at 1000 hPa, geostrophic $\Delta \eta_g$ at 500 hPa (G500), and $\Delta \eta_g$ due to thickness change (GH). Temperature components (b): Total $\Delta T$ of the layer (TOT), $\Delta T$ from horizontal advection (ADV), vertical motion (VERT, or stability), and diabatic processes (DIA); and (c) $\Delta T$ from diabatic processes (DIA), latent heating (LH), sensible heating (SH), and radiation (RAD).
Figure 5.2: Same as in Fig. 5.1 but for the MO NWA composite cyclone.
Figure 5.3: Same as in Fig. 5.1 but for the WE NWA composite cyclone.
Figure 5.4: Same as in Fig. 5.1 but for the ST NCA composite cyclone.
Figure 5.5: Same as in Fig. 5.1 but for the MO NCA composite cyclone.
Figure 5.6: Same as in Fig. 5.1 but for the WE NCA composite cyclone.
Figure 5.7: Same as in Fig. 5.1 but for the ST NEA composite cyclone.
Figure 5.8: Same as in Fig. 5.1 but for the MO NEA composite cyclone.
Figure 5.9: Same as in Fig. 5.1 but for the WE NEA composite cyclone.
absolute vorticity (G500) and thermal wind vorticity associated with the Laplacian of the 1000-500 hPa thickness change (GH). Based on Eqs. (5.21) and (5.22), the relationships between these curves are $\text{TOT} = \text{GEO} + \text{AGEO}$ at 1000 hPa, and $\text{GEO} = \text{G500} + \text{GH}$ for the layer. The GEO curves in all figures are strongly positive during the development stage, roughly between $t_{-30}$ and $t_{-30}$, reaching peak magnitudes during most rapid development between $t_{-12}$ and $t_{0}$. On the other hand, the AGEO effect induced by upper level divergence and PBL friction acts to slow down the low level winds and contributes negatively, appearing as a reflected image of GEO but with smaller magnitude. As a result, total 1000 hPa $\Delta \eta$ (TOT) is also positive but is typically about 50 to 60% the magnitude of GEO ($\text{G500} + \text{GH}$; also Table 5.1). The G500 curves represent 500 hPa geostrophic forcing at 500 hPa mostly from vorticity advection, and are strongly positive with maximum magnitude comparable to GEO for the 1000 hPa level. GH curves generally have smaller magnitudes than other curves, but contribute positively for about 24 to 48 h before $t_{0}$ and negatively afterwards. Therefore, GEO curves usually exceed their G500 counterpart before $t_{0}$, while the opposite is true most of the time after $t_{0}$.

Some fundamental differences among the three sectors are evident when the general timing and shape of these various vorticity change components are compared. For the ST composites, the NWA curves (Fig. 5.1a) tend to have smaller amplitude near their $t_{0}$ maximum but the decline after explosive deepening period is significantly slower such that their magnitude remains large until about $t_{-42}$. The NEA curves (Fig. 5.7a) in contrast tend to reach a greater peak magnitude at $t_{-6}$ or $t_{-6}$, weakening dramatically thereafter such that all processes diminish by $t_{-18}$. The ST NCA curves (Fig. 5.4a) lie between those for
Table 5.1: Total change in various absolute vorticity ($\eta$) and temperature ($T$) components, averaged over a 5 x 5 grid point box centered at the cyclone center, in the vorticity budget analysis for the 1000-500 hPa layer (see text) during the development period from $t_{-48}$ to $t_{-4}$ (top half) and during only the more rapid deepening period from $t_{-24}$ to $t_0$ (bottom half) for the nine categories. Units of vorticity and temperature components are $10^{-5}$ s$^{-1}$ and K, respectively.
NWA and NEA, with intermediate amplitudes and weakening rates. The narrower, higher amplitude, NEA curves indicate rapid spin-up and spin-down processes and a faster life cycle, in agreement with results from the composite analysis (Chapter 4). This tendency is somewhat less obvious when comparing the MO (Figs. 5.2a, 5.5a, and 5.8a) and WE curves (Figs. 5.3a, 5.6a, and 5.9a). These curves will be further discussed in greater detail in the next section.

Middle panels (b) of Figs. 5.1 through 5.9 show (1) 12-h total temperature change \((p\text{-weighted})\) averaged over the same \(5 \times 5\) box for the 1000-500 hPa layer (TOT), (2) the contribution from horizontal advection (ADV), (3) vertical motion (VERT), and (4) all diabatic processes (DIA). Based on Eq. (5.23), \(\text{TOT} = \text{ADV} + \text{VERT} + \text{DIA}\), with the diabatic effect computed as a residual of the other three terms. The averaged 12-h \(\Delta T\) (TOT) is usually slightly positive by \(t_0\) for at least 24 h, having a peak magnitude no more than 1.5 K and afterwards becoming slightly negative from \(t_0\) to \(t_{-66}\). Although GH curves in panel (a) are not a direct result of \(\Delta T\) per se, as mentioned earlier, they and the TOT curves in (b) have qualitatively similar shapes. The contribution from horizontal advection (ADV) is slightly to moderately positive throughout the period, mainly due to the fact that individual storms are constantly moving toward the region of maximum warm air advection (WAA). That is, while thermal advection at the storm center is always nearly zero, during the early portion of each 12-h interval the cyclone center is actually to the west or southwest of the composite origin (Fig. 4.1) and is influenced by WAA. The VERT curves are the stability effect, including vertical transfer and adiabatic effect, and as expected are strongly negative during rapid development (also Table 5.1). The DIA effect
in contrast is strongly positive and usually reaches a maximum at $t_{16}$ (or $t_{18}$ for MO NEA and WE NCA, Figs. 5.6b, 5.8b), and is the major contributor in raising the total $\Delta T$ above zero before $t_0$, thereby producing positive contribution by GH curves in panel (a) to the cyclone vorticity generation. Since DIA is computed as the residual, it also includes computational error and all subgrid scale effects in $\Delta T$ not resolved by grid points. Also, since DIA changes the three-dimensional distribution of $T$ (affecting both the horizontal thermal gradient and vertical lapse rate), the three terms ADV, VERT, and DIA are closely interrelated, and interactions between them are manifested as nonlinear responses in ADV and VERT at the resolvable scale and in DIA at the subgrid scale.

A comparison among categories in panel (b) of the general timing and shape of the curves suggests at least two apparent differences. First, their amplitudes especially for the two major contributors VERT and DIA, tend to be greatest in the NWA sector (Figs. 5.1b to 5.3b) and smallest in the NEA sector (Figs. 5.7b to 5.9b); this difference is most obvious among WE cases and least visible among the ST cases. Second, contrast in amplitudes of these processes (particularly TOT) among intensity classes is largest in NEA sector and smallest in NWA sector. This suggests that net contribution from thermodynamical components is more important for a stronger development toward the NEA sector, but less important over the NWA sector. Further discussion on each of the processes in panel (b) is in Section 5.3.3.

The contribution from diabatic processes is further partitioned in panel (c) of Figs. 5.1 to 5.9 into the 12-h temperature change from three different physical processes. These include radiation (RAD), latent heating (LH), and sensible heating (SH), while DIA = LH
+ SH + RAD (Eq. (5.24)). DIA curves are the same as those in panel (b). The LH curves are the averaged net effect associated with condensational heating or evaporative cooling of the 1000-500 hPa layer inside the 5 × 5 grid point box. They are always positive since the area is dominated by upward motion (Table 5.1), and are usually the primary contributor to DIA but tend to peak after \( t_0 \) for all categories. LH can be added to VERT in panel (b) to give the total 12-h temperature change associated with vertical motion following a pseudo-adiabatic process, with a negative sum indicating an overall conditionally stable environment. The SH curves represent the effect of sensible heat flux, but in practice also include subgrid scale effects, the portion of nonlinear interaction not captured by grid points, and computational error. The SH contribution is typically slightly negative prior to about \( t_{-30} \), becomes positive for a period of about 12 to 24 h and then is negative (or near zero) again after about \( t_{-6} \). The total SH contribution is usually significantly smaller than LH, except for the ST NEA case (Fig. 5.7c, also Table 5.1). The RAD contribution is as expected constantly very small, nearly zero, in all categories, and is largely caused by the fact that some cyclones have their most rapid deepening period from 00 to 12 Z while others have it from 12 to 00 Z (Table 3.2). These diabatic components will be further discussed in Section 5.3.4.

5.3.2 Cyclone Vorticity Generation: Dynamical and Physical Processes

Individual curves shown in panel (a) of Figs. 5.1 to 5.9, contributing to the cyclone vorticity generation or destruction, are further discussed and compared in this section.
The GEO curves (1000 hPa $\Delta \eta_p$) are strongly positive during the development stage and peak at $t_d$ for all categories. Peak GEO magnitudes show very good agreement with storm intensity (ST > MO > WE) in each sector, but are larger in NEA and smaller in NWA sectors for the same intensity class. This tendency is especially clear for ST cases, wherein GEO reaches a maximum of about 7.7, 7.1, and $5.5 \times 10^{-3}$ s$^{-1}$ in the NEA, NCA, and NWA sectors, respectively (Figs. 5.1a, 5.4a, 5.7a). GEO drops to near zero within 24 h after $t_0$ in ST NEA and to a lesser degree in ST NCA, but remains higher for a longer time, to almost to $t_{48}$, in ST NWA. This indicates that the strong NEA and NCA cases cease development and start weakening immediately after $t_0$ while the NWA case continues to deepen, consistent with the composite results. This difference in life cycle, as shown by GEO curves, is not as dramatic among sectors for MO and WE cases, but still noticeable as GEO values always decrease faster toward NEA immediately after $t_0$. As the cyclones weaken, GEO values remain above zero mainly because of positive planetary vorticity tendency ($\Delta f$) since storms also move northward (Fig. 4.1).

The AGEO (1000 hPa $\Delta \eta_{ag}$) component induced by friction, and subsequently the TOT (1000 hPa $\Delta \eta$), are both related to GEO. As a result, the AGEO and TOT curves appear as reverse images of one another, each having reasonably good correspondence among events with different intensities in the same sector (Figs. 5.1a to 5.9a). For cases with similar intensity, peak magnitudes of AGEO and TOT also tend to be greater in the NEA and smaller in NWA sectors, more evident in the ST cases (about $-3.7$ and $-2.8 \times 10^{-3}$ s$^{-1}$ for AGEO, and $4.0$ and $2.8 \times 10^{-3}$ s$^{-1}$ for TOT in NEA and NWA, respectively). Similar to GEO curves, TOT values decline more rapidly after $t_0$ in the NEA sector and in
ST NCA, and more slowly in NWA sector and in weaker NCA cases. The maximum magnitude of TOT (and likely also AGEO and GEO) in ST NWA is somewhat lower than expected (also bottom half of Table 5.1) since the same deepening rate (in Bergeron) at the cyclone center should produce the same increase in geostrophic wind speed (and thus the same \( \Delta \zeta_z \)). One explanation for this is that the NWA storms occur in lower latitudes and have smaller northward movement component near \( t_0 \), so the contribution from \( \Delta f \) is smaller. An additional contributing factor is that the 5 x 5 grid point box for NWA cases covers a larger region, reducing the areal-averaged \( \Delta \eta \) more seriously.

Upper level geostrophic forcing, G500, has slightly larger peak magnitudes in NEA than it does in NWA, reaching its maximum between \( t_{-6} \) and \( t_{-18} \) but tending to be delayed toward NWA. As a result, the magnitude of G500 tends to be greater toward the NEA sector and smaller in NWA (also Table 5.1) before and during the onset of the most rapid development between \( t_{-24} \) and \( t_0 \), consistent with the stronger initial upper level forcing seen in composites toward NEA. After the explosive deepening period, however, G500 tends to remain stronger in the NWA sector, especially for the ST event (Fig. 5.1a). At low levels this produces a longer deepening period during which magnitudes of GEO and thus TOT stay relatively large, in agreement with the mean evolution of sea level storms in Figs. 4.2, and 4.4 to 4.9. The fact that the strength of upper level forcing can be maintained for a longer time in the NWA sector implies that the amount of APE for further storm development should be considerably greater in this region after \( t_0 \), while the rapid decrease in G500 toward NEA after \( t_{-6} \) is consistent with the relatively depleted APE seen in 850 hPa and cross section composites in Chapter 4.
When comparing the two geostrophic components of $\Delta \eta$ at the top and bottom of the air column, G500 and GEO, at least two differences become apparent among the categories. The first concerns the relative timing at which GEO and G500 reach their maxima. Although the coarse temporal resolution of the data set does not allow precise determination of timing of maxima, the relative position of the two features under the GEO and G500 curves can nonetheless be compared. The G500 curves are more in-phase with GEO curves in NE A, tend to shift slightly to the right relative to GEO in NCA, and lag by at least 12 to 24 h in the NWA. This tendency is especially evident for ST events and is related to the stronger negative contribution from the GH curve after $t_0$ toward the NWA sector. Therefore, NWA explosive cyclones tend to have weaker initial upper level forcing but a greater amount of APE and thus stronger baroclinicity. Due to a slower spin-up process it requires more time to produce significant distortion in low level thermal field, leading to a slower upper level wave amplification and a later maximum for G500 curves. NWA cases have a slower life cycle and a longer self-development process, consistent with the composite results, and the weaker initial upper level forcing but stronger baroclinicity suggest that compared with events in NCA and NEA, they are more similar to classical Petterssen's type A cyclones. Toward the NEA, on the other hand, more intense explosive cyclones, are accompanied by stronger initial upper level forcing but overall less APE and weaker baroclinicity, and therefore are more like Petterssen's type B cases. The evolution in G500 and GEO curves (and thus also AGEO and TOT curves) tends to be more in-phase, suggesting a more simultaneous development at upper and lower levels and a more "equivalent barotropic" structure. In these cases the low
level system responds more directly to the upper level forcing in order to maintain thermal wind balance, and they appear to be quite well explained from an IPV viewpoint.

A second difference between GEO and G500 curves concerns their relative peak magnitude and total contribution to development. In the NWA, the G500 peak magnitude and its total contribution tend to be greater than those of GEO (Figs. 5.1a to 5.3a), but in NEA sector the opposite is true (Figs. 5.7a to 5.9a). Difference between GEO and G500 curves are produced by GH, the contribution in 1000 hPa $\Delta \eta_g$ associated with layer thickness changes that are usually positive before $t_0$ but negative afterwards, as mentioned in Section 5.3.1. The GH positive contribution before $t_0$ tends to be larger and long-lasting toward the NEA sector, but smaller and short-lived toward the NWA. The negative contribution after $t_0$, in contrast, tends to be greater in the NWA than NEA. This tendency is again particularly clear in ST cases, where GH contributes positively for a much longer period before $t_0$ toward NEA and the curve drops to below zero only marginally from $t_{-6}$ to $t_{-18}$ (Fig. 5.7a). In the NWA, however, the negative portion of GH after $t_0$ exceeds the preceding positive portion by a significant amount (Fig. 5.1a). The net GH contribution therefore tends to be positive toward the NEA and negative toward the NWA (Table 5.1). There is also a tendency for GH to begin contributing toward the total $\Delta \eta$ (TOT) earlier in the NEA as the curves have positive values as early as $t_{-48}$ for all intensity classes (Figs. 5.7a to 5.9a) while in NWA they only become positive near $t_{24}$ (Figs. 5.1a to 5.3a). Thus, in NEA cases during the early incipient stage (between about $t_{-48}$ and $t_{-30}$) before G500 starts to increase, the thermodynamic component appears to be crucial in maintaining the low level system strength.
The GH curves (the difference between GEO and G500) also indicate the thermal structure of the cyclone. Positive GH values correspond to development toward a warm core structure, and negative values toward a cold core structure. Development of a warm core structure is associated with greater layer warming (or less cooling) at the storm center than over surrounding areas, and the opposite accompanies cold core development. The GH curves indicate that the development toward warm core structure occurs during the explosive deepening period before $t_0$ but this trend reverses afterwards. In the NEA, the overall positive contribution from GH is in good agreement with warm core structure development discussed in Chapter 4, especially for the ST composite (Fig. 5.7a), while in the NWA composite warm core structure is much less prominent and corresponds to very little or even negative overall contribution from GH curve (Figs. 5.1a to 5.3a).

The total G500 contribution toward GEO in all nine categories exceeds that from GH by a substantial amount both for the entire development and for the rapid deepening stage prior to $t_0$, therefore confirming that explosive cyclogenesis is fundamentally driven by baroclinic instability regardless the development intensity. The peak G500 magnitude and its total contribution also correspond well between ST, MO, and WE events in all three sectors, especially in NWA (Figs. 5.1a to 5.3a), suggesting that stronger events tend to be associated with stronger upper level forcing. The positive GH contribution before $t_0$, however, also corresponds to the development intensity reasonably well in all sectors, particularly so in the NEA (Figs. 5.7a to 5.9a), suggesting that the early system strength plays a greater role in determining maximum deepening rates toward the NEA. If the development process is viewed using "IPV thinking" as a nonlinear interaction between
upper and lower level circulations, the early GH contribution (accompanied by
development of a warm anomaly) would enhance the low level PV anomaly, lower the
stability, and promote a more active interaction in which the two components simply
reinforce each other more effectively and lead to a greater G500 contribution. Thus, given
the good correspondence of both G500 and GH to development intensity, it appears that
the maximum deepening rate is determined in general by the strengths of both upper level
forcing and low level system prior to the rapid deepening stage, with the latter becoming
more important toward the NEA sector but less important in NWA.

5.3.3 Thermodynamic Components: Advection, Adiabatic and Diabatic Processes

The curves of averaged 12-h total $\Delta T$ (TOT) in Figs. 5.1b through 5.9b have
shapes qualitatively similar to the GH curves in (a). This similarity suggests that warm or
cold core thermal structure development at the cyclone center can reasonably be inferred
from the temperature change alone. The temperature TOT curve in NEA shows a 48-h
period of significant warming before $t_0$ and a brief cooling period afterwards (Fig. 5.7b)
among the three ST composites. The NWA has only a short warming period before $t_0$ and
a significant cooling period afterwards (Fig. 5.1b). The NCA early warming and
subsequent cooling before and after $t_0$ are of comparable magnitude (Fig. 5.4b). The
magnitude of TOT tends to be quite small for the MO and WE events, while the tendency
of an overall positive contribution from GH in the NEA and negative contribution in the
NWA does not necessarily correspond to an overall NEA warming and NWA cooling.

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The total warming before $t_0$ in ST NEA is considerably stronger than in MO and WE, and the same condition applies to NCA events (Table 5.1). In the NWA, however, the TOT curves in the three intensity classes appear quite similar, as does the positive GH contribution before $t_0$. Overall, the thermal contribution appears to agree with GH curves, and stronger NEA cyclone development is associated with more significant warming in lower to middle troposphere during the incipient stage.

The total $\Delta T$ is broken down to temperature changes from three processes: horizontal advection (ADV), vertical transfer and adiabatic warming or cooling (VERT), and all diabatic effects (DIA). The advection term is positive throughout the period, a reflection of continuous storm movement toward strongest WAA regions. The mean contribution from ADV inside the $5 \times 5$ grid point box is never greater than about 2.5 K for any 12-h interval, confirming that thermal advection is weak near the storm center. ADV is considerably larger in magnitude in the NWA sector than in other sectors (Table 5.1), but this is likely partly related to the larger box size and inclusion of larger area of stronger advection. Nonetheless, it is consistent with its greater amount of APE and stronger baroclinicity in the lower and middle troposphere. The tendency for ADV to peak later at $t_{30}$ or $t_{48}$ (although not as clear in MO case) also agrees with the slower evolution of NWA cyclones. The warming effect of ADV is in general smaller in the NEA and NCA, but tends to reach its maximum early at $t_{30}$ or $t_{48}$, especially in ST and MO cases. This warming, although limited, contributes toward positive TOT values and helps maintain the incipient disturbance. In Chapter 4, the incipient cyclones toward the NEA are seen to develop within a narrow zone with enhanced height and thermal gradients at
850 hPa (Figs. 4.10 to 4.15). Such a configuration suggests that once the incipient disturbance reaches certain strength, and the height gradient is enhanced immediately south or southeast of its center, it tends to maintain itself by advecting more warm air to the region immediately downstream (to its east).

The term VERT represents the effect associated with vertical motion without condensational heating or evaporative cooling. From Eq.(5.10) or (5.23) its value is determined by vertical velocity (i.e., strength of transverse circulation) and dry stability. It strongly cools the air column inside the grid point box, producing at times an averaged temperature drop exceeding 6 K per 12 hours. Since ADV is greater than TOT nearly all the time (except at $r_{18}$ and $r_{28}$ in ST NEA) the cooling from VERT is greater than the warming from DIA. Thus, even with the effect from LH in (c) added to VERT, the total effect associated with vertical motion would still be negative, indicating an overall conditionally stable environment inside the box, at least during rapid development when DIA is greater than LH in all cases. The effect of VERT is also related to horizontal advection, but its magnitude is considerably greater than ADV, indicating the region is mostly dominated by upward motion. In individual sectors, the amplitude of VERT curves corresponds well with intensity classes, since the stability is roughly comparable as shown in Fig. 4.54 and thus the VERT values more or less indicate the strength of transverse circulation. This agreement appears to be better in the NEA and NCA sectors and not as good in NWA, where the difference in VERT curves is relatively small between the three intensity classes. When curves in three sectors are compared, it is clear that the negative effect from VERT is the least in the NEA, intermediate in NCA, and the
strongest in NWA (Figs. 5.1b to 5.9b, and Table 5.1). The difference is so evident that even WE NWA has a VERT curve comparable to ST NCA, which in turn has a greater amplitude than ST NEA. This coincides well with the much lower static stability toward the NEA region during the explosive deepening stage at $t_{12}$ and $t_{0}$ (Fig. 4.54). The incipient stage events toward the NEA sector are associated with a stronger upper level PV anomaly, a lower tropopause and stronger low level system, together with stronger upper level divergence and net mass flux (Chapter 4, Figs. 4.50 to 4.53). This is consistent with a larger G500 magnitude and more contribution from GH (and $\Delta T$ TOT) during the same stage, promoting early destabilization of the atmosphere and leading to low static stability and a reduced negative effect from VERT during the most rapid development. Note in the ST NEA and NCA cases the negative VERT effect from $t_{-54}$ to $t_{-18}$ is considerably stronger than in the NWA, indicative of a more active transverse circulation early on. For the NWA sector, composite results also suggest that the low level system is not initially stronger in ST, and this is supported here by the greater VERT magnitude at $t_{-42}$ and $t_{-30}$ in the MO case (Figs. 5.1b and 5.2b).

DIA is the total 12-h $\Delta T$ from all diabatic effects, a major contributor in producing positive TOT before $t_{0}$ (also bottom half of Table 5.1). In ST composites, the averaged 12-h temperature change can reach almost 5 K. Comparing the three ST cases, most of the contribution from DIA in NWA is after $t_{-6}$ (Fig. 5.1b), and in contrast more is before the same time in NEA (Fig. 5.7b) while the contribution is greater prior to this time in NEA (Fig. 5.7b) and is evenly distributed around $t_{-6}$ in the NCA (Fig. 5.4b), although their total contribution seems comparable. For MO and WE cases, a similar tendency for the
timing of DIA contribution is again apparent but the total contribution becomes increasingly larger toward NWA (Table 5.1), mainly related to more abundant moisture available for condensational heating (LH) to the south. DIA curves for the NEA exhibit very good agreement with intensity classes, with ST having significantly greater contribution from DIA than MO and then WE (Figs. 5.7b to 5.9b). This dependency is less apparent in the NCA, particularly between MO and WE events (Figs. 5.4b to 5.6b). The three curves are surprisingly similar in the NWA and the DIA contribution does not appear to be greater in the ST event (Figs. 5.1b to 5.3b). Therefore, agreement between intensity of development and contribution from DIA is better toward NEA, similar to the situation for VERT curves. This is not surprising because both processes are related to the strength of the transverse circulation, suggesting that the initial low level system strength (and warm anomaly) does appear to be more important toward the NEA for stronger development and less important in the NWA. The incipient system is maintained mainly by the diabatic heating effect, but also by the warm advection.

5.3.4 Diabatic Processes: Latent Heating, Sensible Heating and Radiation

The total contribution from all diabatic effects (DIA) is comparable in the three ST cases. DIA shows a dependency on intensity level in NEA such that its contribution is significantly less in WE cases. In NWA there is no similar dependency such that it contributes nearly the same in WE as in ST events. The DIA contribution is therefore considerably larger toward the NWA for MO and WE events. The total diabatic effect is
broken down to the effects from three different physical processes: radiation (RAD), latent heating (LH), and sensible heat flux plus all other subgrid scale effects not reflected at grid points (SH). As mentioned in Section 5.3.1, the averaged effect from RAD is very limited, with peak magnitudes less than about 0.3 K for any 12-h interval (Figs. 5.1c to 5.9c) and total net contribution nearly zero (Table 5.1). Thus, the DIA term is almost entirely the combined effect from LH and SH.

The LH curves are always positive and represent the averaged warming effect of condensation at the resolvable scale inside the 5 × 5 grid point box where the vertical motion is predominantly upward. They exhibit more gradual changes compared to DIA and SH curves, and can reach peak magnitude of about 3.5 K per 12 h (Figs. 5.1c to 5.9c). For the entire period, the total LH contribution exceeds that from SH by a significant margin (Table 5.1) and is the major contributor to DIA. For individual 12-h intervals, LH also produces more warming than SH in all cases, except for ST NEA during a 22 h period from about \( t_{-32} \) to \( t_{-6} \) (Fig. 5.7c). The warming effect from latent heating has a tendency to reach maximum at \( t_{-6} \), shortly after the most rapid deepening period when vertical velocity (and thus the magnitude of VERT) has become significantly larger, rather than between \( t_{-12} \) and \( t_{0} \). In each intensity class, the LH contribution shows very good dependency on the sectors, and the warming is the greatest toward the NWA and the least in the NEA, especially during the more rapid deepening period. The LH differences in each sector among ST, MO, and WE events are quite limited (Table 5.1). This agrees well with the anticipation that moisture available for latent heat release must
be considerably more abundant toward the southwest in the NWA sector, and least to the northeast in NEA sector (Fig. 3.3).

If the LH curves are inspected more closely, one notices that LH magnitude drops near $t_{18}$ in the four composites of ST and MO in the NEA and NCA (Figs. 5.4c, 5.5c, 5.7c and 5.8c), especially in the ST NEA case. This suspicious feature is more serious when $SH$ tends to be large, and this provides hints of possible underestimations in LH. When the effect of subgrid scale fluxes is large, the validity of the assumption made with Eq. (5.18) in Section 5.1 tends to be weaker. One likely effect is that the $\theta_e$ flux ($\Delta F_e$) would have larger magnitude than the $\theta_u$ flux ($\Delta F_u$), assuming that warmer air generally contains more water vapor. For positive contribution from subgrid scale fluxes ($SH > 0$), when Eq. (5.18) is assumed to be valid a small positive term is left out in the right hand side of Eq. (5.25), and subsequently $\Delta \theta_{ml}$ (and therefore $\Delta T_e$ in Eq. (5.24)) is slightly underestimated. A second source of possible LH underestimation is due to the coarse data set resolution. Because of the loss in features smaller than 2.5° (the current resolution of the ECMWF operational model is about 4 times higher as described in Chapter 3), the advection terms computed at resolvable scale in all three directions in Eq. (5.25), and likely the net difference among them, tend to be underestimated in magnitude. This effect is considered as part of computational error and is likely more serious when $SH$ is large, since it includes the effects of subgrid scale features and nonlinear interaction not reflected by grid points. Numerous modeling studies on explosive cyclogenesis have shown that model resolution is critical in capturing the more detailed structure and true intensity of the storm, as well as its nonlinear deepening process (Sections 2.5 and 2.6).
However, since this reduction in estimated LH is not severe during most of the time period in most cases, Eq. (5.18) in general appears to be a good assumption.

The term SH includes the sensible heat flux, all other subgrid scale effects and the nonlinear interaction not captured by grid points, and computational error. Its value is usually slightly negative before about $t_{-36}$ and after about $t_{-6}$, but positive during the rapid development period so it is helpful in raising DIA to positive values and in the development of a warm core structure (Figs. 5.1c to 5.9c). In the NEA and NCA sectors, SH peaks at either $t_{-18}$ or $t_{-6}$, while in NWA it tends to reach the maximum slightly later, all at $t_{-6}$. Nonetheless, SH curves all peak 12 to 24 h earlier than LH curves. The maximum SH value reaches 3 K at $t_{-18}$ in ST NEA (Fig. 5.7c), while it otherwise is never greater than 2 K. The total contribution from SH is therefore considerably less than LH (Table 5.1) but near its peak SH can become comparable or even exceeds LH for stronger cases toward NEA (including ST, MO, and WE in NEA, and ST NCA). The SH increase during the incipient stage appears to coincide with GEO increase, which represents both low level wind speed (TOT) and convergence (AGEO), consistent with the expectation. Contrasts are large between ST cases during the incipient stage; warming effect from SH is limited in NWA (Fig. 5.1c) but becomes increasingly significant toward the northeast, eventually contributing more toward DIA than LH for a 22 h period before $t_{-6}$ in NEA sector (Fig. 5.7c). The warming from SH in the NWA is never comparable to that from LH (Figs. 5.1c to 5.3c). Although SH is still less than LH in the NCA between $t_{-24}$ and $t_{-6}$, the difference is smaller, especially in the ST NCA case (Fig. 5.4c). In NEA sector, in contrast, SH during the same period is more comparable to LH and even exceeds LH in
ST NEA (Fig. 5.7c). As discussed in previous paragraph, although the magnitude of LH is likely underestimated (and SH overestimated by the same amount) near \( t_{18} \), the relative contribution of SH to LH tends to be considerably higher toward the NEA, especially in ST case, consistent with the higher Bowen ratio at high latitudes.

In the NEA, the contribution from SH between \( t_{24} \) and \( t_0 \) is highest in the strong case (bottom half of Table 5.1) and appears to play a significant role in storm development. During this period, SH is the primary contributor to the high DIA values, which in turn is important in producing a warm core structure and positive contribution from GH that maintains system strength. Since the life cycle of NEA storms is shorter, the initial system strength at low levels (as well as the strength of upper level forcing) becomes crucial in creating a high deepening rate over a short time span. The SH warming effect is from sensible heat transfer from the underlying ocean, and more importantly, from convergence of added latent and sensible heat energy into the storm center by low level winds not resolved by data points both in space and time. Moreover, this process helps destabilize the atmosphere during early stages, which also appears to be important in producing a rapid spin-up and more violent development during the rapid deepening stage in the NEA. It is noted in the previous section that after incipient disturbances reach a certain strength, they are maintained through enhancement of WAA immediately downstream. A similar positive feedback appears to also occur here, in which the incipient system maintains itself by local convergence of latent heat released and surface heat transferred from the ocean near the system center. At the same time, moisture convergence provides an important energy source later as latent heat is released.
upon condensation in ascending air, as shown by the steady LH contribution increase during the rapid deepening stage of all categories (Figs. 5.1c to 5.9c), even though the effect does not immediately contribute toward the system intensification.

5.4 Other Aspects

Section 5.3 presents vorticity budget analysis results for a 5 × 5 grid point box, covering a sizeable area surrounding the storm center for a general description of the evolution of various vorticity and temperature components throughout development in the storm environment. In this section, results for a smaller area near and at the storm center, as well as those for different quadrants relative to the storm center, are described and discussed. The focus is (1) to examine how these components evolve differently directly above the cyclone center and at different quadrants, and (2) to further investigate how consistent some of the findings in Section 5.3 are, when a smaller grid point box is used or when a different quadrant relative to the storm is considered.

5.4.1 Vorticity Budget Near the Cyclone Center

Figures 5.10 to 5.12 present results of vorticity budget analysis for the 1000-500 hPa layer using a 3 × 3 grid point box for ST NWA, ST NCA and ST NEA, respectively. Figs. 5.13 to 5.15 are results for a single grid point (or a 1 × 1 box) located at cyclone center for the same categories. Here, results for ST cases are shown because they exhibit
Figure 5.10: 12-h changes in absolute vorticity ($\Delta \eta$, 10⁻² s⁻¹) and temperature ($\Delta T$, K) averaged over a 3 x 3 grid point box centered at the storm center for 1000-500 hPa layer from $t_{84}$ to $t_{72}$ for the ST NWA composite cyclone. Vorticity and temperature components are the same as in Fig. 5.1. Symbols are plotted in the middle of each 12-h intervals.
Figure 5.11: Same as in Fig. 5.10 but for the ST NCA composite cyclone.
Figure 5.12: Same as in Fig. 5.10 but for the ST NEA composite cyclone.
Figure 5.13: 12-h changes in absolute vorticity ($\Delta \eta$, $10^{-5}$ s$^{-1}$) and temperature ($\Delta T$, K) at the storm center (or a 1 x 1 grid point box) for 1000-500 hPa layer from $t_{-64}$ to $t_{+72}$ for the ST NWA composite cyclone. Vorticity and temperature components are the same as in Fig. 5.1. Symbols are plotted in the middle of each 12-h intervals.
Figure 5.14: Same as in Fig. 5.13 but for the ST NCA composite cyclone.
Figure 5.15: Same as in Fig. 5.13 but for the ST NEA composite cyclone.
the largest contrast among sectors when $5 \times 5$ boxes are used. The $3 \times 3$ box is about 750 \times 750 \text{ km}^2$ while the single grid point represents a $250 \times 250 \text{ km}^2$ area. Note that y-axis range is increased in the figures as smaller box size is used. The amplitude of GEO and AGEO curves increases roughly by about 2.5 times as box size is reduced from $5 \times 5$ to $3 \times 3$, and by another 2.5 times in going to a single grid point. The increase in amplitude is more moderate for $\Delta T$ components, about 1.5 and another 1.3 times respectively for DIA and VERT curves.

The magnitude of all curves related to $\Delta \eta$ increases as the grid point box becomes smaller. The increase in TOT is relatively small however compared to AGEO (which produces low level convergence), such that the ratio between TOT and GEO is slightly reduced from over 50\% to about 40\% as box size goes from $5 \times 5$ (Figs. 5.1a, 5.4a, and 5.7a) to a single point (Figs. 5.13a to 5.15a). The relative G500 increase is gentler than in other curves so the fractional contribution from GH toward GEO ($\text{GEO} = \text{G500} + \text{GH}$) over the whole period becomes larger. The total GH contribution is much smaller than G500 in $5 \times 5$ results (Figs. 5.1a, 5.4a, 5.7a), becomes larger in $3 \times 3$ (Figs. 5.10a to 5.12a), and GH is at least comparable to G500 in the single grid point (Figs. 5.13a to 5.15a). Moreover, GH usually contributes more than G500 during the 12 h of most rapid deepening, indicating the vital role of thermal effects at the cyclone center in producing explosive cyclogenesis. Unlike $5 \times 5$ results in Figs. 10a to 15a, there is almost no negative contribution from GH at any time, clearly indicating the warm core development at the composite cyclone center in all three sectors.

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The total 12-h $\Delta T$ (TOT) in Figs. 5.10b to 5.15b also follows the shape of GH curves closely, suggesting still that the Laplacian of $\Delta T$ can be inferred reasonably well from $\Delta T$ itself. The temperature increase closer to the center during the development period is larger for all sectors as expected, with a peak 12-h warming of over 2 K for $3 \times 3$ box and reaching 5 K at the cyclone center for single point. This actually gives the approximated shape of total $\Delta T$ and its Laplacian, with maximum warming at the storm center and reduced warming at larger radii. Because the composite center is the storm center at the end point of each 12-h interval, this indicates that the system center is constantly moving toward regions of maximum local warming. The largest contributor to this warming is the diabatic effect (DIA) as in $5 \times 5$ results, with peak magnitudes at the center of about 9 K (per 12 h) in ST NWA and ST NCA (Figs. 5.13b and 5.14b) and 12 K in ST NEA (Fig. 5.15b) at the storm center. This enhanced contribution from DIA is largely responsible for the increase in thermal effect in these cases. Effects from ADV and VERT toward total $\Delta T$ are similar to those found earlier with larger grid point box size, with the former producing weak to moderate warming and the latter producing strong cooling. For the diabatic components, the relative contribution by sector is greater from SH, and less from LH, as the box size decreases, in agreement with the larger AGEO magnitude (and thus convergence) at the storm center. The RAD effect remains very limited (Figs. 5.10c to 5.15c).

Comparing results among three sectors reveals differences that are similar to those found in the $5 \times 5$ box analyses. The GEO, AGEO, and TOT curves for $\Delta \eta$ all have greater amplitude in the NCA and NEA sectors than occurs in the NWA, and tend to peak.
earlier (at \( t_{-6} \)) following by rapid weakening. The shape of various \( \Delta \eta \) components is strikingly similar in ST NCA and ST NEA, with peak GEO values near 2.0 and \( 4.5 \times 10^{-4} \) \( \text{s}^{-1} \) and peak AGEO values near \( -1.0 \) and \( -2.7 \times 10^{-4} \text{s}^{-1} \) for the 3 \times 3 box average and at system center, respectively (Figs. 5.11a, 5.12a, 5.14a, and 5.15a). These components in ST NWA peak at \( t_{-6} \) and tend to weaken more slowly afterwards, still suggesting a longer life cycle although G500 curves are no longer delayed as seriously (Figs. 5.10a, 5.13a). Toward the NEA sector, the stronger initial system strength, stronger but shorter-lived upper level forcing, and the more rapid spin-up process are confirmed by larger GEO values between \( t_{-60} \) and \( t_{-24} \), larger G500 values at \( t_{-18} \) and \( t_{-6} \) but smaller after \( t_{-6} \), and the larger GEO values between \( t_{-24} \) and \( t_{0} \), respectively, when they are compared with values in NWA. In NWA sector, the ST case has considerably larger G500 values at \( t_{-6} \) than MO and WE (not shown) for both 3 \times 3 and 1 \times 1 boxes, and confirms that the upper level forcing directly above the cyclone center in ST indeed exceeds those in MO and WE by \( t_{0} \), as suspected in Section 4.5.

In all three sectors and both for 3 \times 3 boxes and at the center, DIA is still the primary contributor to total 12-h \( \Delta T \) changes, especially in NEA after \( t_{-24} \) when the contribution from ADV becomes very limited (Figs. 5.12b, 5.15b). The overall adiabatic cooling effect from VERT is again stronger in NWA, weaker in NCA, and the weakest in NEA, but during the early stage prior to \( t_{-24} \) the opposite is true, indicating stronger initial system strength and early destabilization of the atmosphere toward the NEA. As a result, the negative VERT contribution at the cyclone center is significantly smaller at both \( t_{-6} \) and \( t_{-6} \) in ST NEA (Fig. 5.15b). For ADV, it also tends to peak later at \( t_{-18} \) in NWA with
a larger total contribution (Figs. 5.10b, 5.13b) but much earlier at $t_{30}$ in NEA. ADV shows a significant decrease in the NEA from its early peak (Figs. 5.12b, 5.15b), a reflection of the parent cyclone’s existence to the north or northeast and the strengthening of northerly flow in the vicinity of incipient low (Table 4.2; Chapter 4). The association with a parent cyclone is important in advecting high PV air southward, thereby producing a strong upper and lower level interaction and explosive cyclogenesis. A similar decline in ADV curves is not seen in other sectors, nor does an association with a parent cyclone exist in other sectors.

While the SH contribution relative to LH increases in all sectors as box size decreases, the fractional contribution from SH toward DIA remains highest in the NEA and lowest in the NWA with smaller boxes (Figs. 5.10c to 5.15c). Toward the NEA sector, SH has a particularly high contribution during the development period from $t_{24}$ to $t_0$, when the LH curves decrease to near zero. The magnitude of LH during this period, as discussed in Section 5.3.4, is likely underestimated but by no more than 3 to 4 K (per 12 h), since the largest LH value is less than 5 K at other times when SH is nearly zero. Thus, the warming effect from the low level convergence of latent heat, water vapor, and sensible heat transferred from the ocean, would still contribute considerably more than simply the increase in latent heat release to the development of warm core structure. Compared to DIA curves, the general increase in LH magnitude as the box size is reduced is rather moderate, and may be related to the fact that the strongest ascent is along the warm front slightly to the north or northeast of the surface low center seen in cross sections, instead of at the low pressure center itself.
5.4.2 Vorticity Budget at Different Quadrants of Cyclone

An 11 x 11 grid point box is chosen for the vorticity budget analysis in different quadrants relative to the cyclone center. The box is about 2750 x 2750 km², on the same scale of the mature cyclone (c.f., Figs. 4.10 to 4.15). Each quadrant is one-fourth of the entire box, only slightly larger than the 5 x 5 grid point box. The box is divided by two diagonal lines into four triangular-shaped quadrants lying to the west, south, east, and north of storm center, which are referred to as WQ, SQ, EQ, and NQ, respectively. When numerical values are averaged over a quadrant they are weighted according to the fraction of area represented by grid points that is included in the quadrant, such that the central grid point, points along diagonal lines, and all other quadrant interior points carry weights of 0.25, 0.5, and 1, respectively.

The results for WQ, SQ, and EQ for ST NWA are presented in Figs. 5.16 to 5.18, while those for ST NCA and ST NEA are in Figs. 5.19 to 5.21, and Figs. 5.22 to 5.24, respectively. NQ results are not shown because of their relatively lower significance; the time series curves generally exhibit small changes since most of the quadrant is north of the warm front. Results for MO and WE cases are not shown, again because they exhibit evolution of various components similar to ST cases but with smaller amplitudes. The range of y-axis differs from figures with 5 x 5 boxes, but in general the ΔT components have comparable amplitudes. GEO, AGEO, TOT, and G500 curves for Δη components tend to have smaller magnitudes because quadrant areas are mostly farther away from the
Figure 5.16: 12-h changes in absolute vorticity ($\Delta \eta$, $10^{-5}$ s$^{-1}$) and temperature ($\Delta T$, K) averaged over the western quadrant (WQ) of an $11 \times 11$ grid point box centered at the storm center for $t_{-64}$ to $t_{-72}$ for the ST NWA composite cyclone. Vorticity and temperature components are the same as in Fig. 5.1. See text for the method of averaging. Symbols are plotted in the middle of each 12-h intervals.
Figure 5.17: Same as in Fig. 5.16 but for the southern quadrant (SQ) for the ST NWA composite cyclone.
Figure 5.18: Same as in Fig. 5.16 but for the eastern quadrant (EQ) for the ST NWA composite cyclone.
Figure 5.19: Same as in Fig. 5.16 but for the western quadrant (WQ) for the ST NCA composite cyclone.
Figure 5.20: Same as in Fig. 5.16 but for the southern quadrant (SQ) for the ST NCA composite cyclone.
Figure 5.21: Same as in Fig. 5.16 but for the eastern quadrant (EQ) for the ST NCA composite cyclone.
Figure 5.22: Same as in Fig. 5.16 but for the western quadrant (WQ) for the ST NEA composite cyclone.
Figure 5.23: Same as in Fig. 5.16 but for the southern quadrant (SQ) for the ST NEA composite cyclone.
Figure 5.24: Same as in Fig. 5.16 but for the eastern quadrant (EQ) for the ST NEA composite cyclone.
system center. In contrast GH curves tend to have greater magnitudes mainly due to the
strong thermal advection to be discussed shortly in greater detail.

Figures 5.16 to 5.18 for ST NWA describe some general characteristics in the
evolution from WQ, to SQ, then to EQ that are common to all sectors. GEO values in
WQ throughout the period and in SQ after t₀ are all negative, indicating that 1000 hPa
cyclonic vorticity is being destroyed instead of being generated, since these two quadrants
are to the rear of the low pressure center (Figs. 5.16a, 5.17a). Vorticity is generated
ahead of the cyclone in EQ, especially during periods of significant development (Fig.
5.18a). Relationships between TOT, AGE0, and GEO for all cases are the same as in
previous sections, and thus TOT and AGEO curves will not be further discussed here.
The G500 curves show significant upper level forcing peaking at t₁₈ in WQ, t₆ in SQ, and
t₃₀ in EQ (Figs. 5.16a to 5.18a), an evolutionary sequence in good agreement with the
approaching upper level PV tongue from the west following a hook-shaped path seen on
315 K θₑ surfaces in the composite analysis (Fig. 4.28). The positive contribution from
G500 is generally counteracted by an even larger GH, except in EQ between about t₆ and
t₁₈ when significant warming is taking place. Cyclone vorticity is therefore mostly
destroyed (TOT < 0) in WQ and SQ (except in SQ before t₀) and generated in EQ. The
timing succession of peaks in GH, GEO, and G500 in EQ (Fig. 5.18a) is simply one more
example that downstream from the cyclone the enhancement in upper level forcing is the
result of the self-development process, rather than the cause of it.

The evolution of GH curves can be inferred reasonably well from total ΔT in all
quadrants of all cases (Figs. 5.16 to 5.24). Unlike previous results centered at the storm
center, the warming and cooling from ADV, instead of DIA, is the most important cause for total $\Delta T$ in all quadrants. Significant cold air advection (CAA) peaks at $t_{18}$ or $t_{46}$ in WQ and after $t_6$ in SQ, with maximum cooling reaching almost 6 K per 12 h (Figs. 5.17b, 5.20b), a magnitude much greater than in 5 x 5 results. In EQ, WAA peaking at 3 to 4 K per 12 h dominates the ADV, which follows closely the TOT $\Delta T$ curves. The effect of VERT varies from strongly negative to slightly positive (about -5 to +1 K per 12 h), but tends to be significant only in EQ where it counteracts the WAA (Figs. 5.18b, 5.21b, and 5.24b). DIA usually contributes positively, especially during the development period after about $t_{18}$, but the peak magnitude is only between 2 to 3 K (per 12 h), considerably smaller than in 5 x 5 results. Nonetheless, it reduces the cooling effect of CAA in WQ and SQ (e.g., Figs. 5.16b, 5.17b) and counteracts the VERT effect in EQ (e.g., Figs. 5.18b, 5.19b), sometimes being nearly balanced with the latter (e.g., Fig. 5.24b). In WQ and SQ, the DIA effect is raised mainly because of the latent heat release (LH) near the cold front and increase in SH as much colder air passes over the warmer ocean surface. In EQ, the primary contributor to DIA is LH associated with the warm front. The stability in the three quadrants can be roughly compared by adding the effects of VERT and LH. The WQ and SQ to the rear of the cyclone generally show conditional neutrality or instability. This is likely also related in WQ to the method of producing composites, since the low center is typically moving across this quadrant during each 12-h intervals (c.f., Fig. 4.1). Part of the low stability in SQ may be artificially produced, but cross sections in Chapter 4 do consistently show potential instability in the region. LH is usually not large enough to completely offset VERT in EQ, suggesting that EQ has higher static stability than SQ.
Differences also exist among the sectors. For the upper level forcing, the time needed for G500 curves to peak in WQ, SQ, then EQ tends to be shorter in NEA and considerably longer NWA (Figs. 5.16a to 5.18a, 5.22a to 5.24a). This confirms the faster approach of upper PV tongue prior to $t_0$, and a more rapid evolution after $t_0$ toward NEA discussed in Chapter 4. While the explosive deepening is taking place at $t_{-6}$, the G500 value in SQ is largest in NEA and smallest in NWA, consistent with the composite result that the leading edge of high PV air on the 315 K $\theta_w$ surface moves farther south in the NEA than in the NWA sector (Figs. 4.28, 4.30, 4.32). At lower levels, GEO values in NCA and NEA (Figs. 5.21, 5.24) are considerably larger than in the NWA in EQ at $t_{18}$ and $t_{36}$ (Fig. 5.18), again indicating stronger incipient system toward the NEA. GH and TOT $\Delta T$ in EQ are also larger toward the NEA in the stage prior to about $t_{18}$, and mainly result from the effect of ADV supplemented by DIA, confirming that the incipient low tends to maintain itself through enhanced WAA immediately downstream. Similar tendencies seem to occur in SQ between $t_{48}$ and $t_{24}$, but with contribution from DIA comparable to that from ADV. The faster spin-up process and life cycle toward the NEA sector are again evident, as exemplified by the shorter time needed in the NEA for thermal advection to peak in both SQ and EQ. The negative VERT effect is strongest in the NWA and weakest in the NEA both in SQ and particularly in EQ. This is not so in WQ because stronger CAA toward the NWA tends to produce stronger adiabatic warming. The temperature advection strength in all three quadrants is the strongest in the NWA and weakest in the NEA, agreeing with the contrast in overall baroclinicity and APE among the sectors.
Figure 5.25 shows the strength of WAA and CAA averaged over a 9 × 9 box for all nine categories, with positive and negative values averaged separately. It is clear that toward the NWA, the WAA and CAA are stronger but tend to reach their peak magnitude later near $t_{-12}$, while toward the NEA they are significantly weaker but tend to peak earlier, during the most rapid deepening between $t_{-12}$ and $t_0$. In each individual sector, the CAA strength corresponds well to the intensity of development during a period of 36 to 48 hours near $t_0$. Such an agreement is not as good for WAA, however, except for the NEA between $t_{-36}$ to $t_{-12}$ when stronger WAA is helpful in maintaining the strength of incipient low, an important factor in producing stronger development.

5.5 Conclusion

Results of the vorticity budget analysis of Sections 5.3 and 5.4 are summarized. They are in very good general agreement with results of composite analyses in Chapter 4, and provide quantitative evidence and additional insight into the causes of the observed phenomena. It is confirmed that explosive cyclogenesis is fundamentally driven by baroclinic instability, regardless of the intensity of development, but at a smaller scale near the cyclone center the thermal effect in changing layer thickness during the explosive deepening stage is at least comparable to the 500 hPa geostrophic forcing. The major contributor to this lower to middle tropospheric warming is from diabatic heating, since horizontal advection is limited near the cyclone center (but stronger at larger radii). This diabatic heating is enhanced mainly through the convergence of latent heat released and
Figure 5.25: Mean strength of warm and cold air advection (WAA and CAA, 10^-5 K s^-1) between 1000 and 500 hPa averaged over a 9 × 9 grid point box centered at the cyclone center from t_s4 to t_T2 for (a) NWA, (b) NCA, and (c) NEA sectors. WAA is plotted as solid lines, and CAA plotted as dashed lines. Curves are thick for ST, intermediate for MO, and thin for WE cases, respectively. Values for WAA (CAA) are obtained by averaging all grid points inside the box that have positive (negative) advection.
sensible heat transferred from the ocean by low level ageostrophic winds during the spin-up process, not solely through an increase in latent heat release at resolvable scales. Since low level convergence is strongest near the storm center, diabatic heating has the exact shape needed for efficient warm core development thus contributing significantly to rapid deepening, in a way similar to the tropical cyclone development.

Comparing composite NWA explosive cyclone cases with those of the same intensity class in NCA and NEA, they have weaker upper level forcing between $t_{-24}$ and $t_0$, as well as weaker initial low level system strength, reflected in the amplitude of $G500$, GEO, and TOT $\Delta \eta$ curves. These NWA are broader in shape, indicating a longer cyclone life cycle with slower spin-up and spin-down processes. Longer life cycle is consistent with weaker forcing, a smaller Coriolis parameter, stronger overall baroclinicity and greater amount of APE, stronger static stability and therefore less efficient APE to KE conversion, as demonstrated by stronger but later-peaking cold and warm air advection in the quadrants and stronger adiabatic cooling effects associated with vertical motions near the storm center. The combination of initially weaker upper level forcing but stronger baroclinicity suggests that NWA cyclones are more similar to Petterssen’s classic type A cyclones. In agreement with Chapter 4, low level system strength (GEO) and contributions from the thermal effect (GH) during the early incipient stage do not appear to play a particularly important role in determining subsequent maximum deepening rates achieved by NWA cyclones. Instead, deepening rate is more related to strength of upper level forcing between $t_{-12}$ and $t_0$ and the environmental baroclinicity, as stronger cases in NWA tend to be associated with larger $G500$ values above the developing center and
stronger temperature advection near \( t_0 \). The contribution of thermal effects, although significant to deepening at \( t_0 \) at storm center, are relatively weak during the incipient stage in NWA than in other sectors, consistent with a less pronounced warm core structure. Also, latent heating contributes relatively more toward the total diabatic effect, consistent with more abundant moisture and a lower Bowen ratio in the NWA.

Toward the NEA sector, in contrast, many characteristics observed in NWA cyclones are reversed. Explosive NEA cyclones tend to be associated with \( \Delta \eta \) curves with greater amplitude, and thus stronger upper level forcing (G500) between \( t_{24} \) and \( t_0 \) and early initial system strength (GEO and TOT \( \Delta \eta \)). The evolution in upper level forcing in the quadrant analysis agrees well with the rapid approach of high PV air following a hook-shaped path relative to cyclone center (on the 315 K moist isentrope). The early low level vorticity generation is consistent with early appearance of distinct PV anomalies in the cross sections. In keeping with the rapid spin-up and spin-down processes the generation of cyclone vorticity ceases immediately after \( t_0 \), especially in the ST cases. Rapid evolution agrees well with stronger forcing, a larger Coriolis parameter, weaker overall baroclinicity and static stability (and thus more efficient APE to KE conversion and rapid depletion of APE) as indicated by weaker advection with earlier peaks, and the weak adiabatic cooling effect from vertical motion in the extreme Northeast Atlantic sector. Weaker overall baroclinicity but stronger upper level forcing correspond closely to Petterssen's type B cases. Unlike NWA systems, initial system strength correlates well with intensity of development and therefore appears to also be a factor, along with upper level forcing strength, in determining the subsequent maximum deepening rate. The
contribution from thermal effects is important both in vorticity generation during explosive deepening and in maintaining system strength at low levels (and its associated PV anomaly) during incipient stage. The lengthy NEA thermal component contribution (GH) agrees well with the more pronounced warm core structure, produced largely from strong diabatic heating near the cyclone center. In the NEA, especially for ST events, a large portion of the diabatic heating comes from the subgrid scale convergence of latent heat released, sensible heat transferred from the ocean surface and water vapor, while the rest is from latent heating at the resolvable scale. This is consistent with the faster spin-up process, less available moisture, and larger Bowen ratio in the NEA. Thus, in Table 2.2 as suggested by Emanuel (1986b), we may add that North Atlantic bombs can be grouped according to the maximum deepening position, while cases near Iceland should be placed farther to the right and those near the North America coast farther to the left.

Sanders and Gyakum's (1980) bomb criterion is based on the increase in geostrophic wind speed. To achieve the same deepening rate in Bergeron, NEA cyclones need a greater MSLP drop than NWA cyclones over a similar time period. Therefore, it is not surprising that NEA storms are associated with stronger upper level forcing and considerably larger outward total mass flux during the explosive deepening stage (Fig. 4.53). For strong development, however, it appears that a large deepening rate (e.g., > 2.0 B) would not be easily achieved without a strong incipient low level system since the time for deepening is short. The incipient disturbance is maintained mainly by diabatic heating associated with local convergence of heat and moisture and the heating produces early destabilization further leading to significant lower storm environment stability during
the explosive deepening period. This promotes efficient APE to KE conversion and rapid cyclogenesis with large deepening rates. From an IPV viewpoint, the low static stability allows nonlinear interaction between upper and lower level PV anomalies, facilitating a more violent development. The two ingredients for strong development in the NEA sector include a stronger initial system associated with local PV maximum and lowered static stability, and upon the timely arrival of a strong upper level forcing associated with PV anomaly.
CHAPTER 6

CONCLUSION AND SUMMARY

Major research findings from the composite analysis (Section 4.8) and vorticity budget analysis (Section 5.5) are summarized. It is confirmed that explosive cyclogenesis is fundamentally driven by baroclinic instability, regardless of the intensity of development or the geographic sector in the North Atlantic. At a smaller scale near the cyclone center, however, the thermodynamic contribution to changing layer thickness, largely from diabatic processes is at least comparable to the contribution from upper level (500 hPa) geostrophic forcing during the explosive deepening stage. The diabatic heating is enhanced mainly through convergence of latent heat released and sensible heat transferred from the ocean by low level ageostrophic winds during the spin-up process, not solely through an increase in latent heat release. Since the low level convergence is strongest near the storm center, diabatic heating has the exact shape needed for efficient warm core development and thus contributes significantly to rapid deepening, in a way similar to tropical cyclone development. This point was not adequately addressed in previous studies, but is demonstrated quantitatively in this study through the vorticity budget analysis (Chapter 5).
Differences in characteristics of the evolution and dynamical and thermal structure of composite explosive cyclones are found between events in different North Atlantic sectors. These differences are consistent in a dynamical sense between levels and among variables, and are found in both the composite and vorticity budget analyses. Also, these findings are new since a comprehensive study on a similar topic did not appear in the literature. Composite explosive cyclones of the Northwest Atlantic (NWA) sector off the coast of North America are associated with weaker upper level forcing immediately before and during the most rapid 12-h deepening compared with events in sectors farther east. This is mainly related to the latitudinal adjustment of Sanders and Gyakum's (1980) bomb criterion, such that a lower net column divergence (and thus smaller SLP fall) is needed for cyclones at lower latitudes to reach the same deepening rate in terms of Bergeron. Explosive cyclones in the NWA also tend to have weaker initial low level system strength, higher static stability, and thus stronger compensation of adiabatic effects during development. Their environment is closest to major land-sea boundaries and therefore tends to have stronger overall baroclinicity throughout a greater depth of troposphere. These factors, together with a smaller Coriolis parameter, lead to slower spin-up and spin-down processes, less efficient APE to KE conversion and slower depletion of APE, and result in a slower evolution and longer life cycle. The NWA cyclones therefore tend to continue developing for a longer time period and/or weaken more slowly after explosive deepening, tending to be associated with stronger but later-peaking thermal advection, while their environment tends to exhibit a smaller reduction in vertical wind shear during development. In agreement with the weaker initial system strength, thermal effects
contribute relatively less during the incipient stage. The development of a warm core structure is less pronounced in the NWA cyclones, with no distinct seclusion in 850 hPa composite temperature fields, likely due to the coarse data set resolution. Latent heating at the resolvable scale contributes relatively more toward the total diabatic effects near the storm center in the NWA cyclones, consistent with more abundant moisture in the marine boundary layer and a lower Bowen ratio in this sector.

In the NWA sector itself, cyclones with a larger deepening rate are associated with stronger environmental baroclinicity and stronger upper level forcing during the 12-h period of most rapid development when significant low level potential vorticity (PV) generation takes place. Before the onset of most rapid deepening however, a distinct low level PV anomaly cannot be seen in the composite cross sections, and the strength of the incipient disturbance is not relatively comparable to the storm intensity class, appearing to be unimportant in determining the subsequent maximum deepening rate. It seems therefore that the self-development process in the NWA cyclones is fundamentally the same across all three intensity classes, with components becoming progressively stronger in more violent cyclogenesis, owing to the stronger environmental baroclinicity.

Composite cyclones in the extreme Northeast Atlantic (NEA) sector near Iceland, when compared with NWA, are accompanied by stronger upper level forcing and tend to be stronger during the incipient stage, with a distinct low level PV anomaly visible at least 24 h prior to the onset of most rapid deepening. The static stability in the storm environment is considerably lower during development, so the upward motion only weakly counteracts the storm intensification. The overall baroclinicity is weaker, on the other
hand, with shallower but concentrated horizontal temperature gradients in a relatively confined zone near the system center before rapid deepening. These factors, along with a larger Coriolis parameter at higher latitudes, lead to rapid spin-up and spin-down processes, more efficient APE to KE conversion and earlier APE depletion, and therefore a faster evolution and shorter life cycle. The NEA storms, especially stronger ones, consequently tend to stop further development immediately after the explosive deepening stage and start to weaken, with rapid deterioration in temperature gradient and vertical wind shear, while their associated thermal advection is weaker and tends to peak before or near the end of the explosive deepening period.

Explosive cyclones in the NEA sector are associated with stronger upper level forcing and divergence since a greater SLP drop is required at higher latitudes for the same deepening rate in Bergeron. The forcing on the 315-K moist isentropic surface relative to the cyclone center appears as a tongue of high PV air following a “hook-shaped” path that approaches and eventually overtakes the surface low. Both the PV tongue and its leading edge are stronger in the NEA sector, largely due to proximity to the stratospheric high PV reservoir in the polar region. Because NEA explosive cyclones have a shorter life cycle and limited time for upper and lower level interactions before APE is depleted, the initial low level system strength becomes important together with upper level forcing strength in determining the subsequent maximum deepening rate achieved. During the incipient stage, the low level disturbance is maintained by diabatic heating associated with local convergence of heat and moisture and by enhanced warm air advection, which are both related to upper level divergence. This thermal contribution in
cyclone vorticity generation is also greater in stronger NEA events, producing early destabilization and further leading to significantly lowered static stability during the explosive deepening period. The weak stability promotes efficient APE to KE conversion and interaction between upper and lower PV anomalies, facilitating a more rapid development. Consistent with the lengthier and greater contribution from diabatic processes, NEA storms tend to develop a more pronounced warm core structure. A large portion of the diabatic heating in the NEA sector, especially for strong events, is from the low level convergence of latent heat released and sensible heat transferred from the ocean, while the rest is from latent heating at resolvable scales. This is consistent with the faster spin-up process and a larger Bowen ratio in higher latitudes.

It therefore appears that extreme explosive cyclogenetic events in the NEA sector require both a strong upper level forcing and a stronger incipient disturbance with lowered stability come together in a timely fashion. Such a favorable condition tends to develop when a “parent cyclone” with appreciable strength is located to the northeast of the incipient disturbance. The upper level northerly flow to the west of the “parent cyclone” helps advect lower stratospheric high PV air farther southward toward the incipient low, thus creating a stronger PV anomaly aloft. At low levels, cold air advection to the rear of “parent cyclone” can produce confluence near its boundary and enhance the local temperature gradient, which in turn provides an environment favorable for the maintenance and intensification of the incipient disturbance.

The composite results obtained in this study are of higher confidence in the NWA sector, and of relatively lower confidence in the NCA and NEA due to the more sparse
data coverage there. The degradation of the quality of ECMWF operational analyses, however, is not expected to be serious since the NCA and NEA sectors lie immediately downstream from the NWA sector and the North American continent where the observations are of better coverage. This is in agreement with the considerably higher explosive cyclogenesis frequencies derived from ECMWF analysis over the Greenland-Iceland region, as compared to previous climatologies utilizing conventional weather maps (Section 3.3). Due to more serious problem of data coverage, and also because explosive cyclones toward the NEA sector at higher latitudes have rapid spin-up and spin-down processes and a much shorter life cycle, they tend to be missed in manual analyses if ship reports at the key location and time are not available. Therefore, while it is clear that violent developments do exist in the Greenland-Iceland area (Section 2.1), the frequency maxima in the North-central and extreme Northeast Atlantic were not captured in earlier climatological studies.

An additional point emerged from this study concerns the Bergeron criterion (Sanders and Gyakum 1980) widely used to define explosive cyclogenesis. The criterion is more related to the increase in geostrophic wind speed, and less to the sea level pressure drop itself. While the author does not question its appropriateness given the practical concern of dangerous high seas induced by winds over the ocean, it is necessary however, to point out that the deepening rate expressed in unadjusted surface pressure drop (rather than in Bergeron) is more directly related to the strength of dynamical forcing.

Also, results of this study represent general characteristics of explosive cyclones in different North Atlantic sectors and intensity classes, and variations among individual
cases are expected, as discussed in Section 2.5. In the future, statistical tests will be employed to further investigate the case-to-case variability for selected key variables, as well as for various dynamical and physical processes in their contribution to the rapid storm development.
APPENDIX

BEST ESTIMATE AT 925 hPa

When the finite domain \((31 \times 21)\) composite of a variable is made, the values at all levels are based on all cases in that category except at 925 hPa where the values are based on only those cases after January of 1992. For an arbitrary category, assume that \(N, A,\) and \(B\) are the total number of cases, number of cases after, and before 1992, respectively, and thus \(N = A + B.\) For the composite variable under consideration at 925 hPa, assume that \(T, I,\) and \(E\) represent the true value, the interpolated value from 1000 and 850 hPa, and the difference between the two (error), respectively. For cases after 1992, the subscript \(a\) is used, and the interpolated value deviates from the true value such that

\[
T_a = I_a + E_a. \tag{A1}
\]

For cases before 1992, the subscript \(b\) is used, and the true value is unknown and can only be obtained from interpolation. Therefore, it is assumed that

\[
T_b = I_b. \tag{A2}
\]

The correct (true) value at 925 hPa for the category, \(T_n,\) is

\[
T_n = (A T_a + B T_b)/N = (A I_a + A E_a + B I_b)/N. \tag{A3}
\]
The expression \((A \, I_a + B \, I_n) / N\) is equal to \(I_n\), the interpolated 925 hPa value based on all \(N\) cases. Therefore, substitution yields that

\[ T_n = I_n + \frac{A}{N} \, E_a. \]  

Eq. (A4)

The only unknown term at the right hand side of Eq. (A4) is \(E_a\) which must be estimated from \((T_a - I_n)\) since \(I_a\) in Eq. (A1) is not available. Here, it is assumed that the relationship between \(E_a\) and \((T_a - I_n)\) is

\[ E_a = \frac{A}{N} \, (T_a - I_n). \]  

Eq. (A5)

and the reason is twofold. First, \(I_a\) should be closer to \(T_a\) than \(I_n\), so using \((T_a - I_a)\) directly will tend to overestimate \(E_a\). Second, if in the category \(A\) approaches \(N\), \(I_a\) becomes increasingly better estimated by \(I_n\) and \(E_a\) must approach \((T_a - I_n)\). On the other hand, if \(A\) approaches 0, \(E_a\) must also vanish as \(T_n\) approaches \(I_n\). Therefore, by substituting Eq. (A5) back into Eq. (A4), the “best estimate” \(B_n\) of the true value \(T_n\) at 925 hPa is computed as

\[ B_n = I_n + \left( \frac{A}{N} \right)^2 \, (T_a - I_n). \]  

Eq. (A6)

All variables in Eq. (A6) are known. \(A\) is the number of cases after 1992, \(N\) is the total number of cases in the category, \(I_a\) is the interpolated 925 hPa value based on all \(N\) cases, and \(T_a\) is the 925 hPa composite from \(A\) cases.


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