The Thermodynamic and Microphysical Evolution of an Intense Snowband during the Northeast U.S. Blizzard of 8–9 February 2013

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ABSTRACT
An intense snowband developed across Long Island, New York, to the north and west of the surface cyclone center on 8–9 February 2013. The snowband evolved through three distinct phases during its 12-h lifetime. During phase 1 the band developed in an area of low-to-midlevel frontogenesis and pivoted over central Long Island and southern Connecticut, where it remained for approximately 10 h. The environment surrounding the snowband cooled to <0°C; however, the band was collocated with a 900–700-hPa layer that remained above 0°C for ~5 h. During phase 2 the band exhibited heavy snowfall rates exceeding 7.5–10 cm h⁻¹ with large and aggregated snow, wet-growth hail-like particles, and a radar reflectivity of ~55 dBZ. About 1 h later during phase 3, the snowband reflectivity decreased to near 30 dBZ and was characterized by less dense snow in a colder environment while still maintaining heavy snowfall rates (6.5–6.7 cm h⁻¹). The Weather Research and Forecasting (WRF) Model was used to analyze the band and temperature evolution. Model trajectories terminating within the warmer snowband environment underwent rapid ascent on the east side of the band during which condensation and deposition enhanced the warming before undergoing rapid descent within the band. Analysis of the thermodynamic equation within the band environment revealed that this subsidence warming and upstream condensational heating for trajectories entering the band partially offset the diabatic cooling term, which supported a warmer layer and mixed precipitation during phase 2. Finally, model sensitivity tests showed that melting helped cool low levels and change the microphysical character to all snow during phase 3.

1. Introduction
An extratropical cyclone on 8–9 February 2013 produced 0.3–0.6 m (1–2 ft) of snow across New York City and Long Island, New York, and over 0.9 m (3 ft) of snow in central Connecticut (Fig. 1a). Hurricane-force winds occurred along the coast from Massachusetts to Maine, including an observed gust of 34 m s⁻¹ at Boston Logan International Airport. After the storm, a federal state of emergency was declared for Connecticut and a federal disaster declaration was issued for Connecticut and Long Island. Much of the heavy snow fell within a mesoscale snowband within the comma head of the cyclone with 7.5–10 cm h⁻¹ (3–4 in. h⁻¹) snow rates reported within the band and radar reflectivities exceeding 55 dBZ (Picca et al. 2014). Over the Northeast United States there are mesoscale snowbands within the comma head of ~85% of extratropical cyclones during the winter months (Novak et al. 2004). Well-defined primary bands have been identified as forming north of the surface cyclone center in the region of enhanced midlevel (~700 hPa) frontogenesis and reduced stability (Novak et al. 2004, 2010). For single band formation, Novak et al. (2010) found that development occurred along the northern edge of an upper-level potential vorticity (PV) hook, where there are likely either weakly stable or conditionally unstable conditions. Theoretical and modeling studies have shown that latent heat release is important in the formation and evolution of banded precipitation. Latent heat release on the warm side of a frontal boundary acts to concentrate vertical circulations into a narrow updraft or band (Emanuel 1985; Thorpe and Emanuel 1985). Novak et al. (2009) found that PV generation within the band from the latent heat release decreased midlevel heights northwest of the band and led to larger convergence, tightened the temperature gradient, and increased...
frontogenesis at this level. During band maturity, the stability gradually increased as conditional instability was released. Band dissipation occurred when midlevel frontogenesis weakened when a new diabatic PV anomaly formed a few hundred kilometers to the east of the band, thus reducing the midlevel flow deformation over the original band.

Snowbands can exhibit rapid transitions in ice habit and degree of riming due to the magnitude of vertical motions within different thermal environments. Stark et al. (2013) examined the microphysical evolution (including snow crystal habit and degree of riming) of two snowbands crossing Long Island using vertically pointing radar data and surface observations. Their study highlighted the rapid ascent on the warm frontogenetical side of the band, resulting in light-to-moderate riming and dendritic ice growth, while on the cold side of the band there was less riming and fewer dendrites (more plates). Colle et al. (2014) examined snow characteristics and environments centered around 12 cyclones over three winter seasons for developing and mature East Coast cyclones. Within some of the bands, they found that the heaviest riming occurred closest to the cyclone center with convective cells aloft and strong vertical motions around 800 hPa. The recent field campaign over the Midwest, the Profiling of Winter Storms (PLOWS; Market et al. 2012; Rauber et al. 2014), sampled 17 storms in the 2009/10 winter season and also found cells of vertical motion of 1–2 m s\(^{-1}\) aloft (Rosenow et al. 2014). Similarly, Kumjian et al. (2014) studied cloud-top-generating cells in Colorado and found that the updrafts could maintain supercooled liquid water in the presence of ice crystals.

Picca et al. (2014) used base (0.5°) radar reflectivity and dual-polarization products from the WSR-88D at Upton, New York (KOKX) (cf. Fig. 1b), to document the rapid transitions between snow and sleet as the snowband on 8–9 February moved northward across Long Island. The relatively high (~55 dBZ) reflectivity values around the time of the maximum intensity were attributed to large hydrometeors that likely formed through wet-growth processes within strong updrafts. The environment around the band cooled, but a narrow region of warmer temperatures (≥ 0°C) remained coincident with the band and was responsible for a persistent mixture of rimed snow and sleet. The band reflectivities appeared to weaken to 35 dBZ between 0300 and 0400 UTC 9 February, but surface reports of intense snowfall rates (4–8 cm h\(^{-1}\)) were maintained. The decreased reflectivity values were attributed to smaller cold-type hydrometeors after the environment aloft cooled. For the same storm, Griffin et al. (2014) identified additional features using dual-polarization radar, such as the downward excursion of the melting layer, which they hypothesized was either from melting and evaporative cooling or a localized updraft that produced larger rimed particles with higher fall speeds. They also noted several polarimetric artifacts, such as depolarization streaks indicative of electrification during the band’s maximum intensity.

While many studies have highlighted mesoscale snowbands within the comma head of extratropical
cyclones (e.g., Sanders and Bosart 1985; Wolfsberg et al. 1986; Nicosia and Grumm 1999; Novak et al. 2004; Evans and Jurewicz 2009), their focus has been on the structure and frontogenetical forcing associated with the band, not its detailed microphysical characteristics and thermodynamic budget. The blizzard of 8–9 February 2013 exhibited rapid changes in snow habit and degree of riming, which allows one to build upon previous work on the thermodynamical evolution and microphysics within the comma head of winter storms. The snowbands from these past studies over the Northeast occurred within an entirely subfreezing thermal environment. In addition, the two previous case studies of snowband microphysical observations completed by Stark et al. (2013) were not discussed in terms of the thermodynamic evolution of the bands related to observations nor had dual-polarization radar data available. A more comprehensive temperature budget and model sensitivity tests are needed to elucidate the role of various thermodynamic processes on the maintenance or destruction of the mixed-phase environment of the band.

The microphysical evolution can be compared with the previous knowledge of cyclone-relative dominant snow habits (Colle et al. 2014). Dual-polarization technology provides temporally consistent products to compare with sporadic in situ observations of snow habit and has been used to determine the precipitation type in winter storms (e.g., Trapp et al. 2001; Andrić et al. 2013; Thompson et al. 2014). Picca et al. (2014) provided some polarimetric observations of the 8–9 February 2013 snowband event, but they were not put in context with the full band structural and thermodynamic evolution. Griffin et al. (2014) provided more polarimetric and microphysical insight for the event, but the linkages to the thermodynamic evolution were limited, since the Rapid Refresh (RAP) model that they used failed to accurately capture the vertical temperature structure (i.e., warm layer) around the band.

The purpose of our study is to answer the following questions:

- What was the thermodynamic evolution of the band, and what were the dominant processes that led to the temperature changes at low levels?
- How did the thermodynamic evolution impact the microphysics and snowfall rates?

The rest of this paper is structured as follows. Section 2 discusses the datasets and methodology of this study. Section 3 provides an overview of the event from a large-scale perspective. Section 4 provides details on the microphysical evolution of the snowband based on observations. Section 5 highlights the thermodynamical evolution of the band and the environment within which it evolved. Section 6 examines the temperature evolution with analysis of air parcel trajectories, quantitative comparison of terms of the thermodynamic equation, and sensitivity tests. The study will conclude with a summary and discussion of future work.

2. Datasets and methodology

a. Observational datasets

The KOKX dual-polarization radar was used to observe the apparent intensity and location of the band. In addition to the reflectivity factor at horizontal polarization $Z_H$, the copolar correlation coefficient $\rho_{hv}$ and differential reflectivity $Z_{DR}$ were used to discern regions of liquid, frozen, and mixed-phase hydrometeors, as summarized by Kumjian (2013). The diversity of how the scattering of each particle contributes to the horizontal and vertical return signals is measured by $\rho_{hv}$. For example, spherical particles and uniform particles would result in $\rho_{hv} \sim 1$. Variability of hydrometeor type, shape, and/or orientations can decrease $\rho_{hv}$. Melting particles result in lower $\rho_{hv}$ values because the liquid water may accentuate irregular features. Differential reflectivity $Z_{DR}$ is the logarithmic ratio of the reflectivity factor at horizontal polarization to the reflectivity factor at vertical polarization and is sensitive to the shape, density, and orientation of particles; $Z_{DR}$ is larger for rain than most ice and dry snow, even for the same shape and orientation, because of the higher density of liquid water. The level-II radar data were converted to NetCDF format using the National Center for Atmospheric Research (NCAR) Research Applications Laboratory (RAL) radial radar software package, Radx (http://www.ral.ucar.edu/projects/titan/docs/radial_formats/radx.html), and displayed using NCAR Command Language (NCL) version 6.1.2 (UCAR/NCAR/CISL/VETS 2013).

Ground observations of snow habit, degree of riming, snow water equivalent, and snowfall rate were taken at Stony Brook University (SBNY in Fig. 1b), about 20 km west of KOKX using the methodology presented in Colle et al. (2014) and a stereomicroscope in a cold shed. Satellite data were obtained from the Geostationary Operational Environmental Satellite-13 (GOES-13) via the National Oceanic and Atmospheric Administration (NOAA) National Climatic Data Center (NCDC) Comprehensive Large Array Data Stewardship System (CLASS) and displayed using the Integrated Data Viewer (IDV; Murray et al. 2003). The 13-km RAP (Benjamin et al. 2009) hourly gridded analyses, conventional surface observations, and radiosonde observations
from KOKX and Chatham, Massachusetts (KCHH in Fig. 1b), were used for model verification on the synoptic scale discussed in section 2b.

b. Model setup

The Weather Research and Forecasting (WRF) Model, version 3.4.1 (Skamarock et al. 2008), was used for a 48-h simulation initialized ~17 h prior to band formation at 0000 UTC 8 February 2013. This lead time is within the range (15–21 h) employed by Novak and Colle (2012) for three East Coast banding cases. The simulation was run using the 6-hourly 0.5° Global Forecast System (GFS) data from the 0000 UTC 8 February forecast cycle as the initial and boundary conditions with four one-way nested domains from 36-km down to 1.33-km horizontal grid spacing centered over the band location (Fig. 2c). There were 40 levels in the vertical, with the model top set to 100 hPa. The $\frac{1}{10}$ sea surface temperature data from 0000 UTC 8 February was used to initialize the simulation and obtained from the National Centers for Environmental Prediction (NCEP). Snow cover was initialized from the 0.5° GFS 0000 UTC 8 February analysis file. The Mellor–Yamada–Janjić planetary boundary layer (PBL) scheme (Janjić 1994) and the unified Noah land surface model (Noah LSM; Tewari et al. 2004) were used. The Betts–Miller–Janjić cumulus parameterization (Betts 1986) was applied in the 36- and 12-km domains only, while the Thompson microphysical parameterization scheme was used in all domains (Thompson et al. 2004, 2008). These physics options followed from previous work on simulating snowbands in the region using the WRF Model (Stark 2012), with the nonspherical ice assumption within the Thompson microphyscal parameterization scheme shown to produce realistic simulated reflectivity structures. The 1.33-km domain was output at a temporal frequency of 15 min in order to complete the microphysical and thermodynamic analyses found in sections 4 and 5.

Additional runs were conducted to obtain the most representative simulation by varying the initial and boundary conditions between the 0000, 0600, and 1200 UTC 8 February 2013 NCEP North American Mesoscale Forecast System (NAM), GFS analyses and forecasts, and 3-hourly RAP analysis gridded data while keeping the physics options constant. The results were compared with available observational data (e.g., KOKX radar reflectivity, KOKX and KCHH soundings, various ASOS station data, etc.). Most simulations resulted in larger position and magnitude errors of the snowband than using the GFS forecast data starting at 0000 UTC 8 February (not shown), and some of this is linked to larger position and magnitude errors of the cyclone southeast of...
Long Island. Thus, the 0000 UTC 8 February 2013 GFS forecast initial and boundary conditions were used for the WRF simulation discussed in subsequent sections. This simulation exhibited some error, including with the structure of the snowband, but is sufficiently accurate in the band processes to serve the goals of this study.

3. Large-scale overview

The 8–9 February 2013 cyclone developed when a 500-hPa shortwave trough over the Carolinas strengthened and phased with a larger-amplitude northern stream trough that originated from the Great Lakes region. The original cyclone weakened over the Great Lakes, while cyclogenesis occurred just east of the mid-Atlantic coast as in Miller type-B (Miller 1946) (Fig. 2a). The low off of the East Coast deepened approximately 29 hPa in 24 h during 0600–0600 UTC 8–9 February according to the NOAA/Weather Prediction Center (WPC) 3-hourly surface analyses. The coastal low developed as the poleward exit region of a 300-hPa jet core ≥110 kt (1 kt = 0.5144 m s⁻¹) interacted with a low-level baroclinic zone from 0600 UTC 8 February to 0000 UTC 9 February. The surface cyclone continued to be situated under the equatorward entrance region of a 300-hPa jet core (≥130 kt) as it deepened further (Fig. 2c), and the cyclone occluded as it entered the Gulf of Maine (not shown).

The WRF simulation was verified using available observations and gridded analysis data from the RAP. The WRF realistically simulated the structure and evolution of the midlevel shortwave trough and the interaction with the northern stream trough. This is demonstrated by the similar storm structure between the observed infrared imagery (Figs. 3a,c,e) and the simulated cloud-top temperatures (Figs. 3b,d,f) at 1800 UTC 8 February, 0000 UTC 9 February, and 0600 UTC 9 February. The simulation reproduced the central pressure of the cyclone to within ±2 hPa of the WPC analyzed mean sea level pressure (MSLP) valid at the same time during the critical 12-h time period relevant to the snowband (1800 UTC 8 February–0600 UTC 9 February), whereas the RAP analyses were 2–8 hPa too weak throughout the same time period (not shown). The simulated cyclone track was similar to the observed with the exception of a ~200 km southward shift around 0600 UTC 9 February (forecast hour 33) (Figs. 2b,c).

4. Observed and simulated snowband evolution

a. Phase 1: North–south-oriented mixed-phase transition zone

Picca et al. (2014) and Griffin et al. (2014) used observations from a dual-polarization radar to highlight some of the evolution of this event. We expanded on their analysis of the band life cycle separated into three phases defined by the observed changes in the radar and surface observations. We define phase 1 from 2000 to 2300 UTC 8 February. Phase 1 was associated with a west–east-oriented mixed-phase transition zone that separated the snow to the north of Long Island with the rain to the south. This was characterized by $p_{hv} < 0.95$ and $Z_{DR} \sim 1.0 \text{ dB}$ from rain and mixed-phase precipitation along the southern coast and to the south of Long Island, while the snowband along central Long Island and to the north was characterized by $p_{hv} \sim 1$ and $Z_{DR} \sim 0 \text{ dB}$ (Figs. 4a,d,g), which is associated with hydrometeors with low effective density, such as aggregated snowflakes with little riming. A northwest–southeast cross section (A to A’ in Fig. 4i) at 2129 UTC 8 February shows the melting signature located approximately 20 km southeast of SBNY, given by low values of $p_{hv}$ (<0.90) and large $Z_{DR}$ values (>1.5 dB) below ~1.5 km MSL, which suggests a mixture of hydrometeors, mainly snow melting into rain (Figs. 5a,d,g). Also noted is the height of the 25-dBZ contour to 4.5 km near the snowband. Surface observations taken around 2000 UTC 8 February at SBNY near the western edge of the snowband included large aggregates of mainly colder-type crystals including sideplanes and plates (Fig. 6), with an average snow-to-liquid ratio (SLR) of 10:1 and snowfall rates of 4.0–8.5 cm h⁻¹ (1.6–3.3 in. h⁻¹) within this phase.

The WRF output valid at 2100 UTC 8 February (21 h) simulated an east–west-oriented precipitation band north of the 850-hPa low center (Fig. 7a). A northwest–southeast cross section (B to B’ in Fig. 7a) shows that the WRF realistically simulated the depth of the higher reflectivity values (~40 dBZ) to around 800 hPa (~1.9 km MSL) near SBNY and the simulated cloud depth extended to just above 500 hPa (~5.44 km MSL) (Fig. 7c). The simulated microphysical output reveals a broad area of snow-mixing ratios exceeding 1.5 g kg⁻¹ above 700 hPa coincident with the band (Fig. 7c), as well as snow to the north of SBNY and rain-mixing ratios exceeding 0.35 g kg⁻¹ to the south (Fig. 7e). There is a graupel maximum of 0.30 g kg⁻¹ from the surface to 950 hPa above and ~20 km south of Long Island within the colder lower-level air, indicating either heavy rime resulting in snow converting to graupel or refreezing taking place in the model. The Thompson scheme cannot produce sleet, so rainwater refreezes to graupel (Thompson et al. 2008). Model-derived snow depth and liquid equivalent are output from the Noah LSM, which inputs the precipitation rate and fraction of frozen precipitation (FOFP) from the Thompson scheme. The Noah LSM classifies all precipitation as snow if FOFP
FIG. 3. Geostationary Operational Environmental Satellite-13 (GOES-13) infrared satellite brightness temperature (K, shaded) and 13-km RAP analysis of 500-hPa geopotential height (solid black contoured every 60 m) at (a) 1800 UTC 8 Feb, (c) 0000 UTC 9 Feb, and (e) 0600 UTC 9 Feb 2013. WRF simulated cloud-top temperature (°C, shaded according to scale) and 500-hPa geopotential height contoured every 60 m at (b) 1800 UTC 8 Feb, (d) 0000 UTC 9 Feb, and (f) 0600 UTC 9 Feb 2013.
is >0.5°C or as freezing rain if the air temperature is >0°C but the ground temperature is <0°C. The Noah LSM also accounts for melting and compaction of existing snow with a time-varying and temperature-dependent snow density used in calculating the snow depth (Ek et al. 2003; Tewari et al. 2004). The simulated snow depth values were used to calculate the snowfall rates within each phase. The simulated SLR values were
calculated using the hourly change in liquid equivalent snow on the ground and the snow depth. During phase 1, simulated snowfall rates at SBNY were 5.1 cm h\(^{-1}\) (2.0 in. h\(^{-1}\)). Table 1 provides a comparison between the observed and simulated snow depth and liquid equivalent within each phase and SLR values averaged throughout each phase. Each phase included at least three observed measurements. Simulated snowfall
amounts during phase 1 were 5.0 cm (2.0 in.) less than observed, but liquid equivalent values were 0.5 cm (0.2 in.) larger than observed.

b. Phase 2: Heavy riming and extreme hydrometeor diversity

The snowband evolved from a broad region of smaller bands with enhanced reflectivities (Fig. 4a) into a primary snowband that pivoted to a north–south orientation as the cyclone moved east (Fig. 4b). Phase 2 (2300 UTC 8 February–0200 UTC 9 February) was a period of heavy riming and extreme hydrometeor diversity and was the peak of intensity of the snowband extending from south of Long Island northward into central Connecticut. Reflectivity reached 57.5 dBZ at 0042 UTC in a region of $\rho_{hv} \sim 0.85$ and $Z_{DR} > 3$ dB (Figs. 4b,e,h). This is near the time lightning was observed $\sim 30$ km east of KOKX (D. Stark, NOAA/NWS Upton, NY, 2013, personal communication), which is consistent with electrification implied by the depolarization streaks observed around this time reported in Griffin et al. (2014). A cross section shows $Z_H \sim 40$ dBZ, $\rho_{hv} < 0.90$, and $Z_{DR} \sim 1.5$ dB in the vicinity of SBNY extending to a height of 1.5 km, which is an indication of the hydrometeor diversity at low levels of the atmosphere (Figs. 5b,e,h). Surface microphysical observations at SBNY indicate that phase 2 coincided with heavy riming and sleet (Fig. 6), an average SLR of 6.8:1, and snowfall rates of 1.5–7.6 cm h$^{-1}$ (0.6–3.0 in. h$^{-1}$). During this phase, 10%–20% of observed hydrometeors were classified as miscellaneous ice. Picca et al. (2014) provided an image of ice crystals that experienced heavy riming (their Fig. 8a) possibly due to wet-growth processes within the strong updrafts within a moisture-rich environment with temperatures near 0°C, which supports the same hypothesis proposed by Griffin et al. (2014).

The simulated band at 0100 UTC (25 h) is displaced $\sim 30$ km to the west of the observed band (Figs. 8a and 4b), and the WRF produced another band of higher reflectivities just southeast of Long Island that was not observed. The simulated snowband in the vicinity of SBNY along section B–B’ shows a core of higher reflectivities coincident with $\sim 3$ m s$^{-1}$ upward-directed circulation vectors (Figs. 8c,d), and a horizontal width of the simulated band (<100 km) comparable to what was observed (Fig. 5b). A cross section (B–B’ in Fig. 8a) shows an amplifying thermal wave in the saturation equivalent potential temperature ($\theta_{es}$) field, with the warm axis coincident with an isolated pocket of air $>0$°C from 900 to 700 hPa located above SBNY (Figs. 8b,d), while it was $<0$°C just southeast of Long Island to support graupel-mixing ratios of $\sim 0.25$ g kg$^{-1}$ (Fig. 8b). Meanwhile, farther south over the ocean the temperatures ranged from 0° to 8°C coincident with rainwater-mixing ratio values $\geq 0.25$ g kg$^{-1}$. The heavy snow over central Long Island fell into the 200-hPa-thick layer of air $>0$°C, with some melting to produce
rainwater-mixing ratios <0.25 g kg\(^{-1}\) (Fig. 8e). Over SBNY, the rain refroze in a shallow layer of air <0\(^\circ\)C air between the surface and 900 hPa, producing graupel-mixing ratios ≥0.30 g kg\(^{-1}\) collocated with snow-mixing ratios ≥1.25 g kg\(^{-1}\) (Figs. 8d,e). During phase 2, simulated snowfall rates at SBNY were 5.2 cm h\(^{-1}\) (2.0 in. h\(^{-1}\)) and simulated snowfall amounts were 5.3 cm (2.1 in.) larger than observations (Table 1).
TABLE 1. A summary of observed and simulated snow depth, liquid equivalent, and snow-to-liquid ratio (SLR) values at SBNY for each of the three phases of the snowband: 1) 2000–2300 UTC 8 Feb, 2) 2300 UTC 8 Feb–0200 UTC 9 Feb, and 3) 0200–0800 UTC 9 Feb. Snow depth values are not cumulative and represent the value from the beginning to end of each phase. SLR values are given in fractional form.

<table>
<thead>
<tr>
<th>Phase</th>
<th>Observed snow depth (cm) [in.]</th>
<th>WRF simulated snow depth (cm) [in.]</th>
<th>Observed liquid equivalent (cm) [in.]</th>
<th>WRF simulated liquid equivalent (cm) [in.]</th>
<th>Observed SLR</th>
<th>WRF simulated SLR</th>
</tr>
</thead>
<tbody>
<tr>
<td>Phase 1</td>
<td>20.3 [8.0]</td>
<td>15.3 [6.0]</td>
<td>2.0 [0.8]</td>
<td>2.5 [1.0]</td>
<td>10.0</td>
<td>6.1</td>
</tr>
<tr>
<td>Phase 2</td>
<td>10.2 [4.0]</td>
<td>15.5 [6.1]</td>
<td>1.5 [0.6]</td>
<td>2.3 [0.9]</td>
<td>6.8</td>
<td>6.7</td>
</tr>
<tr>
<td>Phase 3</td>
<td>40.6 [16.0]</td>
<td>12.2 [4.8]</td>
<td>4.3 [1.7]</td>
<td>1.5 [0.6]</td>
<td>9.4</td>
<td>8.1</td>
</tr>
</tbody>
</table>

**c. Phase 3: Transition to colder, less-dense snow aggregates**

The large hydrometeor diversity and +50-dBZ reflectivity signature abruptly ceased around 0230 UTC 9 February and was replaced with more homogeneous snow aggregates and a signature more consistent with snowbands that have been studied in the past (e.g., Novak et al. 2008, 2009, 2010). The reflectivity values decreased to ~30 dBZ within 1 h and the snowband persisted until approximately 0800 UTC 9 February. This change in observed microphysics constitutes phase 3 (0200–0800 UTC 9 February). Figures 4c,f,i show the persistent band at 0340 UTC 9 February with the $\rho_w$ and $Z_{DR}$ fields both showing more uniform values along western Long Island into south-central Connecticut, but given the relatively large coverage of reflectivity values >30 dBZ, heavy snow was still falling across the region. The cross section shows that this transition had occurred throughout the lower atmosphere (Figs. 5c,f,i). Surface observations of ice habit and riming during phase 3 at SBNY show a transition to colder-type crystals with less riming (Fig. 6), while the heavy snow persisted and the SLR increased again to 9.4:1 with snowfall rates of 6.5–6.7 cm h$^{-1}$ (2.6 in. h$^{-1}$).

The simulated snowband at 0400 UTC (28h) showed a similar marked decrease in reflectivity in the horizontal (Fig. 9a) and cross section B–B′ (Fig. 9a), which was greater than observed by <5 dBZ. Meanwhile, the simulated precipitation to the east of the snowband over southeastern Massachusetts was stronger than observed. The vertical profile of temperature over SBNY also cooled by as much as 6°C between 1 and 2 km. Section 6 will examine the reasons for the ~500-m-deep layer above 0°C between 1 and 2 km (900–700 hPa) during the second phase and its abrupt cooling during phase 3. The observed transition to colder, less-dense snow aggregates (Figs. 4c and 5c) is comparable to the simulated transition to all snow and the cessation of rain and graupel (Fig. 9e) and is thus evident in the simulated reflectivity values around 0400 UTC (Figs. 9a,c). Specifically, the rain- and graupel-mixing ratios and cloud water decreased to <0.10 g kg$^{-1}$ and the snow decreased to between 0.50 and 0.90 g kg$^{-1}$ (Figs. 9b,d,e). During phase 3, the simulated snowfall rate at SBNY was 2.4 cm h$^{-1}$ (0.9 in. h$^{-1}$) and the simulated change in snow depth was 36.3 cm (11.2 in.) less than observations (Table 1). This discrepancy is due to the simulated band dissipating ~3 h sooner than the observed band within phase 3 at 0500 UTC 9 February. As such, the simulated snow depth and liquid equivalent amounts were comparable to what was observed during both phases 1 and 2 and any analysis of the simulation within phase 3 will be conducted prior to 0500 UTC 9 February.

**5. Thermodynamic evolution**

**a. Phase 1: North–south-oriented mixed-phase transition zone**

Phase 1 (2000–2300 UTC 8 February) occurred when the simulated band was situated along and parallel to Long Island (Figs. 7a,b) near where 850-hPa frontogenesis was approximately 0.1 K (100 km)$^{-1}$ h$^{-1}$ (Fig. 7f). The band was located in a stable environment above 900 hPa. South and east of the band were multiple narrow snowbands (Figs. 7a,c), but their discussion is beyond the scope of this study.

The 900-hPa thermal structure at 2100 UTC 9 February showed evidence of cold air to the north of the surface cyclone center with easterly winds north of Long Island into southern New England (Fig. 7b). During this phase there was warm advection over southern Long Island and confluence of the flow due to 900-hPa easterly winds decelerating from 75 to 50 kt over ~50 km. At 2100 UTC, a north–south temperature gradient was simulated at 900 hPa with simulated mixing ratios of mostly snow to the north and rain to the south of Long Island, both of which exceed 0.25 g kg$^{-1}$ (Fig. 7b). A cross section from B to B′ taken through the different thermal environments near SBNY shows that a subfreezing layer extended from 800 to 500 hPa, an above-freezing layer from 925 to 800 hPa, a low-level subfreezing layer from 975 to 925 hPa, and a near-surface shallow above-freezing layer extended from the surface to ~975 hPa (Fig. 7d).
b. Phase 2: Heavy riming and extreme hydrometeor diversity

The band exhibited the greatest reflectivity (57.5 dBZ) in the proximity of SBNY during phase 2 (2300 UTC 8 February–0200 UTC 9 February) (Fig. 4b). The area of maximum 850-hPa frontogenesis of 0.8 K (100 km)\(^{-1}\) h\(^{-1}\) was ~60 km southeast of the north–south-oriented snowband (Fig. 8f). The band coincided with a region of weak conditional stability at 0100 UTC 9 February near 600 hPa (Fig. 8d), which is similar to the stability of a snowband studied in Novak et al. (2009).

During phase 2, the 900-hPa temperatures decreased over a fairly broad area around Long Island as winds at this level started to back to the northeast (Fig. 8b). However, there was a narrow north–south-oriented swath of 900-hPa air >0°C maintained over Long Island that collocated with the snowband >45 dBZ (Fig. 8b). The depth of the air >0°C above SBNY extends from 900 to 700 hPa, as shown in cross section B–B’ (Fig. 8a), while the temperature of the environments to the northwest and southeast of Long Island were <0°C (Figs. 8b,d). At this time there was an amplified signal in the \(\theta_{e}\) field, a strong vertical circulation associated with the snowband, and the
pocket of air above 0°C was located below the midlevel subsidence (~750 hPa).

c. Phase 3: Transition to colder, less-dense snow aggregates

By 0300 UTC 9 February (start of phase 3) the 850-hPa frontogenesis at 0400 UTC weakened to 0.5 K (100 km)^{-1} h^{-1} and the snowband persisted in the simulation but decreased in intensity to 30–25 dBZ (Fig. 9a). There was still ascent in the band environment around 750 hPa (Figs. 9c,d), but the immediate band environment exhibited increased stability (Fig. 9d). The strongest ascent over SBNY during this phase decreased to <40 cm s^{-1} and was confined to regions west of SBNY above 2 km (Figs. 9c,d). The 900-hPa temperature around Long Island continued to decrease as winds backed more to the north-northeast (Fig. 9b). The vertical profile of temperature over SBNY also cooled by as much as 6°C between 1 and 2 km compared with the temperature of the band environment during the previous phase (Figs. 8d and 9d). The next section will examine the reason for the ~500-m-deep layer above 0°C
between 1 and 2 km (900–700 hPa) during phase 2 and what led to its abrupt cooling.

6. Examination of thermodynamic environment
a. Trajectory analysis

To diagnose the thermodynamic processes responsible for the warm layer along the band between 900 and 700 hPa during 2300–0200 UTC 8–9 February, backward trajectories were calculated along Long Island from three points starting at 0200 UTC (Fig. 10a). Two trajectories were located to the west and east of the snowband, and one within the temperatures above 0°C at 900 hPa within the band. Although only three trajectories are shown, tens of others were launched in order to find a set of representative trajectories. Backward trajectories using 15-min output from the model were calculated using Read/Interpolate/Plot (RIP), version 4.6, (Stoelinga 2009). Previous work by Novak et al. (2009) used trajectories terminating in a snowband environment to discern the role of the dry slot in destabilizing the environment for snowband development and found that the diabatic heating of the snowband itself acted to stabilize the immediate band environment. A recent study by Fuhrmann and Konrad (2013) employed trajectory analysis to understand the large-scale evolution of air parcels pertaining to cool season extratropical cyclones. Their results indicated that diabatic warming and cooling from precipitation processes contribute significantly to the vertical temperature profile intercepted by trajectories.

Figure 10a shows the potential temperature evolution for the three trajectories ending at 0200 UTC 9 February. The eastern trajectory likely followed the cold conveyor belt as it remained at a nearly constant height (~500 m MSL) and it is much cooler than the other trajectories by approximately 10 K (Fig. 10b). The western and central trajectories followed similar paths, originating from the northeast of the developing cyclone and rotating counterclockwise while undergoing ascent and descent of 0.1 and ~0.25 m s⁻¹ for the central trajectory versus 0.03 and ~0.19 m s⁻¹ for the western trajectory before arriving at their final locations at 0200 UTC; however, the central trajectory is about 4 K warmer than the western trajectory at the same terminal height. The central trajectory underwent ascent between 2300 and 0100 UTC and during that time the water vapor–mixing ratio decreased while the snow and cloud water–mixing ratios increased presumably from deposition and condensation, resulting in a 4-K warming of the parcel’s potential temperature (Figs. 10b,c). The central trajectory underwent descent (~0.25 m s⁻¹) between 0100 and 0200 UTC coincident and just to the west of the snowband (Fig. 8a), and entered an environment where melting and evaporation likely occurred with a parcel relative humidity of 84% and temperature of 2.5°C by 0130 UTC (not shown). Both melting and evaporative processes contributed roughly 7 K of cooling by 0200 UTC of the environment along the course.

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of both the central and western trajectories. Overall, the final potential temperature of the environment of the central trajectory was 4 K warmer than that of the western trajectory because the latent heating allowed for the temperature to increase more before both environments diabatically cooled by $7 \text{K}$.

b. Evaluation of the thermodynamic equation within the band environment

Another way to quantify the thermodynamic changes is to calculate a potential temperature budget as in Eq. (1):

$$\frac{\partial \theta}{\partial t} = - \left( u \frac{\partial \theta}{\partial x} + v \frac{\partial \theta}{\partial y} \right) - \omega \left( \frac{\partial \theta}{\partial P} \right) + \text{Diabatic}. \quad (1)$$

Novak et al. (2009) used this method [their Eq. (2)] to analyze the static stability around the snowband. Our study looks at the spatial structure of the terms during two representative times during the evolution of the band during phases 2 and 3. The left-hand term is the total change over a 15-min interval. The first term on the right-hand side is horizontal advection. The middle term is the vertical advection, which also accounts for adiabatic changes. The term labeled as “Diabatic” is the temperature tendency output from the microphysical parameterization scheme, the radiative schemes, and the planetary boundary layer scheme. The results for the calculation are provided in the following cross section from C to C’ (Fig. 10a) in Fig. 11. Figure 11 shows the contribution from each term at 0030 UTC 9 February, which is just before the $\theta$ pattern amplified by 0100 UTC 9 February. This amplified pattern in the isentropes from moist ascent and compensating subsidence has been shown in another snowband environment in Novak et al. (2008, see their Figs. 12b,d).

The tendency terms are first discussed for what led to the ridge in the isentropes near the eastern edge of the snowband (30 km southeast of SBNY). During this time there is a positive contribution from the horizontal advection term within the environment between 950 and 800 hPa southeast of SBNY ($10.0 \times 10^{-4} \text{K} \text{s}^{-1}$), but the
negative contribution from the vertical advection term is approximately 2.5 times larger \((-23.4 \times 10^{-3}\) K s\(^{-1}\)), likely resulting from strong ascent (Figs. 11b,c). The ascent at this time was frontogenetically forced and extends from 800 to 650 hPa (Fig. 11c). Condensation, deposition, and freezing are likely occurring within the environment to allow the diabatic term to contribute a maximum of \(9.8 \times 10^{-3}\) K s\(^{-1}\) (Fig. 11d). This positive contribution occurs within a much smaller areal extent near the ridge in the isentropes around 650 hPa than both the horizontal and vertical advection terms. The result is a net cooling of \(1.9 \times 10^{-3}\) K s\(^{-1}\), which is attributed mainly to vertical advection (Fig. 11a).

The temperature tendency is now applied to the developing low-level warm anomaly over SBNY, as denoted by an apparent thermal trough in the isentropes. The contribution from subsidence \((5.2 \times 10^{-3}\) K s\(^{-1}\)) covers a larger area and is greater than that of diabatic cooling effects \((-1.4 \times 10^{-3}\) K s\(^{-1}\)), such as the melting of snow. The horizontal advection term in this region is contributing to cooling \((-3.3 \times 10^{-3}\) K s\(^{-1}\)) from 975 to 900 hPa. The net result is a warming of \(2.2 \times 10^{-3}\) K s\(^{-1}\), which is coincident with the warm layer that favored melting and wet-growth processes of hydrometeors discussed in previous sections. To reconcile the results of the evaluation of the thermodynamic equation with those of the trajectory analysis, it is evident that both diabatic effects and subsidence maintained the layer of temperatures above 0°C. The environment warmed by moist air rising through the band that released latent heat via condensation, deposition, and freezing and then rapidly subsided via the descending branch of the frontogenetical circulation along the western edge of the snowband.

Analyzing the contributions from each term of the equation for times during phase 3 (0400 UTC 9 February in Fig. 12) of the snowband life cycle reveals that the vertical advection term weakened to \(-8.6 \times 10^{-3}\) K s\(^{-1}\) within the thermal ridge to the east of SBNY and \(4.4 \times 10^{-3}\) K s\(^{-1}\) in the thermal trough in the isentropes (Fig. 12c). This is likely due to decreased midlevel frontogenesis, which allowed the contributions from diabatic cooling from melting and evaporation \((-2.1 \times 10^{-3}\) K s\(^{-1}\)) and horizontal cold advection \((-3.1 \times 10^{-3}\) K s\(^{-1}\)) to dominate and result in the overall cooling of the column \((-2.1 \times 10^{-3}\) K s\(^{-1}\)), relaxing the amplified signal in the \(\theta\) field. This supports the cessation of mixed-phase processes and the transition to less dense, albeit still heavy snow.

c. Sensitivity tests to phase changes within the band

To better understand the role of diabatic processes in the evolution of the thermal environment and resulting
microphysics, two sensitivity experiments were conducted and compared with the control run (CTRL) discussed in the previous sections. The first simulation was to identify the role of latent cooling that occurred within the band environment. A simulation was run in which the temperature tendency contributions from evaporation, sublimation, and melting after 2000 UTC 8 February (20h) were not included [no latent cooling (NOLC)]. The hypothesis is that the snowband environment will be warmer and produce different microphysical results because there will be no evaporative cooling or cooling by melting acting to erode the above-freezing layer between 1 and 2 km MSL. Studies have highlighted the importance of melting on transitioning precipitation type (e.g., Kain et al. 2000; Lackmann et al. 2002; Market et al. 2006) but that process is not isolated from sublimation and evaporative cooling in this paper in order to study the general aggregate effects in the snowband environment.

The second experimental simulation was similar to the first except that the processes of freezing, deposition, and condensation were isolated. The contributions to the temperature tendency calculation from freezing (including riming), deposition, and condensation were turned off after 2000 UTC so that the thermal environment would only respond to advection, diabatic cooling, and adiabatic vertical motion [no latent heating (NOLH)]. The snowband environment is expected to be cooler and produce less mixed-phase hydrometeors, especially during phase 2 between 2300 and 0200 UTC.

1) NO LATENT COOLING SIMULATION

The main difference between the NOLC simulation and the CTRL simulation was the persistence and magnitude of the layer of air above 0°C between 900 and 700 hPa (0.5–2.0 km MSL). At 2100 UTC 8 February (1 h into experiment) the cross section B–B’ (Figs. 7a, 8a, and 9a) for the NOLC simulation shows only slight (∼2°C) temperature differences from the CTRL simulation. The NOLC simulation shows only slight (∼2°C) temperature differences from the CTRL simulation. A distinct northwest–southeast transition from snow to rain resembles the CTRL simulation (Fig. 13a). During phase 2 (Figs. 13c,d), the magnitude of the above-freezing layer is ∼4°C larger in NOLC than CTRL from 950 to 750 hPa. This environment supports a deeper layer of simulated rainwater-mixing ratios reaching the surface southeast of SBNY. The layer above 0°C persists through 0400 UTC, which altered phase 3 by likely melting most of the simulated snow resulting in less concentrations (0.15 g kg⁻¹) than the CTRL simulation (0.50 g kg⁻¹) reaching the surface at SBNY (Figs. 13e,f and 9e). The NOLC experiment provided evidence that the warm layer that had allowed for hydrometeor diversity was partly removed by the combination of cooling processes of melting, evaporation, and sublimation. An additional sensitivity test that removed the contribution just from evaporation more closely resembled the CTRL simulation (not shown), which suggests that melting was the most important diabatic process contributing to cooling the band environment.

2) NO LATENT HEATING SIMULATION

Figure 14 shows the effect of the warming diabatic processes including freezing, condensation, and deposition in the NOLH simulation on the evolution of the thermodynamic environment in the vicinity of the simulated band. Phase 1 is similar to that of the CTRL simulation with a northwest–southeast-oriented transition zone from snow to rain (Figs. 14a,b). Phase 2 in the NOLH simulation lacks any amplified pattern in the θ_e field, since there is little midlevel frontogenesis, which corresponds with the idea that there is a large diabatic component driving frontogenesis (not shown). The environment of the entire domain is much cooler than that of the CTRL simulation and, most importantly, the entire environment aloft extending ∼40 km south of SBNY is cold enough to support snow (Figs. 14c,d). This environment continues to support snow throughout the beginning of phase 3, but the lack of strong vertical motion from the more stable environment results in less snow and the cessation of the band earlier than the CTRL simulation. The diabatic processes associated with the band itself were critical to the development and maintenance of the band, which supports the conclusions of Novak et al. (2009).

Given the results from the NOLH and the NOLC experiments, a combination of the heating within the updraft just upwind of the observed snowband, the weaker compensating downward motion from weaker ascent, and cooling by melting altered the environment of the snowband near SBNY. Diabatic heating was important in maintaining the thermal gradient that drove the ageostrophic frontogenetical vertical circulation that led to a layer above 0°C from near the surface to ∼700 hPa. This thermal environment led to hydrometeors undergoing mixed-phase microphysical growth processes that ultimately resulted in diverse ground observations of snow habits and ice particles. Diabatic cooling was important in the evolution of the band from a mixed-phased environment to one with a vertical temperature environment over SBNY that supported all snow by phase 3. The importance of diabatic effects to changing precipitation type addressed in this study agrees with previous studies (e.g., Kain et al. 2000; Lackmann et al. 2002). The results support that the simulated band occurrence, intensity, and precipitation type were very sensitive to diabatic effects.
7. Conclusions

The Northeast U.S. snowstorm of 8–9 February 2013 produced an intense snowband that exhibited reflectivity values that were around 20 dBZ higher than has been documented with previous snowband research (e.g., Novak et al. 2004, 2009, 2010; Stark et al. 2013). The dual-polarization radar observations provided insight into the general identification of hydrometeors and mixed-phase transition zones that were verified using ground observations of snow habit, degree of riming, and SLR at Stony Brook University (SBNY). Building
upon Picca et al. (2014) and Griffin et al. (2014), the WRF Model was used to simulate the event to determine 1) the evolution of forcing and stability of the snowband, 2) the thermodynamic evolution of the band, and 3) how the evolution impacted the microphysics and snowfall rates.

The snowband occurred in three distinct phases. Phase 1 was classified by heavy snow with rates of 4.0–8.5 cm s\(^{-1}\) (1.6–3.3 in. h\(^{-1}\)) to the north of a north–south transition zone that was apparent across Long Island, New York. The band developed in a region of weak 850-hPa frontogenesis \(\sim 0.1\) K (100 km\(^{-1}\)) h\(^{-1}\) and in a
region of weak stability. At 2100 UTC during this phase, a distinct northwest–southeast separation between subfreezing and above-freezing air was apparent throughout the low to midlevels in the band environment.

Phase 2 was classified by the largest hydrometeor diversity and was the period during which the highest base reflectivity value of 57.5 dBZ was measured coincident with snow falling at the ground. The band was located near a region of strong frontogenesis [$>0.8 \text{ K (100 km)}^{-1} \text{ h}^{-1}$] in a weakly stable environment. During this phase, the SLR decreased from 13.1 to 4.1 and the snowfall rates decreased to 1.5–7.6 cm h$^{-1}$ (0.6–3.0 in. h$^{-1}$), including increased percentages of observed sleet and unidentifiable ice. Further investigation determined that an approximately 200-hPa layer of temperatures above 0°C was collocated with the snowband, just downwind of a strong frontogenetically induced updraft. The strong updraft led to high snow-mixing ratios aloft that may have never fully melted while descending through the warm layer as well as hydrometeors possibly growing via wet-growth processes in the $\sim$0°C environment.

During phase 3 the environment cooled enough to support snow throughout the snowband environment. This was in part due to the diabatic cooling effects of melting snow into the layer above 0°C which, coupled with the weakened vertical motion from decreased midlevel frontogenesis [$<0.5 \text{ K (100 km)}^{-1} \text{ h}^{-1}$], was able to cool the column over SBNY. Observed SLR increased to 9.6:1 with snowfall rates of 6.5–6.7 cm h$^{-1}$ (2.6 in. h$^{-1}$).

This study discussed the thermodynamic evolution of the band, which showed that diabatic processes, especially condensational heating and cooling by melting, affect the evolution of the band’s observed microphysics, but vertical advections induced by a strong frontogenetical circulation are also important. This case is interesting because the snowband formed in an environment with low to midlevel mixed-phase processes that were dynamic with time. The trajectory analysis and evaluation of the thermodynamic equation provided evidence that the frontogenetically enhanced vertical motion during the band’s most intense phase 2 was important because without it, the environment may have cooled a lot sooner to support an all-snow event instead of the complex hydrometeors that were observed during that time period. Further evidence was provided by the sensitivity tests that showed that latent heating was critical to the maintenance of an environment of decreased stability and narrow updrafts, which agreed with the findings of Novak et al. (2009). In addition to the large-scale horizontal temperature advection, diabatic cooling was important to ultimately cool the band environment to support less-dense snow aggregates. The diabatic effects occurring within the band environment were shown to affect the simulated band occurrence, intensity, and precipitation type, which ultimately changes the storm total snowfall amounts. Future work focusing on the relative magnitudes of adiabatic expansion/compression versus diabatic processes, which are sometimes assumed to equally oppose one another, for a greater number of snowband cases would be enlightening. Additional modeling work could be done to explore the effects of using different microphysical parameterization schemes as well as other model physics variations for this case and other cases on the structure of simulated bands.

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