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Analysis of Banding in 26-27 December 2010 East Coast Blizzard

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ABSTRACT

Blizzard conditions occurred in the New York City metropolitan area and portions of adjacent southern New England on 26-27 December 2010 in conjunction with a strong coastal cyclone. Parts of New Jersey received over 80 cm of snow while Long Island and coastal Connecticut observed wind gusts of over 30 m s\(^{-1}\). The heaviest snow was concentrated along a north-south oriented mesoscale snowband that extended from coastal New Jersey northward through the New York City metropolitan area. This mesoscale snowband, which remained quasi-stationary for approximately 12 h, was associated with strong low and mid-level frontogenetical forcing. In addition to the primary band, 8 secondary bands were observed in the radar imagery which contributed to locally higher snowfall amounts and rates as the bands translated westward across Long Island.

Information regarding mesoscale banding will first be discussed in order to compare and contrast this case with others. A summary of the storm's impacts will briefly be presented in order to demonstrate the storm's significance. An overview of the synoptic environment will be discussed using the 0.5° NCEP GFS analyses focusing on dynamics from a quasi-geostrophic perspective. Mesoscale features such as the persistent band over New Jersey will be identified using radar data and surface observations. The forcing behind this mesoscale band will be examined in greater detail using the GFS and 20-km RUC analyses and soundings from the NWS operational Doppler radar at Upton (KOKX), New York with emphasis on frontogenetical processes. In addition, an analysis of the strong wind gusts as they relate to an overnight spike in the low-level winds and convective mixing will also be presented.
1. Introduction

Mesoscale banding within East Coast winter storms has been a forecasting challenge for decades. The bands are responsible for locally higher snow amounts and rates as compared to that of the non-banded storm environment. A snowband could cause dangerous travel conditions, especially if the public (and forecasters) are caught off guard. Mesoscale banding has been a topic of research (e.g. Sanders 1986; Wolfsberg 1986; Nicosia and Grumm 1999; Novak et al. 2004) to provide forecasters with a better understanding to employ in their forecasts to decrease negative societal and economic impacts relating to the events.

a. Mesoscale band mechanics

Mesoscale banding is a frequent occurrence with extratropical cyclones. A study by Novak et al. (2004) noted that 85% of the total 88 cases reviewed of Northeast U.S. cyclones exhibited banding. Single banded events were found to be the most common forming northwest of the cyclone center in the comma-head portion of developing cyclones (Fig. 1).

![Fig. 1. Location of bands relative to the cyclone center (origin) for single banded events in the observational study of 88 events of Northeast U.S. cyclones. [Adopted from Novak et al. (2004 Fig. 4).]](image-url)

The balance of several factors determine the existence, intensity and areal coverage of mesoscale banding. The factors include, but are not limited to, frontogenetic forcing, moisture availability, small moist symmetric stability (Evans and Jurewicz 2009). Frontogenesis leads to
vertical motion by inducing a thermally direct ageostrophic circulation. An increasing horizontal temperature gradient results in the disruption of thermal wind balance with the existing vertical wind shear; the atmosphere responds by producing a thermally direct ageostrophic circulation transverse to the baroclinic zone. (Nicosia and Grumm 1999)

The coupling of frontogenesis and moist symmetric stability is important regarding the dynamical forcing for mesoscale snowbands. The ageostrophic circulation across a frontal boundary described by the Sawyer-Eliassen equation is affected by a reduction in stability (Holton 2004). It is this reduction of static stability in a moist environment that promotes more vigorous updrafts in the presence of frontogenetically forced ascent on the mesoscale. Emanuel (1985) deduced that ascent associated with frontogenesis is enhanced and constricted to a smaller scale when the symmetric stability is low on the warm side of a developing frontal zone. This relationship is further discussed by Sanders and Bosart (1985) and is shown in Fig. 2a. Due to the reduction of gravitational and inertial resistance to the updraft, the updraft can become stronger but since the downdraft is conserved and mass fluxes must still balance, the updraft is horizontally restricted (Sanders and Bosart 1985). Similarly, Clark et al. (2002) stated that the ascending branch of the direct circulation driven by the frontogenesis will be intensified and concentrated because of the reduced moist potential vorticity (MPV) attributed to horizontal moisture and temperature gradients. Further research by Thorpe and Emanuel (1985) showed that for an environment which has small stability to slantwise convection, frontogenesis proceeds at a much increased rate and the horizontal scale of the ascent in the warm air is considerably reduced. Figure 2b indicates a feedback loop for the relationship between saturation equivalent potential vorticity (EPV); EPV is comparable to moist potential vorticity (MPV) which was
previously discussed. Moore and Lambert (1993) state that negative EPV computed in a cross section is an effective indicator of conditional symmetric instability.

![Diagram](image)

**Fig. 2.** (a) Vertical cross sections of idealized transverse circulations accompanying large-scale frontogenetical confluence (double-shafted arrows): (a’’) uniform potential vorticity and (b’’) vanishing potential vorticity in warmer air. Dashed lines are isotherms. [Adopted from Sanders and Bosart (1985 Fig. 3).] (b) Feedback loop between frontogenesis, potential temperature gradients, EPV, and vertical motion. [Adopted from Nicosia and Grumm (1999, Fig. 19).]

Large-scale forcing can enhance the formation of mesoscale snowbands. Jurewicz and Evans (2004) studied intense snowbands that were related to strong quasi-geostrophic forcing for ascent and rapid surface cyclogenesis. Conceptual models produced by previous studies provide the composite synoptic conditions attributed to mesoscale band formation. Two conceptual models are included in Fig. 3. The first model in Fig. 3a includes frontogenesis within a broad area of deformation with representative dilation axes northwest of the cyclone center with fronts and jet cores. This is a simplified model and does not include an outlined area of EPV reduction as Fig. 3b does. Figure 3b also highlights the dry tongue jet which has been attributed to a reduction in EPV due to its enhancing of horizontal moisture gradients and its introduction of dry air above the moist low-level jet (Clark et al. 2002; Nicosia and Grumm 1999). It is the reduction of static stability in a moist environment that promotes enhanced mesoscale updrafts in the presence of quasi-geostrophic forced ascent at the synoptic scale. From Fig. 3, both conceptual models...
support maximum mesoscale band formation to be northwest of the cyclone center in a region of midlevel frontogenesis with an area of reduced EPV to its south.

**Fig. 3.** Conceptual models of synoptic and mesoscale environments conducive to band formation from (a) Novak et al. (2004, Fig. 15) and (b) Nicosia and Grumm (1999, Fig. 17).

Previous studies highlight the transverse circulations that result in mesoscale snowbands. Therefore, it is important to take the vertical structure of the banded environment into consideration when diagnosing ingredients for band formation. Figure 4 is a composite cross section by Novak et al. (2006) highlighting the main ingredients of slanted mid-level frontogenesis and the corresponding upward branch of the circulation within a region of weak moist symmetric stability, allowing for stronger ascent. The given probable location of a band corresponds with a maximum in mid-level frontogenesis.
b. 26-27 Dec 2010 East Coast blizzard summary

An area of low pressure tracked along the East Coast of the U.S. from the 26-28 December 2010 and produced blizzard conditions in the NYC Metropolitan area and portions of southern New England including snowfall rates $> 7.6 \text{ cm h}^{-1}$ (3 in h$^{-1}$), snowfall amounts $> 80 \text{ cm}$, and peak wind gusts of 35 ms$^{-1}$. Figure 5 shows a snowfall distribution map and the distribution of peak wind gusts over 24 ms$^{-1}$. The higher snowfall amounts recorded west and north of the city were the result of a quasi-stationary mesoscale snowband. The half-width, or lateral distance from the point of maximum snowfall to the point receiving half the amount is approximately 50 km near the NYC metro area (Sanders and Bosart 1985). This snowfall amount gradient over a small distance was devastating given the densely populated nature of the area and wind gusts of
over 24 ms$^{-1}$ caused dangerous travel conditions.

![Figure 5](http://www.erh.noaa.gov/okx/StormEvents/storm12262010.html) and (b) peak wind gusts over 55 mph.

The NYC metropolitan area took weeks to recover from this particular snowstorm. Despite the idea that this was due to political reasons, this event was a considerable forecast challenge. The challenge was rooted in the lack of model consensus regarding the track of the low pressure center. Figure 6 shows the 72-h forecast tracks for 6 models and the observed track. There was inconsistency within the same model (e.g. GFS) regarding the forecast track. Given the study by Novak et al. (2004), banding occurs embedded within the comma-head portion of cyclones which indicates that the cyclone track is critical in forecasting the likely location of the formation of bands. Most of the forecast position of the low pressure center was farther to the east than what verified. Mesoscale banding resulted in snowfall amounts double that of the ambient amounts in highly populated areas. The motivation of this study is to examine the structure and evolution of mesoscale banding during the time of greatest band formation.
2. Data and Methodology

a. Radar identification of bands

WSR-88D radar data from the Upton, NY (KOKX) radar and the Mt. Holly, NJ (KDIX) radar for 1200 UTC 26 December to 0000 UTC 28 December was obtained through the National Climatic Data Center (NCDC) Hierarchical Data Storage System with the radar sites shown in Fig. 7. Level-III data was displayed using the GRLlevel3 program. A band was subjectively identified if a feature satisfied the following criteria: a) quasi-linear structure, b) persisted \( \geq 1 \) h, c) base reflectivity value of \( \geq 30 \) dBZ. Additionally, vertical azimuth display (VAD) wind profiler data were obtained from the KOKX radar.
Fig. 7. Locations of two radar sites used in this study: Upton, NY (KOKX) and Mt. Holly, NJ (KDIX).

b. Gridded data

0.5 degree GFS data and 20 km RUC data were obtained from internal archives and NOAA/NCEP/ESRL, respectively, and analyzed and displayed using the General Meteorology Package (GEMPAK) (desJardins et al. 1991). Two spatial domains shown in Fig. 8 were chosen to display variables to examine synoptic and mesoscale features.

Fig. 8. Synoptic spatial scale (Scale 1) and mesoscale spatial scale (Scale 2) used to display gridded data using GEMPAK.

The synoptic environment was analyzed first by displaying such variables as 300-hPa, 500-hPa, 700-hPa, 850-hPa geopotential heights and winds and sea level pressure for the duration of the event. Quasi-geostrophic forcing for ascent was examined for this case and obtained through
Heather Archambault’s online database (http://www.atmos.albany.edu/student/heathera/all/2010-2011.html).

c. Additional data

Infrared satellite data were obtained from the University Corporation for Atmospheric Research (UCAR) image archive (http://www.mmm.ucar.edu/imagearchive/). METAR reports were extracted from National Weather Service (NWS) first-order stations in the domain to obtain snowfall rate values. Soundings were obtained from the University of Wyoming database (http://weather.uwyo.edu/upperair/sounding.html) for KOKX and KDIX.

3. Results

a. Synoptic environment

Analysis of the synoptic environment associated with this cyclone showed that it was a strongly-forced cyclone. An upper-level trough was evident with a coupled jet feature seen in Fig. 9a and differential cyclonic vorticity advection evident in Fig. 9b provided forcing for ascent by the quasi-geostrophic omega equation to enhance the surface cyclone of this event (not shown). Rapid cyclogenesis with a surface pressure decrease of approximately 8-hPa in 6 h between 2100 UTC – 0300 UTC 27 Dec and a subsequent decrease of 10-hPa in 6 h from 0600 UTC – 1200 UTC 27 Dec can be inferred to be associated with the increasing temperature gradient shown in Fig. 9d due to strong cyclonic circulation with little friction over the ocean and the maintenance of thermal wind balance by strengthening the upper-level jet.
Fig. 9. 4-panel figure for 0000 UTC 27 Dec 2010 displaying (a) 300-hPa geopotential heights (contoured, m), isotachs (shaded, kt), and winds (barbed, kt), (b) 500-hPa geopotential heights (contoured, m), absolute vorticity (shaded, s⁻¹), and winds (barbed, kt), (c) 700-hPa geopotential heights (m, contoured), relative humidity (above 70% and 90%, shaded), and winds (barbed, kt), and (d) 850-hPa geopotential heights (solid contours, m), temperature (°C, red solid contours > 0°C and dashed contours < 0°C), and winds (barbed, kts).

Analysis of the Q-vectors and Q-vector convergence is presented in Fig. 10 for 0000 UTC 27 Dec. There is a maximum in Q-vector convergence near the approximate location of the surface low pressure center (not shown) which implies that the cyclone is strengthening. The Q-vectors northwest of the approximated cyclone center are oriented parallel or across isotherms towards warmer air which is indicative of frontogenesis. Therefore, the Q-vectors, although a diagnostic tool for large-scale ascent, can be related to areas of mesoscale banding which will be discussed.
The infrared satellite imagery was analyzed to observe the cyclone structure. It was useful in noting the location of the “dry tongue” or “dry slot,” a jet intrusion of dry air that has been known to decrease stability and provide favorable conditions for the formation of mesoscale banding. The dry slot is highlighted in Fig. 11, an infrared image for 2315 UTC 26 Dec.
b. Identification of mesoscale bands

This event exhibited 9 mesoscale bands that included one primary band that had a duration of 17 hours, first forming south of Long Island and then rotating cyclonically as the cyclone center translated northward along the East Coast. 7 of the 8 secondary bands formed to the east of the primary band and were caught up in the easterly flow which has been documented by Nicosia and Grumm (1999) to occur. Figure 12 is a schematic that contains the initial and final location of each identified band with a subscript containing its duration.
It was found that the time during which the most bands formed (6 out of 9) was between 2300 UTC on the 26 and 0200 UTC on the 27. This interval was used for further analysis as to why the most bands formed and will be discussed in the next section. The average duration of the mesoscale bands was 4 h (3 h) including (excluding) the primary band from the calculation. The primary band was 30 km wide at its widest and the secondary bands were 10 - 20 km wide. Possible orographic enhancement may be responsible for the strength of the primary band due to convergence with the Ramapo Mountains in northern NJ.

c. Mesoscale analysis of banding
Section 1 highlighted the importance of frontogenesis as a factor of the formation of mesoscale snow bands. 2-D frontogenesis was plotted using the function in GEMPAK for 1000-hPa and 700-hPa shown in Fig. 13. The area of maximum low-level frontogenesis (25 K 100 km$^{-1}$ 3 h$^{-1}$) is just southeast of Long Island with the maximum sloping westward with height and being located at mid-levels across central Long Island and into southern New England. The low-level maximum is coincident with the location of most of the secondary band formation, however the primary band was located to the west of the mid-level maximum.

![Fig. 13. 2-D Frontogenesis (shaded, K 100 km$^{-1}$ 3 h$^{-1}$), potential temperature (K), and wind (barbed, kt) for 0000 UTC 27 Dec for (a) 700-hPa and (b) 1000-hPa.](image)

The area of maximum mid-level frontogenesis corresponds with 700-hPa deformation as seen in Fig. 14a. The axis of dilatation is parallel to the isotherms which is indicative of maximum frontogenesis (Nicosia and Grumm 1999). 1000-hPa convergence in Fig. 14b showed a maximum where most secondary bands formed. The maximum deformation moved slightly westward with the axis of dilatation remaining just to the east of the primary band in subsequent hours (not
According to previous studies, regions of reduced EPV with values < 0.25 PVU on the warm side of the frontal circulation are indicative of strengthening frontogenesis and the enhancement of the updraft of the thermally direct ageostrophic circulation (Jurewicz and Evans 2004). The EPV can be decreased by the presence of the dry tongue as shown in Fig. 11. EPV was calculated using the GEMPAK function for potential vorticity calculated with theta-es. For 0000 UTC 27 Dec, EPV values < 0 PVU were oriented parallel to the primary band orientation at that time with largely negative values located near the approximate location of the dry tongue jet as seen in Fig. 15.
Cross sectional analyses were completed intersecting the location of the primary band. Figure 16 for 0000 UTC on the 27 shows that there is a low-level maximum in frontogenesis eastward of a mid-level maximum with large negative values of omega (< 0.021 hPa s⁻¹) located above the frontogenesis maxima with descent below the mid-level maximum which correlates with the thermally direct circulation. The relative minimum in EPV values (< -1.5 PVU) is collocated with the upward branch of the frontogenetical circulation which corresponds with conditional symmetric instability. The low-level easterly flow may act to translate the location of the primary band (red star) farther westward than expected from the conceptual model provided in Fig. 4. A hypothesis as to why many secondary bands formed around this time (6 bands) but did not persist and translated westward may be convergence associated with the low-level frontogenesis.
maximum with ascent and becoming rapidly caught up in the easterly flow that is in excess of 50 kt.

Fig. 16. Cross section at 0000 UTC 27 Dec intersecting the primary band (red star) from point C to point D with frontogenesis (shaded, K 100 km$^{-1}$ 3 h$^{-1}$), potential temperature (K), omega (10$^{-3}$ hPa s$^{-1}$), EPV* (contoured red, PVU), and wind (barbed, kt). The map in the upper right-hand corner displays the points and cross section axis over a radar image for approximately the same time.

d. Peak wind gust analysis

Peak wind gusts of over 35 ms$^{-1}$ were observed associated with this cyclone recorded between 0000-0700 UTC. Upon analysis of the KOKX sounding for 27 Dec 0000 UTC in Fig. 17a and 27 Dec 0600 UTC in Fig. 17b, it is hypothesized that the low-level winds (60 kt at 925-hPa at 0600 UTC) were mixed down through convective processes. A cross section of theta-exp indicated that
the environment east of central Long Island was convectively unstable as seen in Fig. 18. This location corresponds with the minimum in EPV from Fig. 16 which may indicate that the convectively unstable atmosphere dominated the forcing for band formation as well. The rapid cyclogenesis, as discussed in Section 3.a, increased the pressure gradient and thus increased the low-level winds, especially close to the cyclone center. However, the overnight spike in the low-level winds and resultant peak wind gusts were most likely due to the convectively driven mixing down of stronger wind speeds to the surface. The strong low-level winds are also illustrated by the VAD wind profile image in Fig. 19.

**Fig. 17.** Soundings from KOKX for (a) 0000 UTC 27 Dec and (b) 0600 UTC 27 Dec with a 60 kt measured wind speed at 925-hPa highlighted by the red arrow.
FIG. 18. Cross section at 0000 UTC 27 Dec intersecting primary band (red star) of moist geostrophic potential vorticity ($\text{MPV}_g$) (shaded, PVU) and saturated equivalent potential temperature ($\theta_{es}$) (contoured, K) indicating an area that is convectively unstable inferred by comparison to Fig. 4.

FIG. 19. VAD wind profile from KOKX for 0153 UTC 27 Dec with wind speeds given in knots.

4. Conclusions
The East Coast Blizzard of 26-27 December 2010 caused negative societal and economic impacts to the NYC metropolitan area. The storm was characterized by multiple mesoscale snowbands that caused locally higher snowfall amounts and rates. The storm was poorly forecast in advance due to model discrepancies.

The cyclone was linked to strong quasi-geostrophic forcing for ascent leading to its strengthening as it progressed along the East Coast. This was due to linkage to a coupled jet feature at 300-hPa that provided an enhanced region of upper-level divergence, differential cyclonic vorticity advection inferred from the absolute vorticity distribution at 500-hPa that would force ascent by the quasi-geostrophic omega equation, and an area of enhanced Q-vector convergence which would also lead to ascent by the Q-vector form of the quasi-geostrophic omega equation. With this strongly forced cyclone, a tight temperature gradient was apparent west and north of the cyclone center due to strong warm air advection at low-levels. The reduction of static stability in a moist environment promoted more vigorous mesoscale updrafts in this case.

The mesoscale snowbands were forced by strong frontogenesis sloping westward with height with reduced saturation equivalent potential vorticity on the warm side of the frontal circulation. The primary band was most directly related to the thermally direct ageostrophic frontal circulation. The presence of 8 secondary bands that were relatively short-lived and translated westward were most likely due to strong low-level convergence in an area of maximum low-level frontogenesis with vigorous ascent and ultimately were caught up in the easterly flow.

The overnight spike in the low-level winds was most likely due to the mixing down of greater wind speeds, just above the surface in the boundary layer. This was most likely enhanced by the
fact that the cyclone was experiencing rapid cyclogenesis and the environment was found to be convectively unstable.

The following conceptual model in Fig. 20 for this case was created to display the key ingredients for the formation of mesoscale bands. The main ingredients for band formation were frontogenesis, reduced EPV, and low-level convergence. This model is produced for a representative time period in this study, 0000 UTC 27 Dec.

**Fig. 20.** Conceptual model specific to the East Coast Blizzard of 26-27 December 2010 highlighting the main ingredients for the formation of mesoscale snowbands.

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