A Case Study of Rapid Cyclogenesis over Canada. Part I: Diagnostic Study

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ABSTRACT

A diagnostic analysis is made for the rapid development of two subsynoptic scale cyclones that coexisted over Canada in the spring season, using the Level IIIb First GARP Global Experiment dataset assimilated by the European Centre for Medium Range Weather Forecasts and National Weather Service operational data. Both cyclones started developing at 0000 UTC 25 April 1979. When development began, these cyclones were separated from each other by a distance of only about 1300 km. Nonetheless, the physical processes leading to their initial cyclogenesis are found to be different. One of the cyclones remained a weak, shallow surface low for about 48 h after it formed in a zone of developing upper-level westerly waves. It started developing only when it drifted into a region of the horizontal advection of upper-level potential vorticity anomaly associated with a strong tropopause fold. In contrast, the other cyclone formed in a region of strong surface frontogenesis caused primarily by the velocity deformation. Once formed, it developed rapidly and propagated within a narrow zone of small Richardson number, suggesting that the cyclone developed due to localized baroclinic instability. Eventually the latter cyclone also moved into the region of upper-level potential vorticity advection and absorbed the former to become a major synoptic-scale cyclone. Its deepening rate came close to that of explosive cyclogenesis, while heating by latent heat release was found to be of secondary importance in the rapid development from the heat budget analysis in the core of the cyclone.

1. Introduction

Surface weather maps of the National Weather Service (NWS) indicate that a small, weak low was located at the northwestern corner of Canada (72°N, 130°W) at 0000 UTC 23 April 1979. The central pressure was 1008 mb. During the next 48 h, it remained weak, while drifting east–southward. However, it started developing rapidly around 0000 UTC 25 April; the central pressure dropped 20 mb during the subsequent 24 h period. For a shorter period of time, the central pressure deepened 16 mb in 12 h. Thus, the low developed into a major cyclone with the central pressure of 984 mb, as it reached the Hudson Bay. A well-defined cold front extended from the cyclone center southward to Texas. The southern portion of this cold front swept the Severe Environmental Storm and Mesoscale Experiment–Atmospheric Variability Experiment (SESAME–AVE) observational network during the SESAME–AVE III period (25–26 April 1979) and was intensively sampled. In fact, this case of rapid cyclogenesis came to the lead author’s attention during the course of a study of this cold front (Ogura and Portis 1982).

The purpose of this paper is to present the result of a diagnostic analysis in order to better understand the rapid cyclogenesis in this event. More specifically, the objectives are to identify the physical processes responsible for rapid deepening and to relate the findings to the contemporary understanding of the rapid cyclogenesis observed in maritime “bombs” and polar lows. Following Bergeron’s definition of a rapidly deepening extratropical low, Sanders and Gyakum (1980) defined explosive deepening as surface pressure falls at a latitudinally adjusted rate of at least 1 mb h\(^{-1}\) for 24 h. The latitudinal adjustment for an equivalent pressure fall at 60°N is achieved by multiplying the actual pressure fall by a factor of \(\sin 60° / \sin \phi\), where \(\phi\) is the latitude of the cyclone center. The cyclone under discussion developed around the latitude of 55°N. Thus, the aforementioned rate of deepening comes close to that of explosive deepening. Climatological studies show that explosive cyclogenesis occurs primarily during the cold season, with maximum frequency in January, and is predominantly a maritime event (Sanders and Gyakum 1980; Roeber 1984). Most explosive deepeners intensify within or just poleward of the region of maximum baroclinicity, indicating that baroclinic processes play a significant role in explosive cyclogenesis. This is particularly true when the lower troposphere is weakly stratified because of strong fluxes from the ocean surface (e.g., Anthes et al. 1983; Orlanski and Polinsky 1984; Reed and Albright 1986). In addition, several case studies for explosive cyclogenesis events emphasized the importance of various processes such as strong surface fluxes over the ocean.

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(Bosart and Lin 1984), convection triggered by the fluxes (Bosart 1981), and latent heat release by condensation (Gyakum 1983a,b; Anthes et al. 1983; Reed and Albright 1986; Liou and Elsberry 1987). In the case under discussion, it is hypothesized that surface heating is not an important factor, since the cyclonic development occurred over a continent and a frozen bay. Further, it is hypothesized that latent heat release is of secondary importance in cyclogenesis. Some evidence to support this hypothesis will be presented later when the heat budget in the core of the cyclone is discussed.

The diagnostic analysis in this cyclogenesis study is primarily based on the concept of isentropic potential vorticity (IPV). Much of the basis for the modern cyclogenesis theory was established by the pioneering theoretical and observational work of Charney (1947), Eady (1949), Sutcliffe (1947), Sutcliffe and Forsdyke (1950), and Petterssen (1956). Later, the cyclogenesis theory was formulated in the framework of the geostrophic theory to yield the development equation in which the Laplacian of geopotential tendency is related to the horizontal absolute vorticity advection and the differential thermal advection (e.g., Holton 1979). In recent years, this quasi-geostrophic development equation was restated and generalized in terms of IPV. Beyond that, by virtue of the fact that both potential temperature and IPV are conserved following an air parcel in an inviscid and adiabatic flow, the concept of IPV has been proven to describe the evolution of larger-scale flows in a compact and precise manner. Bleck (1974) described the cyclogenesis process in terms of IPV and later Bleck and Mattocks (1984) stressed the important role that upper-level IPV advection plays in Alpine lee cyclogenesis. Several authors showed that an approaching upper-level high IPV anomaly interacted with a low-level cyclonic circulation (e.g., Uccellini et al. 1985; Uccellini 1986; Reed and Albright 1986; Boyle and Bosart 1986). The reader is referred to an excellent review article of Hoskins et al. (1985, HMR) on the use and significance of IPV maps.

2. Data sources

Data used in this study come from the first version of the First GARP Global Experiment (FGGE) Level IIIb dataset analyzed by the European Center for Medium Range Weather Forecasts (ECMWF) and archived at the National Center for Atmospheric Research. Data are available at the 15 pressure levels: 1000, 850, 700, 500, 400, 300, 250, 200, 150, 100, 70, 50, 30, and 10 mb levels for geopotential height, temperature, horizontal wind components, vertical p-velocity and relative humidity (below 300 mb), and sea level pressure. These data are given at 12 h intervals, with the horizontal resolution of 1.875° lat × 1.875° long. Weather maps issued by NWS are also used. Hereafter, the time and date are expressed using the following format: 0000 UTC 24 April becomes 00Z/24.

3. Synoptic overview

Figure 1 presents the abbreviated version of NWS surface weather maps at 12 h intervals beginning 00Z/24 [synoptic maps times expressed as two digit hours (Z)]. Two separate surface lows are observed and are referred to as cyclones A and B, hereafter. Cyclone A was located at the northwestern corner of Canada (72°N, 130°W) at 00Z/23. It has moved to a position west of the Hudson Bay by 00Z/24 (Fig. 1a), moving southeastward. The birthplace of cyclone B is along a stationary front that extends southeastward from Washington and then northeastward to Minnesota (Fig. 1a). At 12Z/24 (Fig. 1b), there is bulge in the stationary front over the border between North Dakota and Minnesota. Cyclone B is born as closed isobars with the central pressure of 1004 mb at the northern border between Minnesota and Canada by 00Z/25 (Fig. 1c). Meanwhile, cyclone A has moved to the western coast of the Hudson Bay, but its central pressure is not yet deep (1005 mb). At this time (Fig. 1c) the two cyclones are separated about 1300 km from each other.

From 00Z/25 on, both cyclones start developing rapidly, but take different courses of movement; cyclone A moves eastward relatively slowly, while cyclone B moves rapidly northeastward (Fig. 1d). They merged by 18Z/25 at a point over the eastern coast of the Hudson Bay and become a major cyclone with a central pressure of 984 mb at 00Z/26 (Fig. 1e). A cold front that extends south-southwestward from the Great Lakes area in Fig. 1e is analyzed by Ogura and Portis (1982) using the SESAME–AVE data and simulated by Orlanski et al. (1985) and Juang (1988) using regional-scale prediction models. Combined cyclone A/B moves northward, without appreciable deepening of

![Fig. 2. Temporal variations of the central pressure for cyclones A and B based on the NWS surface weather maps.](image-url)
the central pressure (Fig. 1f). The cold front now extends from the center of the cyclone southward to reach Texas.

Time series of the central pressure for both cyclone A and B at sea level are shown in Fig. 2, based on NWS surface weather maps. The rapid deepening occurred in the 24 h period from 00Z/25 to 00Z/26 for both cyclones A and B.

Aloft, at 00Z/25 (Fig. 3), a 500 mb trough extends over central Canada, aligned in the north-northeast/south-southwest direction. A strong baroclinic zone is in place in association with a meandering polar jet.

Fig. 3. Geopotential height (solid lines) with contour intervals of 50 m and isotherms (dashed lines) with contour intervals of 2°K at 500 mb and at 0000 UTC 25 April 1979.

Fig. 4. As in Fig. 3, but at 1200 UTC 25 April 1979.
There is also a broader baroclinic zone that crosses over California and then turns to the northeast to merge with the polar baroclinic zone on the east side of the trough axis. By 12Z/25 (Fig. 4), the 500 mb trough has undergone substantial development. The amplitude of the system has increased and the distance between the trough and the downstream ridge has decreased, indicative of increasing divergence in the southwesterly flow aloft (Boyle and Bosart 1986; Uccellini et al. 1984). Another noteworthy feature in Fig. 4 is the

Fig. 5. As in Fig. 3, but at 250 mb and 1200 UTC 25 April 1979. Shading indicates regions where temperature > 226 K.

Fig. 6. Wind vectors and isotachs with contour intervals of 5 m s\(^{-1}\) at 300 mb and at 1200 UTC 25 April. Shading indicates regions where wind speed > 40 m s\(^{-1}\).
strong warm air advection (as indicated in the out-of-phase relationship between the temperature and geopotential fields) over the Hudson Bay where cyclone A is developing. An important feature of cyclone B is that it develops and propagates within a baroclinic zone that extends from North Dakota northeastward to the southern coast of the Hudson Bay. The significance of this fact will be discussed later.

At 250 mb (Fig. 5), the warm (cold) core structure of the trough (ridge) is evident; the local maximum temperature in the trough is 230 K, whereas the local minimum temperature in the ridge is 216 K. As will be shown later, much of the warm air in the 250 mb trough is stratospheric. Downstream of the 250 mb warm pocket at 12Z/25, there is strong warm air advection overspreading the surface low center A. Figure 6 shows the wind field at 300 mb and at 12Z/25. An important feature is the wind maximum, which extends southwestward from Hudson Bay to South Dakota. When compared with Fig. 1, it is evident that cyclone B developed and propagated beneath this strong southwesterly jet.

4. Evolution of cyclone A

In section 3, we noted that cyclone A was identified as closed isobars on the NWS surface weather maps as early as 00Z/23, but it started developing two days later at 00Z/25. In Fig. 7, the maximum vertical component of relative vorticity associated with cyclone A is plotted in a height–time section. It is important to note that vorticity has its local maximum at 850 mb in the lower troposphere in the predevelopment period, indicating that the incipient cyclone A is a shallow system. However, as the central pressure starts deepening at 00Z/25, the magnitude of vorticity increases not only with time but also with height. This indicates that cyclone A has developed into a deep system.

A discussion regarding possible mechanisms responsible for the formation of the incipient cyclone A (prior to 12Z/23) is beyond the scope of this study.

Our interest lies in describing the transition period from a shallow system to a deep one. To do so, a standard isentropic analysis was made, including computations of isotropic potential vorticity, Montgomery stream-function (M), static stability, and geopotential height (Φ) on each isentropic surface at 5°K intervals from potential temperature θ = 260 K to 360 K for the period from 12Z/23 through 00Z/27. Here isotropic potential vorticity (IPV) is calculated as

\[ IPV = -g(\zeta + f)/(d\rho/d\theta) \]  

(1)

Following HMR, 1 potential vorticity unit (1 PVU) = 10^{-6} m^2 s^{-1} K kg^{-1} will be used. Further, the tropopause is defined as a surface of 2 PVU.

Figures 8 and 9 show several variables on θ = 310 K surface at 00Z/24 and at 00Z/25 to emphasize the differences between the predevelopment period and at a time near the initial development. We observe that surface cyclone A is located in the proximity of the center of a high IPV anomaly (or under the northeastern edge of the steep tropopause slope) at both times. The relative distance between the center of the high IPV anomaly and cyclone A remains about the same. The critical difference lies in the position of cyclone A relative to the region of IPV advection. At 00Z/24 (Fig. 8a), surface cyclone A is in place outside the IPV advection region. In fact, the wind is very weak in the region over cyclone A and the isentropic surface is extremely flat (Fig. 8b). In short, there is no sign of upper-level support for the development of cyclone A at this time.

In contrast, cyclone A is experiencing the IPV advection at 00Z/25 (Fig. 9a), even though the IPV advection maximum is south of cyclone A. Apparently, shallow incipient cyclone A and the upper-level high IPV anomaly have been steered by prevailing winds at different levels. The superposition of the wind field (Fig. 9a) and the M-field (Fig. 9b) indicates that wind in that region is nearly parallel to the M contours, suggesting that the wind is nearly gradient wind balance. The subsequent development of cyclone A will be discussed in section 6. In passing, it is noted that cyclone B is located away from the region of the IPV advection (Fig. 9a).

5. Formation of cyclone B

It was shown above that cyclone A started developing when the upper-level high IPV spread over it, while cyclone B evolved without upper-level support in the form of IPV advection, at least in its formative stage. We have shown in Fig. 1 that the surface front over the northern United States intensified and eventually cyclone B formed along the front. In order to describe this process quantitatively, the frontogenesis function, as defined in Eq. (2), was computed at 850 mb at 12Z/24 and 00Z/25 using FGGE Level IIIb data in pressure (p) coordinates:

![Figure 7](image-url)  

Fig. 7. Time–height section of the maximum vertical vorticity associated with cyclone A in units of 10^{-3} s^{-1}. The central pressure at sea surface level (SLP) is included at the bottom of the diagram in millibars.
Fig. 8. Isentropic analyses for 310 K at 0000 UTC 24 April 1979. (a) Wind vector and potential vorticity with contour intervals of \(0.5 \times 10^{-4} \text{ m}^2 \text{ s}^{-1} \text{ K kg}^{-1}\). Label A indicates the position of surface cyclone A. (b) Montgomery streamfunction (solid, \(310 = 3.10 \times 10^3 \text{ m}^2 \text{ s}^{-1}\)) and geopotential height (dashed) with contour intervals of 600 m; 60 000 = 6000 m.

\[
\frac{d}{dt} |\nabla_H \theta| = -\frac{1}{2|\nabla_H \theta|} \left( \theta_x^2 + \theta_y^2 \right) (u_x + v_y) + 2\theta_x \theta_y (v_x + u_y) - \frac{\theta_p}{|\nabla_H \theta|} \left( \theta_x \omega_x + \theta_y \omega_y \right),
\]

where \(\nabla_H \theta\) is the horizontal gradient of potential temperature and the subscript \(x\), \(y\), and \(p\) denote partial
derivatives with respect to those variables. The first, second, and third terms represent the effects of horizontal convergence, deformation, and vertical tilting on frontogenesis respectively.

The frontogenesis function at 12Z/24 (not shown) has a local maximum at the northern border between Minnesota and North Dakota with a magnitude of 3.0°C (100 km)^{-1} (day)^{-1}. The deformation effect is predominantly large compared to the other two effects. At 00Z/25 (Fig. 10), the local maximum of the frontogenesis function moved northeastward, with a magnitude of 4.5°C (100 km)^{-1} (day)^{-1}. Among the three effects in frontogenesis, the deformation effect contributes about 2.2 in the same units, the convergence effect 1.3, and the tilting effect 1.0, indicating that the vertical circulation associated with the frontogenesis process is now under way. All three terms have the patterns and locations of their local maxima similar to those of the
frontogenesis function shown in Fig. 10. The low-level convergence also helps spin up the initial cyclonic circulation. Note also that the location of the maximum frontogenesis function in Fig. 10 is very close to that of surface cyclone B shown in Fig. 1c. Considering the coarse horizontal resolution of data used, the frontogenesis function in Fig. 10 could have been underestimated.

Once the incipient vortex formed, two aspects of the synoptic conditions favored the further development of the vortex. First, an isentropic analysis at low levels ($\theta = 295$ K) in Fig. 11 clearly shows that winds cross the contour lines of geopotential at large angles at 00Z/25 over the area where cyclone B developed. This indicates both low-level warm advection and ascent, since the $\theta$-surface did not move rapidly in this case. Second, Fig. 12 shows the horizontal distribution of Richardson number (Ri) computed for the layer between 700 and 850 mb levels at 12Z/25. The important feature is the presence of a low Ri zone oriented in the direction of the track of cyclone B. Figure 13 shows the wind speed normal to the vertical plane and the static stability ($\partial T/\partial p$) on the vertical cross section aligned nearly perpendicular to the tract of cyclone B. It is apparent that the horizontal variation of $\partial T/\partial p$ is small, though $\partial T/\partial p$ is relatively small at low levels. It indicates that the small values of Ri in the zone are due to the strong vertical wind shear.

Figure 14 shows the differences in the structure between cyclone A and B as seen at 12Z/25. Cyclone A has the upright and deep structure; the vorticity maxima for cyclone A are collocated at 850, 700, and 500 mb and their magnitudes increase with height. In contrast, the magnitude of relative vorticity decreases with height for cyclone B, indicating that cyclone B is a shallow system. There is a band of large vorticity that extends southward from the center of cyclone A and then extends westward at 500 mb. However, it reflects the southwesterly jet streak shown in Fig. 6, rather than vorticity associated with cyclone B.

At this point, it is of some interest to compare the present situation to the environmental conditions that were reported to be conducive to polar lows, since there are some common features. With the use of radar data, Harrold and Browning (1969) proposed that a polar low formed in the characteristic shallow frontal zone.
found near Iceland and, citing Eady's theory, attributed the small horizontal size to the shallow depth. They also called attention to the existence of deep baroclinicity and a jet stream in the vicinity of the low but argued that these were unlikely to have affected the development, since the propagation speed was indicative of steering by low level winds.

Fig. 12. Richardson number (Ri) for the layer between 700 and 850 mb at 1200 UTC 25 April 1979 in units of natural logarithm of Ri. Line AB indicates position of cross section in Fig. 13.

Fig. 13. Vertical cross section along the line AB shown in Fig. 12 depicting wind speed (solid lines) normal to the plane with contour intervals of 5 m s⁻¹ and ΔT/Δp (dashed lines) with contour intervals of 2°C (100 mb)⁻¹ at 0000 UTC 25 April 1979. Tick marks on the abscissa are separated 92 km apart.
Stimulated by the work of Harrold and Browning, Mansfield (1974) employed the Eady linear inviscid model to find normal mode solutions for unstable disturbances in a shallow (1.6 km) rigidly bounded layer, using the observed mean state in the Harrold and Browning case. Despite the neglect of release of latent heat by condensation, he obtained realistic growth rates (e-folding times of 1–2 days) and realistic sizes (6–800 km) for the most unstable disturbances. Duncan (1977) applied a linear quasi-geostrophic inviscid model to the observed basic states in three cases of polar lows. He found that the presence of small static stability at low levels in conjunction with the existence of low-level baroclinicity resulted in the formation of shallow disturbances that resembled the observed disturbances in size, growth rate, and propagation speed. Orlanski (1986) approached this problem as an initial value problem, using a two-dimensional, dry, inviscid, primitive model and initializing the model with hypothetical soundings. His conclusions are that a flow can be unstable for mesoscale small amplitude disturbances, if $\text{Ri}$ of the flow is $\approx 1$ and that the resulting waves can be shallow, when $\text{Ri}$ is small at low levels. Reed and Duncan (1987) applied the model of Duncan (1977) to the observed background state in the case of the train of four, more or less evenly spaced polar lows. Their computations yielded a wavelength of maximum instability that was consistent with the observed wavelength of 500–600 km, suggesting a possible baroclinic origin for the wave train. The role of static stability in the development of short baroclinic waves has been studied also by Gall (1976), Staley and Gall (1977), and Blumen (1979).

We did not make a modal or nonmodal linear stability analysis for the base state shown in Fig. 13. However, a comparison of the magnitudes of the vertical wind shear and the static stability shown in Fig. 13 to those in the aforementioned work strongly suggests that the base state in this case is baroclinically unstable. It is then hypothesized that the incipient cyclone B of frontal origin developed by baroclinic instability. The ensuing development of cyclone B will be described in section 6.

For completeness of the discussion, it is noted that diabatic heating due to latent heat release was reported to be as important as baroclinic instability in the development of polar lows. The reader is referred to an excellent review article by Reed (1988) on this subject.

6. Development of cyclone B

a. Structure of the PV anomaly

To illustrate the possible upper-level forcing for the subsequent development of cyclone B, the distribution of tropopause height at 12Z/25 is shown in Fig. 15, where the tropopause is defined as $\text{PVU} > 2$ and potential vorticity (PV) in pressure coordinates is calculated as

$$\text{PV} = -g(\partial\theta/\partial p)(\zeta_p + f).$$

The tropopause is nearly at the 500 mb level in the center of the high PV anomaly. Cyclone A remains located under the sloping tropopause, while cyclone B is further away from it. Another tropopause fold is found downstream of the primary high PV anomaly, with the tropopause reaching the 440 mb level. Figure 16 shows the tracks of cyclone A, cyclone B, and the lowest point of the tropopause, which will be referred to as the center of high PV anomaly. The important features are that the center of the high PV anomaly was traveling nearly eastward during the period from 00Z/25 to 12Z/26 and that cyclone B was moving rapidly north-northeastward to cross the track of the high PV anomaly.

Figure 17 depicts the structure of the high PV anomaly in a vertical section taken along the line in Fig. 15 that crosses the center of the high PV anomaly in the west–east direction. In Fig. 17a, a local maximum of PV at low levels represents the southwestern portion of cyclone A. The distribution of $\theta$ relative to the high PV anomaly is similar to that determined by Thorpe (1986) for a combination of an isolated, balanced, circular PV anomaly at upper levels and a cold $\theta$ anomaly at the earth’s surface (his Fig. 5). The characteristics are that an isentropic surface rises as it approaches a
high PV anomaly below the sloping tropopause and a cold dome resides just beneath the folded tropopause. Thus, the structure shown in Fig. 17a is quite similar to those depicted in explosive cyclogenesis events by Boyle and Bosart (1986) and Uccellini et al. (1985). In Fig. 17b, the upper-level jets encompass the sloping tropopause in both sides of the high PV anomaly, again in agreement with Fig. 5 of Thorpe. A difference is that an anticyclonic circulation at low levels in Thorpe’s case is not seen, since his upper-level cyclonic circulation is much weaker than ours. Figure 17b is slightly misleading in the sense that, as can be seen in Fig. 6, the maximum wind speed of \( \approx 50 \text{ m s}^{-1} \) is present on the east side of the high PV anomaly, while the meridional component of wind is weaker than that on the west side in Fig. 17b.

Figure 17c shows the vertical \( p \)-velocity given in the FGGE Level IIIb dataset and indicates a strong ascending motion on the east side of the eastward traveling high PV anomaly. This ascent is expected from the following reasoning. In an adiabatic process, the vertical velocity \( w \) may be given by

\[
w = \left[ -\frac{\partial \theta}{\partial t} - u(\partial \theta/\partial x) - v(\partial \theta/\partial y) \right] / (\partial \theta/\partial z),
\]

(4)

where \( u \) and \( v \) are the velocity components in the \( x \)- (east–west) and \( y \)- (north–south) directions, respectively. For simplicity of argument, the structure of the high PV anomaly shown in Fig. 17 is regarded as representing an isolated, balanced, circular PV anomaly that is traveling with a speed \( U \) in the \( x \)-direction without changing its shape and is embedded in a sheared westerly environment \( \bar{u}(z) \). In this situation, \( \partial \theta/\partial t = -U(\partial \theta/\partial x) \). Consider a region at low levels on the east side of the PV anomaly on the cross section that cuts through the center of the PV anomaly. In this region, \( U - \bar{u} > 0 \), \( \partial \theta/\partial x > 0 \), \( v > 0 \) and \( \partial \theta/\partial y < 0 \), (thermal wind relation corresponding to \( \partial \bar{u}/\partial z > 0 \)), leading to \( w > 0 \). In their discussion of a balanced PV anomaly, HMR noted that one may think of a moving upper air PV anomaly as acting on underlying layers of the atmosphere somewhat like a broad, very gentle
"vacuum cleaner," sucking air upward towards its leading portion and pushing it downward over the trailing portion. The local maximum of ascent is located at the 500 mb level with \(-4.3 \mu \text{b s}^{-1}\) in Fig. 17c. The resulting vertical stretching of an air column in the lower troposphere helps spin up an existing low-level vortex, cyclone B in this case. Subsidence within the high PV anomaly is weak. This is not surprising since the downward extrusion of high PV air of stratospheric origin occurred several days earlier and the high PV anomaly is now mainly drifting horizontally.

b. Quasi-geostrophic consideration

To supplement the discussion based on the PV anomaly, the tendency equation for geopotential height (\(\Phi\)) is considered, which may be written within the quasi-geostrophic framework in spherical coordinates as (Holton 1979):

\[
\nabla_{\sigma} + f_0 \frac{\partial}{\partial \sigma} \left( \frac{1}{\sigma} \frac{\partial}{\partial \sigma} \right) \frac{\partial \Phi}{\partial \sigma} = -f_0 v_g
\]

\[
\cdot \nabla \left( \frac{1}{f_0} \nabla_{\sigma} + f \right) + f_0 \frac{\partial}{\partial \sigma} \left( -v_g \frac{1}{\sigma} \nabla \frac{\partial \Phi}{\partial \sigma} \right),
\]

where all notations are conventional. The first and second terms of the right hand on Eq. (5) represent the forcings due to the horizontal absolute vorticity advection and differential thermal advection respectively, computed geostrophically. Equation (5) may be rewritten as

\[
\nabla_{\sigma} + f_0 \frac{\partial}{\partial \sigma} \left( \frac{1}{\sigma} \frac{\partial}{\partial \sigma} \right) \frac{\partial \Phi}{\partial \sigma} = -J_{\sigma} \left[ \frac{1}{f_0} \nabla_{\sigma} + f_0 \frac{\partial}{\partial \sigma} \left( \frac{1}{\sigma} \frac{\partial \Phi}{\partial \sigma} + f, \Phi \right) \right],
\]

where all notations are conventional. The quantity \(q\) is

\[
q = \frac{1}{f_0} \nabla_{\sigma} + f_0 \frac{\partial}{\partial \sigma} \left( \frac{1}{\sigma} \frac{\partial \Phi}{\partial \sigma} + f \right)
\]

represents quasi-geostrophic potential vorticity introduced by Charney and Stern (1962). Equation (6) indicates that the two forcing terms in Eq. (5) are combined to represent forcing due to potential vorticity advection, and Eq. (6) is a familiar form of the PV conservation in the quasi-geostrophic system.

As noted by HMR, the PV anomaly appears partly as absolute vorticity and partly as static stability. The proportions in which this partitioning occurs depend on the shape of the anomaly. The point is that such partitioning must occur. This means that a region of anomalously high PV is also a region of anomalously high absolute vorticity. Further, from the coordinate transformation, we obtain

\[
\zeta = \zeta + \nabla \theta \times \frac{\partial \nu}{\partial \theta} \cdot k.
\]

Here \(\zeta\) is relative vorticity on a constant pressure surface, \(\nabla \theta\) is the horizontal gradient of \(\theta\) on a pressure surface, \(k\) is the unit vector in the upward direction, and the remaining symbols are conventional. The second term of the right side of Eq. (8) is positive for a geostrophic flow in which the vertical wind shear is rotated 90 degrees counterclockwise from the horizontal potential temperature gradient. Therefore, a region of anomalously high \(\zeta\) is also a region of anomalously high \(\zeta\).

Figure 18 shows the horizontal distributions of the two forcing terms in Eq. (5) at 12Z/25 at 500 mb. Obviously both forcing terms are positive and collocated over Hudson Bay at 12Z/25, resulting in a doubling of the total forcing in magnitude in 12 h. This is the area cyclone B is approaching. Considering the subsynoptic scale of both cyclones A and B, their evolution may not be adequately described by the quasi-geostrophic system. Nonetheless, the result of isentropic analyses for 12Z/25 is found to be consistent with the quasi-geostrophic result in that there is strong IPV advection over Hudson Bay at 12Z/25 (diagram not shown). This may be inferred from Fig. 16 in that the path of the IPV center is to the eastward of the path of cyclone B in the period between 00Z/25 and 00Z/26. Thus, in addition to baroclinic instability, superposition of the upper-level forcing with the already growing low-level perturbation contributed to the dramatic organization of cyclone B between 12Z/25 and 00Z/26.

c. Quasi-Lagrangian heat budgets

Finally, quasi-Lagrangian heat budget calculations are made to get further insight into the difference in the evolution of cyclones A and B. In this approach, the budget is calculated in coordinates following each of the cyclone centers and the contributions of different processes are examined. Following Johnson and Downey (1975) and Liou and Elsberry (1987), the quasi-Lagrangian heat budget equation may be written as

\[
\frac{\delta T}{\delta t} = -\nabla \cdot \nabla T + \nabla_0 \cdot \nabla T - \omega \left( \frac{\partial T}{\partial p} - \frac{\alpha}{c_p} \right) + Q,
\]

where the double bar denotes a mass average over a budget volume and \(\delta T/\delta t\) is the time derivative of temperature following the cyclone movement. The first term on the right side is the horizontal temperature advection due to horizontal wind, the second term is the horizontal temperature advection due to cyclone translation (\(V_0\) = translation velocity), the third term is the adiabatic cooling rate, and the last term is the diabatic heating rate. To concentrate on the inner core of each cyclone, the budget is calculated for a square grid 540 km x 540 km that is centered at the minimum sea level pressure. Figure 19 shows the result of the calculation for cyclone B (except Fig. 19f). The tem-
Fig. 17. Vertical cross section along the line in the east–west direction shown in Fig. 15 of (a) potential vorticity (solid) with contour intervals of 1.0 PVU and potential temperature (dashed) with contour intervals of 3°K, (b) meridional wind component with contour intervals of 5 m s$^{-1}$, and (c) wind vector composed of zonal wind and the vertical $p$-velocity ($\omega$) where $10 \times (-\omega)$ is used at 1200 UTC 25 April 1979. The vector at the upper right corner of (c) represent 49 m s$^{-1}$ for horizontal velocity and 4.9 μb s$^{-1}$ for $\omega$. Broad dashed lines in (b) and (c) indicate the tropopause defined as a 2 PVU contour.
temperature is continuously decreasing (increasing) with time for the period considered in the layer below (above) the 300 mb level (Fig. 19a). It is noted that, during the 36 h period from 00Z/25 to 12Z/26, the geopotential height of 1000 mb at the center of cyclone B deepened 176 m, while that of 100 mb deepened 1170 m to make up the substantial portion of the decrease in thickness of the layer between 1000 and 300 mb. The maximum temperature tendency is as high as -14°C day⁻¹ located around 750 mb (Fig. 19b). There is the warm air advection in midlevels (Fig. 19c), although its magnitude is at most 2°C day⁻¹. On the other hand, the cooling rate due to the cyclone movement is large (Fig. 19d), with the maximum rate of -10°C day⁻¹ located at 750 mb, reflecting the fact that cyclone B moved rapidly to a cold area. Furthermore, the adiabatic cooling rate is as large as -8°C day⁻¹ (Fig. 19e). The diabatic heating rate was computed as a residue and was found to be at most 2°C day⁻¹ (diagram not shown), suggesting that the effect of diabatic heating is of secondary importance in the development of cyclone B.

A similar heat budget analysis was made for cyclone A and Fig. 19f shows the temporal variation of temperature in the core of cyclone A. In contrast to the case of cyclone B, the temperature increases with time continuously in the layer below 300 mb until 00Z/26, the time when cyclone A merged with cyclone B. After that time, temperature starts decreasing with time, as already shown in Fig. 19a. This strongly suggests that the final cyclone center should be regarded as the development of cyclone B. This may be seen also in Fig. 16 where cyclone B traveled steadily toward the north-northeast before, during, and after the rapid deepening, absorbing cyclone A along the way. The heat budget calculation for cyclone A prior to merging with cyclone B shows that the horizontal temperature advection rates due to both horizontal wind and the cyclone movement are positive and the sum of the two terms is large enough to compensate the adiabatic cooling rate, yielding a positive temperature tendency (diagram not shown).

It is of some interest to compare the result in this study to that in a winter storm that developed explosively over the northern Pacific, and was investigated by Liou and Elsberry (1987). They simulated the evolution of this storm successfully, using the UCLA general circulation model and calculated quasi-Lagrangian heat budgets based on both the ECMWF–FGGE Level IIIb dataset and the model prediction. For the purpose of a comparison to our case, references are made to their prediction result. In their case, the temperature tendency was also generally negative in the layer below the 300 mb level for the period when the surface central
km, and yet the dynamical processes leading to their initial development were different. Cyclone A formed as a shallow system in the polar baroclinic zone. In other words, a developing upper-level short wave and an associated tropopause fold were in place in the proximity of surface cyclone A. Nonetheless, cyclone A remained as a weak, shallow cyclone in the next 48 h period. Cyclone A and the region of upper-level high potential vorticity anomaly moved along different tracks. Only when cyclone A moved into the region of positive potential vorticity advection, did it start developing both in terms of its surface central pressure and its vertical extent.

In contrast, cyclone B developed initially without much support from the upper-level forcing. It was born in the region of strong surface frontogenesis. Once formed, it developed and moved within a narrow zone of small Richardson number. The smallness of its horizontal size is consistent with the expectation from a localized baroclinic instability theory for a flow with a small Richardson number. This process has been identified as one of the important processes conducive to rapid cyclogenesis in the earlier studies of polar lows and winter storms over the ocean. However, small Richardson numbers in the lower atmosphere were achieved by weak static stability caused by strong sensible heat flux from the ocean surface in those maritime cyclones. In contrast, it was achieved by the strong southwesterly jet in this case. Eventually cyclone B also moved into the region of upper-level positive potential vorticity advection and was further intensified by the "vacuum cleaner effect" of a moving potential vorticity anomaly. In the traditional geostrophic thinking, this situation corresponded to the one in which the two forcing terms (absolute vorticity advection and differential thermal advection) in the geopotential tendency equation had the same signs and were collocated, and therefore the forcing was strong. While continuing to move toward the north-northeast, cyclone B absorbed cyclone A and became a major synoptic-scale cyclone. Thus, the present work based on isentropic analyses provides a clear view of the complex cyclonic development in this case.

A quasi-Lagrangian heat budget was calculated for the inner core of the cyclone B. During the developing period in which the central surface pressure deepened 20 mb in 24 h, the temperature kept decreasing with time in the layer below 300 mb. This was accounted for by the adiabatic cooling and the cyclone movement to a cold area. The diabatic heating estimated as a residue in the heat budget equation was found to be small compared to the other dominant terms, suggesting its secondary importance in the development of this continental cyclone. In Part II of this paper (Juang and Ogura 1990), numerical simulations of the rapid cyclogenesis event under discussion were made using a regional, primitive equation prediction model. The result was found to be consistent with the finding from

7. Summary and conclusion

In this work, a diagnostic analysis was made for rapid continental cyclogenesis, using the FGGGE Level IIib dataset prepared by ECMWF. An unique feature of this event may be that two cyclones on synoptic scale coexisted, separated by a distance of about 1300 km.
FIG. 19. Quasi-Lagrangian heat budgets for the inner core of cyclone B (except f): (a) temperature as a function of time and height (pressure) with contour intervals of 2°C, (b) quasi-Lagrangian temperature tendency, (c) horizontal temperature advection rate due to horizontal wind, (d) horizontal temperature advection rate due to cyclone movement, (e) adiabatic cooling rate, and (f) as in (a), but for cyclone A. Contour intervals for (b)–(e) are 2°C (day)^{-1}.
the above heat budget calculations in that a simulation without latent heat release and surface heat fluxes reproduced the rapid deepening of cyclone B fairly well. The model prediction with latent heat release was not significantly different from that without moisture.

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