The Role of Moist Processes in the Formation and Evolution of Mesoscale Snowbands within the Comma Head of Northeast U.S. Cyclones

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ABSTRACT

The role of moist processes in regulating mesoscale snowband life cycle within the comma head portion of three northeast U.S. cyclones is investigated using piecewise potential vorticity (PV) inversion, modeling experiments, and potential temperature tendency budgets. Snowband formation in each case occurred along a mesoscale trough that extended poleward of a 700-hPa low. This 700-hPa trough was associated with intense frontogenetical forcing for ascent. A variety of PV evolutions among the cases contributed to midlevel trough formation and associated frontogenesis. However, in each case the induced flow from diabatic PV anomalies accounted for a majority of the midlevel frontogenesis during the band’s life cycle, highlighting the important role that latent heat release plays in band evolution. Simulations with varying degrees of latent heating show that diabatic processes associated with the band itself were critical to the development and maintenance of the band. However, changes in the meso-a-scale flow associated with the development of diabatic PV anomalies east of the band contributed to frontolysis and band dissipation. Conditional stability was reduced near 500 hPa in each case several hours prior to band formation. This stability remained small until band formation, when the stratification generally increased in association with the release of conditional instability. Previous studies have suggested that the dry slot is important for the initial stability reduction at midlevels, but this was not evident for the three banding cases examined. Rather, differential horizontal temperature advection in moist southwest flow ahead of the upper trough was the dominant process that reduced the midlevel conditional stability.

1. Introduction

Although the role of frontogenesis and symmetric stability in band formation in the comma head of extratropical cyclones has been extensively studied (e.g., Emanuel 1985; Thorpe and Emanuel 1985; Xu 1992; Sanders and Bosart 1985; Martin 1998a,b; Nicosia and Grumm 1999; Moore et al. 2005), the life cycle of observed banding events has rarely been investigated. In a case study of the 25 December 2002 snowstorm over the northeastern United States, Novak et al. (2008, hereinafter NCY08) showed that prior to band formation, conditional stability was small while the forcing for ascent increased along a mesoscale trough extending poleward of the 700-hPa low. During band formation the frontal environment was marked by a region of elevated conditional and inertial instability and increasing frontogenesis. Increasing conditional stability and diminishing frontogenesis characterized band dissipation. Documentation of such a band life cycle is important, however, NCY08 did not address why the frontogenesis and stability evolved in this manner.

Moist processes, in particular latent heat release (LHR), are critical to cyclone development (e.g., Reed and Kuo 1988; Davis 1992; Stoelinga 1996), and such
processes may also be critical to precipitation band evolution. For example, on the scale of a midlatitude cyclone, Posselt and Martin (2004) showed that LHR associated with precipitation poleward of the surface low enhances the development of the potential vorticity (PV) treble-cleft structure aloft. The treble-cleft PV structure, or PV “hook,” is characteristic of occluded cyclones (Martin 1999; Thorncroft et al. 1993), and it supports large-scale forcing for ascent in the occluded quadrant of the cyclone through horizontal rotation of the lower-tropospheric baroclinic zone (i.e., rotational frontogenesis; Martin 1999; Posselt and Martin 2004). Han et al. (2007) demonstrated in the modeling of two central U.S. cases that the frontogenetical forcing for ascent due to diabatic processes within the trough of warm air aloft, or trowal (e.g., Godson 1951; Martin 1998a,b), was at least twice as large as horizontal deformation. Furthermore, there is an emerging view that diabatically generated PV anomalies (Raymond 1992) are common features of extratropical cyclones (e.g., Davis 1992; Stoelinga 1996; Wernli et al. 2002; Moore and Montgomery 2004), and that they may affect mesoscale precipitation distribution (e.g., Brennan and Lackmann 2005; Moore et al. 2008). Thus, it is important to understand how the evolution of upper-level and diabatically generated PV anomalies relate to precipitation bands within cyclones.

On the frontal scale, idealized studies have shown that LHR from a precipitation band may influence the band’s formation and evolution (Thorpe and Emanuel 1985; Cho and Chan 1991; Xu 1992; Davies 1999). In a PV framework, the diabatically generated PV anomaly created by the band will increase (decrease) the stability below (above) the level of maximum heating, and induce a horizontal circulation that may be frontogenetical. Although this feedback is presumably active within a snowband of an extratropical cyclone, the role of the band’s latent heating on its own evolution has not been quantified. It is also not apparent if or how this feedback is disrupted during band dissipation.

The vertical and horizontal distribution of moisture may also play a critical role in band evolution by modifying the stability. Conceptual models of the synoptic environment of band formation have highlighted a region of saturation equivalent PV (EPV) reduction southeast of the frontogenesis maximum and associated band (e.g., Nicosia and Grumm 1999, their Fig. 17; Moore et al. 2005, their Fig. 15a), in a layer associated with the midlevel dry slot. This location is consistent with the results of Cao and Cho (1995) and Cho and Cao (1998), who showed that negative EPV can be generated in regions where the thermal wind vector points in the same direction as the moisture gradient. However, banding events can occur well displaced (~200 km) from the dry slot, or form hours before a robust dry slot develops (Clark et al. 2002; NCY08). Thus, it is unclear if the dry slot is even relevant to band formation, and if not, how the stability is reduced.

This paper builds on NCY08’s description of the evolution of an intense banded event to answer the following specific questions:

1) What role do moist processes play in the genesis and evolution of the midlevel frontogenesis maximum in the comma head of banded cyclones?
2) What influence does latent heating associated with the band have on band evolution?
3) What processes are responsible for reducing conditional stability prior to band formation?

NCY08 demonstrated the capability of a triply nested modeling system (36-, 12-, and 4-km grid spacing) to realistically simulate intense banding during the 25 December 2002 northeast U.S. snowstorm. This system will be used to conduct piecewise PV inversions, modeling experiments, and a potential temperature (θ) tendency budget for the 25 December 2002 snowstorm to answer the above questions. The methods used to investigate the 25 December 2002 snowstorm are also applied to two additional cases to further generalize the results. Section 2 describes the methods used in the study. Sections 3 and 4 present analysis of the frontogenetical forcing and stability evolution of the 25 December 2002 snowstorm, respectively. Section 5 presents results from two other major banded snowstorms. A discussion and conclusions are provided in section 6.

2. Datasets and methodology

The realistic simulation of the 25 December 2002 snowstorm by NCY08 is used as the “control run” in this study. The simulation was accomplished using the fifth-generation Pennsylvania State University–National Center for Atmospheric Research (NCAR) Mesoscale Model (MM5; Dudhia 1993) in a nested configuration at 36-, 12-, and 4-km grid spacing. The run was initialized at 0000 UTC 25 December 2002, which is approximately 19.5 h prior to band formation. Further details of the simulation configuration are found in NCY08.

The processes responsible for the midlevel trough evolution and associated frontogenesis evolution are investigated using the piecewise PV inversion technique of Davis and Emanuel (1991), and other modeling experiments. The stability evolution is investigated via application of a θ tendency budget and model trajectory analysis in section 4. These techniques are also applied to two additional cases to explore the generality of the results.
a. Piecewise potential vorticity inversion

This study applied piecewise PV inversion at 12-km grid spacing in the framework of nonlinear balanced dynamics (Charney 1955). NCY08 showed that the first-order attributes of a mesoscale band can be realistically simulated at 12-km horizontal grid spacing. Tests using a 36-km PV inversion showed qualitatively similar results to 12 km (height minima, and wind and frontogenesis maxima in the same locations), except that the 12-km inversion exhibited more realistic shapes and amplitudes of features. To facilitate numerical convergence for the 12-km inversion, the 12-km height field was smoothed using a 25-point smoother and the 12-km domain of NCY08 was expanded (Fig. 1). Sensitivity tests confirm that expanding the domain size had little impact on the band evolution and cyclone depth (within 1 hPa of the NCY08 run).

As applied in McTaggart-Cowan et al. (2006, 1738–1742), PV and \( u \) perturbations are taken relative to a basic baroclinic state consistent with Eady model constraints (Eady 1949), with a constant buoyancy (\( N^2 = 0.015 \text{ s}^{-2} \), where \( N \) is the Brunt–Väisälä frequency), vertical shear (2.5 m s\(^{-1}\) km\(^{-1}\)), and meridional temperature gradient [0.6 K (100 km)\(^{-1}\)]. The PV perturbations (\( q' \)) and \( \theta \) perturbations (\( \theta' \)) are defined by subtracting the basic-state PV and \( \theta \) fields from the modeled PV and \( \theta \) fields, respectively. The \( q' \) field was separated into four parts: upper-level (\( q'_{u} \)), which includes all \( q' \) above and including 500 hPa; lower boundary (\( q'_{b} \)), which includes the \( \theta' \) at 950 hPa; moist interior (\( q'_{m} \)), which includes all \( q' \) between 950 and 550 hPa where the RH was >70%; and dry interior (\( q'_{d} \)), which includes the remaining \( q' \) in the 950–550-hPa layer. The \( q'_{m} \) generally represents PV associated with diabatic process occurring in the 950–550-hPa layer.

To further identify the kinematic contribution of each PV piece to the forcing for ascent, piecewise frontogenesis (Morgan 1999; Korner and Martin 2000) was employed. In this method, the Petterssen (1936) 2D frontogenesis equation:

\[
F_{\text{2D}} = \frac{1}{|\nabla\theta|} \left[ -\frac{\partial\theta}{\partial x} \left( \frac{\partial u}{\partial x} \frac{\partial \theta}{\partial x} + \frac{\partial v}{\partial x} \frac{\partial \theta}{\partial y} \right) \\
-\frac{\partial\theta}{\partial y} \left( \frac{\partial u}{\partial y} \frac{\partial \theta}{\partial x} + \frac{\partial v}{\partial y} \frac{\partial \theta}{\partial y} \right) \right],
\]

is calculated from the induced flow from a particular PV piece and the control run potential temperature gradient. Thus, frontal forcing for ascent can be attributed to specific parts of the PV field. Particular focus is placed on the 700-hPa level, since this was the level of maximum frontogenesis in the 25 December 2002 snowstorm (NCY08), and it is commonly near the level of maximum frontogenesis in northeast U.S. band events (Novak et al. 2006).

b. Dry run experiments

To isolate the effect of moist processes on band evolution, a simulation identical to the control run was integrated in a fake-dry configuration. The “fake dry” run option of the MM5 system sets the temperature tendency from the microphysics and convective schemes to zero while retaining all other effects of moisture (Grell et al. 1995; NCAR 2005). This run is termed the “dry run.” Comparison of the dry and control runs shows the cumulative effect of LHR. A second run (“delayed dry run”) was restarted with the 18-h forecast from the control run (1.5 h prior to band formation) and was integrated in a fake-dry configuration; thus, the difference between the control and delayed-dry runs represents the cumulative effect of LHR between 1800 UTC 25 December and the specified time after 1800 UTC 25 December.

3. Frontogenetical forcing evolution

NCY08 provide a detailed description of the evolution of the 25 December 2002 snowstorm. This section focuses on the frontogenetical forcing evolution of the snowstorm. Cyclone development in the 25 December 2002 snowstorm was similar to a Miller type B cyclogenesis event (Miller 1946), characterized by coastal
redevelopment of a surface cyclone as a 500-hPa trough approaches the eastern United States. A 700-hPa trough developed to the north of the midlevel low, and served as a focus for frontogenesis and subsequent snowband formation.

Figure 2 shows the observed and simulated radar reflectivity as derived in NCY08, 700-hPa height, and frontogenesis evolution during the 0900–1800 UTC 25 December time period. The end of this period precedes band formation by 1.5 h. At 0900 UTC 25 December (Fig. 2a), the 700-hPa low was over Ohio, supporting southwest geostrophic flow over the mid-Atlantic and northeast United States. A region of heavy precipitation was observed in southern Virginia, which moved into northern Virginia and expanded in size by 1200 UTC 25 December (Fig. 2b). A small area of 700-hPa frontogenesis was found in southern Pennsylvania at this time. During the 1200–1800 UTC 25 December period (not shown), the area of heavy precipitation continued to expand and move northward, and was associated with an increase in midlevel frontogenesis over eastern Pennsylvania. By 1800 UTC 25 December, a 700-hPa low had formed over eastern Pennsylvania (Fig. 2c). A sharp trough (trough A) extended from the 700-hPa low center to southern New Hampshire, and served as a focus for frontogenesis and heavy snowfall (Fig. 2c). This trough developed and began to move north during the 1500–1800 UTC 25 December period (not shown). Two other 700-hPa height troughs were evident at 1800 UTC 25 December, one extending southeast from the low center (trough B) and another extending northwest (trough C). Trough C weakened with time as the coastal low developed, and will not be further discussed. The MM5 simulation at 12-km grid spacing exhibited a similar reflectivity, height, and frontogenesis evolution to the observed evolution (Figs. 2d–f).

Band formation occurred near trough A at ~1930 UTC 25 December (NCY08). Figure 3 shows the mature and weakening stages of the band. At 2100 UTC 25 December (Fig. 3a) the intense snowband was evident near the frontogenesis maximum along trough A. The 700-hPa frontogenesis maximum was nearly double of that just 3 h earlier in both the analysis (cf. Figs. 2c and 3a) and model forecasts (cf. Figs. 2f and 3c), associated with a robust cyclonic wind shift along trough A near the band.

By 0000 UTC 26 December, the analyzed and modeled 700-hPa frontogenesis maximum began to weaken as the 700-hPa trough became less defined (Figs. 3b,d). A comparison of the 700-hPa height falls during band formation and dissipation shows that there were large hourly height falls along trough A at 1800 UTC 25 December (Fig. 4a), while height falls had shifted off the New England coast by 0000 UTC 25 December (Fig. 4b). Deformation decreased along trough A during this period as the flow became more symmetric. For example, an average deformation value of $20 \times 10^{-5} \text{ s}^{-1}$ was calculated within a $1.5^\circ \times 2^\circ$ latitude–longitude box along trough A at 1800 UTC 25 December, but was only $12 \times 10^{-5} \text{ s}^{-1}$ along the remnant trough A in western New England at 0000 UTC 26 December (box locations shown in Fig. 4).

Overall, the simulated reflectivity, height, and frontogenesis evolution at 12-km grid spacing was close to the observed evolution throughout the event (NCY08; Figs. 2 and 3). Thus, the genesis and evolution of the midlevel frontogenesis maximum was examined via piecewise PV inversion and model experiments, as described in section 2. Results of the full inversion at 1800 UTC 25 December for the 700-hPa height, wind, and associated frontogenesis are shown in Fig. 5. Since the inversion relies on nonlinear balance and is a smoothing process, an exact correspondence between the simulated and inverted fields is not expected. The full inversion surface cyclone was slightly deeper than the modeled cyclone (~5 hPa), but exhibited a track and evolution that was otherwise nearly identical (not shown). Overall, the inverted 700-hPa low center is ~45 m deeper than the control run; however, both troughs A and B are evident in the inverted height fields, as well as the frontogenesis maximum just north of trough A (Fig. 5). The successful replication of the frontogenesis maximum associated with trough A facilitates piecewise PV analysis of its formation and evolution.

The percentage that each PV piece contributed to the maximum 700-hPa negative height perturbation (i.e., low) during the 1500 UTC 25 December–0000 UTC 26 December time period was analyzed (not shown). The $q'_u$ explained a majority (50%–80%) of the height perturbation, while the $q'_m$ term explained more than 25%. The $q'_d$ and $q'_l$ terms explained less than 10%. Given the dominance of the $q'_u$ and $q'_m$ in explaining the 700-hPa low, further analysis of the contribution of these PV anomalies’ to the forcing evolution is presented at a representative lower (700 hPa) and upper (400 hPa) level.

### a. Band formation

At 0900 UTC 25 December, 10.5 h prior to band formation, the 400-hPa $q'_u$ field exhibited a maximum over western North Carolina (Fig. 6a). Strong positive PV advection was located to the east of this maximum (Fig. 6a), forcing height falls, ascent (e.g., Bluestein 1993; Nielsen-Gammon and Lefevre 1996; Hakim et al. 1996), and associated precipitation over south-central Virginia.
FIG. 2. WSR-88D radar mosaic (reflectivity shaded according to scale starting at 15 dBZ), with the 12-km Eta analysis 700-hPa geopotential height (thick solid, contoured every 30 m), and 700-hPa Petterssen frontogenesis [thin solid, positive values contoured every 2°C (100 km)⁻¹ (h⁻¹) starting at 1°C (100 km)⁻¹ (h⁻¹)] overlaid, valid at (a) 0900, (b) 1200, and (c) 1800 UTC 25 Dec 2002. The 12-km MM5 forecast surface simulated reflectivity, 700-hPa geopotential height (thick solid, contoured every 30 m), and 700-hPa Petterssen frontogenesis [thin solid, positive values contoured every 2°C (100 km)⁻¹ (h⁻¹) starting at 1°C (100 km)⁻¹ (h⁻¹)] valid at (d) 0900, (e) 1200, and (f) 1800 UTC 25 Dec 2002.
The developing precipitation was associated with latent heating\(^1\) in the 400–800-hPa layer (Fig. 7a). Meanwhile, the 700-hPa \(q_m\) field exhibited an elongated maximum in northwestern Ohio associated with the primary cyclone (Fig. 6b), and a small arc in eastern Ohio associated with the parent cyclone’s occlusion.

Three hours later at 1200 UTC 25 December, a small upper PV filament (U1) developed in the 400-hPa \(q_u\) field over eastern Virginia (Fig. 6c). Filament U1 appears to have formed as a consequence of the diabatic redistribution of PV. For example, at 700 hPa, the \(q_m\) field exhibited a newly formed local diabatic PV maximum (L1) in the vicinity of U1 (Fig. 6d). Feature L1 was associated with the developing precipitation shield (Fig. 2e) in the region of enhanced positive upper PV advection and ascent associated with U1 (Figs. 6c and 7b); L1 exhibited perturbation PV values exceeding 2 PVU (1 PVU = \(10^{-6} \text{ K m}^2 \text{ kg}^{-1} \text{ s}^{-1}\)), and large latent heating in its vicinity (Fig. 7b). The associated LHR reduced PV downshear of L1 (\(x = 300–400 \text{ km}\)) in the 300–400-hPa layer, effectively increasing the slope of the tropopause and associated PV advection (Fig. 7b). The inverted 700-hPa height, wind, and frontogenesis

\(^1\) Latent heating calculated in Fig. 7 as heating due to condensation and deposition in regions of explicitly resolved saturated ascent using \(\partial \theta / \partial t = (R_m \partial \theta / \partial p) \omega (\Gamma_m - \Gamma_d)\), where \(R_m\) is the gas constant for moist air, \(g\) is gravity, \(p\) is the pressure, \(\omega\) is the vertical velocity in pressure coordinates, \(\Gamma_m\) is the moist adiabatic lapse rate, and \(\Gamma_d\) is the dry adiabatic lapse rate.
fields associated with the $q_u^q$ and $q_m^q$ indicate that 700-hPa frontogenesis was induced by the $q_m^q$ over northern Virginia (Figs. 8a,b).

During the next 6 h (1200–1800 UTC 25 December), U1 amplified and moved north in tandem with L1 as trough A formed. By 1800 UTC 25 December, just 1.5 h prior to band formation, U1 was prominent, with a “notch” of low 400-hPa $q'$ values over southern Pennsylvania (Fig. 6e), indicating the development of a trough (e.g., Martin 1998a). Positive PV advection was located on the poleward edge of U1 (Fig. 6e), along a similar axis as the newly formed trough A (e.g., Fig. 2f). Feature L1 moved north into northern New Jersey and became deformed along a northeast–southwest-oriented axis by 1800 UTC 25 December (Fig. 6f), corresponding to trough A. At 1800 UTC 25 December, L1 coincided with latent heating and was positioned beneath the region of enhanced PV advection associated with U1 (Fig. 7c). The PV inversion showed that negative 700-hPa height anomalies associated with the $q_u^q$ field had increased along an axis nearly collocated with trough A (Fig. 8c). The wind field associated with the

FIG. 4. MM5 forecast 700-hPa geopotential height (thick solid, contoured every 15 m), wind (1 full barb = 10 kt), and height falls [shading according to scale (m h$^{-1}$)] valid at (a) 1800 and (b) 0000 UTC. Deformation values cited in the text were calculated within the box shown in (a) and (b).

FIG. 5. Comparison of the (a) control and (b) inverted 700-hPa geopotential height (contoured every 15 m), wind (1 full barb = 10 kt), and resulting frontogenesis [shaded according to scale, in units of °C (100 km)$^{-1}$ (3 h)$^{-1}$], valid at 1800 UTC 25 Dec 2002. The control run 700-hPa potential temperature (dashed) is contoured every 2 K in (a) and (b).
Fig. 6. (a) 0900 UTC 25 Dec 2002 400-hPa PV perturbation [shaded according to scale every 0.5 PVU (1 PVU = 10^{-6} K m^2 kg^{-1} s^{-1})], winds (1 barb = 10 kt), and PV advection (thin, contoured where positive every 0.5 × 10^{-3} PVU s^{-1} starting at 0.5 × 10^{-3} PVU s^{-1}). (b) As in (a), but at 700 hPa and with the 70% isohume overlaid (thick solid). (c),(d) As in (a),(b), but at 1200 UTC 25 Dec 2002. (e),(f) As in (a),(b), but at 1800 UTC 25 Dec 2002.
$q_m'$ supported a band of frontogenetical forcing along trough A; however, a majority of the 700-hPa frontogenesis along troughs A and B was associated with the $q_m'$ (Figs. 8c,d).

The role of LHR in the formation of the midlevel frontogenesis maximum is highlighted by a comparison of the control and dry runs at 1800 UTC 25 December (Figs. 9a,b). Troughs A and B were considerably less sharp in the dry run (Figs. 9a,b). Consequently, frontogenetical forcing along trough A was nearly 4 times weaker in the dry run than the control run at this time, and band formation failed to occur. Comparison of the upper PV distribution shows that U1 was absent in the dry run (Figs. 9c,d), which is evidence of diabatic PV redistribution in the control run. However, the upper PV anomaly in the dry run does exhibit a broad elongation toward the southern New England coast (Fig. 9d), contributing to a weak trough A and associated weak frontogenesis (Fig. 9b). The dry run does not exhibit L1 (Fig. 9f), further confirming the diabatic nature of L1.

In summary, strong forcing for ascent ahead of the upper PV anomaly created a small area of precipitation...
over southern Virginia. LHR associated with this precipitation created an elongation in the upper PV distribution (U1) and a small midlevel diabatic PV anomaly (L1). Once U1 and L1 formed, they simultaneously modified the upper and interior PV distributions, leading to mutual amplification and upscale growth. The anomalies exhibited a northeast–southwest orientation that supported larger height falls along this axis and the formation of trough A. Trough A subsequently served as a focus for frontogenesis and band formation.

b. Band maturity

At 2100 UTC 25 December, an intense snowband was present in eastern New York (Figs. 3a,c). Positive PV advection (favoring ascent and height falls) occurred on the poleward edge of U1 over eastern New York at 400 hPa (Fig. 10a). Inversion of the $q_u$ showed negative 700-hPa height anomalies in eastern New York along a similar orientation as trough A (Figs. 10c and 3c). The induced frontogenesis exceeded 2°C (100 km)$^{-1}$ (3 h)$^{-1}$ in eastern New York (Fig. 10c), which was ~40% of the frontogenesis induced by the total balanced flow.

The $q_m$ at 700 hPa shows that L1 oriented along a northeast–southwest axis in eastern New York (Fig. 10b) parallel to trough A and the snowband (Fig. 3c). The L1 remained beneath the region of upper PV advection associated with U1 (Fig. 7d). Frontogenesis induced by $q_m$ increased to values exceeding 3°C (100 km)$^{-1}$ (3 h)$^{-1}$.
FIG. 9. Comparison of the (left) control and (right) dry runs valid at 1800 UTC 25 Dec 2002. (a),(b) Geopotential height (solid, contoured every 15 m), frontogenesis [contoured where positive every 2°C (100 km)^{−1} h^{−1}, starting at 1°C (100 km)^{−1} h^{−1}], and simulated reflectivity (shaded according to scale). (c),(d) 400-hPa PV (shaded according to scale) and winds (1 barb = 10 kt). (e),(f) As in (c),(d), but for 700 hPa.
along trough A in southeast New York (Fig. 10d), which is \( \sim 60\% \) of the frontogenesis induced by the total balanced flow. To isolate the impact the PV anomaly generated by the precipitation band (L1) had on trough A and the associated frontogenesis, the \( q_m' \) in the 800–500-hPa layer in a 3.5° × 4.5° latitude–longitude box centered on the band was inverted (box shown in Fig. 10b). The resultant inverted 700-hPa height and wind field show negative height perturbations exceeding 10 m along the band, with an induced cyclonic circulation (Fig. 11). The resultant frontogenesis accounts for \( \sim 60\% \) of the total \( q_m' \)-induced frontogenesis in southeast New York and \( \sim 35\% \) of the frontogenesis induced by the total balanced flow. These results demonstrate that diabatic heating from the band itself contributed to amplification of the frontogenetical forcing during band maturity.

Figure 12 shows a comparison of the control and delayed dry run at 2100 UTC 25 December. This time is 3 h after latent heating was turned off in the delayed dry run, and 1.5 h after band formation occurred in the control run. A much weaker band occurred in the delayed dry run than in the control run, illustrating the critical role LHR plays in band formation and maintenance, even given an initially favorable flow field just 1.5 h prior to band formation. In particular, the heights in eastern New York in the delayed dry run at 2100 UTC 25 December are \( \sim 15\) m higher than in the control run, consistent with the PV inversion results (i.e., Fig. 11). As a result, the midlevel trough is less defined, and the

**Figure 10.** (a),(b) As in Fig. 6, but valid at 2100 UTC 25 Dec 2002. The region used for the inversion shown in Fig. 11 is outlined in (b). (c),(d) As in Fig. 8, but valid at 2100 UTC 25 Dec 2002.
frontogenesis maximum is a factor of \( \sim 3 \) weaker in the delayed dry run than in the control run. Explicit differences in the wind and temperature (control minus delayed dry) show strong flow convergence in eastern New York, and a local temperature difference exceeding 4°C over southern New England (Fig. 12c). Calculation of normalized frontogenesis [Schultz 2004, his Eq. (3)] reveals that the change in kinematic flow between the control and dry runs accounts for \( \sim 60\% \) of the change in the frontogenesis maximum between these runs, while the change in temperature gradient accounted for the remaining \( \sim 40\% \) of the change in the frontogenesis maximum. Thus, LHR during the 1800–2100 UTC 25 December period altered the flow (creating large deformation and convergence) and enhanced the temperature gradient, resulting in intense frontogenesis in the band region. These results further highlight the critical role of moist processes in the formation and maintenance of the band.

c. Band dissipation

Given that LHR associated with the precipitation band enhanced the frontogentical forcing, what led to band dissipation? As noted in section 3, the largest height falls shifted east of trough A during band dissipation (Fig. 4), creating a more symmetric midlevel height field and reducing the deformation, convergence, and associated frontogenesis. To explore what caused

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**FIG. 11.** The 700-hPa balanced geopotential height (contoured every 5 m), and wind (1 full barb = 10 kt) inverted from \( q_m \) within the box in Fig. 10b at 2100 UTC 25 Dec 2002. The control run 700-hPa temperature (dashed) is contoured every 2 K. The resulting frontogenesis is shaded according to scale, starting at 0.25°C (100 km) \(^{-1}\) (3 h) \(^{-1}\).

**FIG. 12.** Comparison of the (a) control and (b) delayed dry run geopotential height (solid, contoured every 15 m), frontogenesis [contoured where positive every 2°C (100 km) \(^{-1}\) (3 h) \(^{-1}\)], starting at 1°C (100 km) \(^{-1}\) h \(^{-1}\)] and simulated reflectivity (shaded according to scale), valid at 2100 UTC 25 Dec 2002. (c) Temperature (shaded) and wind (barbs) difference (control minus delayed dry run), valid at 2100 UTC 25 Dec 2002.
the eastward displacement of the 700-hPa height falls, the piecewise PV inversion results during the 2100 UTC 25–0000 UTC 26 December period were analyzed.

At 2100 UTC 25 December the $q_u$ field exhibited a small bulge (U2) along the perturbation PV tail off the Atlantic coast (Fig. 10a). The $q_m$ field showed a small anomaly (L2) near U2 (Fig. 10b), suggesting PV was redistributed by LHR. A temperature difference maximum between the control and delayed dry run off the Virginia coast is found near L2 (Fig. 12c), providing further evidence of LHR. Frontogenesis induced by L2 is also evident in this region (Fig. 10d).

Three hours later at 0000 UTC 26 December, U2 had developed into a well-defined filament off the southern New England coast, with an attendant low PV notch (Fig. 13a). Positive PV advection was found on the poleward edge of U2 (Fig. 13a), supporting large height falls in this region. The $q_u$ field was responsible for 700-hPa frontogenesis along the remnant trough A in western New England and near U2 off the coast (Fig. 13c). At lower levels, L2 had grown in amplitude and size over the last 3 h (cf. Figs. 10b and 13b). Accordingly, the inverted 0000 UTC 26 December $q_m$ field exhibited strong frontogenesis near L2 southeast of the New England coast (Fig. 13d). However, frontogenesis associated with the $q_m$ field in western New England was nearly absent (Fig. 13d).

To quantify the effect of L2 on the frontogenesis in western New England at 0000 UTC 26 December, the $q_m$ in a $3.5^\circ \times 4.5^\circ$ latitude–longitude box surrounding L2 was inverted (box shown in Fig. 13b). The L2 was responsible for inducing a frontolytic flow in western
New England (Fig. 14). Separate inversions of $q_{in}$ in the Gulf of Maine and $q_{d}$ in the dry slot (not shown) resulted in negligible frontolysis. Thus, band dissipation was primarily caused by the formation and upscale growth of L2 east of the band. This anomaly created larger height falls off the southeast New England coast than along the former trough A. The result was a more symmetric height and wind field over western New England and associated weaker frontogenetical forcing.

4. Stability evolution

As shown in NCY08, the 25 December 2002 snowstorm exhibited a deep layer of conditional neutrality, with an embedded layer of conditional instability near 500 hPa in the immediate band environment prior to band formation. For example, at 1800 UTC 25 December (~90 min prior to band formation), a region of heavy precipitation and midlevel convergent flow occurred over eastern New York and Pennsylvania (Fig. 15a; NCY08). A cross section through the heavy precipitation region (Fig. 15b) shows a region of near-neutral conditional stability in the 600–350-hPa layer. A spatial view of the 470–600-hPa $\theta_{es}$ difference at 1800 UTC 25 December shows negative values [indicative of conditional instability (CI)] over southeast New York (Fig. 15c). This region of CI was associated with the developing dry slot (not shown). However, an isolated region of negative values was also evident near the New York–Pennsylvania–New Jersey border, near where the snowband would form 90 min later. The conditional stability was restored during band formation and remained stable throughout the subsequent band duration (NCY08).

To explore the stability evolution, a $\theta$ tendency budget was calculated. A $\theta$ budget was calculated as opposed to a saturation equivalent potential temperature ($\theta_{es}$) budget to more readily separate the impact of diabatic heating from temperature advections. The Pearson correlation coefficient (Wilks 1995, 45–50) between $\theta$ and $\theta_{es}$ was 0.99, further justifying use of the $\theta$ tendency as a proxy for the $\theta_{es}$ tendency in this case. The $\theta$ tendency equation [Carlson 1991, his Eq. (1.13c)],

$$\frac{\partial \theta}{\partial t} = -\left( \frac{\partial \theta}{\partial x} + v \frac{\partial \theta}{\partial y} \right) - \omega \frac{\partial \theta}{\partial p} + \frac{Q}{c_p} \frac{\theta}{T}$$  \hspace{1cm} (2)

was evaluated using 5-min output from the model on the 4-km domain with explicit convection, where $u$ and $v$ are components of the vector wind, $p$ is the pressure, $\omega$ is the vertical velocity, and $Q$ is the diabatic heating rate. Term A represents horizontal advection, term B represents vertical advection, and term C represents diabatic heating. Term C is composed of heating/cooling from diabatic processes included in the microphysics, radiation, and boundary layer schemes. An additional diffusional correction term is applied in the MM5, and is considered part of term C for simplicity. On average $\approx$90% of the total magnitude of term C was from latent heating due to cloud processes (derived from the microphysics scheme), and therefore all other heating processes were combined into an “other” category. Thus, the terms assessed in the $\theta$ budget were the local $\theta$ tendency, horizontal $\theta$ advection (term A), vertical $\theta$ advection (term B), latent heating due to cloud processes, and heating due to other processes. Differences between terms at two levels provided a stability tendency budget, which could be assessed through time.

The budget was analyzed for a $4 \times 4$ gridpoint box (256 km$^2$) within the persistent negative EPV region just southeast of the band formation location in the 470–600-hPa layer, where CI was present (e.g., Fig. 15a). Sensitivity tests confirm that shifting the analysis box location two grid points to the northeast or southwest (remaining along the axis of negative EPV) does not change the results. A time series of the 470–600-hPa $\theta$ difference and difference tendency between 1200 UTC 25 December and 2000 UTC 25 December is shown in Fig. 16. Three distinct stability tendency regimes are evident. Destabilization occurred between 1200 and 1400 UTC 25 December as the 470–600-hPa $\theta$ difference decreased
from 7 to 4.5 K. Conditional instability developed in the 500–575-hPa layer during this period (not shown). Between 1400 and 1900 UTC 25 December the weak stability was maintained, with the 470–600-hPa $\theta$ difference remaining around 5.5 K. The stability sharply increased during the 1900–2000 UTC 25 December period, as the band formed at 1930 UTC 25 December.

Analysis of the individual terms in the 470–600-hPa layer during the 1200–1400 UTC 25 December period of destabilization (Table 1) shows that differential horizontal advection contributed to destabilization, while differential vertical advection partially offset the destabilization. Analysis of the individual levels revealed that stronger warm-air advection occurred at 600 than at 470 hPa (not shown), while there was stronger ascent and associated vertical advection of lower $\theta$ at 470 than at 600 hPa. During the period of small stability maintenance (1400–1900 UTC 25 December), terms fluctuated in sign, with differential vertical $\theta$ advection typically opposing differential latent heating (Table 1). During the period of rapid stabilization (1900–2000 UTC 25 December), differential $\theta$ advection (small warm-air advection above, cold-air advection below) was the primary term contributing to stabilization (Table 1), partially offset by differential latent heating and vertical $\theta$ advection. Although on average differential vertical advection was a slight destabilizing term during this period, it was a strong stabilizing term between 1930 and 2000 UTC 25 December, coinciding with the period of band formation. This result is consistent with the release of conditional instability as diagnosed in NCY08.

To view stability processes in a Lagrangian framework, 6-h backward trajectories at the top and bottom of the conditionally unstable layer at the analysis location were also calculated. The trajectories were obtained using a 5-min time step interpolated from 15-min output on the 12-km domain. Trajectories illustrate that the upper and lower parcels arriving at the analysis location during the 1200–1400 UTC 25 December period originated southwest of the analysis location (Figs. 17a,b). Cross sections exhibited a shallow layer of conditional instability between 500 and 600 hPa and a deeper layer of potential instability between 700 and 500 hPa where the parcels originated (not shown). Trajectories ending at 1400 UTC 25 December show that the upper parcel became nearly saturated as the layer experienced
sustained ascent (Fig. 17b). The upper parcel $\theta$ was conserved during its advance (not shown), suggesting dry adiabatic ascent, while the $\theta$ of the lower saturated parcel increased 2.6 K (not shown) in association with LHR. Thus, the advection and lifting of a potentially unstable layer account for conditional stability reduction during the 1200–1400 UTC 25 December period.

Differential moisture advection within the cyclone’s dry slot has been cited as an important destabilization process associated with banded events. Parcels entering the stability analysis box during the 1500–1800 UTC 25 December period arrived generally from the south, and had remained saturated along their entire 6-h trajectories (Fig. 17c). However, the upper parcel entering the analysis box at 2000 UTC 25 December originated within the developing dry slot ~6 h prior (1400 UTC 25 December), while the lower parcel remained nearly saturated during its entire advance to the analysis box (Fig. 17d). Thus, parcels that had advanced through the dry slot arrived in the banding region after the band had formed, and therefore the dry slot did not play a role in the initial stability reduction.

Trajectory analysis of parcels arriving at the stability analysis box at 2000 UTC 25 December shows that the dry slot contributed to the stability evolution after band formation. For example, 6 h prior to the parcels’ arrival at the stability analysis box location (1400 UTC 25 December), the $\theta$ difference between the upper and lower parcel was 11.5 K (suggesting strong stability). Four hours later this difference decreased to 5.6 K (1800 UTC 25 December) as lifting of the potentially unstable layer occurred in the dry slot. If this layer was simply advected to the analysis location, it would have maintained low stability at the analysis location. However, both parcels became saturated as they ascended in the local band environment (Fig. 17d), and the upper parcel experienced slightly greater ascent (~150 m more), leading to a slight increase in the $\theta$ difference to 6.5 K. The parcels arrived at the analysis location with this $\theta$ difference, which was more stable than parcels arriving just 1 h earlier (5.5 K; Fig. 16). Thus, ascent within the dry slot did destabilize the air that arrived in the banding region at 2000 UTC 25 December; however, the stabilizing effects of differential diabatic heating and differential vertical ascent in the immediate band environment offset the destabilization in this case. Presumably, the stability would have been even larger during band maturity in the absence of a dry slot in this case.

5. Other cases

To explore the representativeness of the above results, the methods employed in investigation of the 25 December 2002 snowstorm were applied to the 12 February 2006 and 14 February 2007 snowstorms.

| Table 1. The 25 Dec 2002 potential temperature budget terms ($\times 10^{-4}$ K s$^{-1}$) averaged over the specified time period, including differential horizontal advection (hor adv), vertical advection (vert adv), latent heating (latent), other (see text), and their sum. Positive (negative) values represent a stabilizing (destabilizing) tendency in the layer. |
|-----------------|----------------|--------------|-----------------|-----------------|
| Hor Adv | Vert Adv | Latent | Other | Sum |
| 1200–1400 UTC | -3.1 | 1.7 | 0.3 | -1.5 | -2.6 |
| 1400–1900 UTC | -0.6 | 3.1 | -1.2 | -0.7 | 0.6 |
| 1900–2000 UTC | 2.5 | -0.9 | -0.6 | 0.3 | 1.3 |

FIG. 16. Time series of the potential temperature tendency (black) and potential temperature difference (470 hPa minus 600 hPa; gray) for the stability budget analysis location shown in Fig. 15a during the 1200–2000 UTC 25 Dec 2002 time period. Key periods of destabilization, stability maintenance, stabilization, and band formation (BAND) are labeled.
An intense snowband associated with the 12 February 2006 snowstorm brought a record breaking 68.3 cm of snow to New York City. The 14 February 2007 snowstorm deposited nearly 1 m of snowfall in a narrow band in upstate New York.

As in the 25 December 2002 snowstorm, the analyzed mesoscale evolution of the 12 February 2006 and 14 February 2007 cases featured the formation of a midlevel (~700 hPa) low. This low was evident in the wind field over southwest New Jersey in the 12 February 2006 case (Fig. 18a), and northeast Pennsylvania in the 14 February 2007 case (Fig. 18b). A sharp cyclonic wind shift extended northeast from this midlevel low center in each storm (Figs. 18a,b), associated with a midlevel height trough (not shown). Convergence and deformation were found in the vicinity of this wind shift as strong southeast flow slowed, and backed across the trough (Figs. 18a,b).

Corresponding 4-km MM5 simulations [model configurations are presented in Novak et al. (2007)] show the simulated precipitation bands were within 50 km of the observed location in each case at the time of band maturity (Fig. 18); however, the 12 February 2006 simulation was time shifted 3 h earlier to obtain the best match. Similar to the observations, a sharp wind shift was evident in the model wind forecasts (Figs. 18c,d), associated with intense frontogenesis (not shown). These model simulations are used for further investigation.

a. Forcing evolution

The 400-hPa $q_r$ field 3 h prior to band formation in the 1200 UTC 12 February 2006 case featured a synoptic-scale...
elongation along the eastern U.S. coast (Fig. 19a). The PV values within the upper anomaly were ~2 PVU stronger than in the 25 December 2002 case (cf. Figs. 6c and 19a). Positive PV advection was occurring on the poleward flank of this anomaly, supporting height falls and ascent along the coast. Inversion confirms that the upper PV elongation contributes to 700-hPa trough development along the coast (not shown). The $q_m^{9}$ field (Fig. 19b) shows a PV band along this trough, representing the effects of LHR associated with heavy precipitation developing within the region of strong upper PV advection near the coast. Unlike the 25 December 2002 case, the $q_m^{9}$ anomaly formed and later amplified in place. Similar to the 25 December 2002 case, the induced flow from the $q_m^{9}$ explained more than half (65%) of the total balanced frontogenesis maximum at the time of band maturity at 1800 UTC 12 February (not shown).

The 400-hPa $q_u^{9}$ field 3 h prior to band formation in the 1500 UTC 14 February 2007 case was quite different
from either the 25 December or 12 February cases, with a weak 400-hPa PV anomaly (≤ 2.0 PVU) over Ohio and a weak trough (associated with PV < 1.0 PVU) located along the coast (Fig. 19c). Weak PV advection was occurring on the eastern edge of these anomalies. Inversion confirmed that the $q^*$ field accounted for less than ~10% of the total balanced 700-hPa frontogenesis maximum (not shown). In contrast, the $q^m$ field at the same time exhibited strong elongated anomalies, one associated with the inland system, and another along the coast (Fig. 19d). The coastal anomaly had originated along the southeast U.S. coast around 0000 UTC 14 February. A confluent zone was formed in eastern New York between the two $q^m$ anomalies (see Fig. 19d), as the coastal anomaly induced southeast winds, while the inland anomaly induced south-southwest winds. Inversion confirmed that the $q^m$ field accounted for ~85% of the frontogenesis maximum at the time of band formation (not shown).

Similar to the 25 December 2002 case, both the 12 February 2006 and 14 February 2007 cases exhibited
band dissipation as heights fell more rapidly to the east of the midlevel trough than along the trough. In the 2100 UTC 12 February 2006 case (Fig. 20a), positive 400-hPa PV advection was evident along the poleward edge of the upper PV elongation, which would tend to maintain the trough; however, at lower levels a separate diabatic PV anomaly just off the New England coast developed over the past 2 h (Fig. 20b). This anomaly was associated with heavy precipitation and formed at the tip of the upper PV elongation (Figs. 20a,b). Another PV anomaly exceeding 1 PVU was found at 700 hPa beneath the upper PV elongation in the cyclone’s dry slot (Fig. 20b). Piecewise inversion of these anomalies confirmed that the diabatic PV anomaly in eastern New England was dominant in creating frontolysis (not shown). Frontolysis values induced by the diabatic PV anomaly at 2100 UTC 12 February were ~25% of the total balanced frontogenesis. Thus, as with the 25 December 2002 case, a diabatic PV anomaly helped shift height falls east of the primary band, and weakened frontogenesis in the band region.

During the 14 February 2007 case the coastal diabatic PV anomaly amplified and became dominant over

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**Fig. 20.** (a) 2100 UTC 12 Feb 2006 400-hPa PV perturbation (shaded according to scale every 0.5 PVU where 1 PVU = \(10^{-6} \text{ K m}^2 \text{ kg}^{-1} \text{ s}^{-1}\)), winds (1 barb = 10 kt), and PV advection (thin, contoured where positive every \(2 \times 10^{-5} \text{ PVU s}^{-1}\) starting at \(2 \times 10^{-5} \text{ PVU s}^{-1}\)). (b) As in (a), but for 700 hPa. The 70% isohume is overlaid. (c) As in (a), but for 0600 UTC 15 Feb 2007. (d) As in (b), but for 0600 UTC 15 Feb 2007. Band centroid position shown as an asterisk in (b) and (d).
northern New England. As the coastal anomaly amplified, the upper PV anomaly over Ohio (Fig. 19a) moved to the southern New England coast by 0600 UTC 15 February (Fig. 20c). The PV advection associated with this feature over the Gulf of Maine (Fig. 20c) contributed to height falls east of the primary trough in central New England (not shown). The 700-hPa $q'''$ field at 0600 UTC 15 February exhibited a prominent north–south maximum associated with the primary trough; however, a new PV anomaly was evident along the Maine coast (Fig. 20d). This anomaly formed near a band of heavy precipitation along the cyclone’s surface occlusion (not shown). Piecewise inversion of the $q'''$ within the box shown in Fig. 20 indicated frontolysis in the band region, with values $\sim 30\%$ of the total balanced frontogenesis (not shown). Inversion of the PV in the dry slot (i.e., Fig. 20d) indicated small height falls and little impact on the frontogenesis in the band region (not shown). Thus, the diabatic PV anomaly along the Maine coast helped shift height falls east of the primary band and weakened frontogenesis in the band region.

b. Stability evolution

As in the 25 December 2002 case, the stability evolution of the 12 February 2006 and 14 February 2007 cases were evaluated on the 4-km domain at a representative location (see Figs. 18c,d) within the layer of persistent small conditional stability. Based on cross sections, this layer was between 520–630 and 410–550 hPa in the 12 February 2006 and 14 February 2007 cases, respectively. The stability tendency for each case is shown in Fig. 21. As in the 25 December 2002 case, the conditional stability was reduced prior to band formation, while the stability was partially restored near the time of band formation. During the destabilizing period in all three cases, differential horizontal $u$ advection was the dominant destabilizing term (Tables 1–3). During the stability maintenance period just prior to band formation, differential horizontal $u$ advection was generally small, and the effect of differential vertical $u$ advection was offset by differential latent heating. During the stabilization period (generally during or after band formation), differential horizontal $\theta$ advection was dominant in the 25 December 2002 case (Table 1) whereas differential vertical $\theta$ advection was dominant in the 12 February 2006 and 14 February 2007 cases (Tables 2 and 3, respectively). These results highlight the 3D nature of the stability evolution as well as the role latent heating can play in the stability evolution.

Trajectory analysis of the 12 February 2006 case (not shown) revealed that parcels arriving during the primary destabilization period (0000–0600 UTC 12 February)
originated southwest of the analysis box and were saturated during their advance. After 0600 UTC 12 February, parcels which had experienced stability reduction within the dry slot arrived at the analysis location and contributed to the weak conditional stability during the stability maintenance period (Fig. 21). In the 14 February 2007 case, parcels remained saturated upon their advance to the analysis box at all times. Thus, the dry slot was not related to the initial stability reduction in all three examined cases. Rather, differential horizontal temperature advection in moist southwest flow ahead of the upper trough was the dominant process reducing the midlevel conditional stability.

6. Discussion and conclusions

The processes regulating the frontogenesis and stability evolution in the 25 December 2002 northeast U.S. snowstorm and two other cases were investigated using piecewise potential vorticity (PV) inversions, modeling experiments, and $\theta$ tendency budgets.

Band formation in all three cases was coincident with the sharpening of a midlevel trough to the north of a parent low, and an associated increase in frontolysis in an environment of small conditional stability. Piecewise PV analysis showed that there are a variety of PV evolutions that contribute to band formation and evolution. For example, during the 25 December 2002 case, the diabatic redistribution of PV simultaneously created upper (U1) and lower (L1) PV anomalies. The PV advection associated with these anomalies supported the formation of trough A and associated frontogenesis at 700 hPa. This process began ~8 h prior to band formation and ~500 km south of the band formation location. The dry run showed that LHR was critical for amplifying the trough through the simultaneous modification of the upper and lower PV distribution (Fig. 9). In the 12 February 2006 case, trough formation occurred in situ as the upper PV anomaly forced ascent along the coast and the associated diabatic PV anomaly amplified in place. In the 14 February 2007 case, the midlevel frontogenesis developed in the confluent flow between separate inland and coastal diabatic PV anomalies. Common to all three cases are diabatic PV anomalies whose induced flow accounts for a majority of the midlevel frontogenesis, highlighting the importance of LHR in band evolution.

Band dissipation in all three cases occurred as new diabatic PV anomalies (associated with LHR) formed east of the primary band, creating a more symmetric midlevel flow that weakened the midlevel frontogenesis associated with the band. For example, piecewise PV analysis of the 25 December 2002 case showed that the formation of a new diabatic PV anomaly (i.e., L2) shifted height falls east of trough A, and contributed to frontolysis in the band region (Fig. 14). In the 12 February 2006 and 14 February 2007 cases, new diabatic PV anomalies were also responsible for the eastward shift in height falls and frontolysis. Thus, band duration in each case was partially determined by when these new diabatic PV anomalies formed and how they evolved, illustrating a specific example of how moist processes can influence the lifetime of a band.

Conditional stability reduction occurred several hours prior to band formation in each case (e.g., Figs. 16 and 21). After the initial stability reduction, the stratification remained small until the release of CI during the band formation process. This stability evolution has also recently been shown by Evans and Jurewicz (2009) in a climatological study of heavy snowfall events in central New York, and suggests that the development of CI prior to band formation and subsequent release of CI during the band formation process may be relatively common.

Diagnosis of the processes responsible for destabilization in each case revealed that differential horizontal $\theta$ advection in moist southwest flow ahead of the upper trough was dominant over differential vertical $\theta$ advection or latent heating in all three cases. Trajectories showed that the advection and lifting of a potentially unstable layer (not associated with the dry slot) accounted for the initial conditional stability reduction in the 25 December 2002 case, while differential horizontal $\theta$ advection (in a saturated environment) accounted for the initial stability reduction in the 12 February 2006 and 14 February 2007 cases. The dry slot was not responsible for the initial stability reduction in all three examined cases, counter to past understanding based on coarser temporal and spatial resolution data (e.g., Nicosia and Grumm 1999; Moore et al. 2005).
Piecwise PV inversion and the delayed dry experiment in the 25 December 2002 case showed that LHR was critical to the formation and maintenance of the band. Band formation failed to occur in the absence of LHR in the dry simulation, and was substantially weaker even given a favorable treble-cleft upper PV distribution and lower PV anomaly present just 90 min prior in the delayed dry simulation. These results illustrate the role the band itself plays in its own evolution, via the positive feedback between LHR and frontogenesis (Emanuel 1985; Thorpe and Emanuel 1985; Davies 1999), and frontogenesis and stability (Nicosia and Grumm 1999). In particular, the presence of a band (and associated LHR) locally decreases heights leading to convergence, and locally increases the temperature gradient (e.g., Figs. 11 and 12). The induced convergent flow in the presence of the temperature gradient contributes to frontogenesis. Furthermore, differential latent heating associated with a precipitation band can contribute to reduced stability above the level of maximum heating (Tables 1–3). However, the resulting strong ascent generally leads to a differential vertical advection pattern that stabilizes the environment, serving as a local “brake” on the feedback.

Although the band itself locally modifies the environment via diabatic feedbacks, the band was susceptible to changes in the meso-α-scale flow associated with the formation and subsequent influence of new diabatic PV anomalies. Given that both band formation and dissipation in each case were largely driven by the upscale growth of diabatic PV anomalies, the observed details of which are uncertain (e.g., Zhang et al. 2003, 2007), the predictability of banding events may be greatly limited. Indeed, Evans and Jurewicz (2009) have shown that the correlation between forecasts of band ingredients (frontogenesis, small EPV, and saturation) and observed event total snowfall decrease substantially beyond 12 h in an operational model. They attribute this result to timing and location errors. On the other hand, successful detailed operational short-range predictions of intense bands have been made (e.g., Novak et al. 2006). In the present study the dry run still exhibited frontogenesis (although limited) in the banding region, suggesting a degree of determinism in the potential for band formation, at least for forecast projections <24 h. Thus, it is possible that band occurrence is more predictable than band timing and location. Future investigation of the predictability of band events, including aspects of their occurrence, location, and timing, is planned.

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