Evolution of Mesoscale Precipitation Band Environments within the Comma Head of Northeast U.S. Cyclones

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ABSTRACT

This paper explores the mesoscale forcing and stability evolution of intense precipitation bands in the comma head sector of extratropical cyclones using the 32-km North American Regional Reanalysis, hourly 20-km Rapid Update Cycle analyses, and 2-km composite radar reflectivity data. A statistical and composite analysis of 36 banded events occurring during the 2002–08 cool seasons reveals a common cyclone evolution and associated band life cycle. A majority (61%) of banded events develop along the northern portion of a hook-shaped upper-level potential vorticity (PV) anomaly. During the 6 h leading up to band formation, lower-tropospheric frontogenesis nearly doubles and the conditional stability above the frontal zone is reduced. The frontogenesis increase is primarily due to changes in the kinematic flow associated with the development of a mesoscale geopotential height trough. This trough extends poleward of the 700-hPa low, and is the vertical extension of the surface warm front (and surface warm occlusion when present). The conditional stability near 500 hPa is reduced by differential horizontal potential temperature advection. During band formation, layers of conditional instability above the frontal zone are present nearly 3 times as often as layers of conditional symmetric instability. The frontogenetical forcing peaks during band maturity and is offset by an increase in conditional stability. Band dissipation occurs as the conditional stability continues to increase, and the frontogenesis weakens in response to changes in the kinematic flow.

A set of 22 null events, in which band formation was absent in the comma head, were also examined. Although exhibiting similar synoptic patterns as the banded events, the null events were characterized by weaker frontogenesis. However, statistically significant differences between the midlevel frontogenesis maximum of the banded and null events only appear 2 h prior to band formation, illustrating the challenge of predicting band formation.

1. Introduction

Intense precipitation bands are frequently observed in the comma head of extratropical cyclones over the northeast United States (Novak et al. 2004, hereafter N04). Theoretical work by Emanuel (1985), Thorpe and Emanuel (1985), Xu (1989a,b), and Xu (1992) has shown that intense single cores of ascent can form through a coupled relationship between frontogenesis and moist symmetric stability, whereby the ascending branch of a frontal circulation is narrowed and enhanced when there is weak moist symmetric stability above and on the warm side of a frontal boundary. Consistent with these theoretical results, previous case studies and climatologies over the northeast United States have established that intense precipitation bands are favored in environments of strong frontogenesis, weak moist symmetric stability, and sufficient moisture (e.g., Sanders and Bosart 1985; Nicosia and Grumm 1999; Novak et al. 2006).

Cyclones embedded within diffluent large-scale flow tend to exhibit lower-tropospheric frontogenesis in the

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northwest quadrant of the cyclone (Schultz et al. 1998; Schultz and Zhang 2007). In a composite study NO4 found that band formation was associated with cyclogenesis, as the development of a closed midlevel circulation supported frontogenesis to the northwest of the surface cyclone. However, NO4 was limited to 80-km Eta model analyses at 6-h temporal resolution, and thus could not address the mesoscale aspects of the frontogenesis and stability evolution during the band life cycle.

More recently, Novak et al. (2008, hereafter N08) used dual-Doppler, wind profiler, aircraft, and water vapor observations in concert with model simulations down to 1.33-km horizontal grid spacing to study an intense snowband occurring on 25 December 2002. Band formation was coincident with the sharpening of a midlevel (~700 hPa) trough and an associated increase in frontogenesis. Conditional instability developed above the frontogenesis maximum prior to band formation. Band maturity was marked by increasing conditional stability and frontogenetical forcing, whereas band dissipation occurred as the frontogenetical forcing weakened and the conditional stability continued to increase.

Building on the results of N08, Novak et al. (2009, hereafter N09) examined the role of moist processes in regulating the life cycle of three intense mesoscale snowbands over the northeast United States. They found a variety of upper-level potential vorticity (PV) evolutions contributed to midlevel trough formation. 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 role of latent heating in band evolution. Conditional stability was reduced near 500 hPa in each case several hours prior to band formation via differential horizontal potential temperature (θ) advection in southwest flow ahead of the upper-level trough. Weak stability persisted until band formation, when the stratification generally increased in association with the release of conditional instability.

The band life cycle findings of N08 were based on one event, while the findings of N09 were based on only three events. Hourly moderate resolution (~20 km) analyses from a large set of cases may help establish common mesoscale flow and thermodynamic evolutions associated with banded events, which can be related to cyclone features, such as the trowal (e.g., Martin 1999; Han et al. 2007) and upper-level PV distribution.

For example, deepening cyclones often exhibit a hook-shaped upper-level PV anomaly, with a thin PV anomaly extending equatorward from the poleward PV reservoir and curving cyclonically into a hook (e.g., Thornicroft et al. 1993; Posselt and Martin 2004, their Fig. 2a). The notch of low PV separating the hook from the poleward PV connection has been associated with the trowal (Posselt and Martin 2004). The 25 December 2002 and 12 February 2006 cases studied by N09, as well as the case studied by Martin (1998a,b), exhibited a PV hook, whereas the 14 February 2007 case studied by N09 did not. A large set of cases may help determine whether the PV hook is a common PV distribution that supports band formation.

A spectrum of instabilities have been observed during band evolution, including conditional symmetric instability (CSI; e.g., Schultz and Schumacher 1999), conditional instability (CI; e.g., Trapp et al. 2001; Morales 2008), and inertial instability (II; e.g., Jurewicz and Evans 2004; Schultz and Knox 2007). Additionally, multiple instabilities may coexist, as discussed by Bennetts and Sharp (1982), Jascourt et al. (1988), Schultz and Schumacher (1999, section 5), and Schultz and Knox (2007). However, the frequency of each type of instability during band evolution has not been quantified. Hourly moderate resolution analyses from a large set of cases would allow quantification of the types of instabilities present during band evolution.

How the frontogenetical forcing and stability evolves for a set of nonbanding (null) cases remains an open question. In a study of precipitation patterns of 20 heavy snow cases in the northeastern United States during the 2002–05 cool seasons, Greenstein (2006) found that banded events (his “classic” type) had stronger and deeper frontogenesis and weaker conditional stability than events exhibiting more uniform radar reflectivity patterns. However, this result was based on only 3 classic banded and 9 uniform events using 3-hourly 32-km horizontal grid spacing analyses.

The purpose of this paper is to determine:

- whether the midlevel trough, frontogenesis, and stability evolution presented in N08 and N09 are representative of a larger sample of banded events;
- how the evolution of midlevel frontogenesis and stability relate to features of a mature baroclinic wave, such as the trowal and upper-level PV distribution;
- the frequencies of different types of instabilities during band evolution;
- the types of cyclones that produce intense bands in the comma head; and
- distinguishing characteristics between cyclones with closed midlevel circulations that develop bands in the comma head and those that do not.

To accomplish these goals, a dataset of banded and null (defined in section 2) events during the 2002–08 cool seasons is developed in section 2. Unique to this study is the use of hourly 20-km horizontal grid spacing analyses. These analyses are used to compare the frontogenesis and stability evolution during banded and null events using statistical analysis in section 3.
synoptic classification based on the upper-level PV distribution is presented in section 4. Section 5 details the evolution of the most common banded and null cyclone types through composite analysis. A discussion of the results and summary schematics are presented in section 6.

2. Case selection

For the 2002-08 cool seasons (October–April), heavy precipitation cases over the northeast United States were identified following the methodology of N04 using the same study domain from 36.5°–50°N and 65°–85°W. Precipitation and snowfall information from the National Oceanic and Atmospheric Administration (NOAA) Daily Weather Maps series (available online at http://www.hpc.ncep.noaa.gov/dailywxmap/index.html) were used to identify cases that exhibited 24-h precipitation amounts &gt;25.4 mm of rain or 12.7-mm liquid equivalent in the case of frozen precipitation at any station in the domain. This resulted in the identification of 144 heavy-precipitation cases (Fig. 1), or an average of \( \frac{144}{7} \) cases season\(^{-1} \). This average is comparable to the average found by N04 during the 1996–2001 cool seasons (\( \frac{22}{2} \) season\(^{-1} \)).

The focus of the current study is single-banded events occurring in the comma head of cyclones. Single-banded events are defined as in N04 as the occurrence of a linear radar reflectivity structure 20–100 km in width, &gt;250 km in length, with an intensity &gt;30 dBZ, which is maintained for at least 2 h. An example radar image of a single-banded event is shown in Fig. 2a. More than one single-banded event may occur during the same heavy-precipitation case. N04 found 81% of their sampled single-banded events occurred in the northwest quadrant of the cyclone. To identify cases with likely comma head signatures, the presence of a 700-hPa closed low (using a 3-dam contour interval) served as a proxy for the presence of a comma head. Given that there is no defined threshold depth of a closed circulation that will result in a comma head (as viewed on satellite imagery), this methodology may not identify all cyclones exhibiting comma head satellite signatures; however, a closed 700-hPa low has been associated with previous northeast U.S. banded events (e.g., Nicosia and Grumm 1999; N04). Three-hourly 32-km grid spacing North American Regional Reanalysis (NARR; Mesinger et al. 2006) data were examined to identify cases exhibiting a 700-hPa closed low at some time during their evolution over the study domain. Seventy-five cases met this criterion (52% of the identified heavy precipitation cases; Fig. 1). The remaining 69 cases were not examined further.

WSI Corporation mosaic radar reflectivity factor data was examined for the 75 heavy-precipitation cases with closed 700-hPa lows to identify single-banded events. The data was available at 5-min increments on a 2-km grid. Thirty-six banded events were identified within 30 cases (3 cases exhibited multiple events separated by at least 3 h; Fig. 1). The band formation time (or start time) of each banded event is the time at which the reflectivity structure first meets the single-banded event criteria, and this time will be referred to as \( t = 0 \). Start times were rounded to the nearest hour to allow comparison with hourly gridded datasets. The start times of the banded events are listed in Table 1. Subsequent times prior to or after band formation will be designated as \( t = h \), where \( h \) is the hour relative to band formation (e.g., \( t = +2 \) refers to 2 h after band formation, whereas \( t = -2 \) refers to 2 h prior to band formation). The time of band dissipation will be referred to as \( t = \mathrm{end} \). Band maturity is considered the midpoint (in time) between \( t = 0 \) and \( t = \mathrm{end} \). An example time line for a 4-h duration banded event is shown in Fig. 3a. The mean band duration among the 36 banded events was 5 h; however, two events exhibited band durations of 10 h (Fig. 4).
FIG. 2. (left) Plan-view and (right) cross-sectional examples of (a),(b) banded; (c),(d) transitory banded; and (e),(f) null events valid at (a),(b) 1200 UTC 12 Feb 2006; (c),(d) 0500 UTC 24 Mar 2005; and (e),(f) 0600 UTC 13 Apr 2007. Plan-view fields plotted are WSR-88D radar mosaic [dBZ, shaded according to scale in (a)], 700-hPa geopotential height (thick solid, contoured every 3 dam), and Petterssen frontogenesis [red solid, positive values contoured in a doubling interval, starting at 1°C (100 km)⁻¹ (3 h)⁻¹]. Cross-sectional fields plotted are Petterssen frontogenesis [thick solid, positive values contoured every 1°C (100 km)⁻¹ (3 h)⁻¹ starting at 1°C (100 km)⁻¹ (3 h)⁻¹], saturation equivalent potential temperature (green solid, contoured every 2 K), ascent (dotted, contoured where positive every 10 cm s⁻¹), and the 70% isohume (thick purple). The manually identified values of frontogenesis (F) and conditional stability (S; see text) are shown below the cross section figures.
The remaining 45 cases that exhibited a 700-hPa closed low, but failed to develop a single band in the comma head were analyzed. Transitory null events are defined as in N04 as the occurrence of a reflectivity structure that meets all three single band criteria (i.e., shape, intensity, duration) except one (i.e., usually the duration). For example, a linear reflectivity structure was observed in central New England at 0500 UTC 24 March 2005 (Fig. 2c). This feature met the duration and shape thresholds for a single band, but not the 30-dBZ intensity threshold for a sufficient length (Fig. 2c), and thus it was categorized as a transitory null event. A null event was defined as the occurrence of a radar reflectivity pattern that did not exhibit single or transitory banding. Thus, in null events, the radar reflectivity pattern met at most only one of the three single band criteria. An example of a null event is shown for 0600 UTC 13 April 2007 (Fig. 2e). Twenty transitory null events and 22 null events were identified (Fig. 1). The start time of the transitory null and null events \( t = 0 \) was when the 700-hPa closed low entered the study domain or when the low first closed off. The start times of the 22 null events are listed in Table 2. The end of the transitory null and null events \( t = \text{end} \) was chosen 5 h after their start times to be consistent with the mean duration of the banded events. An example time line for a null event is shown in Fig. 3b. The study focuses on the 36 banded and 22 null events.

### 3. Frontogenesis and stability evolutions

#### a. Cross-sectional analysis methods

Hourly Rapid Update Cycle (RUC; Benjamin et al. 2004) analyses were used to characterize the environmental evolution of the banded and null events. The RUC data were archived on a 20-km horizontal grid with 37 vertical levels (25-hPa resolution). Cross sections of Petterssen (1936) frontogenesis, saturation equivalent potential temperature \( \theta_{eq} \), saturation equivalent potential vorticity (EPV), and RH were taken from \( t = -4 \) to \( t = \text{end} + 1 \). The cross sections were oriented normal to the most intense part of the observed band.

#### TABLE 1. 2002–08 cool-season banded event start times \( (t = 0) \).

<table>
<thead>
<tr>
<th>Start time</th>
</tr>
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<tbody>
<tr>
<td>1800 UTC 16 Oct 2002</td>
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<tr>
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<tr>
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<tr>
<td>1900 UTC 16 Nov 2006</td>
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<tr>
<td>0800 UTC 24 Nov 2006</td>
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<td>2200 UTC 16 Mar 2007</td>
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<tr>
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<td>2000 UTC 1 Mar 2008</td>
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<tr>
<td>2300 UTC 4 Mar 2008</td>
</tr>
</tbody>
</table>

Fig. 3. Example time lines for (a) a banded event lasting 4 h and (b) a null event. Time increases to the right, with each tic mark representing 1 h. The presence of a band and band life cycle stages are labeled in (a).
and followed the movement of the band as in N08. A similar method has been applied by Evans and Jurewicz (2009) and Berndt and Graves (2009). Because the null events did not have a band, cross sections were taken normal to the 700-hPa frontogenesis maximum. If a 700-hPa frontogenesis maximum was not present, the cross section was oriented normal to the 1000–500-hPa thickness gradient through the meso-scale precipitation area in the vicinity of the 700-hPa low. In both the banded and null event samples, the cross-sectional orientation at $t = 0$ was used for the preceding hours (i.e., $t = -1, t = -2, t = -3,$ and $t = -4$). Example radar mosaic images and cross sections are shown in Fig. 2.

The frontogenesis maximum within the 800–500-hPa layer and 100 km of the observed radar reflectivity band was manually identified during the band life cycle ($t = -4$ to $t = \text{end} + 1$) for each event. The 800–500-hPa layer was chosen given the association of bands with midlevel frontogenesis (e.g., Schultz and Schumacher 1999). To quantify the kinematic flow and thermodynamic contributions associated with each frontogenesis maximum, normalized frontogenesis was calculated, as in Schultz (2004):

$$F_{\text{norm}} = \frac{F}{|\nabla \theta|},$$

where frontogenesis ($F$) is divided by the magnitude of the horizontal $\theta$ gradient. Normalized frontogenesis quantifies the kinematic flow contribution to frontogenesis, whereas the magnitude of the horizontal $\theta$ gradient quantifies the thermodynamic contribution to frontogenesis. The normalized frontogenesis and horizontal $\theta$ gradient fields were plotted with the frontogenesis in the cross sections, and the normalized frontogenesis and horizontal $\theta$ gradient values associated with the frontogenesis maximum within each cross section were manually identified.

Conditional stability was assessed in a 200-hPa layer above the identified frontogenesis maximum, as in N08. The difference in $\theta_e$s between the upper and lower level in this 200-hPa layer was manually identified (small positive values represent weak conditional stability) during the band life cycle for each event. Instabilities were also assessed, and will be discussed in section 3c. The identified values of frontogenesis and conditional stability at the time of the example cross sections are shown in Fig. 2.

A time series of the mean frontogenesis and conditional stability for the banded and null events was created. Times from $t = +4$ to $t = +9$ were not included in the banded event time series because less than half of the banded events persisted this duration. Because the frontogenesis and stability samples were not normally distributed, a bootstrapping approach was used to obtain the 90% confidence intervals around the frontogenesis and stability means following Zwiers (1990). For each sample (e.g., $t = 0$ banded event frontogenesis values) a new sample of the same size was generated 1000 times by randomly selecting from the original sample. The 90% confidence intervals around the mean were determined by finding the 5th and 95th percentile of the means of all 1000 resamples. Thus, for example, if the confidence intervals of the banded and null samples

![Fig. 4. Distribution of event duration (h) among the 36 banded events.](image-url)
do not overlap, then they are significantly different at the 90% confidence level.

b. Mean time series

The mean frontogenesis nearly doubles from \( t = -4 \) to \( t = 0 \), reaches a maximum of 6.1 K (100 km)\(^{-1}\) (3 h)\(^{-1}\) at \( t = +2 \), and then decreases to 2.9 K (100 km)\(^{-1}\) (3 h)\(^{-1}\) by \( t = \text{end} + 1 \) (Fig. 5a). This evolution is similar to that found in the case studies of N08 and N09. Although both the mean horizontal \( \theta \) gradient and normalized frontogenesis increase in the leadup to band formation, normalized frontogenesis increases 73% between \( t = -4 \) and \( t = 0 \), while the \( \theta \) gradient only increases 38% (Figs. 5b,c).

Similarly, both normalized frontogenesis and the \( \theta \) gradient decrease between \( t = +2 \) and \( t = \text{end} \); however, the normalized frontogenesis decreases 37% while the \( \theta \) gradient only decreases 26% (Figs. 5b,c). Thus, changes in the kinematic flow dominate changes in the horizontal \( \theta \) gradient in explaining the evolution of the maximum frontogenesis, as found in N09.

The null events have much weaker frontogenesis than the banded events (Fig. 5a). This difference is significant at the 90% confidence level between \( t = -2 \) and \( t = +3 \). Frontogenesis during the null events peaks near 3.0 K (100 km)\(^{-1}\) (3 h)\(^{-1}\) at \( t = 0 \), and then weakens to near 2.0 K (100 km)\(^{-1}\) (3 h)\(^{-1}\) by \( t = \text{end} \). Normalized frontogenesis is stronger in the banded events than the null events from \( t = -1 \) to \( t = \text{end} \) at the 90% confidence level. In contrast there is no statistically significant difference in the horizontal \( \theta \) gradients between the banded and null events. Thus, the midlevel frontogenesis maximum, and in particular the kinematic flow associated with that maximum, is a distinguishing characteristic between banded and null events.

The banded event conditional stability exhibits a relative minimum at \( t = 0 \) [5.0 K (200 hPa)\(^{-1}\)], and generally increases during the rest of the band duration to a maximum of 6.4 K (200 hPa)\(^{-1}\) at \( t = \text{end} \) (Fig. 6). The mean conditional stability of the null events at \( t = 0 \) is 1.1 K (200 hPa)\(^{-1}\) larger than the mean conditional stability of the banded events (Fig. 6); however, this difference at \( t = 0 \) and other times was not significant at the 90% level.

c. Individual event analysis

Individual events were analyzed to explore the variability of the frontogenesis and stability evolutions, and to determine the frequency of each type of instability during band evolution.

1) FRONTGENESIS EVOLUTION

The frontogenesis at \( t = 0 \) among the banded events exceeded 10 K (100 km)\(^{-1}\) (3 h)\(^{-1}\) in 6 events, with a
maximum value of 19 K (100 km)$^{-1}$ (3 h)$^{-1}$. In contrast, the frontogenesis at $t = 0$ among the null events exceeded 10 K (100 km)$^{-1}$ (3 h)$^{-1}$ in only 1 event, with a maximum value of 11 K (100 km)$^{-1}$ (3 h)$^{-1}$. The frontogenesis maximum at $t = 0$ was larger than or equal to the frontogenesis maximum at $t = \text{end}$ in 78% (28) of the banded events, compared to 64% (14) of the null events (not shown). Perhaps of more interest is the remaining 22% (8) of the banded events, where frontogenesis was stronger at the time of band dissipation than at the time of band formation. Five of these eight banded events had stronger conditional stability at the time of band dissipation than band formation, implying that the stronger frontal circulation expected with stronger frontogenesis was limited by the greater conditional stability, as indicated by the Sawyer–Eliassen equation (Sawyer 1956; Eliassen 1962). The remaining three banded events had relatively weak frontogenesis at $t = -1$ and $t = 0$, with values <2 (100 km)$^{-1}$ (3 h)$^{-1}$. However, these events exhibited either CI or weak stability at this time, favoring strong ascent.

2) STABILITY EVOLUTION

Twenty-five of the 36 banded events (69%) exhibited the lowest conditional stability of the band life cycle at $t = -1$ or $t = 0$, consistent with the mean stability evolution (i.e., Fig. 6). The mean conditional stability is positive; however, conditionally unstable layers may be undetected by the use of a 200-hPa layer $\theta_e$ criterion, and other instabilities (such as CSI and II) may be present. Thus, conditional, symmetric, and inertial instabilities were assessed for each cross section in the region bounded below by the frontogenesis maximum, above by the 70% RH contour, and laterally by 2 grid points (40 km) on either side of the frontogenesis maximum. The full wind was used in the calculation of the symmetric and inertial stabilities, as in Novak et al. (2006) and discussed by Clark et al. (2002) and Schultz and Knox (2007). If a frontogenesis maximum was not present within the 800–500-hPa layer, the largest frontogenesis value within the 800–400-hPa layer and within 3 grid points (60 km) of the center of the cross section where RH was >70% was used. Although the degree to which moisture impacts the release of II and associated circulations are unknown (Schultz and Knox 2007, p. 2106), the 70% RH threshold criteria is retained when assessing II to be consistent with the analysis of CI and CSI. Stability states for a given hour of an event were assigned according to Table 3. For example, the hour was considered completely “stable” if $\theta_e$ increased with height at a rate >1 K (100 hPa)$^{-1}$, EPV was >0, and the absolute vorticity was >0. The hour was considered “weakly stable” if $\theta_e$ increased with height at a rate of 0–1 K (100 hPa)$^{-1}$, EPV was >0, and the absolute vorticity was >0. However, if $\theta_e$ decreased with height over a 100-hPa layer, then CI was noted for that hour. Inertial instability could coexist with CSI or CI. Approximately 20% of the banded events exhibited stable conditions prior to band formation (Fig. 7a). Conditional instability was present in ~30% of events prior to band formation, and weakly stable conditions increased from 17% at $t = -4$% to 39% at $t = 0$ (Fig. 7a). The frequency of CI was approximately double the frequency of CSI, highlighting the importance of elevated layers of CI in band formation. The release of such instability is manifest as “elevated convection” (e.g., Colman 1990a,b; Horgan et al. 2007). Inertial instability was also present in ~10% of events prior to band formation. Instabilities may also appear in combination, as eight of the 36 banded events exhibited either a CI–II or CSI–II combination prior to band formation. After band formation, the frequencies of the individual instabilities decrease. At the time of band dissipation, CSI and II were absent in all events, CI was present in only 9% of events, and stable conditions were present in 72% of events (Fig. 7a). These results are consistent with the release of instabilities and overall increase in conditional stability during band life cycle, as diagnosed in N08 and N09.

Although null events may be expected to exhibit more stable conditions than the banded events, the null events have a higher frequency of CSI, II, and especially CI than the banded events near $t = 0$, with similar frequencies at other times (cf. Figs. 7a,b). As a reviewer
noted, the higher frequency of instabilities among the null events may be due to the absence of band formation, which is an efficient mechanism for eliminating instability. Overall, these results reiterate that the stability is not a robust discriminator between banded and null events.

3) SCATTERPLOT OF BANDED AND NULL EVENTS

To illustrate the full spectrum of environments associated with banded and null events, a scatterplot of the stability and frontogenesis values (time averaged, from \( t = -1 \) to \( t = +1 \)) for the 36 banded and 22 null events was developed (Fig. 8). The null event linear regression trend line shows that if a null event has large frontogenesis, this forcing typically occurs in an environment of large stability. If one considers the two null events with frontogenesis greater than 6 K (100 km)\(^{-1}\) (3 h)\(^{-1}\) as outliers, then the null events have little relationship to stability (Fig. 8). This interpretation is consistent with bands failing to form in environments of weak stability and weak frontogenesis and also environments of strong stability and strong frontogenesis. For the banded events, there is little relationship between frontogenesis and stability given the large scatter in stability conditions (Fig. 8). Most notably, the scatterplot shows an overlap between the null and banded event samples, where the frontogenesis and stability are similar, yet one case exhibits banding and the other does not. This result suggests that band occurrence in some cases is particularly sensitive to subtle frontogenesis and stability differences.

4. Synoptic classification

a. Analysis methods

To understand the synoptic evolutions associated with band formation, banded events were subjectively classified according to the upper-level PV distribution at

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<th>EPV</th>
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<tr>
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FIG. 7. Percentage of events meeting stable, weakly stable, CI, CSI, or II thresholds as defined in Table 3 during the (a) 36 banded events and (b) 22 null events.

\( t = 0 \). To assure adequate sampling of all types of banded cyclones when developing the classification, a large number of banded events were desired. Thus, the dataset was extended to include the 1996–2001 cool-season single-banded events (northwest quadrant) from N04. A total of 72 banded events were identified during the combined 1996–2001 and 2002–08 cool seasons. Because the development of a synoptic classification is not dependent on fine temporal and spatial resolution data, the 3-hourly 32-km grid spacing NARR dataset was used. The NARR analysis closest to \( t = 0 \) was selected for each of the 72 events. The classification was then applied to events during the 2002–08 cool seasons, when hourly
20-km grid spacing RUC data was available to analyze the event evolutions.

A 700-hPa trough extending poleward of the 700-hPa low center was often observed within 100 km of the band as in N08 and N09 (not shown). The 700-hPa trough was also evident during many null events near the frontogenesis maximum; thus, a 700-hPa trough-relative composite framework was used to establish common mesoscale flow and thermodynamic evolutions associated with banded events.

In the trough-relative composite framework, the 700-hPa low center serves as a common anchor point. The gridded fields are first moved such that the 700-hPa low center for each event is at 40\textdegree N and 70\textdegree W, which is the mean 700-hPa low position of the banded events at \( t = 0 \). The gridded fields are then rotated about this point such that the mesoscale 700-hPa trough extending poleward of the midlevel low is aligned north–south. This rotation was counterclockwise (cyclonic), and varied from 0\textdegree (trough originally aligned north–south) to 90\textdegree (trough originally aligned east–west). The gridded fields are then composited.

**b. Banded cyclone types**

Three banded cyclone types were identified using the 1996–2008 NARR data: PV hook, cutoff, and lagging upper trough. The PV hook type was most common, representing 74\% of the banded events during the 1996–2008 period. The 25 December 2002 and 12 February 2006 cases studied in N08 and N09 were classified as PV hook. In this type, the band occurs along the northern portion of the PV hook. The composite of these events shows an occluding surface cyclone (Fig. 9a), with 700-hPa frontogenesis exceeding 0.5 K (100 km)\(^{-1}\) (3 h)\(^{-1}\) northwest of the surface cyclone (Fig. 9b). The 700-hPa frontogenesis is oriented along a mesoscale trough extending poleward from the midlevel low as in N08 and N09. A 700-hPa \( \theta \) ridge (dashed blue line in Fig. 9b) is also evident, which is indicative of the trowal in the vicinity of the midlevel trough. An elongated maximum of positive PV advection is found above this region (at 400 hPa), along the northern portion of the PV hook (Fig. 9c). The mean band is found in the left-exit region of an upper-level (400 hPa) jet (Fig. 9c).
The cutoff was the second most common type, representing 14% of the banded events during the 1996–2008 period. This type exhibits an upper-level PV maximum that is isolated from the poleward PV reservoir. Similar to the PV hook type, the cutoff type also exhibits an occluding surface cyclone (Fig. 9d). Consistent with the cutoff classification, the horizontal θ gradients in the cutoff type are weaker than in the PV hook type, resulting in weaker 700-hPa frontogenesis (cf. Figs. 9b,e). The 400-hPa PV field shows that the cutoff is...
asymmetric, with a PV lobe extending northeast into New England (Fig. 9f). Positive PV advection is found on the northern and western flank of this lobe, locally forcing an elongated region of ascent. The mean band is located in the left-exit region of an upper-level jet (Fig. 9f).

The third type was termed “lagging upper trough,” and represented 12% of the banded events during the 1996–2008 period. The 14 February 2007 case studied in N09 was classified as lagging upper trough. In the lagging upper trough type, band formation occurs in the northwest quadrant of the surface cyclone, but more than 400 km to the east of the 400-hPa 2-PVU contour, and within 300 km of a saturated 700-hPa PV maximum (i.e., a diabatic PV anomaly). The composite of these events exhibits a surface cyclone elongated in the north–south direction that is not yet occluded (Fig. 9g). Midlevel frontogenesis northwest of the surface cyclone is focused along the mean band axis (cf. Figs. 9g,h). The mean band occurs between a jet off the mid-Atlantic coast and jet over southern Canada (Fig. 9i). The primary upper-level trough (represented by large positive 400-hPa PV values) is 500–800 km west of the band formation region, although a small extension is found in the vicinity of the mean band (Fig. 9i). Although not explicitly investigated here, such events may be dominated by a diabatic Rossby vortex (Montgomery and Farrell 1991; Wernli et al. 2002; Moore and Montgomery 2004). The diabatic Rossby vortex is an isolated downshear-tilted PV couplet of a positive lower-tropospheric and negative midtropospheric PV anomaly, whose growth and propagation is dependent on diabatic heating and not on the mutual interaction of lower and upper PV anomalies (Montgomery and Farrell 1991; Mak 1998). Indeed, Moore et al. (2008) found that a diabatic Rossby vortex played an integral role in the 24–25 February 2005 snow event over the eastern United States.

To assure events during 2002–08 (when hourly 20-km grid spacing RUC data was available) had a similar climatology, the synoptic classification was applied to both the banded and null events during this period (Table 4). Similar to the 1996–2008 period, a majority (61%) of banded events are associated with the PV hook cyclone type. A majority of the null events (82%) are also associated with the PV hook cyclone type. The terms “banded PV hook” and “null PV hook” will be used to discriminate between these types.

The average central mean sea level pressure (MSLP) was determined using hourly RUC data for all banded and null events as well as each cyclone type during 2002–08. All types exhibit cyclone deepening (Fig. 10). The banded event time series is ~1.5 hPa deeper than the null event time series; however, this result is not significant at the 90% confidence level using a two-tailed Student’s t test (Wilks 1995). The different cyclone types exhibit different deepening rates, ranging from the rapidly deepening lagging upper trough cyclone (12 hPa from \( t = -6 \) to \( t = \) end + 2) to the modestly deepening cutoff cyclone (5 hPa; Fig. 10). The banded PV hook cyclone is approximately 3 hPa deeper than the null PV hook cyclone at all times (Fig. 10); however, this result is not significant at the 90% confidence level.

### Table 4. Number of banded and null events associated with each respective cyclone type during the 2002–08 cool seasons. The percentage of respective total banded and null events is given in parentheses.

<table>
<thead>
<tr>
<th>Cyclone Type</th>
<th>Banded events</th>
<th>Null events</th>
</tr>
</thead>
<tbody>
<tr>
<td>PV hook</td>
<td>22 (61.1)</td>
<td>18 (81.8)</td>
</tr>
<tr>
<td>Cutoff</td>
<td>7 (19.4)</td>
<td>4 (18.2)</td>
</tr>
<tr>
<td>Lagging upper trough</td>
<td>7 (19.4)</td>
<td>0 (0)</td>
</tr>
</tbody>
</table>

5. **Banded and null PV hook event evolutions**

Further comparison of the evolution of the banded PV hook and null PV hook events is motivated by their common upper-level PV distributions and similar surface cyclone pressure evolutions. Composites of hourly 20-km grid spacing RUC data were calculated for events during 2002–08 as described in section 4a. Low positions and rotation angles were determined for each hour of each event. To quantify changes in the composite 700-hPa frontogenesis and conditional stability fields, the cross section results presented in section 3 were recalculated for just the banded and null PV hook events. The
results were very similar to that shown in Figs. 5 and 6 (consistent with the majority of events being PV hook), and thus reference will be made to these figures below.

**a. Banded PV hook event evolution**

The banded PV hook composite exhibits a deepening and occluding surface cyclone from \( t = -6 \) to \( t = \) end (Figs. 11a,d,g). Frontogenesis north of the 700-hPa low center and northwest of the trowal axis nearly doubles between \( t = -6 \) and \( t = 0 \) (Figs. 11b,e). The increase in frontogenesis is consistent with the increase in normalized frontogenesis (i.e., Fig. 5b) as the midlevel trough develops and locally enhances deformation and convergence (e.g., Fig. 11e). Our interpretation follows the concepts of Hoskins and Bretherton (1972) that a large-scale (geostrophic) deformation initiates the frontogenesis, which is followed by ageostrophic (convergence) frontogenesis. The frontogenesis maximum at \( t = 0 \) is found along the newly developed midlevel geopotential height trough and maximum thermal gradient (Fig. 11e). This frontogenesis weakens by \( t = \) end as the thermal gradient and especially the normalized frontogenesis decrease (e.g., Fig. 5). These results reiterate that changes in the kinematic flow dominate changes in the \( \theta \) gradient in explaining the evolution of the maximum frontogenesis.

At upper levels, positive PV advection is maximized on the northern portion of the PV hook, which is in the comma head region of the cyclone (Fig. 11f). The 400-hPa jet strengthens from \(-40 \text{ m s}^{-1}\) at \( t = -6 \) to \( >45 \text{ m s}^{-1}\) at \( t = 0 \), and remains near \( 45 \text{ m s}^{-1}\) at \( t = \) end (Fig. 11i). The band develops and dissipates in the left-exit region of this jet.

N09 showed that changes in the midlevel flow during band dissipation could be attributed to the upscale growth of a remote diabatic PV anomaly (typically east of the band). Evidence of this process occurring in the composite was elusive, which may be a consequence of the diabatic PV anomalies occurring in the same general region (east of the band), but at different specific locations relative to the band. To test this hypothesis, each event was examined for a saturated (RH > 70%), isolated 700-hPa PV maximum exceeding 1 PVU occurring within a 600-km radius to the east of the band and within 2 h of band dissipation. Thirteen of the 22 events (59%) met these criteria. An additional 4 events (18%) exhibited a PV maximum to the east of the band, but the maximum was not completely saturated. Only 5 events (23%) did not exhibit an isolated PV anomaly east of the band.

Cross sections through the band region during the band life cycle (Fig. 12, orientations shown in Figs. 11c,f,i) reveal that the composite midlevel frontogenesis maximum is the vertical extension of the surface warm front (and surface warm occlusion when present). Between \( t = -6 \) and \( t = 0 \) the frontogenesis nearly doubles at all levels, the ascent at 600 hPa near \( x = 240 \text{ km} \) nearly doubles, and the \( \theta_e \) contours steepen in the 500–800-hPa layer near \( x = 360 \text{ km} \) (cf. Figs. 12a,c), which is representative of the destabilization that occurred (e.g., Fig. 6). N09 showed in their case studies that the initial destabilization was primarily due to differential horizontal \( \theta \) advection. The banded PV hook composite also exhibits evidence of this process, with the 450–650-hPa layer conditional stability decreasing in the band region from \(-7.5 \text{ K (200 hPa)}^{-1}\) at \( t = -6 \) to \(-6.0 \text{ K (200 hPa)}^{-1}\) at \( t = 0 \) (Figs. 13a,b). Warm air advection at 650 hPa increases dramatically in the band region during this period to over \( 3.0 \times 10^{-4} \text{ K s}^{-1}\) (Figs. 13c,d), whereas the 450–650-hPa warm air advection is more modest \((-1.5 \times 10^{-4} \text{ K s}^{-1})\) and does not increase (Figs. 13e,f). These composite results are consistent with destabilization due to differential warm air advection.

The cross section at \( t = \) end still exhibits frontal characteristics, with sloping isentropes and active frontogenesis (Fig. 12e). However, the frontogenesis has weakened (Fig. 12). The isentropes are also flatter than at \( t = 0 \) (cf. Figs. 12c,e). N09 showed that stabilization was primarily due to differential vertical \( \theta \) advection associated with the release of instability in their case studies, consistent with the decrease of instability states during band evolution shown in Fig. 7.

Overall, the composite results reiterate that the frontogenesis evolution is dominated by changes in normalized frontogenesis. The initial normalized frontogenesis increase is related to the development of a midlevel trough and associated deformation and convergence. Normalized frontogenesis decreased during band dissipation. Fifty-nine percent of the cases exhibited a saturated 700-hPa PV maximum to the east of the band, suggesting that the flow near the band was affected by the upscale growth of this anomaly during band dissipation, as diagnosed in N09.

**b. Comparison of the banded and null PV hook events**

The null PV hook composite fields at \( t = -6 \) exhibit nearly identical features as the banded PV hook composite (cf. Figs. 11 and 14), including a deepening and occluding surface cyclone, development of a midlevel trough, frontogenesis northwest of the trowal axis, PV advection in the comma head, and a strengthening jet. The only notable difference between the banded and null PV hook composites is the strength of the 700-hPa frontogenesis at \( t = 0 \) (cf. Figs. 11e and 14e).
Cross-sectional analysis quantifies that the null PV hook composite 700-hPa frontogenesis is 40% weaker than the banded PV hook composite 700-hPa frontogenesis at \( t = 0 \) (not shown), similar to the results for the full sample (i.e., Fig. 5). This difference is almost entirely a function of the difference in normalized frontogenesis (e.g., Fig. 5).

Comparison of the banded and null PV hook composite cross-sectional evolution show that the banded PV hook composite frontal environment exhibits deeper and stronger frontogenesis, especially at \( t = 0 \) (Fig. 12). Conditional instability is also less prevalent in the null cross sections, as the isentropes are generally flatter than the banded PV hook composite between 500 and 850 hPa near \( x = 360 \) km. However, the banded and null event conditional stability is not statistically different at any time, as in the full sample (i.e., Fig. 6). Consistent with weaker frontogenesis, the ascent in the null PV hook composite is \(~25\%\) weaker at \( t = 0 \) than in the banded PV hook composite (cf. Figs. 12c,d).

**Fig. 11.** As in Fig. 9, but for the banded PV hook composite at (a)–(c) \( t = -6 \); (d)–(f) \( t = 0 \); and (g)–(i) \( t = \) end. The trowal axis (denoted as a thick dashed blue line) is shown in (b),(e), and (h), and the cross-sectional orientation for Fig. 12 is shown in (c),(f), and (i).
FIG. 12. (left) Banded PV hook and (right) null composite cross section (orientations shown in Figs. 11c and 14c) evolution of frontogenesis (shaded), saturation equivalent potential temperature (green solid, every 3 K), wind into the plane of the cross section (contoured in black dotted every 5 m s$^{-1}$ starting at 30 m s$^{-1}$), ascent (dashed blue, contoured every 2 Pa s$^{-1}$), 70% RH isohume (thick black), and wind in the plane of the cross section at (a),(b) $t = -6$; (c),(d) $t = 0$; and (e),(f) $t = \text{end}$. Mean band position is shown as the black bar along the $x$ axis, and the upper jet axis is labeled as “J.”
Fig. 13. The banded PV hook composite saturation equivalent potential temperature difference (450 – 650 hPa) shaded according to scale every 1 K at (a) $t = -6$ and (b) $t = 0$. Corresponding (c),(d) 650- and (e),(f) 450-hPa potential temperature (dashed, contoured every 2 K), wind (half barb is 2.5 m s$^{-1}$, full barb is 5 m s$^{-1}$, and pennant is 25 m s$^{-1}$), and potential temperature advection (shaded according to scale every $0.5 \times 10^{-4}$ K s$^{-1}$). The cross-sectional orientation for Fig. 12 is labeled, and the mean band position (north–south oriented thick line) is shown at $t = 0$. 
6. Discussion and summary schematics

a. A common banded event evolution

The present climatological and composite results demonstrate a common banded event evolution, which is consistent with the 25 December 2002 and 12 February 2006 case study evolutions explored in N08 and N09. A schematic depiction of the plan- and cross-sectional view of the common banded cyclone evolution is shown in Fig. 15, which synthesizes the results from this study, N08, and N09.

Several hours prior to band formation, an area of weak midlevel frontogenesis is found in diffluent midlevel flow northwest of the developing surface cyclone and downshear of the upper PV anomaly (Fig. 15a). The band development region exhibits an area of frontogenesis sloping into the cold air, associated with the cyclone’s surface warm front (Fig. 15b). However, the frontogenesis is relatively weak, and the conditional instability is limited to the warm sector. Given the weak midlevel frontogenesis and strong conditional stability above the frontal zone, ascent is limited at this time (Fig. 15b).
FIG. 15. Schematic depiction of the banded PV hook cyclone (a),(c),(e) plan-view and (b),(d),(f) cross-sectional evolution. Key features shown in plan-view depiction include the upper jet (dashed thick arrow), the lower PV anomaly (blue hatched outline), the upper PV anomaly (green hatched outline), the midlevel trowal axis (gray dashed), the midlevel geopotential height (thin black), the midlevel frontogenesis (red shading), and the surface fronts and pressure centers. Cross section end points (“A” and “B”) are marked. Key features shown in cross-sectional depiction include frontogenesis (red shading), isentropes (green solid), upper jet (labeled), conditional instability (gray shading), and representative airstream through the ascent maximum in the plane of the cross section (arrows). Hydrometeor growth and drift depicted by snowflake in (d) (not drawn to scale).
During the 6 h leading up to band formation, lower-tropospheric frontogenesis nearly doubles (Figs. 5a and 11b,e) and the conditional stability above the frontal zone is reduced (Fig. 6). As diagnosed by N09, the conditional stability is reduced primarily by differential horizontal $\theta$ advection in a layer centered near 500 hPa ahead of the upper trough. The increase in midlevel frontogenesis is primarily due to changes in the kinematic flow associated with the development of a mesoscale geopotential height trough. The trough extends poleward of the 700-hPa low center and is located along the northwest flank of the trough axis (Fig. 15c). PV diagnostics have revealed that latent heating plays a large role in the development of the midlevel trough and associated frontogenesis (N09), although upper-level positive PV advection along the elongated northern portion of the PV hook also contributes to midlevel trough formation. The latent heat release from the heavy precipitation (not yet organized into a band) creates a diabatic PV anomaly that enhances a frontogenetical circulation. The band develops within this heavy precipitation as frontogenesis increases in a positive feedback between latent heating and associated flow and temperature changes (N09).

The band is found in an elongated region of intense forcing for ascent associated with the midlevel frontogenesis maximum (located along the midlevel geopotential height trough) and upper-level positive PV advection (located along the elongated northern portion of the PV hook; Fig. 15c). The most likely stability state during band formation is either weakly stable conditions or conditional instability, although CSI and II may also be present. The presence of elevated conditional instability above the frontal zone as found in N08 and N09 is shown schematically in Fig. 15d. Hydrometeor lofting and drift may alter the position of the surface band relative to the ascent maximum (N08), as is also shown schematically in Fig. 15d.

During band maturity frontogenesis continues to increase (e.g., Fig. 5a) in response to the latent heating and associated induced circulation of the band itself (N09). However, N09 found the resulting strong ascent generally leads to a differential vertical advection pattern that stabilizes the environment, serving as a local “brake” on the feedback. Consistent with this process, the mean stability increases during band maturity (e.g., Fig. 6). The release of instabilities likely also contributes to the overall stability increase (e.g., Fig. 7).

Band dissipation commonly occurs as a consequence of weakening frontogenesis while the stability remains strong. The frontogenesis commonly weakens as a consequence of kinematic flow changes associated with the upscale growth of diabatic PV anomalies typically to the east of the band (N09; Fig. 15e). The flow in the band region is altered both in terms of wind direction and speed, resulting in reduced deformation/convergence and associated frontogenesis.

Analysis of hourly 20-km grid spacing analyses facilitates an evolutionary view of the band formation and dissipation process. Compared to previous studies and schematics of band formation (e.g., Nicosia and Grumm 1999, their Fig. 17; N04, their Fig. 15a; Moore et al. 2005, their Fig. 15a), features unique to this view include the following:

- a nearly twofold increase in frontogenesis along a developing midlevel (~700 hPa) geopotential height trough during the 6 h leading up to band formation;
- the common occurrence of band formation along the northern portion of the developing upper-level PV hook;
- the positive feedback between latent heat release, frontogenesis, and band development;
- the contribution of remote diabatic PV anomalies to band dissipation via induced changes in flow and associated frontogenesis;
- that the most likely stability state during band formation is either weakly stable conditions or conditional instability, whereas CSI and II are less common.

The relative lack of CSI in the cases is consistent with Schultz and Schumacher (1999, p. 2727), who state that “the nature of banding in clouds or precipitation appears to be only weakly related to measures of CSI.” This study quantifies that CSI is observed one-third to one-half less often than CI near the time of band formation within the comma head of northeast U.S. cyclones (Fig. 7a).

Not all banded events exhibit the above cyclone evolution. Fourteen percent of the sample banded events exhibited a band along the periphery of a closed, isolated upper PV maximum (cutoff events), and 12% exhibited a band to the east of the upper-level PV trough, but within 300 km of a saturated 700-hPa PV maximum (lagging upper trough events). However, even in these scenarios the band was found along an elongated axis of midlevel frontogenesis and upper-level positive PV advection, as in the banded PV hook cases. Similarly not all banded events exhibit the frontogenesis and stability evolution shown in Figs. 5 and 6. However, in each case, the behavior of the band (and associated ascent) is consistent with the coupled frontogenesis–stability relationship described by the Sawyer–Eliassen equation.

b. Distinguishing characteristics between banded and null PV hook events

Banded and null PV hook events exhibit similar synoptic pattern evolutions. However, the midlevel frontogenesis
maximum, and in particular the kinematic flow associated with that maximum, is a distinguishing characteristic between banded and null events. A scatterplot of frontogenesis and stability for all banded and null events shows an overlap of data points in the frontogenesis-stability phase space. Some cases may be clearly banded (strong frontogenesis and weak stability) or nonbanded (weak frontogenesis and strong stability); however, band occurrence in some cases is particularly sensitive to relatively subtle frontogenesis and stability differences. Further evidence of this sensitivity is revealed by the fact that statistically significant differences between the midlevel frontogenesis maximum of the banded and null events only appear ~2 h prior to band formation (e.g., Fig. 5a).

The above results suggest forecasting techniques relying on precursor observational signals to distinguish banded and null events may have very short lead times. The results may also suggest a “tipping point” in the positive feedback between latent heat release and increasing frontogenetical forcing just prior to band formation. Reliably predicting this tipping point beyond several hours appears difficult, given the intrinsic predictability limits of the mesoscale details of moist baroclinic waves (e.g., Zhang et al. 2003; Zhang et al. 2007). Indeed, Evans and Jurewicz (2009) have shown that the correlation between forecasts of band ingredients (i.e., frontogenesis, small EPV, and saturation) and observed event total snowfall decrease substantially beyond 12 h in an operational model. On the other hand, successful detailed operational short-range predictions of intense bands have been made (e.g., Novak et al. 2006), and there is likely case variability in the predictability limit. Future work is needed to define the practical predictability limits of band prediction as well as identify meteorological features that influence band sensitivity.

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REFERENCES


