The Role of Antecedent Surface Vorticity Development as a Conditioning Process in Explosive Cyclone Intensification

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ABSTRACT

We examine the idea that antecedent vorticity development, defined as the surface vorticity spinup in the period prior to a cyclone’s maximum intensification, is an important dynamical conditioning process for explosive cyclogenesis. Previous suggestions from case study research that subsequent intensification may be proportional to the intensity of the preexisting circulation are supported through the systematic study of a large sample of weakly and explosively developing cyclones in the North Pacific and North Atlantic basins.

Additional support for this concept is found with an examination of composite weakly and strongly developing cyclones at the onset of their most rapid intensification period. At this onset, the strongly developing cyclone composite has substantially stronger surface circulation and vorticity than is found in a composite of the weak cases. Ensembles of successive forecasts of an explosive cyclogenesis case during the Experiment on Rapidly Intensifying Cyclones over the Atlantic (ERICAX) suggest similar dynamical behavior, in that small errors in the surface intensity subsequently amplify into larger errors only 12 h later under predominantly similar upper-level conditions.

The temporal evolution of large-scale geostrophic vorticity for 62 cases of cyclogenesis shows that stretching in the presence of relative vorticity is present throughout the life cycle of both the weakly and rapidly developing cases. An examination of 794 cyclones in the North Pacific basin reveals a general trend of increased maximum development as the antecedent deepening increases. Explosively developing cyclones are preferentially characterized by at least 12 h of antecedent development.

We investigate the relationship between the amplitude of the 500-mb quasigeostrophic-ascent forcing and maximum surface cyclone intensification and find a significant positive correlation, as previous studies have shown. However, computations with model-based surface convergence suggest that the response to the upper-level forcing is conditioned by the low-level antecedent vorticity development. Furthermore, variations in successive numerical weather prediction model forecasts of maximum cyclone intensification are well correlated with variations in the initial surface vorticity as well as variations in the 500-mb forcing.

This study suggests that explosive development is typically characterized by a nonlinear interaction between two cyclonic disturbances in the lower and upper troposphere. These disturbances, in some cases, may have formed independently of one another. Thus, the correct simulation of the full life cycle of these cyclones, including the antecedent phase, may be crucial for accurate numerical forecasts.

1. Introduction

The rapidly intensifying extratropical cyclone (referred to as a “bomb” by Sanders and Gyakum 1980 because its central pressure falls at least 1 mb h⁻¹ for 24 h) has been studied extensively in the literature during the past decade. One prominent characteristic of two of the most discussed bombs is their substantial development prior to the onset of their explosive intensification. The QE II storm, studied by Gyakum (1983, 1991) and Uccellini (1986), deepened 7 mb during its first 12 h of existence prior to its extraordinary 24-h pressure fall of nearly 60 mb. The Presidents’ Day cyclone, as documented by Bosart (1981), deepened 13 mb in the 12-h period prior to its extraordinary 6-h pressure fall of 16 mb. Each of these cases was characterized by the surface cyclone being positioned optimally beneath a region of 500-mb cyclonic-vorticity advection during the period of explosive deepening. The authors studying these cases have shown the surface cyclones to have formed and intensified during these antecedent stages, in regions of preexisting surface frontogenesis and heavy precipitation. Gyakum (1983, 1991) has documented 18 m s⁻¹ surface winds and central surface vorticity of $17 \times 10^{-5}$ s⁻¹ in the QE II storm at the onset of its 24-h period of explosive intensification. Uccellini et al. (1987) and Whitaker et al. (1988) have diagnosed numerical simulations of the Presidents’ Day cyclone in terms of two stages of development. These two stages may be described as the antecedent and most rapidly deepening phases.
The dynamical importance of such strong antecedent surface cyclone development for the subsequent stage of most rapid deepening may be seen from the following analysis of the frictionless vorticity equation in pressure coordinates. This may be expressed as
\[
\frac{\partial (\xi + f)}{\partial t} = -V \cdot \nabla (\xi + f) - \omega \frac{\partial \xi}{\partial \rho}
\]
\[- (\xi + f) \nabla \cdot V - k \left( \nabla \times \frac{\partial V}{\partial \rho} \right),
\]
where \(\xi\) is the vertical component of relative vorticity, \(f\) is the planetary vorticity, and \(V\) and \(\omega\) are, respectively, the horizontal and vertical wind components. If we assume that horizontal and vertical advections of vorticity and tilting of the horizontal components of vorticity into the vertical may be ignored at the surface cyclone center, then (1) becomes
\[
\frac{\partial (\xi + f)}{\partial t} = -(\xi + f) \nabla \cdot V.
\]

The assumption of small horizontal vorticity advection is based upon the cyclonic-vorticity maximum being located at or near the low center, while the vertical advection and tilting terms are small given that surface vertical motions are negligible. Therefore, (2) indicates that the absolute-vorticity intensification at the surface cyclone center is the product of the horizontal convergence and absolute vorticity.

Given that \(\xi\) is of the same order of magnitude as the planetary vorticity and is therefore retained on the right-hand side of (2), a simple exponential growth of surface absolute vorticity is indicated for constant surface convergence throughout the phase of most rapid deepening. Since surface convergence typically increases as a cyclone intensifies, the result is an even more rapid increase in surface absolute vorticity during the maximum development phase.

One of the objectives of this study is to demonstrate that stretching of relative vorticity is important throughout the cyclone's explosive intensification phase. The implication of this result is that nonnegligible cyclonic development has already occurred prior to the onset of explosive intensification. Additionally, if we focus upon a time period of most rapid deepening for both explosively and weakly developing cyclones, then this period may be used as a reference to define the antecedent stage of development. Roebber (1984, his Fig. 4), using a 1-yr database, showed that explosive cyclones have a preferentially longer period of deepening than weaker storms. This suggests that vorticity growth during the antecedent stage of the cyclone's life cycle is a conditioning process that may have considerable influence on subsequent intensification. We will examine this question in this paper with respect to a 9-yr sample of North Pacific surface lows.

An illustration of the impact of such a conditioning process is provided by the time integration of (2), shown in Fig. 1. This figure shows the inviscid growth of relative vorticity \(\xi\) integrated from the same initial value \((10.0 \times 10^{-5} \text{ s}^{-1})\) at two different times. During the first 6 h of the integration (the antecedent stage) the convergence increases linearly with time from \(5.0 \times 10^{-5}\) to \(15.0 \times 10^{-3} \text{ s}^{-1}\). During the final 6 h (the development stage) the convergence is held constant at \(15.0 \times 10^{-3} \text{ s}^{-1}\). These vorticity and convergence numbers are similar to values found from a successful operational forecast of a rapidly intensifying cyclone in the North Atlantic, to be discussed in section 6. The case with only a 6-h period of antecedent development has a far greater amplitude at the end of 12 h than the case in which development began at 6 h into the integration, despite identical convergence during the development stage.

The concept of antecedent vorticity development as a potentially important factor in especially rapid cyclone development has been recognized for many years. Pålmen and Newton (1969, pp. 319–320) state:

The reason for most rapid cycloonic development when the upper divergence is superposed over the frontal trough, and not before, is obviously in part related to the circumstance that the increase of vorticity due to convergence is directly proportional to vorticity itself or to the preexisting circulation that is always cyclonic at any surface front.

This same idea has been implicit in several case studies. Hoskins and Berrisford (1988) have described the rapid intensification phase of the October 1987 cyclone that struck the British Isles in terms of an inter-

![Fig. 1. Time integration of (2) for cases in which the relative vorticity \(\xi\) is initialized with a value of \(10.0 \times 10^{-5} \text{ s}^{-1}\) at the times 0 and 6 h. The values of \(\xi (10^{-5} \text{ s}^{-1}, \text{ordinate})\) are plotted as a function of time (h). Convergence increases linearly with time from \(5.0 \times 10^{-5} \text{ s}^{-1}\) to \(15.0 \times 10^{-3} \text{ s}^{-1}\) for 0–6 h, and is held constant at \(15.0 \times 10^{-3} \text{ s}^{-1}\) for the next 6 h.](image-url)
action of a broad upper-level isentropic potential-vorticity (IPV) maximum and an intense small-scale lower-tropospheric IPV center. Bosart and Lin (1984) and Gyakum (1991) have noted substantial surface vorticity development prior to the subsequent explosive intensification phases of, respectively, the Presidents' Day and QE II cyclones. DiMego and Bosart (1982) have suggested that the vorticity-rich surface boundary layer of Tropical Storm Agnes enhanced its redevelopment during its transformation into an extratropical system.

However, a systematic verification that antecedent surface vorticity development may be preferentially associated with rapidly developing cyclones has not been performed. Indeed, much of the research published during the past ten years has been explicit in its emphasis on the role of the 500-mb trough in forcing explosive cyclone intensification. For example, the schematic picture of cyclogenesis presented by Hoskins et al. (1985) and widely cited elsewhere shows a cyclonic upper-air IPV anomaly arriving over a preexisting low-level baroclinic zone without substantial circulation. As pointed out earlier, Hoskins and Berrisford (1988) have more recently suggested the importance of the low-level circulation for its interaction with the upper disturbance in the rapid intensification process. The positive relationship of 500-mb forcing to explosive surface cyclogenesis has been documented by Sanders (1986) in his study of western Atlantic cases. Clearly, the development of the upstream 500-mb trough is critical in determining how rapidly a surface cyclone will deepen. However, the surface response to the baroclinic forcing, as measured by dynamical processes occurring at 500 mb, has been characterized as exceptional in cases of rapid cyclogenesis (Sanders 1986; Hadlock and Kreitzberg 1988). The factor most often cited as a likely candidate for amplifying the response is weak static stability in the vicinity of the cyclone, a condition that has been observed by Gyakum (1983) and Rogers and Bosart (1986).

However, given that explosive cyclogenesis is characterized by some extraordinary response and in light of the foregoing discussion of (2), we propose antecedent development of surface vorticity as another conditioning factor. To investigate these ideas, we will consider the relationship between antecedent development and 500-mb forcing, with subsequent intensification.

A novel aspect of this research is that we will examine cyclone developments of all intensities for a large sample of North Pacific and North Atlantic cases. Such a perspective that includes weakly developing cyclones is necessary if we are to understand the processes associated with the more rapidly developing systems.

Though our focus on examining the role of antecedent surface cyclone vorticity development is strictly a dynamical one, the implications for such research include thermodynamic processes. As an example, surface heat fluxes may act frontogenetically in the vicinity of a developing surface cyclone. This particular frontogenesis would drive a direct vertical circulation that implies surface convergence and cyclonic vorticity development in the warm air. Such a development may be associated with the antecedent development stage of the cyclone's life cycle. These processes have been discussed by Kuo et al. (1991a) for the early (antecedent) stages of surface cyclone development, and may also be important factors in several cases included in this paper. Other mechanisms of incipient development, including cyclonic-vorticity advection effects from a strong upper trough, may be more dominant in other cases of explosive intensification.

The issue of the amount of time allotted for antecedent development of surface cyclones will be examined in this study. Figure 1 suggests the importance of the time scale of development for the especially rapid deepeners. Gyakum (1991) has suggested that one means by which this extended time of intensification is accomplished is through the surface low's initial independence of the upper-level vorticity maximum that subsequently is associated with the explosive development. Long development periods could also be associated with a coupling of one upper-level trough with one surface system throughout the surface system's development. The hypothesis to be examined in this study is whether the degree of a surface low's antecedent intensification, regardless of its specific mechanism, plays a role in that system's subsequent maximum intensification period.

2. Database and methodology

The data for this research include surface cyclone compilations of central pressure, latitude, longitude, day, and time for regions of the North Pacific and North Atlantic basins. The compilations are made at 12-h intervals (0000 and 1200 UTC) for each cyclone analyzed within a specified region on the Northern Hemisphere surface final analyses produced by the National Weather Service. Individual systems were tracked for continuity using the 0600 and 1800 UTC charts. Though cyclogenesis undoubtedly begins and ends at times not precisely at 0000 and 1200 UTC, this 12-h resolution was used because of the availability of both surface and upper-air reports for these times. The particular region chosen for the Pacific basin includes the area from 20° to 70°N and from 120°E eastward to 120°W. The sample includes nine cold seasons (1 October–31 March) beginning on 1 October 1975. Further details of the North Pacific cyclone dataset may be found in the study by Gyakum et al. (1989).

We examine a subset of the North Pacific cases with the National Meteorological Center (NMC) Northern Hemisphere analyses in section 3, interpolated to a 381-km (at 60°N) octagonal grid, at sea level and 500
mb. These analyses are available on a compact disk, and they are described further by Mass et al. (1987). The full North Pacific sample is studied from the perspective of pressure-change data in section 4. An analysis of the vorticity equation (2) with observations and model data is performed in section 5.

The North Atlantic study region includes the area from 30° to 60°N, bounded by the east coast of North America to the west and extending approximately 20° longitude eastward. The sample includes all cyclones for the period November 1988–March 1989 that underwent a maximum in analyzed 12-h central pressure fall within these boundaries. This resulted in a total sample of 52 events. Further details concerning this sample may be found in Roeber (1990). Gridded initial analysis and forecast data for a subset of these cases have been obtained from the Canadian Meteorological Centre (CMC) Regional Finite-Element (RFE) Model digital archive. This model is discussed by Benoit et al. (1989) and Tanguay et al. (1989). The model uses finite elements to solve the equations, which result in a variable horizontal grid resolution. However, the polar stereographic model domain is centered on North America, and the resolution in this central window is fixed at 100 km (at 60°N). For the purposes of this study, the model output is first interpolated to a 100-km polar stereographic grid before processing. However, the region of interest in the North Atlantic falls well within this central window, so that we expect interpolation errors to be negligible. These data, used to compute gridded diagnostic quantities, are discussed in section 6. The concluding discussion is found in section 7.

3. Western North Pacific sample

Past studies of explosive cyclogenesis have consistently shown frequency maxima located in the western regions of the North Pacific and North Atlantic oceans, and in one area located east of the date line in the central North Pacific (see Sanders and Gyakum 1980, Fig. 3; Roeber 1984, Fig. 7; and Gyakum et al. 1989, Fig. 12). We focus on the 9-yr cold-season dataset of Gyakum et al. (1989) to obtain a large number of cases in one limited geographical region. We choose the western North Pacific region from this database because of its good surface data coverage and its proximity to the upstream network of conventional rawinsonde observations in eastern Asia.

We find 31 cases whose maximum 24-h central pressure fall commenced within the 5° × 5° latitude–longitude region defined such that 35°N ≤ φ1 < 40°N and 145°E < λ1 ≤ 150°E, where φ1 is the latitude in this region and λ1 is the longitude. We call this set of 31 cases sample 1. We also find another independent set of 31 cases, defined in the same manner, to be located in the 5° × 5° region to the west of sample 1, such that 35°N ≤ φ2 < 40°N and 140°E < λ2 ≤ 145°E. We refer to this set of cases as sample 2.

All 24-h pressure falls are adjusted geostrophically to 45°N, so that

\[ \Delta p_{45} = \Delta p_\phi \frac{\sin 45^\circ}{\sin \phi}, \]

where \( \Delta p_{45} \) is the adjusted pressure change, \( \Delta p_\phi \) is the observed 24-h central-pressure change, and \( \phi \) is the latitude of the cyclone center 12 h into this period of deepening. An adjusted central-pressure fall of 19.6 mb would correspond to 1 bergeron, as defined by Sanders and Gyakum (1980), a rate geostrophically equivalent to a 24-h central-pressure fall of 24 mb at 60°N. The beginning time of this maximum adjusted pressure fall (or the minimum adjusted pressure rise) will be denoted as \( T_0 \).

We illustrate the basic large-scale features associated with surface cyclones of varying maximum deepening rates with the following procedure. Seven different classifications of 24-h intensification are defined for each of the 31-case samples, whose adjusted 24-h deepening rates range from 4.9 to 39.9 mb (24 h)\(^{-1}\). We choose here to focus on two classifications of sample 1: the strong nonbomb [11–19 mb (24 h)\(^{-1}\), denoted as SN1] and moderate-bomb [24–35 mb (24 h)\(^{-1}\), denoted as MB1] classes. We use these classes because they represent cases that are well separated by intensification, and each class consists of nearly the same number of cases: nine for SN1 and eight for MB1. We perform this stratification in order to produce composite fields of sea level pressure, 500-mb height, and temperature for surface lows located in the same 5° × 5° region. Similar smoothing of structure would be expected in composing similar numbers of cases. These fields are produced from the NMC gridded analyses.

Figures 2 and 3 show composites of the two surface cyclone classes at the onset of maximum intensification. The first class, SN1, shows a sharp 500-mb trough approximately 1000 km to the west of the composite SN1 cyclone. Baroclinity, extending through the 1000–500-mb layer is also prominent in the region of this trough. Figure 3, showing the moderate-bomb composite, MB1, shows an apparently similar 500-mb pattern, though with a sharper trough upstream of the composite surface cyclone. Such a result shows clearly that both weakly developing and rapidly developing surface cyclones, at the onset of maximum intensification, are associated with a finite-amplitude midtropospheric trough and are fundamentally baroclinic. This finding is in agreement with the observational work of Sanders and Gyakum (1980), Sanders (1986), and the theoretical results of Farrell (1984).

The sea level pressure fields, shown for each composite cyclone, show a more robust circulation for MB1 than for SN1 at the onset of the system's most rapid deepening. Indeed, the composite geostrophic relative vorticity as computed on the NMC mesh is 1.4 × 10^{-2} \, s^{-1} for MB1 versus only 0.6 × 10^{-4} \, s^{-1} for SN1. Such a result suggests that substantial boundary-
Fig. 2. Composite (a) 500 mb, (b) sea level pressure (solid, contour interval of 4 mb) analyses for the nine cyclone cases of SN1, found in the region bounded by 35°–40°N, 145°–150°E, at the onset of the surface cyclone 24-h maximum intensification period. Geopotential heights (solid, contour interval of 6 dam) and temperatures (dashed, contour interval of 10°C) are shown at 500 mb. The 1000–500-mb thickness (dashed, contour interval of 6 dam) is shown on the sea level pressure panel. Part (c) shows fields of $2\nabla \cdot Q \left(10^{-19} \text{mb}^{-1} \text{s}^{-1}\right)$ for 500 mb.
layer vorticity development has occurred during the
time prior to the onset of most rapid development in
both systems, but more so for MB1. Therefore, it ap-
pears that surface vorticity development may be an
important conditioning process that preferentially en-
hances the surface development for bombs. Addition-
ally, the MB1 composite shows a stronger ridge east of
the low, allowing for enhanced downstream southerly
goestrophic flow.
To understand further the horizontal structure of
the 500-mb forcing, we use the composite height and
temperature fields to compute quasigeostrophic \( Q \) vec-
tors, as described by Hoskins et al. (1978) and Hoskins and Pedder (1980). The authors have shown that the quasigeostrophic $\omega$ equation, in its adiabatic, inviscid form, may be written as
\[
\left( \sigma \nabla^2 + f \frac{\partial^2}{\partial \phi^2} \right) \omega = -2h \nabla \cdot \mathbf{Q}, \tag{4}
\]
where $h = (R/p)(p/p_0)^{\kappa}$, $\kappa$ is the ratio of the gas constant to the specific heat at constant pressure, $\sigma = -h \times \partial \theta / \partial p$, $p_0 = 1000$ mb, $\theta$ is the potential temperature, $\nabla$ and $\nabla^2$ are evaluated on a constant-pressure surface, and $\mathbf{Q} = -(\partial V_x / \partial x) \cdot \nabla \theta - (\partial V_y / \partial y) \cdot \nabla \theta$. The right side of (4) includes the traditional horizontal vorticity and temperature-advection forcing.

Although quasigeostrophic theory does not fully explain the development of these cyclones, it does aid in a physical understanding of the large-scale environment. Kuo et al. (1991b) have successfully used quasigeostrophic vertical velocities to understand their more accurate primitive equation computations.

Figures 2c and 3c show the spatial fields of the right-hand side of (4), without the minus sign, for the respective SN1 and MB1 composites. As noted, both composites are associated with finite-amplitude troughs at 500 mb, whose effects are seen in $\mathbf{Q}$-vector convergence fields and forcing for ascent. The horizontal structures are different, however, in each composite. The SN1 shows the strongest $\mathbf{Q}$-vector convergence occurring upstream of the surface low, with divergence of $\mathbf{Q}$ maximized to its west in a region centered between Japan and the Korean peninsula. The MB1 composite shows larger values of ascent forcing associated with the composite surface low. This forcing also encompasses a region downstream of the surface system toward which it travels. The results of Figs. 2 and 3 provide us with motivation to examine the surface vorticity conditioning process further.

To learn more about the antecedent behavior of the 62 cyclones in the western Pacific, we locate their 12-h positions from 48 h prior to the beginning of the maximum deepening period through 12 h after the beginning of this period. Generally, these lows had not formed as early as two days prior to maximum development—only one of these 62 cases formed 48 h before its maximum deepening onset $T_0$. The rest of these cases formed from 36 to 0 h before the onset of maximum development. The tracks of these surface cyclones are shown by the black solid lines for samples 1 and 2 (each containing 31 cases) in Figs. 4a and 5a, respectively. The open circles show the locations of each cyclone at the beginning of its 24-h maximum deepening period. Additionally, we locate the 500-mb vorticity maximum, derived from the NMC analyses, that is associated with each surface cyclone at the onset of its maximum intensification. We trace this vorticity maximum at 12-h intervals to 48 h prior to and 12 h after ($T_{12}$) this onset. We had no difficulty locating such 500-mb systems for the entire 60-h period of each case. These tracks are also shown for each of the samples in Figs. 4a and 5a, and their positions at time $T_0$ are indicated by open squares. We note the long antecedent tracks of most of the 500-mb systems from the continental regions of Asia before interaction with a surface cyclone. Such behavior has been noted by Sanders (1988) in his study of mobile 500-mb troughs.

To examine whether there are differences between the explosively developing and weakly developing surface cyclones and their associated 500-mb vorticity maxima, we stratify each sample accordingly and present these results in parts (b) and (c) of Figs. 4 and 5. Though the 500-mb tracks appear similar for both weak and explosive developments, there are differences in the surface tracks. The most distinguishing characteristic of the explosively developing cyclone is its history of antecedent tracking. Note that the antecedent tracks for the total sample (Figs. 4a and 5a) are almost entirely specified by the rapidly developing systems (Figs. 4b and 5b). The cases of weak development are nearly devoid of antecedent tracking (Figs. 4c and 5c). Their incipient formations are clearly associated with the approach of an upstream 500-mb trough and its cyclonic-vorticity advection, characteristic of the Petterssen and Smeybe (1971) type B cyclogenesis. Though 8 of the 30 bombs did form at $T_0$, the others were characterized by antecedent propagation into the region of rapid deepening. As we will see at the end of this section, one of these bombs formed in association with a 500-mb trough that was distinct from the trough locking onto the surface low during its explosive deepening phase.

The tracks shown for the rapidly developing cases in Figs. 4b and 5b appear similar to that shown for the QE II storm by Gyakum (1991, Fig. 2) in which the formation of the surface vortex is followed by 12 h of development and subsequent interaction with a well-developed 500-mb trough in the polar jet. Ogura and Juang (1990, Fig. 16) showed a case of rapid cyclone intensification that was also characterized by an interaction with a developing shallow vortex with a high potential vorticity anomaly in the upper troposphere. As we will see later, some, but not all, of these cases form independently of the upper trough that subsequently interacts with the low during the explosive development. The independent development of two disturbances, one in the middle troposphere and one in the lower troposphere, and their subsequent coupling have been studied numerically by Takayabu (1991). He finds this coupling to be an efficient means of achieving rapid cyclone intensification, and illustrates the result with a case of two disturbances whose tracks (his Fig. 18) are similar to those seen in Figs. 4b and 5b.

We now investigate the intensification characteristics of these 62 cases. Figure 6 shows the central pressure traces of the 30 bombs and the 32 weakly developing
Sample 1 surface cyclone and 500 mb tracks (31 cases) (a)

Rapidly-developing cyclone and 500 mb tracks (16 cases) (b)

Weakly-developing cyclone and 500 mb tracks (15 cases) (c)

Fig. 4. Tracks of 500-mb vorticity maxima (gray) and surface cyclones (solid black) for the sample 1 cases for the period from 48 h prior to the onset of maximum deepening through 12 h after this onset. Circles and squares represent the respective surface cyclone and 500-mb vorticity maximum positions at $T_0$. The panels include (a) the total sample of 31 cases, (b) the subset including only explosively developing cyclones (16 cases), and (c) the subset including only those cases that did not develop explosively (15 cases).
Sample 2 surface cyclone and 500 mb tracks (31 cases) (a)

Rapidly-developing cyclone and 500 mb tracks (14 cases) (b)

Weakly-developing cyclone and 500 mb tracks (17 cases) (c)

Fig. 5. As in Fig. 4, except for the sample 2 set of cyclones, including (a) the total sample of 31 cases, (b) the subset of 14 explosively developing cyclones, and (c) the subset of the 17 other cases.
where $\bar{p}$ is the mean of the surrounding four sea level pressure values, located a distance $\Delta$ north, south, east, and west of the low center, and $p_c$ is the central pressure. The density used in these calculations is 1.2 kg m$^{-3}$, and the Coriolis parameter $f$ is that shown for the latitude of the cyclone center. Given a $3^\circ$ latitude computational mesh, centered at $37.5^\circ$N, a 1-mb central-pressure error would yield an error in $\xi_g$ of $0.34 \times 10^{-4}$ s$^{-1}$. We regard this number as the noise level of our geostrophic-vorticity calculations.

Figure 7 shows the time traces of $\xi_g$ for the rapidly and weakly developing cases. Since the geostrophic relative vorticity represents the circulation divided by the area, we also show in Fig. 7 the tangential geostrophic wind component, averaged around the surface low at a distance $\Delta/2$. This tangential wind is thus $\xi_g$ times $\Delta/4$. These averaged geostrophic values may well be substantial overestimates of actual mean surface winds.

The greater central pressure falls and generally longer periods of development, prior to $T_0$, are associated with bombs. While central pressure has traditionally been used as a parameter in classifying cyclones, it is more meaningful dynamically to examine the vorticity field during the cyclone life cycle. To do so, we examine the geostrophic relative vorticity, as measured from the sea level pressure field. We use the NMC final Northern Hemisphere surface analyses to find the sea level pressures at the cyclone center and at four surrounding points located $3^\circ$ latitude north, south, east, and west. Such a computational spacing is comparable to that used in the NMC gridpoint analyses shown in Figs. 2 and 3. The geostrophic vorticity $\xi_g$ is computed with the finite-difference form of the expression

$$\xi_g = \frac{4(\bar{p} - p_c)}{\rho \Delta^2},$$

(5)
and vorticities, because of cyclonic air-parcel trajectories and frictional effects. The time trace of $\zeta$ is a more relevant proxy for the dynamical behavior of these surface systems than is seen simply from the central pressure traces (Fig. 6). Consequently, Fig. 7 may be compared with the idealized time trace of $\zeta$ shown in Fig. 1.

The rapidly developing systems are generally characterized by values at $T_0$ ranging from $1.0 \times 10^{-4}$ to greater than $3.0 \times 10^{-4}$ s$^{-1}$. Even the weakly developing cases show $\zeta$ values ranging from $1.0 \times 10^{-4}$ to $2.0 \times 10^{-4}$ s$^{-1}$. Since these geostrophic vorticities are calculated on a relatively large-scale grid mesh, these results suggest that throughout the life cycle of a surface cyclone, $\zeta$ should not be neglected in comparison with the Coriolis parameter in the vorticity equation (2).

To view these cases from the perspective of other rapid cyclogenesis cases published in the literature, we compute geostrophic relative vorticities on the same 3° latitude mesh at $T_0$ for the QEI cyclone (Gyakum 1983), the eastern Pacific case of 13 November 1981 (Reed and Albright 1986), and the Experiment on Rapidly Intensifying Cyclones over the Atlantic (ERICA) cyclones of 14 December 1988 and 4 January 1989 (Sanders 1990). The respective values of $\zeta$ (10$^{-4}$ s$^{-1}$) from these research analyses are 1.1, 2.9, 2.0, and 2.1, all within the range shown in Fig. 7a.

There are six cases that formed at least 36 h prior to the onset of most rapid development. As shown in Fig. 7, only one of these cases was not a bomb. The five bombs in this category developed to vorticity values ranging from $2.2 \times 10^{-4}$ to $3.7 \times 10^{-4}$ s$^{-1}$ at $T_0$. Such antecedent development is not observed in any of the weakly developing cyclones. We investigate whether there was any association between the incipient formation of these systems and the same 500-mb vorticity maximum associated with the 24-h period of most rapid deepening. One of the bombs was associated with one 500-mb trough, associated with the subtropical jet, for the entire antecedent and maximum deepening phases. The other four bombs were associated with subtropical troughs during at least part of their antecedent development phase. Their intensification after $T_0$ was associated with separate 500-mb cyclonic disturbances associated with the polar jet. Such important antecedent development that was not directly related to the relevant 500-mb trough during the most rapid development period has also been observed in the QEI (Gyakum 1991) and Presidents’ Day cyclones (Bosart 1981; Uccellini et al. 1984, 1985). The one case of weak development was initially associated with a disturbance in the subtropical jet; the surface cyclone apparently propagated ahead of the downstream 500-mb ridge, thus explaining its lack of intensification prior to $T_0$ (Fig. 7b).

The presence of these five cases of explosive intensification suggests that even relatively weak surface convergence and interaction with one or more upper troughs, when integrated over a relatively long time period, may be present in some bombs. Figure 7, along with (2), also suggests that other cases of rapid intensification may be more accurately characterized as having large surface convergence occurring over shorter time periods.

An illustration of the synoptic-scale environment for one of the six cases whose initial formation occurred at least 36 h prior to $T_0$ is seen in Fig. 8. The case is one in which the maximum 24-h intensification was nearly 1.5 berghorns. All of the central pressures noted in this discussion are derived from the NMC final manual analyses that benefit from all available surface data. Such analyses are truncated considerably to form the NMC gridded analyses shown in Fig. 8. The time of most rapid deepening onsets $T_0$ is 1200 UTC 20 March 1984. As shown by Fig. 6, the 48-h period of antecedent deepening is characterized by a slow but steady deepening from 1007 to 986 mb. The 1007-mb center 48 h prior to $T_0$, $T_{-48}$, is shown by Fig. 8a to be located within a surface trough extending from the subtropical latitudes. It is also located downstream of a 500-mb vorticity maximum (indicated by the X as $9 \times 10^{-4}$ s$^{-1}$). The surface system, having deepened to 1001 by $T_{-36}$, is still located downstream of the southernmost 500-mb vorticity maximum (Fig. 8b). By 24 h prior to $T_0$, the 1000-mb center is nearly collocated with the 500-mb trough axis. During the subsequent 12 h, the second vorticity maximum, seen as the $16 \times 10^{-4}$ s$^{-1}$ center in Fig. 8c, begins to interact with the surface cyclone as it deepens to 994 mb by $T_{-12}$ (Fig. 8d). The first 500-mb trough, evidently responsible for the system’s initial formation, has moved to 40°N, 146°E, a position to the northeast of the surface low by this time. The 986-mb center at $T_0$ (Fig. 8e) continues to be located in a region of 500-mb cyclonic-vorticity advection associated with the second 500-mb trough. The surface cyclone begins to deepen rapidly to 972 mb by $T_{12}$ in response to the 500-mb trough that has also amplified to $21 \times 10^{-4}$ s$^{-1}$. This sequence of events has been shown to illustrate one means by which a relatively long period of antecedent surface development is accomplished. The surface system benefits from its interaction with two separate 500-mb vorticity centers for 72 h, a characteristic not observed for any of the weakly developing systems.

4. Statistical evaluation of the North Pacific sample

To consider whether surface vorticity development is an important conditioning process for subsequent cyclone intensification, we examine all cases from the North Pacific surface cyclone sample (Gyakum et al. 1989) whose entire life cycle was contained within the domain and whose existence could be documented for at least 12 h prior to the onset of the 24-h period of maximum deepening. The number of such cases is 794 for the 1975–84 period. As a proxy for the relative
vorticity, $\zeta$, we define the *antece dent deepening* as the difference between the cyclone central pressure at its initial time of appearance and its central pressure at $T_0$. This particular deepening is adjusted geostrophically as in (3), except that the latitude, $\phi$, is the average during the cyclone's antecedent phase. We use this antecedent deepening as a predictor for the cyclone's subsequent 24-h maximum deepening.

The result, shown in the form of a scattergram and a linear regression for the 794 cases, is shown in Fig. 9a. A positive relationship between the antecedent and maximum 24-h deepening is shown for these cases. The linear correlation coefficient $r$ is .62, so that 39% of the total variance can be accounted for by the relationship. The implication of this result is that stronger 24-h cyclone intensification tends to be associated with stronger antecedent intensification. In fact, if we knew where a particular surface cyclone deepens maximally in a 24-h period, and if we knew the low's prior central pressure history for at least 12 h, then we could use the antecedent deepening as a viable forecasting tool. We also find that the positive relationship between antecedent deepening and 24-h maximum deepening holds for regions where rapidly developing cyclones are especially prominent (south of 50°N) and for those regions where bombs are nearly absent (north of 50°N). Figure 9b shows a subset of the 794-case sample, defined by including those cases existing for only 12 h prior to their maximum intensification. This subset is chosen to show the antecedent deepening rates. Though the linear correlation coefficient is virtually unchanged, the slope of the regression line is larger for this sample. We see that the vast majority of antecedent deepening is nonexplosive, even for those systems that subsequently develop explosively.

Though Fig. 9 does show a positive relationship be-
between the antecedent and subsequent deepening, there is a large number of cases without as much as 12 h of existence prior to the maximum 24-h intensification. Indeed, the number of cyclone cases whose maximum 24-h development commences “immediately” (using 12-h tabulations) is 1093. However, this sample has a mean maximum 24-h central-pressure fall of only 8.0 mb, with a standard deviation of 10.9 mb. The sample of 794 cases with a history of at least 12 h prior to the onset of maximum deepening has a mean 24-h pressure fall of 14.5 and a standard deviation of 11.4 mb. The application of the Student’s t-test reveals there to be less than a 1% chance that these two samples were randomly drawn from the same population of surface cyclones. We conclude that the sample of cases with a history of existence prior to maximum deepening is characterized by significantly larger maximum deepening rates. This result provides us with an a posteriori justification for considering the antecedent deepening rates as predictors of subsequent intensification.

We also find that, of the 1887 cases of cyclones, 397 (21.0%) had a maximum deepening rate of at least 1 bergeron. This bomb (Sanders and Gyakum 1980) criterion is met for only 151 (13.8%) of the 1093 cases with less than 12 h of existence prior to maximum deepening, while this criterion is met for 246 (31.0%) of the 794 cases characterized by at least 12 h of antecedent existence. We may apply the binomial test to estimate the probability of rapid cyclone intensification; our sample sizes are sufficiently large so that the binomial distribution approaches the normal distribution. The null hypothesis, based upon the total population of 1887 cases, is that the probability of a bomb event is 21%. If we apply this hypothesis to each of the two samples, we find the expected number of bombs to be 230 and 167 for the 1093-case and 794-case samples, respectively. The sample standard deviations are, respectively, 13.5 and 11.5. The sample with less than 12 h of existence prior to maximum deepening contains 151 bombs; this number is 5.9 standard deviations
fewer than the expected 230. Conversely, the sample of cases with at least 12 h of antecedent existence has 246 cases of rapid cyclone intensification; that is, 6.9 standard deviations larger than expected. The probability of such observations occurring with the given null hypothesis is less than 0.01%, and we reject the null hypothesis that the probability of rapid cyclone intensification is 21%. Alternatively, each of the two samples must be considered as part of two distinct populations: the sample of cases with less than 12 h of existence has only 13.8% of its number rapidly intensifying, and the sample with at least 12 h of history prior to maximum deepening has 31% intensifying rapidly. We conclude that, although bombs do sometimes occur immediately (within 12 h of initial detection), the probability of such an occurrence is significantly less than that expected from a sample of cases with a history of development prior to maximum deepening.

Figure 10 summarizes these results and shows a stratification of the 794-case sample into age at the onset of maximum deepening. The numbers of both cyclones and bombs monotonically decrease with age. The probability of rapid cyclone intensification for each of the five categories of finite history ranges from 26.5% for 12 h to 36.2% for 48 h. We find there is no substantial difference in the probability of explosive cyclone intensification among the varying times (of at least 12 h) of antecedent development. Our results, showing preferentially longer times of antecedent development for rapidly developing cyclones, are consistent with our earlier analyses of a smaller sample of cases.

5. Analysis of the vorticity equation following the low center
To fully analyze the inviscid forcing for absolute-vorticity change (2), we must consider both the effect
of surface convergence and absolute vorticity during the period of development. The objective of this section is to show that each of these effects is germane to development, and to discuss the role that friction may be playing in these systems.

Data points representing the cyclone dynamics at a more mature stage of development (where frictional influences presumably become important) might be expected to differ substantially from those obtained during a previous stage. As an example, we show in Fig. 11 the 6-h divergence and geostrophic vorticity change data from the Canadian operational numerical weather prediction model forecasts (the Regional Finite-Element Model, or RFE) and analyses for the particular case of the ERICA intensive observation period (IOP) 2 cyclone (see Sanders 1990). Analyzed geostrophic vorticity was computed on a comparable scale to the model output data (available on a polar stereographic grid with 100-km resolution at 60°N). In the figure, the relationship between the boundary-layer divergence and vorticity change is markedly different for the points associated with the mature stage of the cyclone (corresponding to times at and after the end of the explosive development—1200 UTC 14 December 1988).

Accordingly, we stratify the 62 Pacific cases defined in section 3 into two samples of 31 cases on the basis of an evaluation of the parameter $(\tilde{\zeta}_{g12} - \tilde{\zeta}_{g0})/\tilde{\zeta}_{g0}$, where $\tilde{\zeta}_{g0}$ and $\tilde{\zeta}_{g12}$ refer, respectively, to the geostrophic relative vorticities at the times $T_0$ and $T_{12}$. One sample consists of the 31 cases with the largest values of this parameter, representing systems with strong development relative to the initial intensity. The other sample contains 31 cases with the smallest values, representing systems in which the development is relatively weak with respect to the initial strength of the cyclone. As we are not able to evaluate directly the surface convergence at the low center for this sample of cases, we use the right-hand side of (4), the $\mathbf{Q}$-vector divergence at 500 mb, as a proxy for the surface convergence. The surface convergence $-\nabla \cdot \mathbf{V}$ may be expressed from the continuity equation as $\partial \omega / \partial p$. Though it would be desirable to use $\mathbf{Q}$-vector divergences at lower levels, it is not possible to resolve adequately the vertical tilt of the cyclone between the surface and, say, 850 mb, with the NMC gridded dataset over this data-sparse marine area. Given that forcing for surface cyclogenesis is related to the equivalent barotropic effect of upward increase in cyclonic-vorticity advection, the 500-mb level is often assumed to be approximately the level of maximum ascent. Additionally, Sanders (1986) has used the 500-mb vorticity advection and maximum central pressure falls as respective proxies for the right- and left-hand sides of (2), and found an excellent relationship between this 500-mb vorticity-advection effect and explosive surface cyclone intensification over the western North Atlantic basin. Though the ascent maxima may be considerably lower, diabatic processes may be important, and surface response to any forcing is modulated by static stability, we gain an a posteriori assessment of the relevance of the 500-mb level to surface intensification with this exercise.

We focus our attention on the 12-h time period subsequent to $T_0$. For this time period, we compute both the right-hand side of (4) and $\tilde{\zeta}_g$ for each case. An arithmetic mean of each quantity is found from the two times, $T_0$ and $T_{12}$. We compute the right side of (4) at 500 mb on the NMC octagonal grid, and then interpolate the value over the cyclone center, defined by the location on the NMC final surface analyses, at each of the times, using a 16-point Bessel interpolation scheme.
in Fig. 12a is likely due to random errors in the 500-mb height analysis, as suggested by Sanders, the evidence presented thus far suggests that some of this variability can be explained by variations in surface relative vorticity. Such variability would not be seen in a composite regression. Indeed, our correlation of .58 is even larger than Sanders’ finding of .45 (1991, personal communication) for individual bombs over the western North Atlantic basin.

The results for the weak-response sample, shown in Fig. 13, show only negligible positive correlations between both 500-mb Q-vector forcing and $\zeta_s$ and the low development. Such a result suggests that frictional processes may be playing a role in mitigating the surface low’s development in response to 500-mb Q-vector convergence or to its intensity. That there are no discernible differences between the values of either 500-mb forcing or surface $\zeta_s$ in Figs. 12 and 13 suggests that friction may be playing a more important role in the weak-response sample.

6. Model results

We examine in this section a set of North Atlantic cyclogenesis forecasts produced by the RFE model. We use the model as an objective analysis tool yielding dynamically consistent, high-resolution datasets in order to refine certain aspects of the investigation of the idea of antecedent vorticity conditioning presented in previous sections. Such an approach has been discussed by Keyser and Uccellini (1987).

Implicit in the concept of vorticity preconditioning is interaction between the cyclonic low-level vorticity field and the upper-level disturbance. Thus, one might expect to see an improved explanation of vorticity

Figure 12 shows the scattergrams for the strong response sample. The Q-vector forcing and geostrophic vorticity both represent good diagnostic parameters for the development of the system. We judge the respective linear correlation coefficients of .58 and .69 to be significantly different from 0 at the 99% level of confidence, using the analysis of variance technique (Panofsky and Brier 1958). The result shown in Fig. 12a is consistent with the findings of Sanders (1986), though our interpretation is different. His composite regression analysis of explosively developing cases shows a nearly perfect linear fit between the magnitude of 500-mb vorticity advection and the associated central pressure fall. Indeed, we also get a similarly excellent linear fit when we composite the cyclones according to deepening rate. Such compositing, however, masks the case-to-case variability. Though some of this variability seen

**Fig. 11.** Scattergram of the 6-h geostrophic vorticity change versus 1000-mb divergence for the RFE model simulations (black dots) of ERICA IOP 2. Squares are taken from the Sanders (1990) surface pressure analyses and the corresponding RFE initial analyses of 1000-mb divergence. The linear regression is based upon all of the points except those that are labeled. These labels show the time (day and hour) at the end of the 6-h period and either range of forecast or "Anal" for analysis.
Fig. 12. Scattergram of geostrophic vorticity change ($10^{-5}$ s$^{-1}$) from $T_0$ to $T_{12}$ as a function of (a) time-averaged $Q$-vector convergence ($10^{-10}$ mb$^{-1}$ s$^{-1}$), as defined in the text, and (b) time-averaged $\ell^* (10^{-4}$ s$^{-1}$). The linear regression line, along with the associated correlation coefficient squared $r^2$, are also shown. The 31-case sample represents a strong response for the strength of the low.

change (pressure change) when this interaction is taken into account. In order to study this problem, a subset of the 52 Atlantic cases discussed in Roeber (1990) for which gridded forecast data were available from the RFE digital archive was used to form the analysis sample in this section. We perform these analyses using the boundary-layer divergence (computed directly from the gridded wind data), since this quantity is directly tied to the dynamics through (2). The diagnosed quantities are averaged in a $3 \times 3$ grid box (approximately 300 km on a side at 60°N), centered on and moving with the surface cyclone.

Since the model data were obtained for the period corresponding to the 12-h maximum analyzed deepening, a sampling problem can occur; namely, because of errors in model timing of the development, the diagnostics may include data prior to or following the time of maximum model cyclogenesis. As shown in section 5, data points representing the cyclone dynamics at the mature stage can differ substantially from those obtained during the development stage. Although it was not possible to assess directly such timing problems using the 12-h model output available for the full model sample, detailed examination of a particular run from ERICA IOP 2 showed the forecast to lead the observations by as much as six hours. Accordingly, the parameter $(\ell_{12} - \ell_0)/\ell_0$ was used to restrict further the 12-h model data to the development stage for the full sample, where $\ell_0$ and $\ell_{12}$ refer, respectively, to the model forecast relative vorticity at $T_0$ and $T_{12}$. Only

Fig. 13. As for Fig. 12, except for the 31-case sample representing a weak response for the strength of the low.
those data corresponding to parameter values within the upper 50th percentile were included in the analysis that follows. This approach has the effect of removing those points for which the 12-h vorticity change was small for a given level of initial vorticity; cyclones with large values of initial vorticity must experience correspondingly large surface vorticity increases to be included in the sample. For example, the outliers highlighted in Fig. 11 for ERICA IOP 2 are all removed by observance of this procedure.

Figure 14 shows the distributions of the upper (500 mb, upstream of the surface disturbance) and lower (1000 mb) cyclonic-vorticity maxima at the onset of most rapid deepening. These data indicate that the disturbances are substantial at both levels relative to the planetary vorticity in many cases, providing further justification for retaining $\delta$ in (2).

We show in Fig. 15 the cyclone 12-h vorticity change as a function of the divergence, the product of 1000-mb absolute vorticity, and the divergence and the product of the planetary vorticity and the divergence, averaged over the 12-h period. Large-scale baroclinic forcing (as indicated by the divergence alone) is clearly important (Fig. 15a), but the magnitude of the non-negligible relative vorticity at the onset of development in the lower atmosphere appears to positively condition the response to that forcing (Fig. 15b). We note that the sample model data include cyclones positioned at latitudes ranging from 30° to 60°N, representing a considerable range of planetary vorticity. Consequently, one might expect that the improved results in Fig. 15b might reflect conditioning by the planetary vorticity alone. However, when such a regression is computed using the product of the planetary vorticity only and the divergence (Fig. 15c), the relationship deteriorates substantially. These results are not dependent upon the existence of extreme values of either convergence or of the product of convergence and vorticity.

Roebber (1990) found that considerable variability can exist in successive model forecasts of cyclogenesis, and suggested that uncertainty in initial conditions might play a crucial role. It might be expected that some of this variability is related to differences in upper-level forcing and static stability between model runs, which in turn is dynamically tied to the boundary-layer divergence. Here, we wish to examine the relationship of the variability more generally to variations in the antecedent vorticity conditioning. Ensembles of the four successive forecasts of 12-h cyclogenesis at time lags of 0–12, 12–24, 24–36, and 36–48 h were compiled for those cases that met the development criteria outlined above. The standard deviation of the four forecasts of 12-h 1000-mb vorticity change is plotted versus the standard deviation of the successive forecasts of 1000-mb divergence (Fig. 16a) and the standard deviation of the successive forecasts of 1000-mb relative vorticity at the onset of the 12-h period of development (Fig. 16b). The variations in the forecast 12-h vorticity changes are related to variations in both the antecedent vorticity development and the boundary-layer divergence. Since the boundary-layer divergence is in part a manifestation of upper-level forcing (such as 500-mb vorticity advection), this relationship between variable forecasts of development and variations in the vorticity conditioning is consistent with the view of the cyclogenetic process as an interaction between cyclonic disturbances at low- and midlevels of the troposphere. Both disturbances contribute to the cyclogenesis and variations in the forecasts of their magnitudes and consequent interactions result in differing forecasts of surface cyclone development.

As a final comment concerning these modeling results, we wish to refer briefly to the forecasts for a specific case, the cyclone of ERICA IOP 2. As noted by Roebber (1990), this cyclone represents one of the most extreme cases for the 1988/89 western Atlantic cold season, both from the point of view of analyzed 12-h cyclone central pressure fall (30 mb) and in terms of the variability of the sequence of successive model
forecasts of its development. The details of this case will be presented in a subsequent paper. Here, we will use this case only as a means of illustrating some of our main points concerning the concept of antecedent conditioning.

We show in Fig. 17 time traces of surface geostrophic relative vorticity and 1000-mb convergence for this storm, beginning 12 h prior to the onset of rapid deepening (0000 UTC 14 December 1988). The analyzed geostrophic vorticity was derived from manual surface analyses (see Sanders 1990 for a description), using a 1.5° latitude mesh. The model geostrophic vorticity was computed using twice the horizontal grid of the model (representing a mesh of approximately 174 km at this latitude). In both cases, the surface pressure fields were converted hydrostatically to 1000-mb height, and the vorticity was computed as

$$\zeta_0 = \frac{4g(Z - Z_c)}{f\Delta^2},$$

where $[Z]$ is the arithmetic average of the four surrounding grid values of 1000-mb height, $Z_c$ is the geopotential at the storm center, $\Delta$ is the grid-mesh length, and $g$ is gravity.

Notable is the substantial vorticity present in the analysis 6 h prior to the onset of rapid deepening, with a geostrophic vorticity of more than three times the Coriolis parameter. The large difference in analyzed versus model vorticity by 1200 UTC 14 December is largely due to a highly concentrated central-pressure core in the analysis that was not completely represented in the smoothed model data. However, as noted by Sanders (1991, personal communication), the large geostrophic vorticity diagnosed from the analysis is somewhat uncertain, owing to the lack of wind observations within 250 km of the center at this time.

The model traces show that, although the differences in relative vorticity between the 0–24-h and 12–36-h forecasts were initially small (approximately 0.47), the discrepancies grew to nearly three times the Coriolis parameter by 0000 UTC 14 December. Overall, the profiles of surface vorticity and convergence are remarkably similar between these two runs, with the differences in relative vorticity arising from consistently higher low-level convergence in the 12–36-h forecast in the 12 h leading up to the time of maximum development (concurrent with the main interaction of the surface cyclone with the upper-level trough). The vorticity profiles of the 12–36-h and 0–24-h forecasts are broadly similar in character to the idealized, inviscid vorticity growth shown in Fig. 1. This suggests that an understanding of the evolution of these forecasts is dependent upon considering the 12 h preceding the rapid deepening, when this antecedent vorticity development occurred.

The model 500-mb and surface fields for these same forecasts at 0000 UTC 14 December are shown in Fig. 18. Qualitatively, the 500-mb flow is similar in these
two forecasts. Both forecasts at this time display a sharp 500-mb trough with its associated absolute vorticity maximum that has crossed the coast near Cape Hatteras. Indeed, root-mean-square errors in the forecast 500-mb height fields, relative to the initial analysis for the region shown in the figure, ranged from 13.6 to 16.3 m. Since estimates of the precision of measurement at 500 mb of a typical radiosonde range from 10 to 12 m (Benjamin 1989; Dey and Morone 1985), these differences are near the noise level of the analysis in some of the runs.

However, it is clear that some details contained within the respective 500-mb fields are quite different. Most prominent is the highly ageostrophic cyclonic-

anticyclonic vorticity couplet downstream of the trough, directly above the surface cyclone and its associated inverted trough. The cyclonic-vorticity maximum was produced largely through vertical advection of vorticity from lower levels in the strong ascent region near the surface cyclone center, while the vorticity minimum developed in response to upper-level ridging associated with the hydrostatic effects of horizontal advection and latent heat release between the surface and 500 mb. We note that a similar vorticity couplet was initialized in the nested-grid model (NGM) 00-h forecast at 0000 UTC 14 December. Owing largely to the strong vorticity gradient at 500 mb between the upstream trough and the downstream short-wave ridge associated with this vorticity couplet, the inferred cyclonic-vorticity advection resulting from the 500-mb geostrophic flow along the middle Atlantic coast is somewhat stronger in the 12–36-h forecast. During the rapid deepening stage of development in the first few hours of 14 December, vertically averaged values of Q-vector divergence between 700 and 500 mb, averaged horizontally in $3 \times 3$ grid boxes centered on and mov-

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Fig. 18. RFE output at 0000 UTC 14 December 1988 for (a) 500-mb 24-h forecast, (b) surface 24-h forecast, (c) 500-mb 12-h forecast, and (d) surface 12-h forecast. The 500-mb charts show lines of constant absolute vorticity (dashed, interval of $5 \times 10^{-5}$ s$^{-1}$) and geopotential height (solid, 6-dam interval). The surface charts show sea level pressure (solid, 4-mb interval), temperature (dashed, 5°C interval), 700-mb relative humidity greater than 70% (light stipple), and precipitation (dark stipple). Wind bars are shown using the standard meteorological convention [pennant = 25 m s$^{-1}$, full (half) barb = 5 (2.5) m s$^{-1}$].

Thus, the effects of the antecedent conditioning include the direct effects of the integrated convergence and implied surface developments prior to the interaction with the 500-mb trough shown in Fig. 18, and the indirect effects resulting in the amplification of the forcing from that upper trough during the rapid deepening stage. Both of these effects are illustrated in Fig. 19, which shows the fields of 1000-mb relative vorticity and divergence at 0000 UTC 14 December for the two model forecasts. The strongly convergent flow (maximum $22.7 \times 10^{-5}$ s$^{-1}$) superimposed upon an envi-
Fig. 19. RFE output at 0000 UTC 14 December 1988 for (a) 24-h forecast of 1000-mb relative vorticity, (b) 24-h forecast of 1000-mb convergence, (c) 12-h forecast of 1000-mb relative vorticity, and (d) 12-h forecast of 1000-mb convergence. Contour interval is $5 \times 10^{-3}$ s$^{-1}$ for both quantities. Stippled regions indicate areas of anticyclonic relative vorticity and divergent flow.

Environment characterized by substantial cyclonic relative vorticity (maximum $23.9 \times 10^{-3}$ s$^{-1}$) depicted in Figs. 19a,b results in a spinup much more intense than is the case for the weaker low-level convergence (maximum $15.0 \times 10^{-3}$ s$^{-1}$) operating on the more weakly cyclonic field of vorticity (maximum $15.3 \times 10^{-3}$ s$^{-1}$) of Figs. 19c,d.

Finally, we note that both forecasts captured the gross characteristics of the sea level pressure analysis at 0000 UTC 14 December (Sanders 1990), but missed the complexity of the mesoscale sea level pressure field. It may be that not all details are equal as far as cyclogenesis is concerned. A key to understanding the model behavior is to determine which processes were particularly relevant to the subsequent development, and so determine how the model responds to some details while remaining relatively insensitive to others. This problem has particular relevance to the idea of model sensitivity to initial conditions; the introduction of additional observations may have a significant impact on the subsequent forecast provided it helps to determine a relevant physical process during the antecedent stage.

These model results suggest that forecast development can be conceptually viewed as an interaction between upper and lower cyclonic disturbances and that the often exceptional response to a particular magnitude of upper forcing may be related in part to the conditioning of the lower atmosphere in terms of cy-
clonic vorticity. Furthermore, the variability in model forecasts is itself tied to variations in the representation of features at both these levels.

7. Concluding discussion

We have investigated the idea that antecedent surface vorticity development is an important conditioning process for explosive cyclone intensification. While the research focus on bombs has traditionally been the 12- or 24-h period of most rapid development (Sanders and Gyakum 1980; Roebber 1984), we have concentrated our study on the time period prior to most rapid development. An examination of the vorticity equation (2) following a low pressure center shows that, in the presence of a preexisting surface disturbance, a case with only 6 h of antecedent, inviscid forcing (surface convergence) yields a far greater subsequent development rate than is seen for another case whose formation and development was delayed by 6 h (Fig. 1).

Though such a theoretical result is not new, the systematic identification of such surface vorticity conditioning in observations and modeling is novel. Qualitatively similar dynamical behavior has been shown for some cases of explosive cyclone intensification in the western North Pacific (Fig. 7a) and for ensembles of successive forecasts of a case of cyclogenesis during ERICA (Fig. 17). The latter result suggests that small errors in the surface circulation of a low may amplify into significant forecast errors only 12 h later.

Additional evidence is presented to support the hypothesis that antecedent vorticity development is germane to subsequent explosive development. A composite examination of eight cases of rapid, compared with nine cases of weak, cyclone intensification over the Kuroshio Current in the western North Pacific basin shows that the former composite cyclone has substantially stronger circulation and vorticity at the onset of explosive intensification than is shown by its weaker composite cyclone counterpart in the same region. The preferential tendency for explosively developing cyclones to form, develop, and travel into the region of most rapid development is shown for the 62-case sample in the western North Pacific (Figs. 4 and 5). Such behavior is generally not observed in the weakly developing cases. Central-pressure traces of these same 62 cases show the storms to be preferentially associated with antecedent development. Time traces of geostrophic relative vorticity, computed on a 3° latitude mesh, show five bombs with at least a 36-h history of antecedent development to greater than $2.0 \times 10^{-4} \text{s}^{-1}$ at the onset of most rapid intensification. Such behavior is not observed in an independent sample of weak cases occurring in the same region. One of these bombs, with 48 h of antecedent development, is shown to be associated with two separate 500-mb vorticity maxima (Fig. 8) during its 72 h of development.

Figure 7 also shows that, even in those cases that lack at least 12 h of development prior to maximum deepening, $\zeta_k$ cannot be ignored in comparison with the Coriolis parameter in (2). This result suggests that either the low forms in a background clonic flow or that much of the development occurred on a time scale less than 12 h. Evidently, stretching in the presence of relative vorticity is present throughout the life cycle of both the weakly and strongly developing cases.

An examination of 794 cyclones in the North Pacific basin, whose entire life cycle could be documented, reveals a general trend of increased maximum development as the antecedent deepening increases. Though there is considerable scatter about the least-squares-fit line in this regression, the correlation coefficient of .62 is significant and cannot be ignored. Owing to the strong relationship between these parameters and the fact that most of the antecedent intensification is not explosive (Fig. 9b), such a result may be useful in forecasting explosive developments.

To examine in more detail whether antecedent development is a particularly prominent characteristic of more rapidly developing cyclones, we focus upon all of the 1887 cases in the Pacific that existed for at least 24 h. We find that, even though the majority of cases do develop most rapidly during the first 24 h of existence (1093), this sample of cases is characterized by significantly smaller maximum deepening rates than are found for the sample of cases characterized by at least 12 h of development prior to most rapid intensification. Furthermore, an application of the binomial test demonstrates that a significantly higher percentage (31.0%) of explosively developing cyclones occurs when there is at least 12 h of antecedent development than for those cases (13.8%) with less than 12 h of prior development. Such a robust result for a large number of cases suggests that random analysis errors are not contributing to our conclusion of the importance of antecedent surface vorticity conditioning.

A detailed examination of the relationship between 1000-mb convergence and 6-h geostrophic vorticity increase, as computed from the RFE operational forecast model (Fig. 11), shows an excellent relationship between the two quantities, except for outliers. These latter points apparently represent more limited development, though in the presence of strong convergence, as a consequence of frictional processes. We also find that both 500-mb forcing (Q-vector convergence) and surface-low strength (\(\zeta_k\)) relate well decorrelationally to the low's intensification, for a strong response parameter designed to select cases not strongly influenced by friction (Fig. 12). That such a relationship is weaker for cases in which the response parameter is small (Fig. 13) suggests that friction is acting to mitigate development in the presence of convergence and vorticity comparably strong to those cases seen in Fig. 12.

The modulation of the midtropospheric forcing by the intensity of the surface circulation is seen in Fig. 15, in which the product of convergence and the absolute vorticity specifies the 12-h development better.
than the product of convergence and planetary vorticity. The importance of the modulation of development by the initial surface circulation ($\zeta_0$) is shown by the strong relationship between variability in $\zeta_0$ and variability of the subsequent development (Fig. 16).

The rapid amplification of low center vorticity differences (Fig. 17) during the first 12 h in a case of rapid cyclone intensification during ERICA shows the importance of accurate forecasts during this antecedent phase of development. That the 500-mb forcing is acting on a stronger system at the end of this antecedent period (0000 UTC 14 December 1988) in the better forecast (12–36 h) cycle (Figs. 18a,b and 19a,b) than is seen in the 0–24-h forecast cycle (Figs. 18c,d and 19c,d) suggests the importance of antecedent surface vorticity development in the successful forecast of this case.

Though there have been suggestions in the literature that antecedent intensification may be a process crucial to explosive cyclogenesis (see, for example, Bosart 1981; Uccellini et al. 1987; and Whittaker et al. 1988 for the Presidents’ Day cyclone), such a general assessment has not appeared previously in the literature. Examples of other specific case studies in which the importance of lower-tropospheric vorticity at the onset of rapid deepening has been noted include the eastern Pacific cyclone studied by Kuo and Reed (1988), the QE II storm (Gyakum 1991), the cyclone of 16–18 November 1972 (Rogers and Bosart 1991), and the renewed development of Tropical Storm Agnes in an extratropical environment (DiMego and Bosart 1982).

Our results suggest that explosive cyclogenesis is best characterized by a nonlinear interaction between two well-developed waves that have fundamentally different origins and may initially develop independently of one another. The growth of the low-level vorticity prior to the time of the maximum development represents a conditioning process by which a subsequent extraordinary response to a given baroclinic signal can be achieved.

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