Interaction of Potential Vorticity Anomalies in Extratropical Cyclogenesis. Part I: Static Piecewise Inversion

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ABSTRACT

The relative importance of various potential vorticity (PV) perturbations and their mutual interactions associated with the superstorm of 12–14 March 1993 are investigated by applying a piecewise PV inversion diagnostic system to a 36-h simulation of the storm. It is shown that the contributions from all PV anomalies to the surface development increase with time, although their relative significance varies during the rapid deepening stage. In general, the upper-level dry PV anomalies contribute the most to the rapid deepening of the storm, followed, in order, by the lower-level thermal anomaly and latent heat release.

Comparing the PV anomalies and their inverted circulations reveals that there exists a favorable phase tilt between the upper- and lower-level anomalies that allows lower- and upper-level mutual interactions, in which the circulations associated with the upper-level PV anomalies enhance the lower-level anomalies and vice versa. In addition to the vertical interactions, lateral interactions are also present among the upper-level PV anomalies and the background flow. It is also found that the background flow advection dominates the vortex–vortex and vortex–background flow interactions in the deepening of the storm. The vortex–vortex interactions of the two upper-level positive PV anomalies cause the negative tilt of the main upper-level trough during the rapid deepening period.

1. Introduction

Since the seminal work of Hoskins et al. (1985), a growing number of studies have used the potential vorticity (PV) concept to gain a better understanding of extratropical cyclones and the interactions among different processes or entities leading to cyclogenesis. Indeed, the PV concept provides an alternative approach to envisaging the influence of upper-level PV anomalies on the downstream development of a surface cyclone (Boyle and Bosart 1986; Uccellini et al. 1987; Whitaker et al. 1988; Bleck 1990; Reed et al. 1992), in contrast to the conventional synoptic thinking of upper-level troughs influencing surface cyclones (Petterssen 1956; Sanders 1986, 1988).

The PV concept is especially attractive to use because of (a) its conservative property in the absence of diabatic heating and friction, and (b) its invertibility principle; that is, a knowledge of the PV distribution and boundary potential temperature is sufficient to infer the meteorological fields (e.g., winds, temperatures, and geopotential heights) subject to some balanced flow constraints. Numerous studies have used various piecewise PV inversion schemes to examine the dynamical effects of upper- and lower-level PV anomalies on surface cyclogenesis (e.g., Davis and Emanuel 1991; Black and Dole 1993; Davis et al. 1993; Hakim et al. 1996). For example, Davis and Emanuel (1991) proposed a piecewise PV inversion technique to isolate the relative contributions of upper- and lower-level PV anomalies to surface cyclogenesis. The same technique was later utilized to examine the importance of initial structures and diabatic heating in an observed cyclogenesis event (Davis 1992), and the integral effect of condensational heating in a simulated winter storm (Davis et al. 1993). These studies indicate that (i) the piecewise PV inversion technique is a useful tool to diagnose the interac-
tions between different PV anomalies in a cyclone system, (ii) the presence of an upper-level PV anomaly and its position relative to surface disturbances are both critical, and (iii) the primary effect of condensation is simply to superpose a positive PV anomaly onto a baroclinic circulation and the cyclogenesis is basically driven by baroclinicity in nature. Recently, Hakim et al. (1996) utilized a piecewise quasigeostrophic PV (QGPV) inversion scheme to study the lateral interactions between two upper-level PV anomalies and background flow in a trough-merger cyclogenesis case. They identified and quantified three types of interactions: vortex–vortex, advection of background QGPV by vortex-induced flows, and advection of vortex QGPV by background flows. They found that the rapid cyclogenesis occurs as the two upper-level vortices achieve their closest proximity. With a piecewise PV tendency diagnosis, Nielsen-Gammon and Lefevre (1996) examined the contributions of different processes to the development of an upper-level mobile trough. Their approach differs from the previous ones in that only instantaneous effects could be evaluated.

The PV inversion techniques have also been applied to the improvement of model initial conditions in simulating extratropical cyclones. For instance, Huo et al. (1998) used the piecewise PV inversion developed by Davis and Emanuel (1991) to include the influence of observed surface temperatures over the Gulf of Mexico into the model boundary layer by treating it as a surface PV surrogate. With the improved initial conditions, they obtained a simulation of the 12–14 March 1993 superstorm, also referred to as “the storm of the century,” that is superior to the simulation without the corrected boundary layer in terms of both the track and intensity.

The purpose of this study is to provide a better understanding of the interaction of upper-level PV anomalies in the development of the 12–14 March 1993 superstorm using the more realistic simulation as presented in Huo et al. (1998). Specifically, Huo et al. (1995), Kocin et al. (1995), and Uccellini et al. (1995) have shown that intense low-level baroclinicity and upper-level jet streaks, strong surface fluxes from the ocean, the merging of two upper-level short-wave troughs or PV anomalies, and intense diabatic heating all contributed to the rapid deepening of the superstorm. However, uncertainty still remains concerning the relative importance of these processes as well as their interactions in the rapid cyclogenesis. In particular, what are the roles of the two troughs? How do they interact with the surface processes in intensifying the storm? Thus, the objectives of this study are to (i) gain insight into the vertical and lateral interactions of various major PV anomalies during the development of the storm, and (ii) quantify the effects of different dynamical and physical processes on the surface cyclogenesis using the piecewise PV inversion technique developed by Davis and Emanuel (1991). This diagnostic system is based on the invertibility principle of PV, which supports a dynamic partition of the total PV field into many significant portions. The circulation associated with each portion of the PV field can be deduced individually. In principle, such an approach provides a means to (i) calculate the contributions of selected PV perturbations to the instantaneous cyclone intensity, and (ii) quantify the dynamical interactions between discrete PV features in the flow field and their influences on the flow’s subsequent evolution. The results so obtained will provide useful background information for Part II of this series of papers (Huo et al. 1999) in which the impacts of the two troughs and their interactions with diabatic heating and large-scale baroclinicity on the surface cyclogenesis will be examined by treating their associated PV anomalies as an initial-value problem.

The next section provides a brief overview of the structures and evolution of PV associated with the storm during its rapid cyclogenesis stage from the 36-h model simulation given in Huo et al. (1998). Section 3 contains a brief summary of the piecewise PV inversion system and the procedures used for the present case study. Section 4 presents the analysis of the contributions of various PV anomalies to the cyclone’s intensification. Section 5 shows the dynamical interactions between upper- and lower-level PV anomalies, whereas section 6 examines the lateral interactions between the upper-level PV anomalies. A summary and concluding remarks are given in the final section.

2. Structures and evolution

For the convenience of subsequent discussions, we review briefly the PV structures of the superstorm; see Huo et al. (1995) for more details. At 0000 UTC 13 March (i.e., 12 h into the integration), hereforth 13/00–12, there are two distinct features of PV at 400 hPa: a northern tongue of large PV (>3.5 PVU, 1 PVU = 10^{-6} \text{ m}^2 \text{ K} s^{-1} \text{ kg}^{-1}) over the central United States and a southern PV maximum over the western Gulf of Mexico (Fig. 1a). The northern (southern) PV tongue corresponds approximately to a northern (southern) upper-level trough (see Fig. 2c in Huo et al. 1995). An examination of the vertical cross section (Fig. 1d) shows that the two PV maxima result from the descent of a dry stratospheric high-PV reservoir, which is further evidenced by the model-generated cloud-free region associated with the PV tongue and nearly saturated conditions to its east (Fig. 1a). The intensification of the two midlevel troughs is a consequence of continued descent of the stratospheric dry and warm air (Fig. 1d). The flow vectors in the vicinity of the storm suggest that the PV at the tip of the southern depression is being advected toward the cyclone center by southwesterly currents. This advection of high PV would increase the cyclonic vorticity of the storm (Hoskins et al. 1985) and provide an important forcing for the rapid cyclogenesis in the prior 12 h when the cyclone travels from the northwestern Gulf of Mexico to the south of the Mis-
sissip River delta. At 850 hPa, an elongated zone of PV is distributed downstream of the upper-level PV anomaly with a maximum value exceeding 1 PVU located to the northwest of the cyclone center. Most of this lower-level PV concentration is created by an upward increase in latent heat release in cloud regions. In the upper levels above the layers of latent heat release, however, heating destroys PV and creates the tight PV gradient around the outskirt of the “comma shaped” cloudiness. Note that the lower- and upper-level PV anomalies are in a favorable phase for cyclonic development. The lower-level broad baroclinic zone (see Fig. 1 in Huo et al. 1995), which occurred 12 h earlier (i.e., at 12/12±00), also experiences rapid changes, namely, strong cold (warm) advection behind (ahead of) the surface cold fronts begins to generate a large cold (warm) anomaly (Fig. 2a).

By 13/12±24, the southern PV anomaly has moved to the western Florida peninsula, and begun to experience the influence of southwesterly flows with weak upward motion (Fig. 1b). Subsequently, the southern PV anomaly weakens with time while being advected toward the surface cyclone center. The northern PV anomaly to the west, on the other hand, has moved rapidly into the base of the short-wave trough, enhancing its associated cyclonic circulations. The two PV anomalies merge into one strip 6 h later and only a single short-wave trough becomes evident (not shown). By comparison, the lower-level PV anomaly, intensifying with time, expands along the warm front near the cyclone center with increased overlap with its upper-level counterpart. Similarly, the low-level thermal wave continues to amplify as a result of advection, causing marked increases of the thermal gradient near the triple point of the cold/warm fronts (Fig. 2b).

At 14/00±36, the upper-level PV ribbon has wrapped around the short-wave trough (Fig. 1c). The surface cyclone is being overtaken by the leading portion of the merged PV ribbon; so this time marks the end of the storm’s explosively deepening phase. The vertical cross section of PV taken through the cyclone center shows two distinctive PV concentrations in the vertical: one is the merged PV anomalies at upper levels in the cloudless dry air, and the other is the diabatically produced PV anomaly centered in the 900–800-hPa layer within the cloudy air. Descending motion prevails within the upper-level dry PV region except to its northern edge where upward motion occurs as a result of latent heat release along the warm front. We can see that the dry stratospheric large-PV air, denoted by the 1-PVU contour, penetrates downward as low as 700 hPa not far behind the cyclone center. The descending tropopause large-PV air above the surface low implies that the cyclonic circulation induced by the PV anomaly may have assisted the surface cyclone development. In the following sections, we will use a piecewise PV inversion diagnostic system as developed by Davis and Emanuel (1991) to quantitatively assess the interactions between the upper- and lower-level PV anomalies and their importance in the surface cyclogenesis.

3. Isolating PV anomalies

In this section, we describe procedures on how to compute balanced flows associated with each of the above-mentioned anomalies. First, we need to define a mean state to isolate each PV perturbation. The mean state for the present case is defined as the time average between 0000 UTC 11 March and 0000 UTC 15 March 1993, which approximately corresponds to one synoptic-scale wave period. Given the mean state, the total PV anomaly is computed simply as the departure from the time average.

The next step is to partition the total PV anomaly field in a dynamically meaningful way. The philosophy of partitioning the total PV perturbations is to isolate distinct perturbations of different origins and to examine their interactions with each other and with the mean PV. In the present case, these are perturbations from the tropopause depression, the surface baroclinicity, and the interior troposphere that is associated with latent heat release. It is evident from Figs. 1 and 2 that the stratosphere-related PV perturbations \( P_{\text{d}} \) could be defined as positive PV anomalies lying in the dry air with relative humidity less than 30%, which includes the upper boundary and positive dry anomalous PV from 200 to 800 hPa, mainly of stratospheric origin. The lower-tropospheric PV perturbations associated with conditional heating \( P_{\text{c}} \) are defined as positive PV anomalies with greater than 70% relative humidity and in layers below 500 hPa. The subsaturated threshold value of 70% relative humidity is chosen to include PV that may be advected out of the precipitation region. The surface potential temperature perturbation can also be regarded as equivalent to a concentrated PV anomaly contained in a thin surface layer. Because of the surface heat flux, the lower-level interior PV is strongly influenced by the boundary layer, and the circulation associated with the lower-level PV anomaly opposes that at the surface. Therefore, we will follow the procedure of Davis (1992).
and group the lower boundary and the lower-level interior PV at 900 hPa (except where the lower-level PV anomaly coincides with larger than 70% relative humidity), forming the effective lower boundary ($\theta_{\text{eff}}$). Finally, the remainder of the interior PV perturbations, consisting of primarily the negative PV perturbation associated with the upper-level waves, will be referred to as $Q_r$.

The piecewise PV inversion technique of Davis and Emanuel (1991) is used to perform three-dimensional inversions of $Q_d$, $Q_h$, $Q_r$, and $\theta_{\text{eff}}$, using a standard successive overrelaxation technique. The inverted winds, temperature, and height fields will help reveal how the different PV anomalies interact and what their relative contributions are to the surface cyclogenesis. When inverting each anomaly, we set the other anomalies equal to zero, assuming homogeneous lateral boundary conditions. The PV inversion is performed every 6 h from 1200 UTC 12 March to 0000 UTC 14 March.

4. Cyclogenesis attributions

Table 1 shows the area-averaged contributions from the aforementioned four PV elements to the 1000-hPa height perturbation at the cyclone center during the whole period of cyclogenesis. It is evident that the magnitudes of all the contributions increase with time, an indication of the cyclone’s intensification. The relative contribution of diabatic heating $Q_h$ (18%–24%) is the smallest and does not change much during the deepening period. However, those of $Q_d$ and $\theta_{\text{eff}}$ behave differently during the different stages of the cyclogenesis. For example, the relative contributions of $Q_d$ and $\theta_{\text{eff}}$ between 12/12–00 and 13/00–12 are about 38%–40%. After 13/00–12, the $Q_d$ contribution grows faster than that of $\theta_{\text{eff}}$ owing to the approaching and subsequent merging of the northern PV anomaly into the southern one. Consequently, $Q_d$ becomes the dominant factor in determining the cyclogenesis during the second half of its life cycle, that is, >50%. The different behaviors between $Q_d$ and $\theta_{\text{eff}}$ occur because after entering the occlusion stage the surface cyclone center is usually located away from the warm sector (cf. Figs. 2b,c) and the $Q_d$ begins to overtake the surface cyclone center. As mentioned in the previous section, $Q_d$ is mainly made up of the negative PV anomaly associated with the midto upper-level ridge. Its negative contribution to the cyclone’s depth opposes the positive contributions from $Q_d$, $Q_h$, and $\theta_{\text{eff}}$. The increase of $Q_d$ can be seen as part of the upper-level wave amplification and is strongly

Fig. 2. Evolution of 900-hPa potential temperature (solid, every 5 K) and its perturbations (dashed, every 4 K), superposed with 900-hPa wind vectors for (a) 0000 UTC, (b) 1200 UTC 13 Mar, and (c) 0000 UTC 14 Mar 1993.
Table 1. The magnitudes (dam) and the relative contributions (%) to the 1000-hPa height perturbation as inverted from the upper-level dry PV anomaly ($Q_d$), lower-level moist PV anomaly ($Q_r$), effective bottom boundary ($\theta_{eff}$), and the remaining PV component ($Q_e$). They are averaged over an area of 300 $\times$ 300 km$^2$ at the cyclone center.

<table>
<thead>
<tr>
<th>Day/h</th>
<th>$Q_d$ (dam)</th>
<th>%</th>
<th>$Q_r$ (dam)</th>
<th>%</th>
<th>$\theta_{eff}$ (dam)</th>
<th>%</th>
<th>$Q_e$ (dam)</th>
<th>Total</th>
</tr>
</thead>
<tbody>
<tr>
<td>12/12</td>
<td>-11.0</td>
<td>(40.1)</td>
<td>-6.0</td>
<td>(21.9)</td>
<td>-10.4</td>
<td>(37.9)</td>
<td>10.9</td>
<td>-15.5</td>
</tr>
<tr>
<td>12/18</td>
<td>-12.2</td>
<td>(37.2)</td>
<td>-6.3</td>
<td>(19.2)</td>
<td>-14.3</td>
<td>(43.6)</td>
<td>15.2</td>
<td>-17.6</td>
</tr>
<tr>
<td>13/00</td>
<td>-19.1</td>
<td>(39.7)</td>
<td>-10.3</td>
<td>(21.4)</td>
<td>-18.7</td>
<td>(38.9)</td>
<td>25.1</td>
<td>-23.0</td>
</tr>
<tr>
<td>13/06</td>
<td>-27.8</td>
<td>(47.0)</td>
<td>-13.4</td>
<td>(22.6)</td>
<td>-18.0</td>
<td>(30.4)</td>
<td>31.5</td>
<td>-27.7</td>
</tr>
<tr>
<td>13/12</td>
<td>-39.0</td>
<td>(52.1)</td>
<td>-16.4</td>
<td>(21.9)</td>
<td>-19.4</td>
<td>(25.9)</td>
<td>41.7</td>
<td>-33.1</td>
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<td>-26.1</td>
<td>(29.0)</td>
<td>52.7</td>
<td>-37.4</td>
</tr>
<tr>
<td>14/00</td>
<td>-53.4</td>
<td>(53.0)</td>
<td>-19.2</td>
<td>(19.1)</td>
<td>-28.1</td>
<td>(27.9)</td>
<td>60.3</td>
<td>-40.4</td>
</tr>
</tbody>
</table>

In order to gain further insight into the results of static PV inversions, Fig. 3 presents the contributions to the cyclone’s 6-hourly deepening rates from the above four different PV anomalies. By comparing all the traces, one can see that the contribution from $Q_d$ (Fig. 3b) dominates the cyclone’s deepening rates. While the direct contribution from $Q_r$ (Fig. 3c) is small throughout the genesis period, its impact on the storm development is by no means less important than the other contributions. In fact, a sensitivity run, in which the latent heat release is turned off (not shown), shows that all the anomalies become much weaker, indicating the importance of the $Q_r$ contribution from the upper-level PV anomalies, since their induced height perturbation is almost in phase with the $\theta_{eff}$ anomaly (Fig. 4c). Nevertheless, the essence of the piecewise PV inversion is well demonstrated; that is, the contribution from each single PV identity can be quantified.

5. Vertical interactions

We have seen in the preceding section that the piecewise-inverted height and flow fields are in three dimensions and they extend almost into the entire inversion domain. Since these winds are free to interact with other nearby PV elements, both vertical and lateral interactions will occur as the system deepens. In this section, we discuss the possible interactions between the upper- and lower-level PV anomalies and attempt to explain how the interactions could amplify the PV perturbation in both intensity and volume coverage.
Fig. 3. The contributions to 1000-hPa height change (dam/6 h) from (a) the total PV anomaly (i.e., $Q_d + Q_h + Q_r + \theta_{eff}$), (b) the upper-level dry PV anomaly $Q_d$, (c) the lower-level moist PV anomaly $Q_h$, (d) the effective bottom boundary anomaly $\theta_{eff}$, and (e) the remaining PV anomaly $Q_r$.

Solid (dashed) lines are for positive (negative) values. Cross sections are taken along line $AB$ as given in Fig. 1c.

Fig. 4. Vertical cross sections of the balanced geopotential height perturbations, at intervals of 4 dam, inverted from (a) the upper-level dry PV anomaly $Q_d$, (b) the remaining PV anomaly $Q_r$, (c) the low-

level moist PV anomaly $Q_h$, (d) the effective bottom boundary anomaly $\theta_{eff}$, and (e) the total PV anomaly (i.e., $Q_d + Q_h + Q_r + \theta_{eff}$). Solid (dashed) lines are for positive (negative) values. Cross sections are taken along line AB as given in Fig. 1c.
anomalies. For the convenience of the subsequent discussions, $Q_d$ and $Q_r$ will be combined as upper-level PV anomalies associated with the trough–ridge system while treating $Q_h$ and $\theta_{\text{eff}}$ as the lower-level PV anomalies. Such a grouping is based on the inverted height perturbations that show clearly two different levels of maximum Laplacians (see Figs. 4a–d). This methodology is, however, not necessarily unique because they are all closely related. For example, a cold–warm $\theta_{\text{eff}}$ couplet often implies a strong trough–ridge system in the upper levels, while a high $Q_h$ associated with intense latent heat release may cause the buildup of an upper-level ridge. In the next subsection, we will first examine the influence of the upper-level PV anomaly-induced flows on the lower-level PV anomalies, and then study how the winds associated with the lower-level anomalies can modify the upper-level anomalies.

**a. Influence of upper- on lower-level PV anomalies**

As suggested by Reed et al. (1992), the surface thermal anomaly can be treated as a surrogate PV; it is generated mostly by the horizontal advection of the potential temperature $\theta$. Thus, the $\theta$ tendencies due to the horizontal advection by 900-hPa winds inverted from the upper-level PV anomalies (i.e., $Q_d + Q_r$ and $Q_h$) are given in Figs. 5a,b, which show an organized cyclonic–anticyclonic circulation couplet associated with the upper-level trough–ridge system. The induced flow increases in intensity toward the mature stage and its...
speed reaches over 25 m s⁻¹ by 14/00–36 (Fig. 5b). Since the extent of the surface warmth is determined by both the induced circulations and the thermal gradient, the upper-level trough–ridge system helps build up a warm anomaly in southerly flows, which is maximized in the vicinity of the surface cyclone center (cf. Figs. 2c and 5b). This positive warming tendency is clearly favorable for the spinup of the system, as shown in Fig. 4. In contrast, the development of the cold anomaly in northerly flow is slow and about a half wavelength away from the cyclone center. Thus, the induced cold anomaly by the upper-level trough–ridge system does not contribute significantly to the surface development.

The results indicate that upper-level processes leading to the intensification of low-level southerly flows tend to have more important effects on the surface cyclogenesis than the northerly or northwesterly flows. With the induced circulation structures shown in Figs. 5a,b, we can infer a pattern of moisture transport similar to the tendencies toward the warm front and cyclone center for latent heat release, which would in turn increase the diabatic generation of lower-level positive PV anomalies (cf. Figs. 1a,b). Since the $Q_h$ anomalies are mainly distributed along the warm front and maximized near the cyclone center, they induce a single cyclonic circulation around the storm with intensity decreasing outward (Figs. 5c,d). The so-induced flows help enhance warm anomalies along the warm front and cold anomalies behind the cold front, thus leading to the amplification of thermal perturbations near the cyclone center. If the thermal tendencies are inverted to the rate of change of height perturbations, the $Q_h$-induced circulation would contribute little to the deepening rate of the system through the effective boundary. Thus, latent heating tends to have direct effects on the cyclogenesis by warming the midtroposphere and lowering the pressure below.

Finally, the $\theta_{eff}$-induced balance winds, when overlaid on the 900-hPa $\theta$ field (not shown), are seen to act to propagate the thermal perturbation downstream along the baroclinic zone, similar to that discussed in Davis (1992). With all the above influences superposed, the low-level thermal wave is deemed to increase, particularly the thermal ridge ahead of the surface low, which itself represents an important part of the surface cyclone development.

b. Influence of lower- on upper-level PV anomalies

Although $Q_d + Q_r$ and $Q_h + \theta_{eff}$ are different in origin, each contributes to circulations at all levels (see Figs. 4a–d). Thus, the circulations from the lower-level PV anomalies ($Q_d$ and $\theta_{eff}$) must have influences on the development of the upper-level dynamics. To see this point, let us examine the PV advection by the 300-hPa winds induced by $Q_h$ and $\theta_{eff}$ (Fig. 6). In general, the induced winds are much smaller in magnitude than the low-level counterpart induced by the upper-level PV anomalies; the maximum wind is about 5 m s⁻¹ (see Fig. 6b). The advection of mean PV by the $\theta_{eff}$-induced winds is mostly positive in the trough and negative over the downstream ridge, acting to amplify the upper-level anomalies throughout the rapid deepening period. Similarly, the winds induced by $Q_h$ also amplify the upper-level positive and negative PV anomalies during the rapid deepening stage (Figs. 6c,d). In addition, increasing $Q_h$ implies rapid latent heat release in the midtroposphere, which assists the amplification of the negative anomaly in the ridge by (i) transporting the upper-level positive PV down to the lower levels, and (ii) creating divergent outflow such that the negative PV anomaly expands in area coverage. It follows that although the upper-level winds induced by the low-level PV anomalies are weak, the development of $Q_h$ and $\theta_{eff}$ facilitates the amplification of the upper-level baroclinic wave.

In summary, the favorable phase relationship between the upper- and lower-level disturbances during the rapid deepening stage allows different anomalies to interact constructively with each other. In this regard, cyclogenesis may be viewed as a mutual interaction of the upper- and lower-level PV anomalies, as also discussed by Davis and Emanuel (1991), in which the circulations associated with the upper-level PV anomalies enhance the lower-level anomalies, and vice versa.

6. Lateral interactions

Huo et al. (1995) showed that there are two upper-level PV anomalies or troughs contributing to the surface cyclogenesis. They are separated by more than 3000 km prior to the genesis over the northwestern Gulf of Mexico. They first approach and finally “merge” as the surface cyclone deepens. The scenario fits the description of the trough-merger process by Gaza and Bosart (1990), who performed a climatological study of the trough-merger events in North America. They found that (i) two-thirds of the 21 trough-merger events they examined are associated with explosive cyclogenesis; (ii) the meridional tilt of the principal 500-hPa height trough axis changes from positive to negative prior to the rapid cyclogenesis; and (iii) the two 500-hPa vorticity maxima amalgamate into a single maximum with larger amplitude than either of them. Therefore, it is of interest to examine how these troughs or PV anomalies interact to form the upper-level jet and the surface cyclogenesis. To study the lateral interactions of the two troughs, we must isolate the related PV anomalies as done in the previous sections. In the present case, they include two positive dry PV anomalies associated with the upper-level troughs (i.e., $Q_d$ for the northern trough and $Q_d$ for the southern trough), and one negative anomaly ($R$) representing the downstream ridge. The negative PV anomaly ($R$) is determined by $\pm 0.2$ PVU at all levels above 500 hPa. The rest of the PV perturbations (i.e., other than $Q_d$, $Q_d$, and $R$) and the mean PV will be called the background PV—an approach similar to that
FIG. 6. The PV advection (PVU/6 h) by the balanced flows at 300 hPa that are inverted from (a) and (b) the effective bottom boundary PV anomaly $\theta_{\text{eff}}$ and (c) and (d) the low-level moist PV anomaly $Q_d$. Solid (dashed) lines are for positive (negative) values. Left and right panels are for 0000 UTC 13 Mar and 0000 UTC 14 Mar, respectively. Thick-dashed lines denote the upper-level trough axes.

of Hakim et al. (1996). In this way, $Q_{d1}$ and $Q_{d2}$ can be cleanly separated up to 13/06–18.

Following Hakim et al. (1996), we use Fig. 7 to illustrate qualitatively the expected dynamic interactions for the upper-level PV anomalies ($Q_{d1}$, $Q_{d2}$, $R$) and the background flow at 13/00–12. Let us discuss first the vortex–vortex interaction among $Q_{d1}$, $Q_{d2}$, and $R$. Between $Q_{d1}$ and $Q_{d2}$, there exist advections by their induced circulations and the main contribution of the instantaneous tendencies is cyclonic relative motion of the vortices such that the southern system moves eastward with the background flow, while the northern system moves westward against the background flow (Fig. 7a). The instantaneous tendencies caused by the interactions of $Q_{d1}$ with $R$ and $Q_{d2}$ with $R$ force a northward motion of all three anomalies (Fig. 7b). In addition to the vortex–vortex interaction, there are also interactions of individual vortices with the background flow. The advections of the background PV by $Q_{d1}$ and $Q_{d2}$ tend to induce instantaneous height falls (rises) west (east) of the vortices and cause the vortices to propagate westward (Fig. 7c). Likewise, the advections of the background PV by $R$ forces a height rise (fall) west (east) of $R$, and therefore $R$ will move westward against the background flow as well (Fig. 7c). This is essentially the Rossby wave propagation mechanism. Finally, all these vortices are advected eastward by the background flow (Figs. 7d). For the present case, the advection of
Fig. 7. Schematic illustration of geopotential height tendencies associated with (a) and (b) vortex–vortex interaction, (c) advection of background PV by vortex-induced flows, and (d) advection of PV anomalies by background flows. Shaded and open circles represent positive and negative upper-level PV anomalies, respectively. Positive and negative signs indicate the sense of geopotential height tendencies. Thick arrows show the instantaneous motion of the anomalies and thin arrows show the background flow motion.

Fig. 8. Inverted 400-hPa balanced geopotential height from the northern PV anomaly \((Q_{d1}, \text{dashed})\), southern PV anomaly \((Q_{d2}, \text{dashed})\), the upper-level negative PV anomaly \((R, \text{thick solid})\), and the background PV (thin solid) at intervals of 6 dam for (a) 1800 UTC 12 Mar and (b) 0600 UTC 13 Mar. Solid circles show the surface cyclone center.

Quantitative presentations of the lateral interactions are given in Fig. 8, which shows the significant overlap of the inverted balanced heights associated with \(Q_{d1}\), \(Q_{d2}\), \(R\) and background PV at 400 hPa during the intensifying stage (i.e., 12/18–06 to 13/06–18). The three perturbations all have pronounced influences on the cyclone, depending on their magnitudes and position with respect to the cyclone center. At 12/12–00, \(Q_{d2}\) is located west of \(Q_{d1}\) and the two PV anomalies are separated by more than 3000 km (not shown). Therefore, the interactions between \(Q_{d1}\) and \(Q_{d2}\) are weak and the background flow dominates the evolution, acting to advect \(Q_{d1}\) southeastward and \(Q_{d2}\) eastward.

Because the 400-hPa background flow is slightly diffluent downstream of the Rocky Mountains, northerly flows behind the trough axis decrease southward in intensity. This flow structure tends to advect \(Q_{d1}\) faster than \(Q_{d2}\) toward the cyclone center. The approaching of the two positive PV anomalies coincides with the rapid intensification of the system (Huo et al. 1995; Kocin et al. 1995). On the other hand, the \(R\)-induced flow, which is diffluent northwestward, assists further the merging of \(Q_{d1}\) and \(Q_{d2}\), but slows their eastward movements. The net result is that by 12/18–06, the two anomalies are brought together within a distance of less than 2000 km (Fig. 8a); so the vortex–vortex interactions become significant. Clearly, the \(Q_{d2}\)-induced circulation tends to force \(Q_{d1}\) to move westward away from the genesis area while enhancing the moisture transport and the vorticity advection toward the cyclone center. Similarly, the \(Q_{d1}\)-induced circulation tends to advect \(Q_{d2}\) and a colder air mass toward the...
genesis area. Because of the complicated interaction, it is not possible to determine the relative integral effects of $Q_d$ and $Q_d$ on the cyclogenesis. Their individual influences could only be examined by treating $Q_d$ and $Q_d$ separately as an initial-value problem. This will be discussed in Part II (Huo et al. 1999).

By 13/06–18 (Fig. 8b), the two positive PV anomalies are being advected farther toward the surface cyclone and the system begins to reach its most rapid deepening stage. Of interest is that both the R- and the $Q_d$-induced circulations increase, owing to the midlevel rapid latent heat release along the warm front and the continued descending of stratospheric air behind the trough axis. However, the $Q_d$-induced circulation weakens from −20 to −16 dam during the prior 12 h. As shown in Fig. 1, $Q_d$ continues to intensify as it moves southeastward whereas $Q_d$ weakens as it moves northeastward, causing changes in the orientation of the upper-level troughs. The weakening of $Q_d$ and strengthening of $Q_d$ can be understood as the interaction of background PV with the $Q_d$ and $Q_d$-induced circulations. Specifically, the $Q_d$-induced winds are southerly (northerly) at $Q_d$ (Fig. 1), thus weakening $Q_d$ (strengthening $Q_d$) by advecting background PV. In the height field, this is the time when the two upper-level troughs start to merge together and produce a negative meridional tilt of the trough axis. Thus, the transition of the trough axis from the positive to negative tilt, a characteristic of the trough-merger climatology (Gaza and Bosart 1990), may be a manifestation of the interaction between the two upper-level positive anomalies.

To quantify the relative importance of these different interactions in the anomaly development and surface cyclogenesis, we calculate the instantaneous PV tendencies induced by each of the above-mentioned advective processes. The results for 13/00–12 are given in Fig. 9, which supports the conceptual model given in Fig. 7 but shows significantly varying magnitudes and structures of individual anomalies. For example, the $Q_d$-induced flow advects the background PV to create a positive (negative) PV tendency to its west (east), acting to propagate itself westward (Fig. 9a). For the same reason, the $Q_d$-induced flow also acts to propagate itself westward, but much weaker than $Q_d$ (Fig. 9b). The advection of the background PV by the R-induced anticyclonic circulation is more complicated (see Fig. 9c) due to the presence of local maxima in the background PV field (e.g., south of the Great Lakes and off the coast of Nova Scotia; see Fig. 8). On average, the R-induced flow causes PV increases (or height falls) to its east and PV decreases (or height rises) to its west.

Figures 9d–f show the tendencies due to each of the $Q_d$, $Q_d$, R-induced flows advecting the other two PV anomalies. Although these tendencies are localized, the magnitudes are much larger than those self-propagation terms as shown in Figs. 9a–c (note the different contour intervals between Figs. 9a–c and 9d–f). We see that the $Q_d$-induced flow forces $Q_d$ to move eastward and R to move northward (Fig. 9d), which is consistent with the results obtained from the static PV inversion (Fig. 8b). Similarly, the $Q_d$-induced flow acts to move R northwestward and $Q_d$ westward against the background flow (Fig. 9e). Like the case in Fig. 7b, the R-induced flow tends to advect both $Q_d$ and $Q_d$ northward.

The more pronounced dynamic forcing is associated with the advection by the background flow, which acts to advect all the PV anomalies downstream: $Q_d$, southeastward and $Q_d$ eastward, both toward the surface cyclone (Fig. 9f). This results in a broad area of positive PV tendency between positive ($Q_d$ and $Q_d$) and negative (R) anomalies. More importantly, this large area of positive PV tendency coincides closely with the surface cyclogenesis area, and its magnitude dominates all the vortex–vortex interactions. Therefore, the sum of all the elements yields the total PV tendency (Fig. 9h), which is dominated by background advection and modulated by the vortex interactions.

Based on the above analysis, it is evident that in the absence of both $Q_d$ and $Q_d$ (or the upper-level shortwave troughs) the surface cyclogenesis could not take place, at least during the 36-h integration period despite the presence of intense baroclinicity. What would happen to the surface cyclogenesis in the case of only having $Q_d$ or $Q_d$? Based on the above discussion, we may infer that without $Q_d$, the advection of $Q_d$ toward the cyclogenesis area would be reduced due to the decreased eastward component of the flow associated with $Q_d$. Thus, the eastward movement of the surface cyclone (if developed) would be slow. Without $Q_d$, the background flow would tend to advect $Q_d$ faster toward the surface cyclone for its deepening. These hypotheses will be tested in Part II of this series of papers through a series of sensitivity experiments (see Huo et al. 1999).

Finally, let us examine the formation of the upper-level jet streak (Fig. 1) in the context of the PV concept since its transverse circulations have been indicated by Huo et al. (1995) and Kocin et al. (1995) to play an important role in the rapid development of the storm. Here we may visualize the upper-level outflow jet streak as the superposition of circulations associated with $Q_d$, $Q_d$, R, and background PV as given in Fig. 8. Two vertical cross sections of the inverted height normal to the jet streak are taken to see their relative contributions: one near the entrance region (Figs. 10a–d) and the other near the R-induced circulation center (Figs. 10e–h). In the entrance region, the $Q_d$, R, and background PV-induced circulations contribute approximately equally to the upper-level jet streak (i.e., the winds in the jet core region); $Q_d$’s contribution is slightly negative and negligible. In the core region, only the R-induced and background flows are important whereas $Q_d$ and $Q_d$’s contribu-
Fig. 9. The PV advection at 1300–12 by the balanced flows that are inverted from (a) $Q_{d1}$, (b) $Q_{d2}$, and (c) $R$ advecting the background PV; (d) the northern PV anomaly $Q_{n}$, (e) the southern PV anomaly $Q_{s}$, (f) the negative PV anomaly $R$ advecting the other two PV anomalies;
7. Summary and conclusions

The contributions of various potential vorticity perturbations to and their mutual interactions on the superstorm of 12–14 March 1993 are investigated by using a piecewise PV inversion diagnostic system. This study is aimed at the vertical and lateral interactions among various distinct PV anomalies: upper-level dry positive PV anomalies consisting of a northern and a southern PV anomaly, the upper-level negative PV anomaly, the lower-level moist PV anomaly, and the surface thermal anomalies. It is shown that the absolute contributions (to the surface cyclone depth) from the above anomalies increase with time, but their relative importance varies during the genesis stages. At the mature stage, the upper-level dry PV anomaly contributes the most to the surface cyclone depth (53%), followed by the surface thermal anomaly or thermal advection (28%). The low- to midlevel moist PV anomaly contributes the least (19%) to the genesis, and its relative significance remains nearly unchanged during the life cycle of the storm.

By comparing the PV anomalies and their inverted circulations, we found that there exists a favorable phase tilt between the upper- and lower-level anomalies that allows lower-level–upper-level mutual interactions. That is, the circulations associated with the upper-level PV anomalies enhance the lower-level anomalies, which in turn feedback positively to the upper-level perturbations. In addition to the vertical interactions, lateral interactions are also present among the upper-level PV anomalies and the background flow. It is found that the background flow advection dominates the vortex–vortex and vortex–background flow interactions. The vortex–vortex interaction of the two upper-level positive PV anomalies (or two troughs) causes the negative tilt of the main upper-level trough during the rapid deepening stage. The negative PV anomaly associated with an upper-level ridge plays an important role in the formation of the upper-level outflow jet. To isolate the impacts of the upper-level troughs and their interaction with diabatic heating and large-scale baroclinicity on the cyclogenesis, we have to find a way to separate these perturbations first in the model initial conditions and then evaluate their impacts by treating them as an initial value problem. This will be presented in Part II of this series of papers through the analysis of various sensitivity experiments to the above-mentioned PV perturbations.
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REFERENCES


