Baroclinic Transition of a Long-Lived Mesoscale Convective Vortex

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ABSTRACT

The period 5–15 June 2003, during the field phase of the Bow Echo and Mesoscale Convective Vortex (MCV) Experiment (BAMEX), was noteworthy for the wide variety of mesoscale convective systems (MCSs) that occurred. Of particular interest was a long-lived MCV that formed in the trailing stratiform region of an MCS over west Texas at 0600 UTC 10 June. This MCV was noteworthy for its (i) longevity as it can be tracked from 0600 UTC 10 June to 1200 UTC 14 June, (ii) development of a surface cyclonic circulation and attendant $-2$ to $-4$-hPa sea level pressure perturbation, (iii) ability to retrigger convection and produce widespread rains over several diurnal heating cycles, and (iv) transition into a baroclinic surface cyclone with distinct frontal features. Baroclinic transition, defined here as the acquisition of surface fronts, occurred as the MCV interacted with a remnant cold front, left behind by a predecessor extratropical cyclone, over the Great Lakes region. Although the MCV developed well-defined frontal structure, which helped to focus heavy precipitation, weakening occurred throughout the baroclinic transition process. Energetics calculations indicated that weakening occurred as the diabatic and baroclinic energy conversion terms approached zero just prior and during baroclinic transition. This weakening can be attributed to (i) an increase in environmental wind shear, (ii) the development of a downshear tilt and associated anticyclonic vorticity advection over the surface low center, and (iii) the eastward relative movement of organized convection away from the MCV center.

1. Introduction

a. Motivation

During the field phase of the Bow Echo and Mesoscale Convective Vortex Experiment (BAMEX), five mesoscale convective vortices (MCVs) were sampled during intensive operations periods (IOPs) 1, 4, 5, 8, and 15 (Davis et al. 2004). A detailed analysis of the structure of these MCVs, using observation data collected during their respective IOPs, is presented in Davis and Trier (2007) and Trier and Davis (2007). The MCV sampled during IOP 8 (1600–1900 UTC 11 June 2003) is particularly noteworthy for its (i) longevity as it can be tracked from 0600 UTC 10 to 1200 UTC 14 June, (ii) development of a surface cyclonic circulation and attendant $-2$ to $-4$-hPa sea level pressure perturbation, (iii) ability to retrigger convection and produce widespread rains over several diurnal heating cycles, and (iv) transition into a baroclinic surface cyclone with distinct frontal features. The focus of this paper will be the investigation of the structural evolution of this long-lived MCV from the development of a surface cyclonic circulation over Oklahoma on 11 June 2003 through the transition to a baroclinic surface cyclone on 13 June 2003. Baroclinic transition is defined in this study as the...
MCV’s acquisition of surface cold and warm fronts. Investigation of the physical processes involved in surface frontal development and attendant intense precipitation, and weakening subsequent to baroclinic transition, will receive particular attention.

b. Background

1) Brief Discussion of MCVs

MCVs are warm-core systems (in the middle and lower troposphere) that have been documented to form in the trailing stratiform region of MCSs (e.g., Johnston 1982; Zhang and Fritsch 1987; Menard and Fritsch 1989; Bartels et al. 1997; Johnson and Mapes 2001). Cyclonic and anticyclonic relative vorticity maxima are characteristically found in middle and upper levels, respectively, in association with a diabatic heating maximum. An anticyclonic, or weakly cyclonic, relative vorticity maximum can also be found near the surface in association with a lower-tropospheric cold pool. There are occasions, however, in which a cyclonic circulation of appreciable intensity reaches the surface (e.g., Fritsch et al. 1994; Rogers and Fritsch 2001). In MCVs over tropical oceanic regions, the development of a surface cyclonic circulation is instrumental for initiating air–sea interactions (Emanuel 1986; Rotunno and Emanuel 1987), which can lead to tropical cyclogenesis in some cases. The physical processes important in MCV penetration to the surface are beyond the scope of this study, but previous studies have suggested that the generation of low-level cyclonic vorticity beneath the midlevel MCV results in a deep vortex with a near-surface vorticity maximum (e.g., Davis and Trier 2002; Davis and Galarneau 2009). This low-level cyclonic vorticity appears to be generated by (i) the transport of vorticity rearward [including line-end vortices (e.g., Weisman and Davis 1998)] from the leading convective line to beneath the midlevel center, and (ii) the development of a divergence profile (low-level convergence) that favors vorticity generation near the surface during the early morning hours when the environment approaches moist neutral. Other studies have also demonstrated that a wide variety of divergence profiles can be present in low levels on a case-to-case basis, and that tilting of horizontal vorticity (associated with the rear-inflow jet in some cases) and the vertical advection of vorticity can be important in the generation of low-level cyclonic vorticity (e.g., Zhang 1992; Chen and Frank 1993; Cram et al. 2002; Kirk 2003; Knievel and Johnson 2003; Kirk 2007).

Although most MCVs dissipate as, or shortly after, the parent MCS decays, there are occasions in which the MCV can last long after the parent MCS has dissipated and can even reinitiate convection (e.g., Cotton et al. 1983; Bartels and Maddox 1991; Trier and Davis 2007). Long-lived MCVs, which have been documented occur ~20–30 times per year east of the Rockies during the summer months (Davis et al. 2002), can occasionally produce excessive rains over large geographical regions by reorganizing convection over several diurnal heating cycles (e.g., Bosart and Sanders 1981; Menard and Fritsch 1989; Fritsch et al. 1994; Schumacher and Johnson 2008). Approximately 5% of all long-lived MCVs are associated with a well-defined surface cyclonic circulation (e.g., Davis et al. 2002). The longevity and maintenance of an MCV with an attendant surface cyclonic circulation appears to be dictated by the character of the low-level vertical wind shear (Davis and Trier 2007; Davis and Galarneau 2009). While strong low-level shear appears detrimental to the maintenance of low-level vortices (Davis and Weisman 1994), it has been linked to the generation of vertical vorticity through tilting (e.g., Knievel and Johnson 2003). Strong low-level wind shear can also act to trigger convection through balanced motions in a very moist environment with weak instability by readily enabling air parcels to reach the level of free convection [see Fig. 1 in Trier and Davis (2007) for conceptual model; see also, e.g., Raymond and Jiang (1990); Jiang and Raymond (1995); Conzemius et al. (2007)]. This convection can help the MCV resist the debilitating effects of strong low-level shear by generating low-level cyclonic vorticity and reducing environmental deep-layer wind shear through momentum rearrangement.

2) Vortices Approaching Baroclinic Zones

The interaction between a vortex and a baroclinic zone can lead to the development of distinct frontal zones within the circulation of the vortex (e.g., Fig. 8 in Keyser et al. 1988). Examples of this interaction include the extratropical transition [ET; see Jones et al. (2003) and references therein for review of ET] of poleward-moving tropical cyclones (TCs), and less frequently, the interaction of MCVs with baroclinic zones (Zhang and Harvey 1995). The importance of these developing frontal structures lies in their ability to act as a focusing mechanism for intense precipitation. For example, nearly 250 mm of rain fell at Wellsville, New York, during the ET of TC Agnes [1972; Bosart and Carr (1978), see their Fig. 2], and 200–500 mm of rain fell from North Carolina to southeast New York during the ET of TC Floyd (1999; Atallah and Bosart 2003; Colle 2003).

Along with the acquisition of the surface frontal structure, whether a vortex intensifies after acquiring frontal structure has implications for quantitative
precipitation forecasting. In the case of ET of recurring TCs for example, previous studies have shown that post-ET intensification occurs if the interacting trough is negatively tilted and the remnant TC is intense, while post-ET weakening occurs if the interacting trough is positively tilted and the remnant TC is weak (Agustí-Panareda et al. 2004, 2005; Hart et al. 2006). This link between post-ET intensification and trough tilt agrees with Atallah et al. (2007), who characterized strong ET and widespread heavy precipitation left of track with a strong upstream trough (likely negatively tilted) and strong warm-air advection, and weak ET and widespread precipitation right of track with a strong downstream ridge (likely weak positively tilted up-stream trough in this case) and weak warm-air advection. Zhang and Harvey (1995) examined a rarely documented case of extratropical cyclogenesis evolving from a MCV. They found that the intensification of the extratropical cyclone on 10–11 June 1985, examined in observations and their numerical simulations, occurred in the presence of an upstream neutrally to weakly negatively tilted 500-hPa trough. While the frequency of the extratropical transition of continental MCVs is unknown, the pathway by which the MCVs transition is likely quite similar to what is more commonly observed with diabatic Rossby vortices over oceans (e.g., Wernli et al. 2002; Moore and Montgomery 2005), and the ET of TCs (e.g., Jones et al. 2003). With these comparisons in mind, the Zhang and Harvey (1995) results suggest that the ET paradigm may have some applicability to warm-core vortices of midlatitude continental origin that interact with baroclinic zones and mid- to upper-tropospheric troughs.

c. Outline of paper

The remainder of this paper is organized as follows. Section 2 describes the dataset used, and the methods behind the calculations and diagnostics shown in the paper. Section 3 establishes the presence of a tropospheric-deep MCV with an attendant surface cyclonic circulation, and section 4 provides an overview of the MCV’s interaction with its environment. Section 5 presents the energetics calculations, while section 6 describes the evolution of the surface boundaries and
attendant heavy precipitation during baroclinic transition. Section 7 discusses the key results and provides concluding remarks.

2. Data and methods

The detailed observational analysis performed in this paper involved the synthesis of several datasets including: (i) standard hourly Automated Surface Observing System (ASOS) surface observations [obtained from the University at Albany, State University of New York, surface data archive (not online)], (ii) 5-min temporal resolution Oklahoma mesonet surface data [obtained from the Oklahoma Climatological Survey (see online at http://www.mesonet.org/)], (iii) 6-h temporal and 4-km horizontal resolution gridded precipitation data [obtained from the National Precipitation Verification Unit (see online at http://www.hpc.ncep.noaa.gov/npvu/)], (iv) Weather Surveillance Radar-1988 Doppler (WSR-88D) level-II and level-III data [obtained from the National Climatic Data Center (NCDC; see online at http://www.ncdc.noaa.gov/oa/radar/radarresources.html)], and (v) National Oceanic and Atmospheric Administration (NOAA) Profiler Network hourly wind observations [obtained from the National Center for Atmospheric Research (NCAR; see online at http://dss.ucar.edu/data-sets/ds359.0/)]. Additional datasets, including 1-min temporal resolution ASOS surface observations and BAMEX IOP-8 dropsonde observations and satellite imagery, were obtained from the BAMEX field catalog (see online at http://catalog.col.ucar.edu/bamex/index.html).

Fig. 2. Manual analysis of sea level pressure (solid contours every 2 hPa) and potential temperature (dashed contours every 4°C) overlaid on WSR-88D level-III base reflectivity (dBZ; shaded according to the color bar) for (a) 0600, (b) 0900, and (c) 1200 UTC 11 Jun 2003. (d) NOAA profiler winds (half barb = 2.5 m s⁻¹; full barb = 5.0 m s⁻¹; pennant = 25.0 m s⁻¹) at 3-km MSL at 0600 (black), 0900 (blue), and 1200 (red) UTC 11 Jun 2003. The KTUL ASOS station is marked with a “T” in (a). The MCV position is marked with an “X.” The sample surface station model is shown in (d).
Dynamic tropopause (DT) and other standard synoptic analyses and diagnostic calculations were generated using the National Centers for Environmental Prediction (NCEP) Global Forecast System (GFS) final analyses available at 6-h intervals and at 1.0° × 1.0° horizontal and 50-hPa (25 hPa in lowest 100 hPa) vertical resolution. Quasi-Lagrangian time series of selected model diagnostics are presented in sections 3 and 4, and were azimuthally averaged to a radius of 200 km from the MCV center, except for 850–200-hPa wind shear that was averaged to a radius of 500 km to better represent the background environment. Specifically, the resultant deformation E was calculated as

\[ E = \left( E_{st}^2 + E_{sh}^2 \right)^{1/2}, \]  

(1)

where \( E_{st} = \partial u/\partial x - \partial v/\partial y \) is the stretching deformation and \( E_{sh} = \partial u/\partial x + \partial u/\partial y \) is the shearing deformation (Bluestein 1992, 99–107). The DT was defined at the 1.5 potential vorticity (PV) unit surface (PVU; where 1.0 PVU = 1.0 \times 10^{-6} \text{ m}^2 \text{ s}^{-1} \text{ K kg}^{-1}). The diabatic residual term, \( \partial \theta/\partial t_{\text{DIAB}} \), is calculated as a residual from the local tendency and horizontal advection of potential temperature on the DT, defined as \( \partial \theta/\partial t + \mathbf{V} \cdot \nabla \theta \). Here, \( \partial \theta/\partial t \) is defined as \( \theta \) at time \( [t + 6 \text{ h}] \) minus \( \theta \) at time \( t \) and \( \mathbf{V} \cdot \nabla \theta \) is defined as \( [(\mathbf{V} \cdot \nabla \theta)_{\text{time}=t+6\text{h}} + (\mathbf{V} \cdot \nabla \theta)_{\text{time}=t}] / 2 \). A limitation of the \( \partial \theta/\partial t_{\text{DIAB}} \) calculation is that it represents heating at the DT and not the vertical heating gradient that the MCV actually responds to. While recognizing this limitation, the utility of the \( \partial \theta/\partial t_{\text{DIAB}} \) calculation here is that it indicates simply the presence (or absence) of deep moist convection within the MCV circulation as nonconservation of \( \theta \) on the DT.

The energetics calculations presented in section 5 follow the approach originally developed by Lorenz (1955), and reformulated by Muench (1965) and Norquist et al. (1977). Following the notation of recent studies by Moore and Montgomery (2005) and Conzemius et al. (2007), the time rate of change of eddy available potential energy (APE) is given by

\[ \frac{\partial A_E}{\partial t} = C_A - C_E + G_E, \]  

(2)

where \( A_E \) is the eddy APE, \( C_A \) is the conversion from basic-state APE to eddy APE, \( C_E \) is the conversion of eddy APE to kinetic energy, and \( G_E \) is the diabatic source of eddy APE. The following definitions of these terms rely on area mean quantities (denoted by \( \langle \rangle \)), zonally averaged values (denoted by \( \langle \rangle \)), perturbations from the area mean (denoted by \( \delta \)), and perturbations from the zonal mean (primed values):

\[ AE = \int_{p}^{p_1} \frac{\left[ \mathbf{T}^2 \right]}{2\sigma} dp, \]  

(3)

\[ CA = -\int_{p}^{p_1} \frac{\partial \theta/\partial t [\mathbf{T}]}{\sigma} dp - \int_{p}^{p_1} \frac{\mathbf{V} \cdot \nabla \theta}{\sigma} dp, \]  

(4)

\[ CE = -\frac{1}{g} \int_{p}^{p_1} \frac{R}{p} \left[ \mathbf{\omega} \cdot \mathbf{T} \right] dp, \]  

(5)

\[ GE = \int_{p}^{p_1} \frac{Q \mathbf{T}}{\epsilon_p \sigma} dp, \]  

(6)

Variables in these equations have their usual meteorological meanings, where the static stability (\( \sigma \)) is defined following Norquist et al. (1977) as

\[ \sigma = \frac{1}{g} \left( \frac{\mathbf{T}}{\epsilon_p} - \frac{p}{R} \frac{\partial \mathbf{T}}{\partial p} \right). \]  

(7)

The Lagrangian derivative of potential temperature is used to compute the diabatic heating \( (Q) \) in Eq. (6) because the 6-h temporal resolution of the 1.0° NCEP GFS analysis data leads to large computational residuals in the Eulerian calculation (Moore and Montgomery 2005). Diabatic heating is therefore calculated as

\[ Q = \frac{d \theta}{dt} \epsilon_p \left( \frac{p}{p_0} \right)^{R/\epsilon_p}, \]  

(8)

where \( d \theta/\partial t \) is computed directly using the 12-h air parcel trajectory centered on each grid point in the domain of interest at every analysis time. Linear temporal interpolation is undertaken at 2-h intervals during the trajectory calculation in an attempt to mitigate errors associated with flow evolution between analysis times. This trajectory analysis strategy has been shown by McTaggart-Cowan et al. (2007) to be sufficiently accurate in capturing the effects of both radiative cooling and latent heating on parcel potential temperature values.

In this study, the area of interest is an 1100 × 1100 km² box, with 1100 km being approximately one wavelength, centered on the MCV at each analysis time from 0000 UTC 11 June to 1200 UTC 14 June 2003 (cf. MCV track in Fig. 1). The MCV center was defined as the 600-hPa relative vorticity center in the 1.0° NCEP GFS final analyses. The limits of integration in Eqs. (3)–(6) are \( p_s = 925 \text{ hPa} \) and \( p_1 = 250 \text{ hPa} \). The upper limit matches that used by Conzemius et al. (2007) in their idealized representation of this MCV, while the lower limit is raised to eliminate uncertainties introduced by diabatic calculations within the boundary layer of the 1.0° NCEP.
GFS analysis dataset and the intersection of the ground with the next lower model analysis level. Sensitivity tests of the energetics calculations to modification of the upper and lower bounds of the integrals showed no qualitative impact on the results.

3. MCV genesis and amplification

At 0000 UTC 10 June 2003, an upper-level disturbance embedded along the subtropical jet (STJ) axis crossed the Rockies and enhanced diurnally forced convection over central New Mexico (Galarneau and Bosart 2006; Hawblitze et al. 2007). This convection became organized into a mature MCS over western Texas by 0600 UTC 10 June (Fig. 1). A midlevel MCV formed along the northern end of this mature MCS during 0300–0600 UTC 10 June, and is evident in the radar imagery southwest of Amarillo and north of Midland, Texas, by 0600 UTC 10 June. The midlevel MCV and upper-level STJ disturbance became associated with a subsequent episode of convection over western Oklahoma at 0000 UTC 11 June (not shown). This convection subsequently organized into a trailing stratiform MCS over central and eastern Oklahoma by 0600 UTC 11 June (Fig. 1).

The 3-km NOAA profiler winds (Fig. 2d) show that the Oklahoma MCS (cf. Fig. 1) at 0600 UTC 11 June is embedded in broad cyclonic flow as shown by the \(-10 \, \text{m} \, \text{s}^{-1}\) northwesterly, westerly, and southwesterly winds at Vici (VCI), Purcell (PRC), and Haskell (HKL), Oklahoma, respectively. A time–height section at PRC (Fig. 3a) shows the passage of the MCS near 0300 UTC 11 June, with deep warm-air advection and backed near-surface winds apparent during 2300 UTC 10 June–0200 UTC 11 June. The midlevel MCV is evident near 5 km, and has an apparent westward tilt as shown by the northerly winds growing upward above 5 km with time after 0300 UTC 11 June. The time–height section at HKL (Fig. 3b) shows a deeper cyclonic circulation extending from 1 to 10 km, with a westward tilt above 7 km, as compared to PRC. The deeper circulation at HKL versus PRC is likely because the vortex passed directly over (north of) HKL (PRC). Regardless of the possible incomplete sampling of the vortex at PRC, the HKL profiler wind observations establish the presence of a near-tropospheric deep vortex, suggesting that the development of convection over Oklahoma likely contributed to the deepening of the MCV leading up to the HKL wind observations near 0600 UTC 11 June.

Although the MCV is a deep vortex embedded within the trailing stratiform region of the eastern Oklahoma MCS, the surface analysis at 0600 UTC 11 June (Fig. 2a) does not indicate an obvious cyclonic circulation over eastern Oklahoma. An \(\sim1006\,\text{hPa}\) surface trough is apparent ahead of the MCS, while a bubble high structure near \(\sim1010\,\text{hPa}\) is indicated just west of the leading convective line. An attendant cold pool, with surface potential temperature values near \(20^\circ\text{C}\), is juxtaposed with the bubble high structure and represents a \(-2^\circ\) to \(-3^\circ\text{C}\) perturbation relative to the environment east of the MCS. The meteogram from Tulsa (KTUL), Oklahoma (Figs. 4a,b), shows an \(-10^\circ\text{C}\) decrease in temperature, \(-5\,\text{hPa}\) increase in sea level pressure, and \(-11\,\text{m} \, \text{s}^{-1}\) northerly gust with the passage of the leading convective line at 0230 UTC 11 June associated with the MCS. These temperature and sea level pressure perturbation values are slightly larger in magnitude than indicated on the manual analysis at 0600 UTC 11 June (Fig. 2a) and is likely due to the higher temporal resolution available for the meteogram and/or slight weakening of the MCS. By 0900 UTC 11 June, evidence for a developing surface cyclonic circulation is apparent over northeast Oklahoma and northwest Arkansas with a weak low pressure center (\(-1006\,\text{hPa}\)) shown east of KTUL (Fig. 2b). The surface cyclonic circulation and low pressure center (\(-1008\,\text{hPa}\)) became better defined by 1200 UTC 11 June over northwest Arkansas as the MCS has virtually decayed (Fig. 2c).

During this first convective retriggering cycle over Oklahoma during 0000–1200 UTC 11 June, the \(850–700\,\text{hPa}\) layer-averaged relative vorticity increased from \(2.5 \times 10^{-5} \, \text{s}^{-1}\) to \(5.3 \times 10^{-5} \, \text{s}^{-1}\), while potential temperature on the DT increased from 337 to 345 K during the same period, likely due entirely to diabatic heating \([\partial \theta / \partial t]_{\text{DIAB}}. \, +15–20 \, \text{K} (24 \, \text{h}^{-1})\) associated with the Oklahoma MCS (Figs. 5a,b). This increase in \(850–700\,\text{hPa}\) layer-averaged relative vorticity, concurrent with an increase of DT \(\theta\) through \(\partial \theta / \partial t]_{\text{DIAB}}\), suggests an inherent link between MCV amplification and convective processes. Zonal cross sections of absolute vorticity and vertical motion (not shown), accompanied by radar imagery (not shown), suggested that the growth of \(850–700\,\text{hPa}\) cyclonic vorticity, and the attendant development of a negative sea level pressure perturbation (Figs. 2a–c), may have been a response to stretching \((\partial \omega \partial \rho \gg 0)\) in the lowest 200 hPa of the troposphere near the leading convective line, and possibly the rearward MCS-relative movement of a line-end vortex on the poleward end of the leading convective line during 0000–0600 UTC 11 June.

A second convective retriggering cycle began at 1800 UTC 11 June over the Mississippi Valley southeast of the MCV center (Fig. 1) in a region of conditional instability and enhanced vertical wind shear (Trier and Davis 2007). During 1800 UTC 11 June–0000 UTC 12 June, the \(850–700\,\text{hPa}\) layer-averaged relative vorticity
increased from $4.5 \times 10^{-5} \text{ s}^{-1}$ to $6.0 \times 10^{-5} \text{ s}^{-1}$, while potential temperature on the DT increased to near 347 K by 0000 UTC 12 June. This increase in potential temperature on the DT likely occurred in response to diabatically induced ridging aloft [$\frac{\partial \theta}{\partial t_{\text{DIAB}}} \sim +20-25 \text{ K (24 h)}^{-1}$], which resulted in weaker flow aloft with an attendant decrease in the environmental 850–200-hPa wind shear to <7.0 m s$^{-1}$ (Figs. 5a,b).
4. MCV interaction with its environment

At 0000 UTC 12 June, the MCV was located over southern Illinois along the equatorward jet-entrance region of an upper-level jet streak located over the Great Lakes and southern Ontario (Fig. 6a). The relative position of the MCV and the jet-entrance region is consistent with the shift of precipitation to the north side of the MCV center by 0000 UTC 12 June (Fig. 1). The equatorward jet-entrance region moved eastward to over northern New York and New England by 0000 UTC 13 June (Figs. 6a–c). This eastward movement placed the equatorward jet-entrance region farther east of the MCV center relative to its position at 0000 UTC 12 June. The relative eastward movement of the jet-entrance region, concurrent with the development of warm-air advection east of the MCV as it approached the baroclinic zone over the Great Lakes...

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**FIG. 4.** Meteogram of (a) sea level pressure (hPa), air temperature (°C), dewpoint (°C), and (b) wind speed (m s⁻¹) and direction (meteorological degrees) at 1-min temporal resolution for KTUL during 0000–1200 UTC 11 Jun 2003.

**FIG. 5.** MCV-following time series of (a) 850–700-hPa layer-averaged relative vorticity ($\zeta; 10^{-5}$ s⁻¹), 925–850-hPa layer-averaged resultant deformation ($E; 10^{-5}$ s⁻¹), and the diabatic heating residual term $[\partial \theta / \partial t]_{DIAB};$ K (24 h)⁻¹ and (b) dynamic tropopause potential temperature ($\theta; K$) and 850–200-hPa wind shear (m s⁻¹). The 850–200-hPa wind shear is (all other parameters are) azimuthally averaged to a 500-km (250 km) radius.
FIG. 6. Sea level pressure (solid contours every 2 hPa), 1000–500-hPa thickness (dashed contours every 3 dam), and 850–200-hPa wind shear (shaded according to the grayscale; m s$^{-1}$) at (a) 0000 UTC 12 Jun, (b) 1200 UTC 12 Jun, (c) 0000 UTC 13 Jun, (d) 1200 UTC 13 Jun, (e) 0000 UTC 14 Jun, and (f) 1200 UTC 14 Jun 2003. The MCV position is marked with an “X,” and the coastal cyclone is marked with an “L.”
associated with the predecessor extratropical cyclone over northern Maine and southern Ontario (Figs. 7a–c), is in agreement with the relative eastward movement of precipitation during this period. This change in precipitation areal coverage suggests that the warm-air advection, and associated ascent, driving the heavy precipitation (12-h precipitation totals approach 32–64 mm over parts of Pennsylvania, Ohio, New York, and New England; Figs. 7a–d) were moving away from the MCV core with time, which contributed to the MCV’s subsequent weakening. By 0000 UTC 14 June, the equatorward jet-entrance region was located over Maine and just off the east coast where a coastal redevelopment was under way (Fig. 6e). By 1200 UTC 14 June, a 1010-hPa coastal low was developing in the favorable equatorward jet-entrance region while the remnant MCV, located over the St. Lawrence Valley in northern New York, continued to weaken (Fig. 6f).

The DT perspective reveals the presence of an upper-level disturbance embedded in the STJ (labeled “A”) just west of the MCV, and an upper-level jet entrance region north of the MCV, at 0000 UTC 12 June (Fig. 8a). This STJ disturbance A was the original system that helped trigger the convection over eastern New Mexico and west Texas at 0600 UTC 10 June, which led to the initial development of the MCV (Fig. 1). By 1200 UTC 12 June, STJ disturbance A was located directly over the MCV center, and by 0000 UTC 13 June it, along with the upper-level jet entrance region, had moved east of the MCV center (Figs. 8b,c). A sequence of cross sections showing potential vorticity, potential temperature, and relative humidity from 0000 UTC 12 June to 0000 UTC 13 June show the relative movement of STJ disturbance A from a position upstream of the MCV at 0000 UTC 12 June to a location downstream of it by 0000 UTC 13 June (Figs. 9a–c). This change in the

![Fig. 7. 850-hPa temperature (thin solid contours every 3 K), relative vorticity (thick solid contours every $4.0 \times 10^{-5} \text{ s}^{-1}$ starting at $8.0 \times 10^{-5} \text{ s}^{-1}$), wind (barbs as in Fig. 2), and 12-h total precipitation (shaded according to the color bar; mm) centered on the time of analysis at (a) 0000 UTC 12 Jun, (b) 1200 UTC 12 Jun, (c) 0000 UTC 13 Jun, and (d) 1200 UTC 13 Jun 2003. The MCV position is marked with an “X,” the predecessor extratropical cyclone is marked with an “L,” and the cold front is marked with an “F” (see also Table 1).]
FIG. 8. Dynamic tropopause potential temperature (shaded according to the color bar; K), wind (barbs as in Fig. 2), and 850-hPa relative vorticity (solid contours every $4.0 \times 10^{-5}$ s$^{-1}$ starting at $8.0 \times 10^{-5}$ s$^{-1}$) at (a) 0000 UTC 12 Jun, (b) 1200 UTC 12 Jun, (c) 0000 UTC 13 Jun, (d) 1200 UTC 13 Jun, (e) 0000 UTC 14 Jun, and (f) 1200 UTC 14 Jun 2003. Subtropical jet disturbances are marked as “A” and “B,” and the Canadian upper-level trough is marked with a white-dashed line. Cross-sectional orientations for Fig. 9 are indicated.
Fig. 9. Cross sections of potential vorticity (shaded according to the grayscale; PVU), potential temperature (solid contours every 3 K), and relative humidity (dashed contours every 10%) at (a) 0000 UTC 12 Jun, (b) 1200 UTC 12 Jun, (c) 0000 UTC 13 Jun, (d) 1200 UTC 13 Jun, (e) 0000 UTC 14 Jun, and (f) 1200 UTC 14 Jun 2003. The subtropical jet disturbance is marked with an “A,” and the Canadian upper-level trough is marked with a white-dashed line. Cross-sectional orientations are indicated on Fig. 8.
tropospheric-deep tilt of the MCV–STJ disturbance A system, and the increase of 925–850-hPa resultant deformation from $\sim 3.0 \times 10^{-5} \text{s}^{-1}$ at 1800 UTC 11 to $\sim 4.0 \times 10^{-5} \text{s}^{-1}$ after 0600 UTC 12 June (Fig. 5a) associated with the interaction of the MCV with the baroclinic zone over the Great Lakes (note elongation of MCV in Figs. 7a–c), occurred concurrently with the weakening of the MCV vortex by 0000 UTC 13 June.

By 1200 UTC 13 June, the MCV was located just west of Buffalo, New York, beneath a potential temperature ridge on the DT (Fig. 8d). Farther west, however, an upper-level trough was located over southern Canada near 93°W and another STJ disturbance (labeled “B”) was located over Missouri. Both of these upper-level disturbances moved eastward toward the remnant MCV, and by 1200 UTC 14 June were located just west of the remnant low-level MCV (Figs. 8e,f and 9e,f). A cyclonic vorticity maximum formed over the Gulf of Maine by this time as the remnant MCV dissipated and coastal redevelopment occurred in response to the approach of the Canadian upper-level trough and STJ disturbance B (Figs. 8f and 9f).

In summary, environmental conditions were not favorable for rapid MCV decay through 1200 UTC 12 June as STJ disturbance A was positioned on the western periphery of the MCV and the upper-level jet entrance region was positioned poleward of the MCV, which helped generate precipitation within the warm-air advection region over and just northeast of the MCV center. Subsequent to 1200 UTC 12 June, however, the environment became more favorable for MCV decay as STJ disturbance A at the upper-level jet entrance region moved downstream away from the MCV. This placed the MCV in a region of anticyclonic vorticity advection as the system tilted downshear, and displaced the warm-air advection region, and much of the attendant precipitation, well north and east of the MCV center and toward the equatorward entrance region of the downstream jet.

5. Energetics

Conzemius et al. (2007) conducted a set of idealized numerical simulations of an MCV with similar vertical structure (cf. Figs. 9a,b with their Fig. 5b) embedded in an analogous background environment (cf. Fig. 6a with their Fig. 1d) as presented here. They estimated the relative contributions to MCV intensification from diabatic and baroclinic processes by calculating the diabatic ($G_E$) and baroclinic ($C_A$) energy conversion (to eddy APE) terms as in Moore and Montgomery (2005). The results from their control run showed that initial MCV amplification was due to diabatic heating $G_E$, which was the dominant source of the eddy APE (their Fig. 11a). After this initial growth phase, the MCV was in a quasi-steady state for a short period, followed by a rapid intensification phase as the baroclinic $C_A$ and diabatic $G_E$ terms increase in tandem. They suggest that the rapid increase in the baroclinic conversion term $C_A$ was the result of excitation by the larger-scale baroclinic system.

Given the similarity in structure between the idealized case used by Conzemius et al. (2007) and the case presented here, their results provided motivation to calculate the energy conversion terms, $C_A$ and $G_E$, during the amplification and subsequent baroclinic transition of the MCV in this study (Fig. 10). The mechanics of the calculations were described in section 2. The results show that diabatic heating $G_E$ ($0.5$–$1.0 \text{ kg s}^{-3}$) was the dominant source for eddy APE, particularly at 1200 and 1800 UTC 11 June, which corresponds to deepening of the MCV to $\sim 1005 \text{ hPa}$ while convection was active within the MCV’s circulation (see also Fig. 1).

The diabatic conversion term $G_E$ approached $\sim 0 \text{ kg s}^{-3}$ at 0000 UTC 12 June and thereafter, corresponding to the displacement of diabatic heating to the northeast away from the MCV center in association with the development of stratiform precipitation, with embedded banded structures, in the region of warm-air advection along the northeast flank of the MCV (see also Fig. 1). Baroclinic energy conversion $C_A$ was slightly positive, near 0.2 $\text{ kg s}^{-3}$ [comparable initially to Fig. 11a in Conzemius et al. (2007)], at 0000 and 0600 UTC 12 June, corresponding with sea level pressure values remaining nearly constant ($\sim 1005 \text{ hPa}$). The subsequent $\sim 4.5$-hPa increase in sea level pressure during 1200 UTC 12 June–0600 UTC 14 June corresponds with the diabatic $G_E$ and baroclinic $C_A$ conversion terms hovering between slightly negative and zero. The slightly negative value ($\sim -0.2 \text{ kg s}^{-3}$) of the baroclinic energy conversion term $C_A$ after 1200 UTC 13 June was likely associated with STJ disturbance A moving downstream away from the MCV (see also Figs. 8b,9c and 10b,c). This change in tropospheric-deep “tilt” between the MCV and STJ disturbance A, combined with the displacement of widespread precipitation and associated heating away from the MCV center ($G_E \sim 0 \text{ kg s}^{-3}$) to counteract the debilitating effects of increasing low-level deformation as the MCV approached the cold front F over the Great Lakes (cf. Fig. 7), likely contributed to the weakening of the MCV after 1200 UTC 12 June. This change in vertical tilt of the MCV and the loss of convection near the MCV core is the reason for the inconsistency with the results from Conzemius et al. (2007, see their Fig. 11a) where the MCV continued to strengthen after 160 h in
their simulation. The large increase of $G_E$ after 0600 UTC 14 June was associated with convection in the warm sector of the developing coastal cyclone (cf. Fig. 6f) that was occurring along the eastern side of the calculation domain, and explains why the MCV continued to weaken despite this large increase of $G_E$.

6. Evolution of surface boundaries

a. Initiation of fronts and convective retriggering

At 1800 UTC 11 June, the MCV was located over southwest Missouri while convection reinitiated along meridionally oriented bands southeast of the MCV center over the Mississippi Valley, western and central Kentucky, and eastern Mississippi (Fig. 11a). This convection formed in the vicinity of surface boundary 1 (cf. Table 1), stretching east-southeast from the MCV center through central Tennessee (Fig. 11a). This convection separated a slightly cooler maritime tropical (mT) air mass with surface $\theta$ values near 23°C north, from a warmer mT air mass with surface $\theta$ values near 30°C south, of the boundary. Profiler wind data from Conway, and Bloomfield, Missouri (Figs. 12a,b), indicate the presence of a strong ~15–25 m s$^{-1}$ southwesterly flow in the 1.0–5.0-km layer just east of the MCV center during 0500–1000 and 1200–1900 UTC 11 June, respectively. This strong low-level flow represents a westerly/southwesterly jet along the southern flank of the MCV (note ~23 m s$^{-1}$ wind at Dequeen, Arkansas (DQU) at 0600, 0900, and 1200 UTC 11 June; Fig. 2d), that was likely maintained by the geopotential height gradient, at and below 850 hPa, between the northeastward-moving MCV and the persistent anticyclone over the southeast United States (not shown). This low-level jet is oriented approximately perpendicular to surface boundary 1, with the nose of the jet near the boundary (verified with 20-km Rapid Update Cycle analyses; not shown).

The presence of the strong low-level southwesterly flow above near-surface southerly and southeasterly flow suggest that the convection organized within a corridor of enhanced low-level wind shear just east and south of the MCV center (cf. Fig. 11a over southeast Missouri with Figs. 12a,b). Selected dropsonde observations from BAMEX IOP-8 support the inference of an enhanced low-level wind shear corridor. For example, dropsonde 1 showed nearly 90° of veering in the lowest 200 hPa (Fig. 13a) just east of the vortex center in southeast Missouri (Fig. 14), while dropsonde 3 sampled the environment south of surface boundary 1 (Figs. 11a and 14) and indicated minimal directional wind shear in the lowest 200 hPa. However, the wind shear magnitude

![Fig. 10. MCV-following time series of the minimum sea level pressure (hPa; dotted contour), conversion of basic-state APE to eddy APE ($C_A$; kg s$^{-1}$; dashed contour), and conversion of diabatic heat energy to eddy APE ($G_E$; kg s$^{-1}$; solid contour), computed for an 1100 × 1100 km$^2$ box centered on the MCV.](image-url)
in dropsonde 3 was still near \(\sim 15 \text{ m s}^{-1}\), and the presence of steep lapse rates in the 800–700-hPa layer and warmer surface temperatures (near 24°C) relative to dropsonde 1 resulted in CAPE values over 1200 J kg\(^{-1}\) (Fig. 13a).

By 0000 UTC 12 June, the precipitation around the MCV had become less convective, although bands of stratiform precipitation existed in a region of weak warm-air advection [note region of coincident isentropic ascent at 1730 UTC 11 June in Trier and Davis...
Table 1. Summary of all the labeled features, where applicable, on the surface analyses (Figs. 11 and 15) and summary schematic (Fig. 17).

<table>
<thead>
<tr>
<th>Symbol</th>
<th>Position of MCV center</th>
</tr>
</thead>
<tbody>
<tr>
<td>W</td>
<td>Warm front (former cold front F) moving northward on northeast side of MCV</td>
</tr>
<tr>
<td>cP1</td>
<td>Continental polar air mass north of F that was originally associated with the predecessor extratropical cyclone</td>
</tr>
<tr>
<td>cP2</td>
<td>Region of cP1 air mass that was relatively cooler because of sensible cooling from Lakes Erie and Ontario</td>
</tr>
<tr>
<td>mT1</td>
<td>Maritime tropical air mass that was being advected northward from the south and east flank of the MCV</td>
</tr>
<tr>
<td>mT2</td>
<td>Region of mT1 air mass that was slightly drier because of subsidence on the west side of the MCV</td>
</tr>
<tr>
<td>mT3</td>
<td>Maritime tropical air mass that was slightly rain cooled relative to mT1. This air mass was “left over” on the south side of F, originally associated with the predecessor extratropical cyclone</td>
</tr>
</tbody>
</table>

(2007), see their Fig. 12i] along the north side of the MCV. This banded precipitation occurred between surface boundaries 1 and 3, marked by distinct cyclonic wind shifts, on the north side of the aforementioned southwestern low-level jet concurrent with an enhanced corridor of southwest geostrophic flow (note tightening of sea level pressure gradient southeast of the MCV center between 1800 UTC 11 June and 0000 UTC 12 June; Figs. 11a,b). A narrow line of intense convection also formed along a cold front (labeled F; see Table 1), trailing southwestward from the aforementioned predecessor extratropical cyclone (Fig. 7a). This convection focused over central Illinois and Indiana where east-southeasterly mT flow associated with the MCV’s circulation over southeast Illinois met cool continental polar (cP) northerly flow over northern Illinois and Indiana. Several brief tornadoes, rated F0 on the Fujita scale (Fujita 1981), were reported in association with this narrow convective line (NCDC 2003).

Meanwhile, precipitation was much less extensive west of the MCV (Fig. 11b) where deep northerly flow and weak cold advection, as indicated by backing winds below 500 hPa in dropsondes 21 and 22 at 1846 and 1854 UTC 11 June, respectively, and in the Conway, Missouri, profiler after 1700 UTC 11 June (Fig. 12a), likely contributed to drying associated with subsidence throughout much of the troposphere (see dropsonde 21 in Fig. 13b). This subsidence likely facilitated the development of clear skies southwest of the MCV center (Fig. 14), which in turn created surface boundary 2 (Table 1), enhanced via differential diabatic heating between clear and cloudy air, just west of the Mississippi Valley, that divided the rain-cooled air mT to the east from the warmer and slightly drier cold air over Missouri and Arkansas (Fig. 11b).

b. Baroclinic transition

By 0600 UTC 12 June, the MCV was located over the southern Illinois–Indiana border (Fig. 15a). Surface boundaries 1–3, observed at 1800 UTC 11 June and 0000 UTC 12 June, were still apparent by 0600 UTC 12 June, running east, southwest, and northeast, respectively, from the MCV center. While the MCV center was embedded in a quasi-homeogeneous surface temperature field at this time, the northern periphery of the circulation was interacting with cold front F stretching from central Illinois to northern Ohio. Surface potential temperature values north and south of cold front F were near 15°–18° and 21°–22°C, respectively. Similarly, precipitation became focused along cold front F during 1800 UTC 11 June–0600 UTC 12 June north and east of the MCV center (Figs. 11a,b and 15a). The MCV center remained isolated, however, in the mT air mass on the south side of cold front F (Fig. 15a) with surface temperature values remaining steady near 21°C during the passage of the MCV at Evansville, Indiana (EVV; Fig. 16a). A sea level pressure perturbation of ~4 to ~5 hPa, with a gradual veering of surface winds from southeasterly to northeasterly, was also observed at EVV during the passage of the MCV at 0600–1200 UTC 12 June.

By 1200 UTC 12 June, the MCV center became fully embedded within the cold front F over northern Ohio and central Indiana while surface boundaries 3 and 4 were absorbed by cold front F (Fig. 15b). Much of the precipitation near the MCV center had dissipated by 1800 UTC 12 June, while new precipitation had developed north of a warm frontogenesis region from northwest Ohio eastward along the southern shore of Lake Erie (cf. Figs. 15b,c). This warm-frontal precipitation developed in a region of enhanced isentropic ascent where the low-level jet (~20 m s⁻¹) east of the MCV center interacted with cold front F (not shown). Meanwhile, the leading edge (cold front 5; see Table 1) of slightly cooler cP air, likely due to sensible cooling from Lakes Erie and Ontario (not shown), north of the cold front F was located over extreme northwest Ohio and southern Michigan (Fig. 15c). By 0000 UTC 13 June, the MCV was positioned over the southwestern shore of Lake Erie (Fig. 15d). Precipitation was enhanced along the warm front (former cold front F) over the southern
shore of Lake Erie, while scattered convection formed over and east of surface boundary 2 over eastern Ohio and western Pennsylvania. Some light precipitation was occurring near the weakening cold front F over western Ohio, while the stronger cold front 5 was still located over the Michigan–Ohio border. Cold front 5 passed through Toledo, Ohio (KTDZ; just west of the MCV center), during 1915–1930 UTC 12 June and was associated with a ~4°C decrease in temperature as the surface wind veered from southwesterly to northerly (Figs. 16c,d).
c. Summary of surface boundary evolution

The evolution of surface boundaries during the baroclinic transition of the MCV are summarized in Fig. 17. Initially (time $t$), surface boundaries 1–3 separated an mT air mass that was unmodified (mT$_1$), adiabatically warmed in conjunction with subsidence (mT$_2$), and rain cooled (mT$_3$). Cold front F was located north of the MCV, and separated the three mT air masses around the MCV center from a cP air mass. At an intermediate time ($t + \delta t$), cold front F absorbed surface boundary 1 and became embedded in the MCV’s cyclonic circulation. Cold front 5 formed along the leading edge of a region of cP air cooled via sensible heating over Lakes Erie and Ontario (cP$_2$). At the final time ($t + 2\delta t$), cold front 5, cold front F, and surface boundary 2 were located west, southwest, and southeast, respectively, of the MCV center. Cold front F and surface boundary 2 were initiating precipitation. Warm front W, northeast of the MCV center, was moving northward as former cold front F. As a result, widespread stratiform precipitation formed in a region of warm-air advection near the warm front over Ohio, Pennsylvania, and New York by 1200 UTC 12 June (Figs. 7b and 15b).

7. Discussion

The MCV presented in this paper began as a midlevel cyclonic circulation along the poleward end of an MCS over eastern New Mexico and western Texas at 0600 UTC 10 June as shown in Galarneau and Bosart (2006; also see the current Fig. 1). This parent MCS was triggered by an eastward-moving upper-level disturbance embedded along the STJ axis. Hawblitzel et al. (2007) demonstrated, using ensemble forecasting techniques, that the amplitude of the STJ disturbance was important in triggering the initial MCS that led to the development of the MCV. The MCV triggered another MCS over western Oklahoma by 0000 UTC 11 June (Fig. 1; see Fig. 2d in Davis and Trier 2007). The MCV grew vertically and amplified in the presence of the Oklahoma MCS, evolving into a near-tropospheric-deep vortex with an attendant surface cyclonic circulation by 1200 UTC 11 June while over northwest Arkansas (Fig. 2c). The development of low-level cyclonic vorticity beneath the midlevel MCV may have been associated with stretching in the lower troposphere and the rearward MCS-relative advection of a line-end vortex, which is consistent with previous studies by Davis and Trier (2002). A forthcoming study by Davis and Galarneau (2009) will examine the physical processes important in low-level cyclonic vorticity generation for this MCV case, and the 6 July 2003 MCV case also during the BAMEX field program, through numerical simulations. Energetics calculations (as in, e.g., Conzemius et al. 2007) indicated that vortex amplification over Oklahoma was strictly due to diabatic processes associated with convection (Fig. 10), which is consistent with
previous studies showing that MCVs amplify during convective retriggering cycles (e.g., Fritsch et al. 1994; Rogers and Fritsch 2001).

A second convective retriggering cycle was observed at 1800 UTC 11 June, during BAMEX IOP-8, over the Mississippi Valley south and east of the MCV center (Fig. 11a). The convection formed in a region of favorable instability (CAPE values \( \sim 1000 \text{ J kg}^{-1} \)) and low-level wind shear (\( \sim 15-20 \text{ m s}^{-1} \) 0–6-km wind shear) associated with the leading edge of a low-level jet along the southern flank of the MCV (Figs. 12b and 13a; see also Fig. 17 in Trier and Davis 2007). This convection was also concentrated along surface boundaries on the south and east flank of the MCV near 1800 UTC 11 June (Fig. 11a; see also Fig. 14d in Trier and Davis 2007), in a broad region of isentropic upward motion (see Fig. 12i in Trier and Davis 2007). The presence of isentropic ascent on the downshear side of an MCV is consistent

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**Fig. 14.** GOES-12 1-km visible imagery (shaded according to the grayscale) at 1739 UTC 11 Jun 2003. The flight track (thin gray contour) and Learjet dropsonde locations (triangles) for BAMEX IOP-8 are overlaid. The MCV position is marked with an “X.” This figure was obtained from the BAMEX field catalog.
with the Raymond and Jiang (1990) conceptual model associated with an interior PV anomaly embedded in an environment with weak westerly shear (see Fig. 1 in Trier and Davis 2007).

By 0000 UTC 12 June, convection and widespread stratiform precipitation became focused on the north and east flank of the MCV, just north of a surface boundary extending southeast from the MCV center (Fig. 11b). This precipitation was driven by isentropic ascent and weak warm-air advection (see also Fig. 5d in Davis and Trier 2007) along the nose of a low-level jet along the southeast side of the MCV center. Meanwhile, multiple surface boundaries were evident within the MCVs circulation, prior to completion of baroclinic transition, by 0000 UTC 12 June (Fig. 11b; Table 1). The evolution of these surface features during baroclinic transition is summarized in Fig. 17 and Table 1.

The focusing of precipitation in the warm-air advection (Figs. 7b and 11b) and equatorward jet-entrance regions (Figs. 8a,b) likely precluded significant low-level cyclonic vorticity generation associated with diabatic heating near the MCV center. This vorticity generation was needed to resist the debilitating effects of low-level deformation and increasing environmental wind shear on the vortex (Figs. 5a,b). The right-of-track precipitation distribution in a region of warm-air advection, and weakening of the MCV after baroclinic transition, is consistent with the post-ET decay paradigm as described, for example, in Hart et al. (2006) and Atallah et al. (2007). An interesting aspect of this MCV case is that despite a nontropical origin and a much weaker vortex than a remnant TC, there are similar rainfall patterns as with landfalling ET cases with right-of-track precipitation. This evolution of the vortex and attendant precipitation contrasts, however, with the Zhang and Harvey (1995) study of extratropical cyclogenesis from an MCV. Their case was essentially an ET of a warm-core MCV that interacted with an amplifying...
upstream trough of approximately neutral horizontal tilt (see their Figs. 2 and 12), and a low-level baroclinic zone that was enhanced by an extensive cold pool generated by the parent MCS. Post-ET intensification as a growing baroclinic wave was observed in their case, which is consistent with the upshear vertical tilt of the trough axis (see their Figs. 17 and 18), and in agreement with the post-ET intensification and left-of-track precipitation described in Hart et al. (2006) and Atallah et al. (2007).

Along with contributing to the weakening of the MCV after baroclinic transition, other consequences from a lack of convection within the MCV’s circulation include a negligible contribution of diabatic energy conversion to eddy APE, and cessation of PV reduction aloft. The importance of PV reduction aloft, concurrent with momentum redistribution, is that it may have reduced the environmental wind shear and steering flow over and east of the MCV, which helped to keep STJ disturbance A from moving downstream away from the MCV. The implications of the movement of STJ disturbance A downstream of the MCV during and after the baroclinic transition was that it resulted in a downshear-tilted system, which is considered unfavorable for baroclinic development because of the presence of anticyclonic vorticity advection over the surface low center [Figs. 9a–d; cf. the Holton (2004) quasigeostrophic two-layer model], and verified by small contributions to eddy APE by baroclinic processes $C_A$ (Fig. 10). This change in tilt and eastward movement of STJ disturbance A is concurrent with secondary coastal redevelopment over the Gulf of Maine. This evolution is
analogous to the type B coastal cyclogenesis as depicted in Miller (1946, his Fig. 3), but with the MCV replacing the primary cyclone over the Ohio and St. Lawrence Valleys that contributes to the frontal wave that evolved into the deepening coastal cyclone particularly as the Canadian upper-level trough and STJ disturbance B approached.

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