The Possible Influence of Upstream Upper-Level Baroclinic Processes on the Development of the *QE II* Storm

LOUIS W. UCCELLINI

Laboratory for Atmospheres, NASA/Goddard Space Flight Center, Greenbelt, MD 20771

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ABSTRACT

An analysis of the *QE II* storm of 9–11 September 1978 presents evidence for the existence of upper-level baroclinic processes upstream of the rapidly developing cyclone. The analysis shows that a deepening short-wave trough was located 400 to 500 km upstream of the site of the storm 12 h prior to rapid cyclogenesis. The trough was associated with 1) a polar jet marked by 65 m s^{-1} winds in its core and significant vertical and horizontal wind shear, 2) positive vorticity advection and divergence at the 300 mb level, and 3) an intense frontal zone that extended from 300 mb down to the surface. It also appears that a tropopause fold likely extruded stratospheric air down to the 700–800 mb level, 400–500 km upstream of the surface low and 12 h prior to the explosive development phase of the cyclone. These findings raise questions about Gyakum's assertion that the *QE II* storm developed in an area in which the baroclinic support was confined to the lower troposphere and the related assertion by Anthes et al. that upper-level forcing upstream of the area of rapid cyclogenesis was weak and apparently not important in this case.

1. Introduction

Many recent studies on cyclogenesis have shown the importance of diabatic processes in general, and latent heat release in particular, in the evolution of the storm system (e.g., Danard, 1964; Tracton, 1973; Johnson and Downey, 1976; Gall, 1976; Sanders and Gyakum, 1980; Chang et al., 1982; Atlas, 1984; and Smith et al., 1984). These studies have stressed the role of latent heat release in amplifying the vertical motion fields and enhancing the development rates of the cyclone, while also noting that baroclinic processes associated with trough/jet systems that extend through a deep portion of the troposphere are clearly linked to the development of surface cyclones. Theoretical studies, including Staley and Gall (1977), also show that the baroclinic growth rates are very sensitive to decreased static stability in the lower troposphere, indicating that sensible heating in the planetary boundary layer could be an important element in rapid cyclogenesis. These results generally conform with Sanders and Gyakum's (1980) climatological survey of oceanic "bombs," in which they point to the critical importance of diabatic processes for rapid cyclogenesis and also to the existence of baroclinic waves 400–500 km upstream of the incipient surface low. It appears from these and other studies that baroclinic processes throughout the troposphere exert a strong influence on the initial evolution of cyclones, with diabatic processes acting to accelerate and/or amplify the low-level development, especially with respect to surface deepening rates.

The *QE II* storm, which occurred on 9–11 September 1978 (analyzed by Gyakum, 1983a,b), clearly meets the criteria for oceanic "bombs" that are influenced by diabatic processes. Furthermore, the numerical experiments presented by Anthes et al. (1983) for the same case convincingly demonstrate the sensitivity of mesoscale model simulations of the *QE II* storm to diabatic processes associated with convective latent heating, grid-scale latent heating, and sensible heating in the boundary layer. The inclusion of each of these processes in the numerical model appears to contribute positively to the rapid deepening phase of the simulated cyclone. Nevertheless, the authors conclude "that no one process could be identified as obviously more important than others and no combination produced an entirely satisfactory forecast" (p. 1188).

An important aspect emphasized by Gyakum and Anthes et al. is that, unlike other rapidly deepening cyclones, the explosive development of the *QE II* storm apparently occurred without any noticeable upper-tropospheric forcing upstream of the surface low. Gyakum repeatedly states that the *QE II* storm developed within a region in which relatively weak baroclinic support was confined to the lowest levels and that this cyclone developed analogously to a hurricane, driven by heating on the cumulus scale. Based on the analysis used to initialize the numerical model, Anthes et al. summarize this viewpoint by noting that an upper-level trough is located *downstream* of the incipient surface low at 1200 GMT 9 September, which is not a climatologically favorable location for cyclogenesis. They then state that, in the analysis used to initialize the model simulations at 1200 GMT 9 September, the vorticity advection over the surface low was nearly zero at the 300 mb level "which supports Gyakum's (1983a,b) conclusions that the early cyclogenesis was a shallow phenomenon not associated with strong dynamical effects in the upper troposphere" (p. 1178).

The purpose of this paper is to present evidence that 1) a deep frontal zone associated with a short-wave trough and polar jet was indeed present upstream of the incipient cyclone and 2) baroclinic processes associated with this trough/jet system might have contributed to the rapid development of the *QE II* storm *in addition to* the diabatic processes emphasized by Gyakum and Anthes et al. A synoptic analysis is presented in section 2, which identifies the upper-level short-wave trough/polar jet system and associated intense frontal zone upstream of the surface low at 1200 GMT 9 September 1978. Vorticity advection and divergence analyses from a global analysis scheme are also presented along with evidence for the existence of a tropopause fold and stratospheric extrusion upstream of the developing storm system. Possible implications of these findings with respect to the model simulation of the storm are discussed in section 3. The results are summarized in section 4.

2. Characteristics of upstream conditions for 9 September 1978

The *QE II* storm began to develop after 0000 GMT 9 September 1978 near 40°N, between 70° and 75°W (Gyakum, 1983a). By 1200 GMT 9 September 1978, the low was situated near 40°N, 70°W (Fig. 1a) and was already beginning to show signs of significant development as revealed by the Seasat wind analysis (Gyakum, 1983a). Between 1200 GMT 9 September and 1200 GMT 10 September the low deepened explosively as it moved to a position well off the Newfoundland coast (Fig. 1c). As shown by Gyakum (1983a), the surface system not only was under analyzed by the National Meteorological Center (NMC), but was also badly forecast by the operational Limited Area Fine Mesh (LFM) numerical model.

To illustrate that there is upper-level support for the development of the *QE II* storm, the NMC 500 mb analysis is shown for 0000 GMT 9 September (Fig. 2)

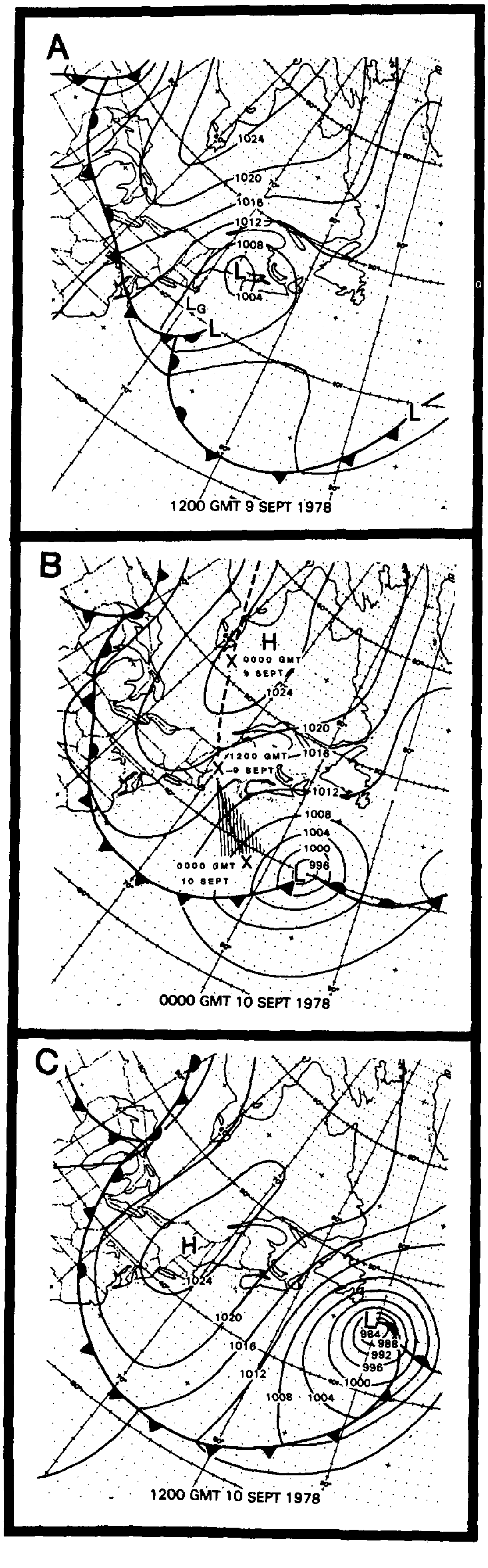

FIG. 1. National Weather Service operational surface analyses for (a) 1200 GMT 9 September; (b) 0000 GMT 10 September; and (c) 1200 GMT 10 September 1978. L_G on (a) is position of surface low analyzed by Gyakum (1983a) using Seasat wind analysis. Indicated central pressures for *QE II* cyclone are too high for 10 September analyses as Gyakum emphasizes. The ×'s on (b) show position of observed 500 mb height fall maximum for short wave for 12 h periods ending at 0000 and 1200 GMT 9 September, while shaded region indicates extrapolated position for height fall maximum during 12 h period between 1200 GMT 9 September and 0000 GMT 10 September.

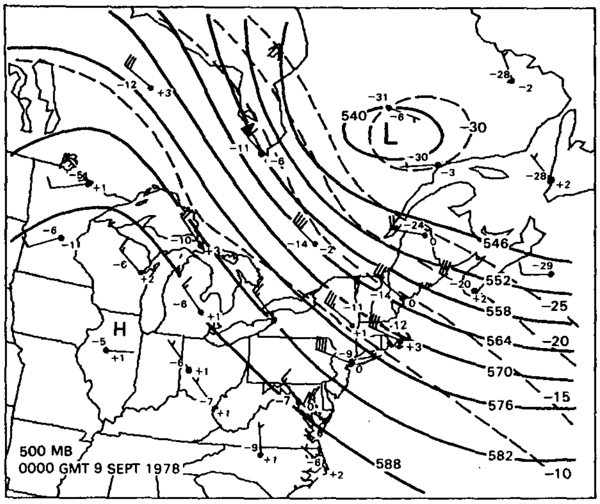

FIG. 2. National Weather Service 500 mb analysis for 0000 GMT 9 September 1978. Temperature (°C) is dashed; geopotential height (dam) is solid. Plotted data at stations include temperature (°C) at upper left and 12 h height tendency (dam) at lower right; wind barbs indicated by standard notation (m s^{-1}).

and the 300, 500, 700 and 850 mb analyses from NMC are shown for 1200 GMT 9 September (Fig. 3). At 0000 GMT 9 September, a 500 mb short-wave trough is evident in southeast Canada with 60 m height falls and wind speeds of at least 60 m s^{-1} near the trough axis just southeast of the Hudson Bay (Fig. 2). Gyakum's 250 mb analysis (see Fig. 6 in Gyakum, 1983a) indicates that this jet streak extended toward the New England Coast and that the initial surface development occurred on the anticyclonic side of the exit region of the jet.

By 1200 GMT 9 September, the trough had moved into northern New England and was located 400–500 km upstream of the region in which the explosive development phase of the *QE II* storm commenced. There are indications that the 300, 500 and 700 mb maximum height falls associated with this short-wave trough may have been amplifying as the system propagated rapidly southeastward, although the data density in southeastern Canada and over the Atlantic Ocean is not adequate to fully resolve the structure and temporal evolution of the system.[1] At 300 mb, the trough was centered over the mouth of the St. Lawrence River with the short-wave feature located over northern New England marked by 160 m height falls and wind speeds near 60 m s^{-1} (Fig. 3a). At 500 mb, the 12 h height falls were at least 110 m in southeastern Maine (Fig. 3b), while at 700 mb, the height falls were at least 80 m in the same region (Fig. 3c). These height falls were occurring near or beneath the axis of the jet streak, where wind speeds likely exceeded 65 m s^{-1} in the core (Fig. 5b), a sign that a baroclinically unstable regime existed throughout the entire troposphere upstream of the *QE II* storm 12 h prior to the explosive development stage. The 850 mb cold air advection over New York and 850 mb warm air advection above the surface low position just off the coast (Fig. 3d) are additional indications of the baroclinic nature of the short wave propagating rapidly toward the region where cyclogenesis occurred over the next 24 h. Thus, the NMC analyses reveal that a short-wave trough/jet system was propagating rapidly over southeast Canada on 9 September to a position within 400 to 500 km upstream of the surface low at 1200 GMT 9 September and moving directly toward the region of subsequent explosive surface development (see extrapolated track in Fig. 1b).

As an indication of the upper-level forcing associated with the short-wave trough, the 300 mb vorticity advection and divergence fields were computed for 0000 and 1200 GMT 9 September (Fig. 4) using a model-based global analysis system described by Baker et al. (1984). The analysis cycle was initialized at 0000 GMT 7 July 1978 and continued through a 3-month period, ending on 10 October 1978. Thus, the vorticity advection and divergence fields in Fig. 4 are influenced by the continuous assimilation of radiosonde, surface, aircraft, and satellite data (including Seasat wind fields) into the global model during the analysis cycle with the heaviest weight going to the most recent observations at any given time. In the analysis, a region of positive vorticity advection (PVA) could be diagnosed just north of Vermont at 0000 GMT 9 September (Fig. 4a) with a corresponding area of 300 mb divergence in southeastern Canada (Fig. 4c). Both of these fields are just downstream of the trough axis located over the Hudson Bay (Fig. 2). By 1200 GMT 9 September, the area of PVA (Fig. 4b) and 300 mb divergence (Fig. 4d) moved to a position off the coast, downstream of the axis of the short-wave trough over New England (Fig. 3a–c) and over the position of the surface low (Fig. 1a). The close proximity of this short-wave trough/jet system to the *QE II* storm and the superposition of the 300 mb PVA and divergence over the deepening surface low at 1200 GMT 9 September appears to place this storm in a category which is similar to other mid-latitude oceanic cyclones as discussed by Sanders and Gyakum (1980).

Although the vorticity advection and divergence fields provide evidence for upper-level forcing, the 4° × 5° grid of the global analysis system likely underestimated the magnitude of these fields and the intensity of this short-wave/jet system. An indication of the in-

[1] The data void region in Canada and over the ocean poses severe limitations in resolving the temporal evolution of the short-wave trough and the structure of the jet for this case, especially the along-stream variations in the wind field. This variation is an important factor in defining the strength of transverse circulations and associated vertical motions near the jet streak (Bjerknes, 1951; Uccellini and Johnson, 1979; Brill et al., 1985).

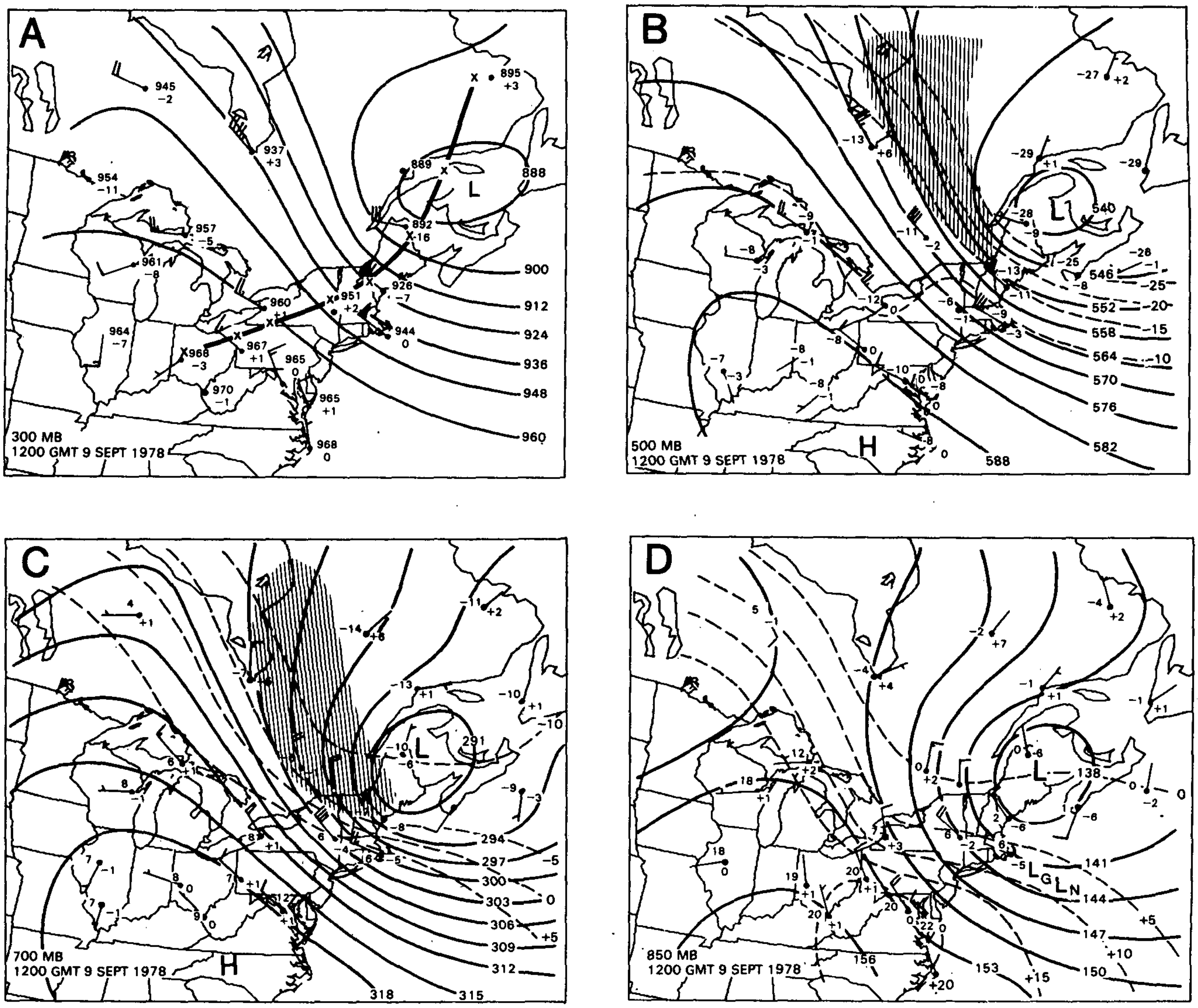

FIG. 3. National Weather Service analyses for 1200 GMT 9 September 1978 for (a) 300 mb; (b) 500 mb; (c) 700 mb; and (d) 850 mb. In (a) geopotential heights (937 = 9370 m) and height tendency plotted at stations. See Fig. 2 for convention for plotting station data in (b), (c) and (d). Thick solid line on 300 mb map indicates position of cross section shown in Fig. 4. Shading in 500 and 700 mb analyses indicates approximate region in which cold air advection coincides with area of cyclonic geostrophic wind shear, a contributing factor to a frontogenetical ageostrophic circulation (see text). L_N on 850 mb map denotes position of surface low from NMC analysis, and L_G denotes position as analyzed by Gyakum (1983a), that eventually developed into *QE II* storm.

tensity of the upper-level system upstream of the *QE II* storm is the existence of a strong midtropospheric frontal zone across northern New England at 500 mb (Fig. 3b) and across central New England at 700 mb (Fig. 3c). To illustrate the magnitude of the thermal gradients in the upper, middle and lower troposphere and the vertical extent of this front and associated jet, a vertical cross-section of potential temperature (θ) and total wind speed was constructed from Goose Bay, Newfoundland (YYR) to Dayton, Ohio (DAY) (Fig. 5). The cross-section reveals an intense frontal zone with a 20°C temperature gradient and 40 m s^{-1} wind difference concentrated in a 150 to 200 km wide band at the 500 mb level between Caribou, Maine (CAR) and Portland, Maine (PWM). The base of the front extended downward from 350 to 300 mb just north of CAR, to 600 mb near PWM, 750 mb near Albany, New York (ALB), and to the surface near Pittsburgh, Pennsylvania (PIT; Fig. 5a). Furthermore, 1) the 65 m s^{-1} wind speeds at PWM, 2) the large vertical and horizontal wind shears within the frontal zone, and 3) the missing wind data above 200 mb at CAR, above 270 mb at PWM, and above 700 mb at ALB (where the wind speeds had already reached 40 m s^{-1}) are all signs

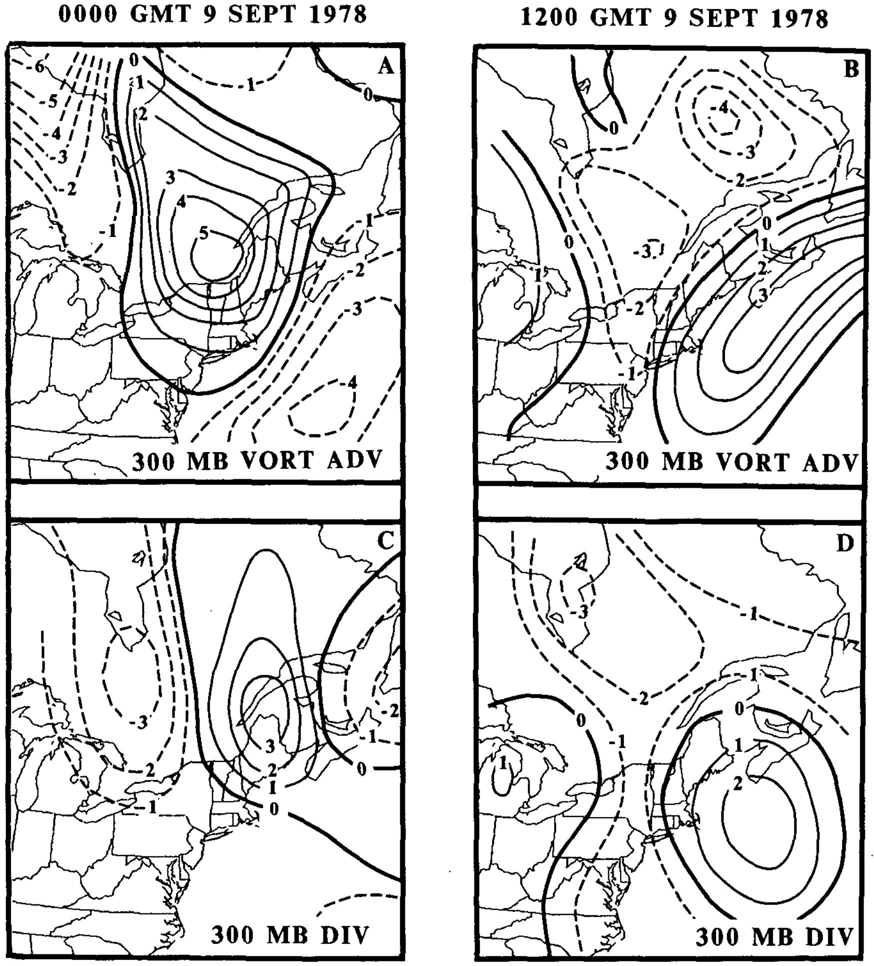

FIG. 4. (a), and (b): 300 mb vorticity advection ($3 = 3 \times 10^{-9}$ s^{-2}); (c), and (d) 300 mb divergence ($2 = 2 \times 10^{-5}$ s^{-1}) at 0000 GMT and 1200 GMT 9 September 1978, respectively. Analyses were derived on a 4° latitude by 5° longitude grid using a model-based global analysis system described by Baker et al. (1984) which assimilated all available satellite, surface, aircraft and radiosonde data through a three month period beginning at 1200 GMT 7 July 1978.

that a remarkably intense jet streak was associated with the short-wave trough imbedded within the generally northwesterly flow upstream of the *QE II* storm. The NMC analyses, vorticity advection and divergence diagnostics from the global analysis, and the cross sections derived directly from the radiosonde data all indicate that relatively strong baroclinic forcing did exist within the relatively straight flow field despite the lack of significant curvature in the geopotential fields. It again appears that strong forcing for upper-level divergence and associated vertical motion fields can exist for relatively straight flow regimes, as recently noted by Uccellini (1984) and shown by Brill et al. (1985).

A tropopause fold is defined by Reed (1955) and Reed and Danielsen (1959) as an extrusion of stratospheric air within an upper-tropospheric baroclinic zone that slopes downward from a normal tropopause level to the middle or even lower troposphere. The concept of a tropopause fold complemented the studies of Reed and Sanders (1953) and Newton (1954), which pointed to a growing recognition of the importance of subsidence in the upper and middle troposphere as a mechanism contributing to upper-level frontogenesis. Bosart and Lin (1984) and Uccellini et al. (1985) show that an intense upper-level frontal zone, polar jet and tropopause fold appeared to influence the rapid development phase of the Presidents' Day cyclone. The folding process extruded dry stratospheric air (marked by high values of potential vorticity) down toward 700 mb, 1500 km upstream and 12–24 h prior to the explosive development of the cyclone. The stratospheric air mass then descended toward 800 mb as it moved toward the East Coast and was nearly collocated with the surface low as the explosive surface development

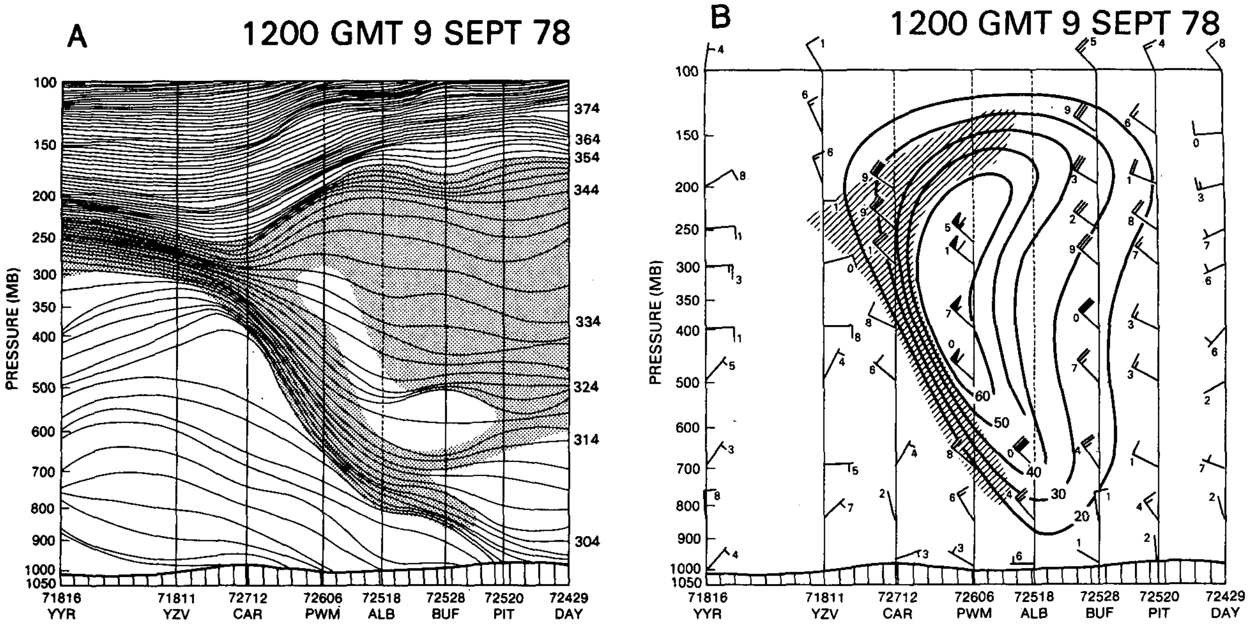

FIG. 5. Vertical cross section from Goose Bay, Newfoundland (YYR) to Dayton, Ohio (DAY), for 1200 GMT 9 September 1978 (see thick solid line in Fig. 3a for position of cross section). Dashed portions of vertical lines at CAR, PWM and ALB indicate levels at which wind data were not available. (a) Isentropes (K, solid); shading represents area where relative humidity is less than 20%. (b) Subjective isotach analysis for total wind speed (solid, m s^{-1}); wind barbs plotted with last digit of wind speed observation. Wind analysis over ALB derived from cross checking with NWS analyses on 200, 300, 500 and 700 mb surfaces. Shading indicates region where potential vorticity {computed from selected points using data restricted to plane of the cross section and the expression $[-(\partial u/\partial y_\theta + f)(\partial\theta/\partial p)]$} exceeded 10 $\times$ 10^{-6} K s^{-1} mb^{-1}, values typical of stratospheric air.

commenced. The analysis by Uccellini et al. shows that the vertical stretching of the stratospheric air mass appears to play a role in the cyclogenesis, although the relative importance of the various dynamic and diabatic processes could not be assessed given the limitations of the operational observing network.

The analyses for the *QE II* storm thus shown (Figs. 3 and 5) all provide evidence that, like the Presidents' Day storm, tropopause folding occurred prior to and upstream of the site of explosive cyclogenesis. At 1200 GMT 9 September, there are indications[2] that "forcing" for strong subsidence (and related frontogenesis) associated with confluence of the geostrophic wind and cold advection in the region marked by cyclonic geostrophic shear (Shapiro, 1981; Keyser and Pecnick, 1985a,b) is evident with this system. At 300, 500 and 700 mb geostrophic confluence is quite noticeable north of the United States–Canada border and upstream of the frontal zone in New England (Figs. 3a–c). In the same area, 300 mb convergence was diagnosed from the global analysis extending from the same area toward the middle Atlantic states (Fig. 4d). Furthermore, strong cold air advection coincides with a region marked by significant cyclonic shear in the geostrophic wind at 500 and 700 mb (see shaded area in Figs. 3b and 3c). This feature is an important factor that shifts the direct circulation associated with the geostrophic confluence such that the descent is concentrated on the warm side of the front and thus acts to strengthen the frontal zone through the frontogenetical tilting effect (Shapiro, 1981; Keyser and Pecnick, 1985a,b). An indication of subsidence on the warm side of the front is the change in the 500 mb temperature at Albany, New York (ALB), and Chatham, Massachusetts (CHH), between 0000 and 1200 GMT 9 September. At ALB (where there was little or no horizontal temperature advection on the 500 mb surface at 0000 GMT and weak cold advection at 1200 GMT) the 500 mb temperature increased from −11°C to −6°C. At CHH, the 500 mb temperature increased from −12°C to −9°C. While the temperature ahead of the front was increasing, the 500 mb temperature behind the front decreased at CAR from −24°C to −28°C, again where the temperature advection was apparently weak especially at 1200 GMT 9 September (Fig. 3b). Therefore, the upper-level frontogenesis was probably related to the distribution of vertical motion across the front as simulated by Keyser and Pecnick (1985a,b) for idealized flow situations that include both the confluence and horizontal shear and diagnosed by

[2] Given the lack of data over Canada and the Atlantic Ocean, a detailed diagnostic analysis using the Petersen (1979) isentropic analysis scheme could not be completed for this case, as was done for the Presidents' Day storm.

Uccellini et al. (1985) for the Presidents' Day cyclone. The likelihood of subsidence along the frontal zone combined with the intensity and depth of the front (as displayed in the cross section in Fig. 5) indicates that tropopause folding likely extruded stratospheric air down toward the 700 mb level by 1200 GMT 9 September immediately upstream of the region where the *QE II* storm underwent explosive deepening. The dry air extending down from the 400 mb level at CAR to below 800 mb at BUF within the frontal zone (Fig. 5a) is one sign of a stratospheric extrusion. Between 0000 and 1200 GMT 9 September, the 700 mb dewpoint depression at both ALB and CHH increased to greater than 30°C[3]. Another indication of tropopause folding is the large static stability ($\partial\theta/\partial p$) in the frontal zone (Fig. 5a) which, combined with the large cyclonic shears north of ALB at 700 mb (Fig. 3c) and near PWM and CAR between 700 and 350 mb, would yield large values of potential vorticity (Fig. 5b). The static stability and cyclonic wind shears analyzed in the plane of the cross section yield potential vorticity values exceeding 10×10^{-6} K s^{-1} mb^{-1} which extend down toward the 800 mb level within the front (illustrated by the shading in Fig. 5b)[4] with values greater than 20×10^{-6} at 700 mb and 500 mb, 30×10^{-6} at 400 mb and 40×10^{-6} at 300 mb. Although Gyakum (1983b; Fig. 6) shows values of potential vorticity greater than 10×10^{-6} immediately over the surface low center, his figure shows no evidence of a continuous extrusion of stratospheric values extending along the intense frontal boundary toward the lower troposphere.

Again, the data void areas in Canada and over the ocean on either side of the cross section in Fig. 5 preclude a detailed analysis of the three-dimensional flow associated with the front upstream of the developing *QE II* storm. Nevertheless, it appears that intense tropopause folding associated with deep tropospheric frontogenetical forcing occurred upstream of the *QE II* storm prior to its rapid development. Given the wind speeds within the front, the stratospheric air could be near the location of the surface low at 0000 GMT 10 September and could therefore have influenced the explosive development phase of the *QE II* storm (along with the diabatic processes) as discussed for the Presidents' Day cyclone by Uccellini et al. (1985). The existence of dry stratospheric air descending toward the developing storm may also explain the cloud free dry slot that extended from the west-southwest to the northeast over the storm center as revealed by the satellite images at 0359 GMT and 0859 GMT 10 September (shown in Gyakum's, 1983a, Figs. 15b and 15c).

3. Discussion

The model sensitivity study conducted by Anthes et al. (1983) for the *QE II* storm shows the importance of diabatic processes in weakening the static stability in the oceanic planetary boundary layer and also in contributing to more rapid cyclogenesis. However, questions remain concerning the control experiment used by Anthes et al. in that the upper-level forcing may have been underestimated at the initial time. Specifically, the initial analyses used in the control experiment at 1200 GMT 9 September contained little if any PVA at 300 mb (Anthes et al., 1983) compared with a distinct pattern of PVA and divergence diagnosed using the global analysis (Figs. 4c and 4d) described by Baker et al. (1984).[5] Furthermore, the east-to-west cross section shown by Anthes et al. (1983, Fig. 5) derived from the initial model analysis at 1200 GMT 9 September as well as the cross section presented by Gyakum (1983b, Fig. 6) for the potential vorticity distribution near the *QE II* storm both indicate overly smoothed tropospheric fronts. Since potential vorticity computations are very sensitive to static stability and wind shear measurements (Staley, 1960; Bleck and Mattocks, 1984; Uccellini et al., 1985), the vertically smoothed θ analysis shown in the cross sections by Gyakum and the coarse vertical resolution of the numerical model used by Anthes et al. likely underestimate the magnitude and vertical extent of the stratospheric reservoir of high potential vorticity that existed upstream of the *QE II* storm. As such, the potential contribution of upstream dynamic processes associated with the jets and fronts was probably underestimated in both studies. The model forecast, however, did generate a 60 m s^{-1} jet core and attending upper-level front immediately upstream of the developing surface cyclone 12 to 24 h into the numerical simulation, although the simulated frontal zone is weaker than the observed front [see Figs. 9 and 10 in Anthes et al. (1983)]. Questions then arise as to 1) whether the model was attempting to produce the structure of the deep tropospheric front that was in fact present at the initial time, 2) whether this attempt by the model to generate an upper-level frontal zone and associated baroclinic

[3] The dewpoint values were derived directly from the radiosonde reports listed on the Northern Hemispheric data tabulations from the National Climate Data Center.

[4] Reed and Danielsen (1959) define 10×10^{-6} K s^{-1} contour as a value for potential vorticity that separates stratospheric and tropospheric air masses near upper-level fronts. The potential vorticity computations showing large values in the shaded region in Fig. 5b likely underestimate the actual magnitude near the front, given the poor resolution of the data set and the exclusion of curvature vorticity by using only the data in the plane of the cross-section. Since the cross section is nearly normal to the front and jet streak (Fig. 3a), the $\partial u/\partial y_\theta$ in the plane of the cross section represents the shear vorticity.

[5] This discrepancy between the analyses may be related to the use of the NMC analysis by Anthes et al. as a first guess field for their initial analysis at 1200 GMT 9 September. At the time of the *QE II* storm, NMC used the Hough analysis which was designed to filter features with a wavelength of 15° longitude or less.

forcing for surface cyclogenesis was too little and too late for a more accurate simulation of the cyclone, and 3) how the diabatic processes simulated by the model during the cyclogenetic period would have interacted with a better defined initial baroclinic structure.

Recent preliminary numerical experiments for the Presidents' Day storm (Petersen et al., 1985) show that the evolution of the upper-level front and surface cyclogenesis was sensitive to a more detailed initial analysis of the fronts and associated jet streaks and the increased vertical resolution for the numerical model. The model initialization using the Petersen (1985) isentropic scheme contained the vertical and horizontal structures of the wind and temperature fields that better defined the trough/jet system and developing tropopause fold upstream of the area where the Presidents' Day cyclone ultimately developed. The numerical experiments show that the upper-level frontogenesis, associated sinking motion, tropopause folding, and the *initial* surface cyclogenesis were forecast better when the upper-level polar jet/trough system was analyzed more accurately at the initial time. While it is difficult to generalize model impacts from one case to another, these results for the Presidents' Day storm indicate that questions remain for the model simulation of the *QE II* storm concerning the role and/or relative importance of the very intense upper-level front and associated tropopause fold in the rapid development of the cyclone. Unfortunately, these questions will be hard to resolve for the *QE II* storm given the large data void areas in Canada and over the ocean that make a detailed 3-dimensional analysis of the polar jet and associated front difficult.

4. Summary

Evidence is presented in this paper that suggests that the rapid development of the *QE II* storm was marked not only by diabatic processes, but also by the presence of a deepening short-wave trough/jet streak system. The short-wave trough was associated with 1) a polar jet with 65 m s^{-1} winds in the core and significant vertical and horizontal wind shears, 2) positive vorticity advection and divergence at the 300 mb level, 3) a very intense frontal zone that extended from a normal tropopause level of 300 to 400 mb down to below 800 mb, and 4) a tropopause fold that extruded stratospheric air down to the 700–800 mb level by 1200 GMT 9 September upstream and prior to the explosive development phase of the cyclone. Although this evidence does not detract from the importance of diabatic processes emphasized for this case by Gyakum (1983a,b) and Anthes et al. (1983), these observations 1) point to the existence of upper-tropospheric forcing that is conducive to surface cyclogenesis and 2) are in contrast to the numerous assertions by Gyakum and Anthes et al. that upper-level baroclinic forcing was either nonexistent or weak at 1200 GMT 9 September just prior to the rapid development phase of the *QE II* storm, and that the cyclone developed explosively "subsequent to 1200 GMT 9 September in association with an initially *shallow* lower tropospheric short-wave trough" (Gyakum 1983a, p. 1153).

It also appears that the intensity of the upper-level trough, jet streak and associated front may not have been captured well by the initial analysis used for the numerical experiments conducted by Anthes et al. (1983) although the model did attempt to generate an upper-level frontal zone upstream of the storm center by 1200 GMT 10 September, 24 h into the simulation. Given that preliminary numerical experiments for the Presidents' Day cyclone indicate that the simulations of the initial period of cyclogenesis is positively impacted by a better initial analysis of the upper-level fronts and associated vertical circulations, it may be that the control run for the *QE II* model experiments might have been improved if these features were accounted for more completely.

It seems that the explosively developing cyclones like the Presidents' Day and *QE II* storms are related to the interaction of dynamical and diabatic processes over the entire extent of the troposphere, especially those that influence the intensity of short-waves and attending frontal zones, the strength of upper- and lower-level jets and associated transverse ageostrophic circulations, and the magnitude of the static stability in the oceanic boundary layer. Unfortunately, the resolution of the current oceanic data base is totally inadequate to resolve the basic structure of some of these features.

The poor resolution of oceanic data bases has permeated many studies of oceanic cyclones, such as the Petterssen et al. (1962) study, in which they claim to have identified the so-called type A cyclone in straight flow aloft with no appreciable upper-level forcing. It is possible that the oceanic data base was too sparse to resolve the structure (and perhaps even the existence) of upper-level jet/trough systems for the cyclones analyzed by Petterssen et al., much less the relative importance of the various physical processes. An example of the data analysis problems associated with the study of oceanic cyclones is illustrated in the *QE II* case, where two different analysis schemes apparently yielded noticeably different upper-level vorticity advection fields near a coastline where there were at least 7 radiosonde reports. These discrepancies would likely be compounded over the ocean where there are no radiosonde reports to provide the vertical detail needed to resolve the deep baroclinic processes that could significantly influence rapidly developing storms, even if the flow is relatively straight. Future field experiments are required to provide the datasets with the vertical, horizontal and temporal resolutions needed to properly diagnose the synoptic and mesoscale interactions of diabatic and dynamical processes that produce explosive cyclogenesis.

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