Observations of 250-km-Wavelength Clear-Air Eddies and 750-km-Wavelength Mesocyclones Associated with a Synoptic-Scale Midlatitude Cyclone

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ABSTRACT

Satellite images of a decaying synoptic-scale cyclone over the North Pacific reveal two distinct types of multiple mesoscale cyclonic disturbances. The approximate positions of these disturbances within the synoptic-scale cyclone are determined using analyses from an operational global-scale numerical model.

One type, a set of four 250-km-wavelength eddies, occurred in clear air and represent perturbations within the cyclonic shear side of a 40–45 m s$^{-1}$ upper-level jet–front system. Their propagation at 28.5 ± 3.5 m s$^{-1}$ roughly matched the wind speed and direction within the jet at their position and provided evidence of stretching deformation along their axis. Their growth is documented over 18 h, and is measured in terms of horizontal displacements of a preexisting moisture boundary in water vapor imagery. Their e-folding time increased from 6 to 9 h as horizontal displacements exceeded 100 km and horizontal billows indicative of wave breaking formed. The billow-like structures most likely represent areas of enhanced mixing of stratospheric and tropospheric air by means of a quasi-horizontal process acting in the vicinity of a tropopause fold. Because they developed in a region of significant horizontal shear and because the absence of clouds suggests vertical motions were small or absent, the behavior of these eddies is consistent with barotropic instability on a 50-km-wide shear zone or potential vorticity strip.

The other type, a set of five 750-km-wavelength cyclonic disturbances (mesocyclones), is also evident in the satellite images, but because they modulated the cloud field they appear in both the WV and infrared images. They wrapped fully around the synoptic-scale cyclone, forming a wavenumber 5 perturbation that later became a wavenumber 4 perturbation propagating cyclonically about its center at 19 ± 1 m s$^{-1}$ and likely formed on an occluded front. Two of these waves are distinguished by the development of their own comma clouds, indicating they had deep vertical circulations and suggesting that moist baroclinic instability or CISK was active. These 750-km-wavelength mesocyclones most likely affected the deformation and shear along the upper front, which could have modified the barotropic stability of the region, and thus influenced where the 250-km-wavelength eddies formed.

1. Introduction

The appearance of distinct sets of two types of mesoscale cyclonic disturbances in satellite images of a decaying synoptic-scale cyclone over the North Pacific Ocean on 30 March–1 April 1994 provides an opportunity to document their evolutions and their positions within the larger cyclone. Although very few in situ observations are available for this study, the clarity of the features in the satellite imagery makes the event a useful addition to the limited existing documentation of such phenomena. Although the lack of experimental data restricts this study in many ways, some important characteristics of the disturbances can still be measured directly from the satellite data, including wavelength, propagation, and growth rate. Also, analyses from an operational global-scale numerical model are used to infer the positions of the disturbances within the synoptic-scale cyclone, and some plausible, but tentative, conclusions about dominant growth mechanisms can be made.

The smaller perturbations are evident only in water vapor (WV) images and have wavelengths of 250 km. A somewhat similar event was used by Weldon and Holmes (1991), along with many other types of events, in their extensive and useful description of the interpretation of water vapor imagery. In their case (see their Fig. 56c), a set of four 450-km-wavelength undulations appear along an upper-air moisture boundary similar to the boundary shown here in Fig. 1a. Although they commented that “similar eye features often form when upper-air moisture boundaries around a cyclone weaken, become perturbed, and break down into cyclonic eddies,” they did not examine the event in any detail. The 1000-km-wavelength disturbances along a potential vorticity (PV) streamer studied by Appenzeller and Davies (1992) and other similar disturbances documented by Appenzeller et al. (1996) are also related to the 250-km eddies studied here in that they result from barotropic instability near the tropo-

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pause, although they are much longer and are in a different synoptic-scale environment. They suggested that such eddies may be important regions of mixing of stratospheric and tropospheric air and that the mixing occurs primarily on the 100-km scale, an inference that is better supported by the case studied here. A single vortex documented by Browning (1993), which in some ways resembles this event and the unanalyzed case presented in Fig. 54 of Weldon and Holmes (1991), is the most complete study to date of a mesoscale vortex evident in WV images. Browning used a quasi-operational forecast model to show that the vortices seen in satellite images formed along a line of strong horizontal shear and corresponded to localized PV maxima on 315- and 330-K isentropic surfaces. Although Browning’s study focused on the structure and evolution of the most prominent vortex, a set of vortices was seen to develop with about 900-km spacing (another had 300-km spacing), and it was concluded that the 900-km wavelength was broadly consistent with the wavelength of instabilities that could grow along such a PV strip (Joly and Thorpe 1990). The event presented here extends earlier work by determining the eddy growth rate, showing that they can occur on smaller scales (i.e., 250 km versus 1000 km) and in a significantly different synoptic scale setting, as well as more clearly illustrating their development from initial growth through wave breaking.

The larger disturbances have wavelengths of 750 km, are associated with deep clouds, and occur within the synoptic-scale cold dome. They appear to be most closely related to the multiple mesoscale cyclones (mesocyclones) studied by Forbes and Lottes (1985), Bond and Shapiro (1991), and Ralph et al. (1994). Such mesocyclones can significantly affect the mesoscale distribution of precipitation, strong winds, and large ocean waves (Bond and Shapiro 1991; Ralph et al. 1994). The event presented here complements these

**Fig. 1.** Satellite images providing a preliminary view of the 250-km-wavelength clear-air eddies (labeled a–d) and the 750-km-wavelength mesocyclones (labeled 2–5): (a) 1200 UTC 31 March, WV/IR (MB enhancement); (b) 1200 UTC 31 March, IR (linear gray scale); (c) 0000 UTC 1 April 1994, WV/IR (MB enhancement).
and other earlier studies by documenting the evolution, propagation, and position within a synoptic-scale cyclone of five 750-km-wavelength mesocyclones. These mesocyclones are examined after first describing the data used in this study. This is followed by documentation and analysis of the 250-km-wavelength clear-air eddies and a discussion of their possible relationship to the mesocyclones.

2. Data and methodology

This study uses satellite images from Geostationary Operational Environmental Satellite-7 (GOES-7) such as in Fig. 1 and large-scale analyses from the National Meteorological Center’s (NMC) aviation (AVN) model, which is a spectral (T126) global-scale numerical model that incorporates aircraft reports and cloud-track winds over the data-sparse North Pacific Ocean. These data were viewed using an experimental real-time data display system, the Denver AWIPS (Advanced Weather Interactive Processing System) Risk Reduction and Requirements Evaluation (DARE) workstation, developed and operated at the National Oceanic and Atmospheric Administration’s (NOAA) Forecast Systems Laboratory (FSL). The event studied here was first identified during inspection of events as they occurred and was later reconstructed for review on the DARE workstation, which allowed model-analyzed fields to be overlaid on satellite imagery and then animated. The satellite images, upper-air data, and AVN analyses were all obtained directly from the DARE workstation and drafted into the figures presented here.

Three types of satellite images are used in this paper. The visible channel is centered on 0.7-μm wavelength and has 1-km horizontal resolution, while the infrared (IR) channel is centered on 11.2 μm and has 8-km horizontal resolution. The WV channel is centered on 6.7 μm and has a horizontal resolution of 16 km. However, for this case study, the DARE workstation displays all of the images at 16-km horizontal resolution. While interpreting visible and IR images is relatively straightforward and well known, interpreting WV images is more complicated because the intensity of the radiation measured by the satellite at 6.7 μm depends critically on the vertical distributions of both moisture and temperature. [See Weldon and Holmes (1991) for a description of the use of WV imagery, including quantitative analyses of the radiative transfer behavior that determines the observed radiance in the WV channel.] Because atmospheric water vapor is a good absorber, and hence a good emitter, of 6.7-μm radiation, moisture at upper levels will absorb upwelling radiation and re-radiate it at an intensity determined by the temperature, pressure, and water vapor mixing ratio at that level. Because temperature usually decreases with altitude in the troposphere, regions containing moisture aloft will appear as cool areas (light gray). Where very little moisture is present at upper levels, there is very little absorption and reradiation from those levels, making such areas appear warm (dark) in WV images. On average, the layer centered on 400 mb is the primary source of radiation observed by the satellite at this wavelength, although the radiation can come from anywhere between 800 and 200 mb, depending on circumstances (Muller and Fuelberg 1990). Rarely does this channel see conditions below 800 mb. Temperature inversions and thin layers of moisture in the upper troposphere can also complicate matters (Weldon and Holmes 1991). Although there are important exceptions, it is possible to consider that the brightness in a WV image approximately represents the height of the top of a moist layer. This is most accurate when no clouds are present and it is dry above the moist layer.

In the water vapor images, the regions where high clouds are present are identified by replacing the radiance measured in the 6.7-μm WV channel by the IR brightness temperature T_B, measured at the same location, when T_B is less than a certain threshold. This threshold is −41°C for the MB enhancement used in Figs. 1a, 1c, 3, 5a, and 6. Pixels with −41°C > T_B > −52°C are white and those with −52°C > T_B > −58°C are black. The regions of 6.7-μm radiances that remain in these images are shown using a linear gray scale. These images are referred to as WV/IR images. Figures 1b, 4, and 5b are pure IR images, which use a linear gray scale for regions where T_B > −32°C, and all points with T_B < −32°C appear as white.

3. The synoptic environment

At 0000 UTC 30 March 1994 a synoptic-scale cyclone was located near 40°N, 160°W and had developed a comma cloud that looked typical of a mature, recently occluded midlatitude cyclone (e.g., Houze 1993). Analyses from NMC’s operational AVN model provide a synoptic-scale perspective of the event, which is portrayed in Fig. 2 for conditions at 1200 UTC 31 March. The sea level pressure pattern (Fig. 2a) indicates that the cyclonic geostrophic circulation near the surface had a diameter of approximately 2500 km. In addition, the 1000−500-mb thickness pattern (Fig. 2a) shows deep cold air had already wrapped around the south side of the cyclone, suggesting it was in a mature stage and may have begun decaying. Although there is significant uncertainty in the AVN surface pressure analyses over such a data-sparse region, the analyses support the conclusion that the cyclone is decaying because the analyzed central pressure rose from 987 mb at 0000 UTC 31 March to 993 mb 24 h later, which is roughly the time interval of interest in this study. Because the AVN model only marginally resolves the 750-km-wavelength disturbances and cannot resolve the 250-km eddies aloft, it is used here only to provide an approximate representation of the large-scale conditions and is not used to characterize the structure of
either type of disturbance. Although aircraft reports at upper levels are incorporated into the model analyses, the overall sparsity of observations in the region containing the cyclone implies that even this large-scale perspective may contain significant uncertainties. However, comparison with satellite images indicates it does not grossly misrepresent the synoptic-scale conditions.

4. Observations of 750-km-wavelength mesocyclones

By 1500 UTC 30 March, IR satellite images began showing systematic perturbations to the distribution of
mid- and high-altitude clouds associated with the synoptic-scale cyclone. Temporal loops of hourly images show these initially small perturbations amplified over the next 15 h into the five disturbances labeled 1–5 in Fig. 3b. These evolved into a wavenumber 4 perturbation (Figs. 3c–e) as evidence of disturbance 1 disappeared. [The positions marked by numbers on Figs. 1–4 are based on the center of the warmest (driest) region of each disturbance, as seen in the WV channel, although the positions are offset in some images so as not to obscure important features.] Most of their evolution is portrayed in Figs. 3 and 4, using WV/IR and IR images, respectively. By 1200 UTC 31 March, disturbances 3 and 4 had developed mature comma clouds of their own (Figs. 1b, 3c, and 4c) and thus are referred to as mesocyclones. Although the cloud features associated with disturbances 1 and 2 appear in IR images to have been shallower and less well organized than were disturbances 3 and 4, and disturbance 5 appears in the IR as only a slight arc in middle and high clouds, each one has a relatively distinct signature in the WV/IR images (Fig. 3). The less well-defined disturbances also had morphologies indicative of cyclonic vorticity maxima and moved similarly to the more well-developed perturbations. Considered as a whole, disturbances 1–5 appear as a continuous packet with nearly uniform 750-km spacing, as measured along an arc passing through each disturbance, but only disturbances 3 and 4 developed mature comma clouds. The sparsity of data in their vicinity precludes determining if all had significant circulations and fronts near the surface, although the clouds marking 3 and 4 suggest they did.

Tracks of mesocyclones 2–5, based on their WV signatures every 3 h between 0600 UTC 31 March and 0000 UTC 1 April, are shown in Fig. 2d. While disturbances 3 and 4 were easily followed in the IR images throughout this interval, 2 and 5 appeared best in the WV/IR images, and 1 emerged out of a region of complex cloud structures that was present earlier. Subtle features in the sequence of WV/IR images (Figs. 3a–c) suggest that disturbances 1 and 2 evolved from a set of disturbances that initially had scales of roughly one-half of their final 750-km spacing and that these precursors persisted somewhat even after disturbances 1 and 2 had developed. On the basis of the positions of disturbances 2–5 in the WV/IR images, it is possible to estimate their motions throughout this interval (disturbance 1 was not identifiable after 1500 UTC 31 March). All moved cyclonically about the approximate center of the synoptic-scale cyclone (Fig. 2d). After adjusting for the motion of the synoptic-scale cyclone, assumed to be that of the surface low-pressure center in operational surface analyses (6 m s⁻¹ toward 20° during this time), the motion of mesocyclones 2, 3, 4, and 5 was 20, 19, 18, and 19 m s⁻¹, respectively, relative to the larger vortex. This yields an average of 19 ± 1 m s⁻¹ and indicates the disturbances propagated cyclonically and uniformly about the center of the larger-scale vortex in which they were embedded.

Not only was the propagation of the disturbances rather uniform, animation of the IR satellite images from 1500 UTC 30 March to 1800 UTC 31 March suggests the disturbances all grew at nearly the same time, although some developed more deep cloud than others (as described above), and there is evidence that precursor signatures of mesocyclones 3–5 were present in the southwesterly flow located southeast of the synoptic-scale cyclone center (e.g., follow perturbation 5 in Fig. 3). Comparison with upper-air analyses from the AVN model at 1200 UTC 31 March, after the mesocyclones had become relatively mature, indicates that the model had some indication that they were present. This is seen best in the 500-mb absolute vorticity field (Fig. 2b), where the four major disturbances appear correlated with four lobes of locally enhanced absolute vorticity. This is consistent with the fact that the AVN model, which is a spectral model, has approximate horizontal resolution of 150 km, which is marginally adequate to resolve the 750-km disturbances. Although the extended warm sector of the occluded synoptic-scale cyclone and the area of warm advection are well correlated with the clouds east and north of the cyclone center and with disturbances 4 and 5, disturbance 2 appears to be in the cold sector that had wrapped around to the south of the cyclone center (Fig. 2a). However, the fact that disturbances 2 and 3 are in the same larger comma cloud containing disturbances 4 and 5, combined with the uncertainties in the analyses due to the sparseness of data in the region, suggest that disturbances 2 and 3 could also have been associated with an occluded front or bent-back warm front. This is further supported by comparison with the detailed study (Bond and Shapiro 1991) of two similar-looking 500-km-wavelength polar lows. In the Bond and Shapiro (1991) event, it was shown that the polar lows developed along a zone of strong horizontal shear separating warm air near the cyclone center from colder air wrapped around to its south. The presence of strong low-level baroclinicity in the Bond and Shapiro case, documented by aircraft, and the similarities between that case and the one studied here suggest that baroclinic processes may have contributed to the mesocyclone growth in this event. This is also supported qualitatively by the fact that some of the disturbances developed deep cloud indicative of vertical circulations, a characteristic of baroclinic instability. However, the strong horizontal shears in the Bond and Shapiro case suggest barotropic instability may also have been active. Another study (Ralph et al. 1994) recently suggested that barotropic instability along occluded fronts had produced several documented cases of multiple mesocyclones within synoptic-scale cyclones. These events were characterized by short wavelengths (<300 km), the absence of deep clouds, and clear evidence they were within a synoptic-scale
FIG. 3. WV/IR satellite images at (a) 0000, (b) 0600, (c) 1200, and (d) 1800 UTC 31 March and (e) 0000 UTC 1 April 1994, showing both the 250-km-wavelength clear-air eddies [labeled a–d in panels (d) and (e)] and the 750-km-wavelength mesocyclones (labeled 1–5). A linear gray scale is used with MB enhancement in which white areas represent clouds with cloud-top temperatures colder than −41°C.
cyclone. In the same study, other multiple mesocyclone events were noted that did have deep cloud, were of somewhat longer wavelengths, and thus more likely involved baroclinic instability. Although the inference that deep vertical motions were present in this case supports the baroclinic instability hypothesis, it is also possible that CISK (convective instability of the second kind) could have produced a similar signature, as has been noted in earlier studies of polar lows (e.g., Rasmussen 1979). These issues are left unresolved in the current study but may be addressed in the future if an adequate simulation can be performed using a model capable of representing these processes with limited data for initial conditions.

It should be noted that the overall cloud patterns associated with the group of mesocyclones in this case is remarkably similar to an event presented briefly in Forbes and Lottes (1985), including the fact that both appear as a wavenumber 4 perturbation on a larger vortex (compare this paper’s Fig. 4c with their Figs. 1 and 4). In their event the wavelength was about 400 km. More recently, satellite images of a synoptic-scale cyclone just off the United States west coast on 22 and 23 March 1995 revealed wavenumber 4 perturbations that had wavelengths of 400–500 km. Thus, the behavior seen in the 31 March 1994 event is not unique to this case. However, the fact that Forbes and Lottes (1985) found only one similar event over the northeastern Atlantic Ocean during the 36-day interval they
studied suggests these relatively axisymmetric mesocyclones may only rarely grow to observable amplitude.

5. Observations of 250-km-wavelength clear-air cyclonic eddies aloft

In addition to the 750-km-wavelength mesocyclones documented in the previous section, a set of four 250-km-wavelength disturbances, marked as “a”–“d” in Figs. 1, 2, 3, and 6, developed between 0600 UTC 31 March and 0000 UTC 1 April. These are evident only in the WV images, as is seen by comparison of Figs. 3c and 3d with Figs. 4c and 4d. They appear along a boundary between moist and dry air aloft that is similar to one on 7 April 1988 documented by Weldon and Holmes (1991) and referred to as a synoptic-scale head boundary (see their page 66). Such boundaries are not uncommon and often represent the westward displacement of upper-tropospheric moisture around the north side of a synoptic-scale cyclone (Weldon and Holmes 1991). The boundaries extend from northwest to southwest of the synoptic-scale cyclone and approximately mark a large-scale deformation zone aloft where cyclone-induced easterly flow opposes the background westerlies (Rotunno et al. 1994). Although only a few upper-air observations are available in this case, several found between 50°–56°N and 145°–161°W (Fig. 2c) clearly show that winds in the 300–200-mb layer had a substantial easterly component north of the synoptic-scale cyclone center and were coincident with upper-level moisture in the WV image. [The AVN analysis at 300 mb and data from between 300 and 200 mb are used for comparison with the satellite images because the WV channel is most indicative of conditions near 400 mb and aircraft reports are most common in the 300–200-mb layer.] Consistent with the examples shown in Weldon and Holmes (1991), the upper-air moisture boundary in this event coincided roughly with a deformation zone at 300 mb with a col point northwest of the center of the synoptic-scale cyclone, as inferred in Fig. 2c from the geopotential height field and the few wind observations available. Also, the portion of the 4000-km-long moisture boundary where the upper-level eddies developed is characterized by cyclonic shear associated with a 40–45 m s⁻¹ 300-mb jet.

The narrow dark zone west of and adjacent to the moist region in the WV images is also fairly common and represents a region where subsidence had displaced moisture downward to warmer temperatures. This has been inferred by Weldon and Holmes (1991) on the basis of temporal changes in brightness temperature in conditions very similar to those in this study. Several other studies have found that dry bands in WV images are well correlated with a lowered tropopause (Ramond et al. 1981; Muller and Fuehberg 1990; Appenzeller and Davies 1992; Browning 1993) and with the axis of an upper jet (Ramond et al. 1981). Muller and Fuehberg (1990) used a mesoscale numerical simulation, with a radiation scheme that simulated WV channel observations, to find that warm or dry areas in the WV images did not always represent regions of active subsidence but that these areas were coincident with a lowered tropopause. On the basis of these earlier studies, and the position of the eddies relative to the 300-mb jet (Fig. 2c), it is plausible to infer that the moisture boundary in the vicinity of the eddies occurs within the entrance region of a moderate-intensity (40–45 m s⁻¹) upper-level jet–front system with its associated stratospheric intrusion [for a review of jet–front systems and the tropopause, see Shapiro and Keyser (1990)]. This conclusion is consistent with the recent findings of Rotunno et al. (1994), where it was shown that the region northwest of a cutoff synoptic-scale cyclone is often characterized by deformation frontogenesis aloft resulting from the interaction of the circulation of the synoptic-scale cyclone with the larger-scale background westerlies. They also showed that this process can lead to the development of a narrow “streamer” of enhanced PV. Although evidence for the presence of an upper-level jet–front system is fairly strong, and by inference this may have included a tropopause fold, it remains uncertain whether or not a true tropopause fold was present. It should also be noted that a WV image represents a view of the moisture-weighted temperature profile, and hence the position of the dark band will be affected by how moist the upper troposphere is on either side of the upper jet, on the steepness of the tropopause fold, and on the temperature difference across the upper front [often 10°–15°C at 400 mb (e.g., Palmén and Newton 1969; Shapiro 1978)]. It also appears that a trough aloft, with its own subsidence zone, had moved in from the west and was deformed as it encountered the stronger synoptic-scale cyclone’s circulation on 30 March. The remnants of this region of subsidence may have contributed to the sharpness of the moisture boundary that was observed in the water vapor imagery. Although not as evident in upper-level charts, the moisture boundary in the Weldon and Holmes (1991) event (Fig. 5c) may have also been affected in a similar way. The position of the upper jet in the case studied here, relative to the WV image, clearly indicates that the upper-tropospheric air within the synoptic-scale cold pool is more moist than on the warm side of the upper front.

Comparison of WV/IR, IR, and visible images at 0000 UTC 1 April (Fig. 5) establishes that the waves found along the moisture boundary were present in clear air and were not associated with clouds at that time. Although the visible image (Fig. 5c) reveals scattered clouds near the latitude and longitude of the bilow-like waves in the WV/IR image, they are not organized in any way clearly related to the billows, and the IR image shows they are at relatively low altitude. This suggests the eddies were restricted to altitudes well above the surface. Although a visible image at
1800 UTC (not shown) shows more overcast conditions in the low clouds near the growing eddies, the IR image once again shows that these clouds were below the eddies (compare Figs. 3d and 4d, and 3c and 4c).

Detailed views of the waves are presented in Fig. 6, where their quasi-simultaneous evolution from small undular perturbations (Fig. 6b) at 1200 UTC 31 March to large billow-like eddies (Fig. 6d) at 0000 UTC 1 April is apparent. Unlike the 900-km-wavelength vortices presented by Browning (1993) the eddies in this case appear to consist only of features resulting from horizontal advection. The secondary moist region, suggested by Browning (1993) to result from lifting, is absent in this case. The eddies had a rather uniform wavelength of 250 km, measured from the three crests of moisture marked "a," "b," and "c." As seen at earlier times as well, the curl of the eddies at 0000 UTC 1 April (Fig. 3e) indicates that air parcels on the south side of the moisture boundary moved eastward more quickly than air parcels on the north side, a pattern indicating that the eddies grew in a region of cyclonic shear, which is consistent with the 300-mb analysis (Fig. 2c).

The images can also be used to measure the growth of the waves by measuring the temporal changes in horizontal displacement of the moisture boundary perpendicular to its initially smooth state, as shown in Fig. 7a. Horizontal displacement is measured from crest to trough, and represents the peak-to-peak amplitude ($A$). Between 0600 and 1800 UTC, $A$ increased from less than 10 km to about 85 km, and by 0000 UTC 1 April it had reached 150 km. This evolution is plotted in Fig. 7b. If the growth is assumed to be exponential, then the growth rate can be estimated from $A = A_0 e^{\sigma t}$, where $A_0$ is the amplitude at an arbitrary initial time (as long as there is a measurable amplitude at that time), $\sigma$ is the growth rate, $t$ is time, and $A$ is the amplitude at time $t$. These calculations show that $\sigma = 0.16 \pm 0.03 \text{ h}^{-1}$ from 0600 to 1800 UTC 31 March and $\sigma = 0.11 \pm 0.05 \text{ h}^{-1}$ from 1500 UTC 31 March to 0000 UTC 1 April. The growth rates given here are averages of at least six separate growth-rate estimates from times within the given intervals, based on 3-h sampling. The ranges represent the extremes of these estimates. These correspond to e-folding times of 6.2 and 9.4 h, respectively. Because wave breaking represents a transfer of perturbation energy back to the mean flow, the temporal decrease in growth rate is qualitatively consistent with the interpretation that the curling over of the billows in the satellite images is indicative of wave breaking. However, because wave breaking is an inherently nonlinear process, the simple formulation used to determine growth rate becomes less representative during the wave-breaking interval, and the later growth rate should be considered a rough estimate.

The eddies moved east-southeastward between 1200 UTC 31 March and 0000 UTC 1 April (Fig. 2d). The velocities of eddies a, b, c, and d were 25, 27, 30, and 32 m s$^{-1}$, respectively. This variation could be the result of large-scale, but weak, strain of roughly 7 $\times$ 10$^{-6}$ s$^{-1}$ oriented along the moisture boundary in their vicinity. Both the 28.5 m s$^{-1}$ average velocity of the eddies and their direction of motion (Fig. 2d) were nearly equal to the wind speed and direction at 300 mb (Fig. 2c) at their position. In addition, the tendency for the westernmost eddies to move more slowly than the easternmost is consistent with the fact that wind speed increased from west to east along the axis of the eddies.

Several characteristics of the upper-air eddies are consistent with what should be expected for barotropic shear instability. These include the indication that they resulted from an instability, based on the simultaneous growth of four perturbations, and the favorable comparison between their morphologies and their growth rates. They were in a position characterized by cyclonic shear, based on comparison with the 300-mb isotach.
motion, whereas baroclinic instability does. Under the assumption that barotropic shear instability is responsible for the development of the eddies, it is possible to use their wavelength (250 km) in the theoretical relationships between shear-zone width and the most unstable wavelength to estimate the width of the shear zone. While Haurwitz (1949) showed that the wavelength should be eight times the shear-zone width, more recently Joly and Thorpe (1990) and Schär and Davies (1990) showed that instabilities five times as long as the width of a PV strip would be produced by frontal instability. These suggest that the PV strip, or shear zone, was 30–50 km wide, a width that is unobservable from the sparse data in the region or from the course model analyses (Fig. 2c). Joly and Thorpe (1990) concluded that the shorter wavelengths are dominated by energy conversions characteristic of barotropic instability and that they had relatively small vertical extent (e.g., 3–4 km). However, the minimum wavelengths emphasized in those papers were much longer (700–1000 km) and the growth rates much slower (e-folding times of 1–2 days) than those observed in this case (250-km wavelength and 6-h e-folding time).

The eddies developed on a 1000-km-long westward bulge that formed a portion of the 4000-km-long moisture boundary (Fig. 6a). The northern and southern ends of this bulge remained approximately in phase with two of the 750-km-wavelength mesocyclones (Fig. 3). Although it is unclear what the horizontal circulations associated with the 750-km-wavelength mesocyclones were like at levels that could have affected the moisture boundary, which is aloft, their proximity in the satellite images suggests a possible relationship. The phase relationship is such that mesocyclone 2 could have contributed to the outward deflection of the southern end of the bulge and the inward deflection to its south, where “inward” is toward the center of the synoptic-scale cyclone. Conversely, mesocyclone 3 was oriented so as to contribute to the inward deflection of the moisture boundary in a way that created the northern end of the bulge. It should also be expected that the horizontal shear and deformation along the initial moisture boundary would be modulated by the mesocyclones, and hence could affect the stability of the inferred horizontal shear zone to perturbations such as those that appeared. The influence of a larger-scale deformation, or strain field, on the stability of a shear zone has recently been the focus of much attention and was examined by Dritschel et al. (1991), Bishop and Thorpe (1994), and Juckes (1995).

6. Summary and discussion

Two types of mesoscale cyclonic disturbances were identified in satellite images of a decaying synoptic-scale cyclone. Each consisted of a packet of at least four distinct waves, which provided direct measure-
ments of wavelength. A group of disturbances with average wavelength of 750 km formed a complete ring around the center of the synoptic-scale cyclone. At the same time a set of 250-km-wavelength clear-air eddies developed along an upper jet–front system. This jet–front system was marked by a moisture boundary in WV imagery that developed from the pattern of horizontal moisture advection associated with a large-scale deformation zone and vertical advections due to ageostrophic vertical circulations associated with the upper front and possibly the remnant region of subsided air associated with a separate upper-level trough. All disturbances were evident in WV/IR images, but only the 750-km-wavelength disturbances were associated with clouds. On the basis of the development of comma clouds with some of the 750-km-wavelength disturbances and the implied vertical motions, it was suggested that baroclinic processes or CISK may have contributed to their growth, and hence they are referred to as mesocyclones. They represent a clear example of multiple mesocyclones within a synoptic-scale cyclone. They appear closely related to the polar lows studied by Bond and Shapiro (1991) in that both their scale and cloud morphology are similar, and they likely developed on a low-level baroclinic zone, possibly an occluded front, within a larger cyclone. In contrast, the 250-km-wavelength eddies occurred without clouds, grew very quickly (e-folding time of 6 h), and appeared in association with horizontal shear in the upper troposphere. These attributes point toward the likely role of barotropic instability as their source. Although recent observational and theoretical studies have explored the structure and origins of mesoscale eddies such as these, the 250-km wavelength documented here is significantly smaller than the 800–1000-km-wavelength disturbances that had been examined and the growth rate is significantly greater. In addition, the evolution of the 250-km-wavelength eddies in the satellite imagery uniquely illustrates growth through an instability that leads to wave breaking. On the basis of the nearly constant phase relationship between two of the 750-km-wavelength mesocyclones and the region where the 250-km-wavelength eddies developed, and on their proximity to one another, it is likely that the mesocyclones significantly affected the development of the smaller eddies by modulating the deformation along the shear zone on which the 250-km-wavelength eddies grew.

Because of the limited data available for this analysis, the conclusions concerning growth mechanisms must be considered tentative. To address this uncertainty, it is hoped that a numerical simulation of the event can be performed and that appropriate diagnosis of the model output could test these inferences. However, such a simulation faces significant difficulties: there are few observations upon which an accurate initial condition could be determined or data assimilation performed; the narrowness of the inferred PV strip upon which the clear-air eddies grew indicates the need for relatively fine spatial resolution (<10 km); and the possibility that the 750-km-wavelength mesocyclones determined which part of the shear zone became unstable indicates that the mesocyclones must be accurately simulated in order to simulate the clear-air eddies.

Although earlier detailed aircraft observations provided evidence that vertical transport across the tropopause occurs predominantly on 10-km horizontal scales (Shapiro 1980), it appears that the roll up of the 250-km-wavelength eddies into billows could represent a process of quasi-horizontal mixing of stratospheric
and tropospheric air in the vicinity of a tropopause fold. As pointed out recently by Appenzeller and Davies (1992) and Appenzeller et al. (1996), it is possible that such eddies may contribute significantly on a global scale to the important exchange of chemicals and aerosols across the tropopause.

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**REFERENCES**


