Tornadoes in the New York Metropolitan Region: Climatology and Multiscale Analysis of Two Events

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ABSTRACT

This paper describes the climatology of tornadoes around New York City (NYC) and Long Island (LI), New York, and the structural evolution of two tornadic events that affected NYC on 8 August 2007 and 16 September 2010. Nearly half (18 of 34 events from 1950 to 2010) of NYC–LI tornadoes developed between 0500 and 1300 EDT, and August is the peak tornado month as compared to July for most of the northeast United States. A spatial composite highlights the approaching midlevel trough, moderate most unstable convective available potential energy (MUCAPE), and frontogenesis along a low-level baroclinic zone. Shortly before the early morning tornadoes on 8 August 2007, a mesoscale convective system intensified in the lee of the Appalachians in a region of low-level frontogenesis and moderate MUCAPE (~1500 J kg^{-1}). Warm advection at low levels and evaporative cooling within an elevated mixed layer (EML) ahead of the mesoscale convective system (MCS) helped steepen the low-level lapse rates. Meanwhile, a surface mesolow along a quasi-stationary frontal zone enhanced the warm advection and low-level shear. The late afternoon event on 16 September 2010 was characterized by a quasi-linear convective system (QLCS) that also featured an EML aloft, a surface mesolow just west of NYC, low-level frontogenesis, and a southerly low-level jet ahead of an approaching midlevel trough. The QLCS intensified approaching NYC and generated mesovortices as the QLCS bowed outward. These cases illustrate the benefit of high-density surface observations, terminal Doppler radars, and sounding profiles from commercial aircraft for nowcasting these events.

1. Introduction

a. Background

Early in the morning of 8 August 2007, a large mesoscale convective system (MCS) produced two tornadoes in the New York City, New York (NYC), area (Fig. 1). At 1022 UTC, a tornado classified as a category 1 storm on the enhanced Fujita scale (EF1) occurred on Staten Island, New York, and at 1035 UTC, the same mesocylcone produced an EF2 tornado in Brooklyn, New York. The tornadoes occurred less than an hour after sunrise (1030 UTC), well before the maximum heating of the day. It was the first documented tornado in Brooklyn in over 100 yr. On 16 September 2010, a quasi-linear convective system (QLCS) developed over eastern New Jersey and tracked east across New York City and Long Island. The QLCS spawned two tornadoes (EF0 and EF1) in Brooklyn and Queens, New York, respectively, during the late afternoon hours, causing one fatality [National Climatic Data Center (NCDC) *Storm Data*].

Wurman et al. (2007) remarked on the potential catastrophic damage that could occur if a violent tornado impacted Manhattan, New York, stating that the damage "would far exceed the economic cost of the destruction of high-rise structures on 11 September 2001, though the number of fatalities is difficult to estimate." The 8 August 2007 event caused tens of millions of dollars in damage, and the associated severe flash flooding [8.8 cm in 4 h at John F. Kennedy International Airport (JFK) in southern Queens] disrupted subway and surface transportation throughout the city. Although there have been only nine reported tornadoes in

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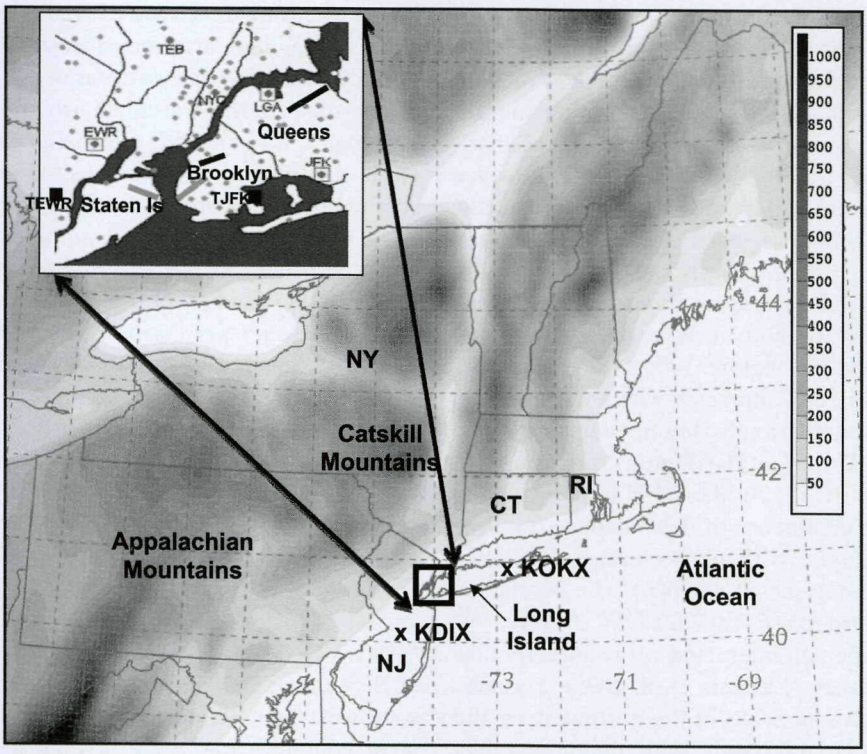

FIG. 1. The northeastern U.S. and NYC domains (see black region in inset panel) used in this study. NYC regional map, with the inset map, shows the available surface stations (gray dots) and terminal Doppler radars at the black box locations (TJFK and TEWR). The EF1–2 tornado track for the 8 Aug 2007 track is in gray and the EF0 and EF1 tracks for the 16 September 2010 case are in black. Terrain is gray shaded every 50 m.

the New York metropolitan region in the past 58 yr (NCDC *Storm Data*), anticipating the potential for tornadoes around NYC can save lives and help protect property.

Several studies have highlighted conditions favorable for tornadic activity over the northeastern United States. Tornadoes in this region tend to occur between June and August, with the peak in activity in July (David 1977; Giordano and Fritsch 1991; Johns and Dorr 1996; Brooks et al. 2003). Johns and Dorr (1996) found that significant tornadoes [some ranked as category 2 on the Fujita scale (F2) and greater] occur primarily between 1700 and 0000 UTC for New England and eastern New York. Across the mid-Atlantic region, there is a bimodal distribution for the development of strong (F3 or greater) tornadoes (Giordano and Fritsch 1991), with one peak between 1700 and 2200 UTC and the second maximum between 2300 and 0300 UTC. Lombardo and Colle (2011) also found a late afternoon peak for tornadoes over the northeastern United States, and they showed that most tornadoes in this region developed from isolated cellular storms and QLCSs. Meanwhile, near the southern New England coast, Lombardo and Colle (2011) showed that tornadoes formed equally from isolated cells and quasi-linear and nonlinear structures.

The percentage of tornado events over the northeastern United States has been categorized by the midlevel (500 hPa) flow patterns (Giordano and Fritsch 1991; Johns and Dorr 1996). Tornadoes occurred with midlevel flow from the northwest (33%), southwest (55%), and west (12%). Separating the data into two flow regimes, David (1977) illustrated that a surface pressure trough exists in the lee of the Appalachians under both northwesterly (270°–350°) and southwesterly (180°–270°) low-level flow. During the southwesterly flow regime, there is often a cold front located to the west of the lee trough to help lift a parcel to its level of free convection (LFC). Northwest flow events tend to be confined to the warm season, with the maximum frequency occurring between May and August. There is a peak in July, due to the seasonal lag in the availability of low-level moisture for this flow regime (Johns 1982). During periods of northwest flow, there is an average of ~15° of directional shear between 850 and 500 hPa over the northeast United States (Johns 1984). For the other

wind patterns, tornadoes develop with little (<5°) directional shear in this layer.

Most of the analysis of tornadogenesis over the northeast United States has focused on terrain influences. Wasula et al. (2002) linked the spatial distribution of tornadoes over New England to the influence of terrain for different 700-hPa flows. More specifically, terrain flow channeling has been suggested to play an important role in tornadogenesis by accelerating the low-level flow and thus increasing the shear in the lowest kilometer. The F4 tornado at Windsor Locks, Connecticut, developed just before 1900 UTC 3 October 1979, as a convective cell interacted with a warm frontal boundary associated with a subsynoptic-scale low (Riley and Bosart 1987). The Great Barrington, Massachusetts, F3 tornado at 2146 UTC 29 May 1995 occurred in association with the confluence of flow channeled up the Hudson Valley merging with an accelerated flow down the Catskill Creek (Bosart et al. 2006). The Mechanicville, New York, tornado (F2) at 2022 UTC 31 May 1998 developed as a supercell interacted with southerly flow up the Husdon Valley (LaPenta et al. 2005). Tornadic activity in Pennsylvania peaks in the northwestern and southeastern regions of the state, with a minimum over the terrain in the central region of the state (Nese and Forbes 1998). Warm, moist low-level air may be channeled east and west of the Appalachian crest, thus inhibiting the development of severe storms over central Pennsylvania.

b. Motivation

While most previous studies of tornadic convection over the northeast United States have investigated terrain influences, we focus on the area around NYC and the surrounding coastal region. There have been no formal case studies of tornado events around NYC, a highly populated region with over 10 million residents. To more accurately warn the public during these events, forecasters require an understanding of the conditions that favor the tornado development in this coastal area. A few key synoptic-scale ingredients have been highlighted for the entire northeast United States (David 1977; Giordano and Fritsch 1991; Johns and Dorr 1996). More recently, Lombardo and Colle (2011) provided some regional composites for all severe weather associated with convective cells and linear and nonlinear convective systems along the northeast U.S. coast. They highlighted the importance of the Appalachian lee trough and low-level thermal ridge for severe convective cells, while linear severe systems develop with relatively large vertical wind shear along a frontogenetical boundary just west of the coast. They also showed that severe nonlinear systems have relatively low most unstable convective available potential energy (MUCAPE) but large quasigeostrophic forcing. The synoptic and mesoscale patterns of evolution supportive of tornadoes in this coastal environment have not been thoroughly investigated. Given the importance of the meso- and microscale processes for severe storms, it is crucial to obtain observations at small scales. In the NYC region, the Aircraft Communications Addressing and Reporting (ACARS) profiles, two Terminal Doppler Weather Radars (TDWRs), as well as a regional surface mesonet were utilized (Fig. 1). The benefits of these additional observations will be shown.

This paper describes the temporal climatology and synoptic conditions associated with tornadoes in this urban coastal region. Our paper builds on the results of Lombardo and Colle (2011) by focusing on the average synoptic and regional patterns associated with coastal tornado events to help improve forecaster pattern recognition. Some of the large-scale features and any variations from this climatology are illustrated using two NYC tornado events. The 8 August 2007 and 16 September 2010 events fit within the northwest and southwest flow regimes, respectively, which are the two most common flow regimes for tornadic convection over the northeast United States (Giordano and Fritsch 1991; Johns and Dorr 1996). The higher-resolution TDWR, ACARS, and surface observations are presented to help improve forecaster understanding of the mesoscale structures during these two tornado events. Neither of these events was well anticipated by forecasters. For example, the area forecast discussion by the National Weather Service (NWS) Forecast Office for New York City issued just a few hours before the September 2010 event mentioned a potential for strong storms, but the probability for severe weather was considered to be quite low. There were also no severe thunderstorm or tornado watch or mesoscale convective discussions issued by the Storm Prediction Center for these events. A major goal of this paper is to highlight the synoptic and mesoscale evolution associated with NYC tornadoes by addressing the following motivational questions:

- What synoptic-scale conditions support tornado development around NYC?
- What is the role of the low-level baroclinic zones and elevated mixed layers?
- How important are the mesonet and terminal Doppler radars in nowcasting these events?
- What is the monthly and diurnal climatology for tornadoes around the NYC region?

Section 2 will describe the data and methods used to examine the NYC tornado events. A monthly and

TABLE 1. Dates for the 20 tornado events for the NYC–LI region (inset in Fig. 2a) used in the NARR composites.

1800 UTC 10 Aug 1979	1200 UTC 8 Aug 1999
1800 UTC 5 Aug 1981	0900 UTC 15 Sep 2000
1800 UTC 25 Aug 1982	0300 UTC 30 Jun 2001
1500 UTC 30 Aug 1985	1800 UTC 11 Jul 2001
1500 UTC 5 Oct 1985	0000 UTC 13 Aug 2005
1500 UTC 19 Aug 1991	1500 UTC 25 Aug 2006
2100 UTC 23 Jul 1995	1200 UTC 18 Jul 2007
2100 UTC 26 Jun 1997	0900 UTC 8 Aug 2007
1200 UTC 30 Jun 1998	1800 UTC 25 Jul 2010
1800 UTC 9 Jul 1998	2100 UTC 16 Sep 2010

diurnal climatology and synoptic composites will be presented in section 3. Two case studies are used to highlight some of the composite structures and mesoscale details for the August 2007 and 16 September 2010 events in sections 4 and 5, respectively. A discussion and summary will be provided in section 6.

2. Data and methods

The events for the coastal tornado climatology were identified utilizing both the Storm Prediction Center (SPC) and NCDC storm reports archives from 1950 to 2010. To evaluate the synoptic and thermodynamic conditions that support tornadoes over the coastal zone, spatial composites were constructed for the 20 tornadoes around NYC, Long Island, and southern Connecticut from 1979 to 2010 (Table 1) using the North American Regional Reanalysis (NARR) at 32-km grid spacing (Mesinger et al. 2006). The closest 3-h NARR time prior to the tornado report was used. For this study, MUCAPE uses a parcel with the maximum equivalent potential temperature θ_e in each 30-hPa layer from 0 to 180 hPa above the ground.

The synoptic environments for the August 2007 and September 2010 NYC tornado events were examined using the Rapid Update Cycle (RUC) analysis grids available hourly at 80-, 20-, and 13-km grid spacing every 3 h. The 80-km grid was used for the large-scale analysis and quasigeostrophic diagnostics, while the 20- and 13-km grids were applied to the regional and thermodynamical analyses for the 2007 and 2010 events, respectively. Hourly surface observations were used, including surface data from the MesoWest mesonet archive (Horel et al. 2002). Standard radiosonde soundings were utilized in conjunction with the ACARS profiles from LaGuardia Airport (LGA), Newark Liberty International Airport (EWR), and JFK (cf. Fig. 1).

Several different ground-based radar datasets were used. The National Operational Weather Radar (NOWrad), provided by Weather Services International (WSI) and the National Center for Atmospheric Research (NCAR), illustrated the regional convective evolution patterns. The Weather Surveillance Radar-1988 Doppler (WSR-88D) level II reflectivity data from Upton, New York (KOKX), and Fort Dix, New Jersey (KDIX), provided a more detailed look at the convection at key times during the storm's evolution. To analyze the evolution of the tornadic mesocyclones and precipitation structures around NYC, the TDWRs located near JFK and EWR airports (TJFK and TEWR) were used (Fig. 1), which sampled reflectivity and radial velocity data every minute at a 0.3° elevation scan for TEWR and 0.5° at TJFK.

3. Climatology and composite evolution

A climatology and composite analysis of NYC area tornado events is presented to highlight monthly and diurnal frequency as well as some important synoptic precursors. The monthly and diurnal distributions of tornadoes for NYC–LI region were compared with those of the northeastern United States and the southern Connecticut region (Fig. 2a). The southern Connecticut region is similar in size to the NYC–LI region and illustrates how the climatology can change over a relatively small distance to the north of the Atlantic coast.

The monthly distribution of 34 documented [National Oceanic and Atmospheric Administration (NOAA) storm reports] tornado events for the NYC–LI area from 1950 to 2010 is shown in Fig. 2a. Tornadoes in this area occurred mainly from June through October, with the maximum number of events occurring in August. In contrast, areas farther inland across the Northeast have a peak in tornadic activity during July (David 1977; Johns and Dorr 1996; Nese and Forbes 1998; Wasula et al. 2002; Brooks et al. 2003). Over coastal Connecticut (46 cases within approximately 30 km of the coast) a July maximum is also observed (Fig. 2a).

Lombardo and Colle (2011) showed that the coastal northeastern U.S. tornadoes are skewed slightly toward the morning and early afternoon hours, but their analysis included the full northeast U.S. coastal region. To illustrate some of the diurnal variations from the coast to slightly inland, the diurnal analysis was repeated for two smaller areas (coastal Connecticut versus NYC–LI). While tornadoes over coastal Connecticut can occur at any time of the day, there is a clear maximum in the number of reports from 2100 to 0000 UTC (Fig. 2b). In contrast, for NYC–LI, only ~25 km to the south of Connecticut, there is a preference for tornado development during the early morning through early afternoon hours

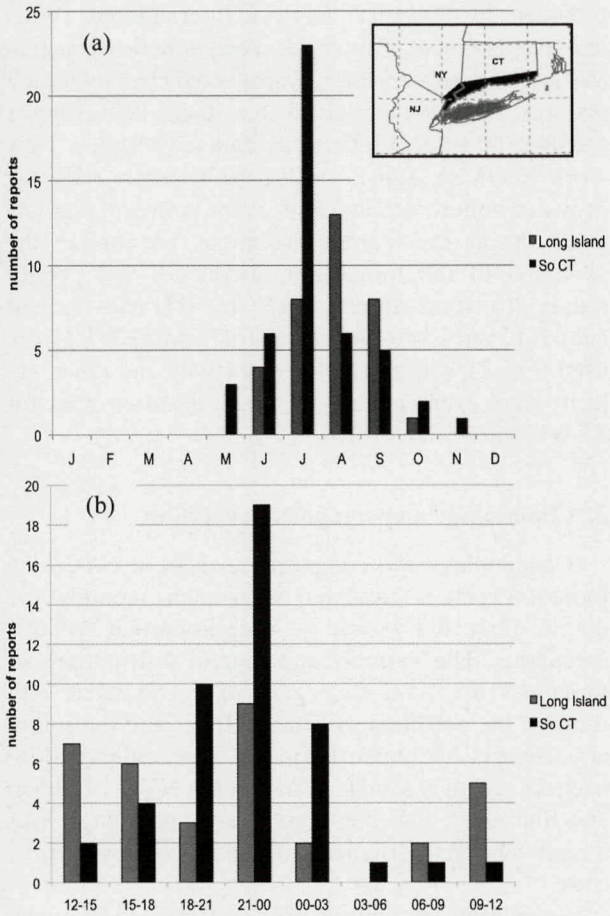

FIG. 2. (a) Monthly distribution of tornado events from 1950 to 2010 for southern CT (black) and LI (gray), with the two regions shown in the inset panel. (b) The diurnal tornado distribution every 3 h for NYC–LI (gray) and southern CT (black) for the 1950–2010 period.

(0900–1800 UTC). The difference between the mean number of southern Connecticut tornadoes that develop between 1800 and 0600 UTC was tested against the mean number of LI tornadoes that form during the same 12-h time period using a two-sided t test. This difference was significant to the 95% level. The differences in these means were also tested using a bootstrap analysis technique, yielding a statistical significance of 90%. Overall, the Connecticut late afternoon maximum in tornado development is significantly different from the NYC–LI morning to early afternoon maximum.

Spatial composites using the NARR were created for the NYC–LI (20 cases of NYC–LI events from 1979 through 2010 in Table 1) and southern Connecticut regions. The composite was created for a group of cases by averaging a particular field of interest at each grid point on the NARR grid. Only the NYC–LI region results are presented since the differences were not statistically significant between the NYC–LI and Connecticut regions. A closer examination of the 20 NYC–LI cases reveals that there were 7 Appalachian lee trough events, 6 cold front events, 5 warm front events, 1 stationary front event, and 1 tropical cyclone event (Hurricane Bob 1991). Although Long Island tornadoes develop under a variety of scenarios, several important ingredients are common to many of the events.

In the composite, at 1 day prior to the tornado event ($T - 24$ h), a 300-hPa westerly jet extends across the northeastern tier of the United States (Fig. 3a). By the time of the event ($T - 0$ h), the upper-level jet becomes amplified to 25 m s^{-1}, with NYC near the equatorward jet entrance region, which is a favored region for large-scale ascent (Fig. 3b). The axis of the upper-level trough is located along the Eastern Seaboard into Canada. Meanwhile, a 700-hPa shortwave embedded in west-northwesterly flow is located over the upper Midwest around Wisconsin. There is associated 700-hPa upward motion just east of this trough axis (Fig. 3a). This shortwave trough amplifies as it moves eastward over the Appalachians during the next day, with enhanced mid-level ascent over NYC (Fig. 3b).

At the surface, 1 day prior to the tornado events (Fig. 4a), there is a westward extension of the subtropical high over the western Atlantic into the southeast United States. Meanwhile, there is a weak trough extending north–south across the upper Midwest. Over the NYC–LI area, there are relatively light 1000-hPa westerly winds (<3 m s^{-1}) that are nearly parallel to a west–east zone of relatively large θ_e differences. The reservoir of MUCAPE is south of the baroclinic zone over the central United States and the mid-Atlantic.

The surface composite at $T - 0$ h reveals the presence of a trough in the lee of the Appalachians, and this trough extends northward into a parent cyclone over northern New York. The southwesterly wind ahead of the trough advects higher values of 1000-hPa θ_e northward toward NYC–LI. Consequently, the average MUCAPE over the region increased from 500 to over 1000 J kg^{-1} (Figs. 4a,b).

The composites show that the 0–1-km shear was ~10 m s^{-1} on average around NYC at $T - 0$ h (not shown), which falls within the range expected for tornadogenesis (Markowski et al. 2003). Also, previous studies have shown that a relatively low lifted condensation level (LCL) favors tornadogenesis (Rasmussen and Blanchard 1998), since a relatively strong cold pool from evaporation can rapidly cutoff and weaken the mesocyclone from the warm and unstable air. The composite LCL is 920 hPa at $T - 0$ h, which is consistent with the mean LCL for the tornado events over the central United States (Rasmussen and Blanchard 1998).

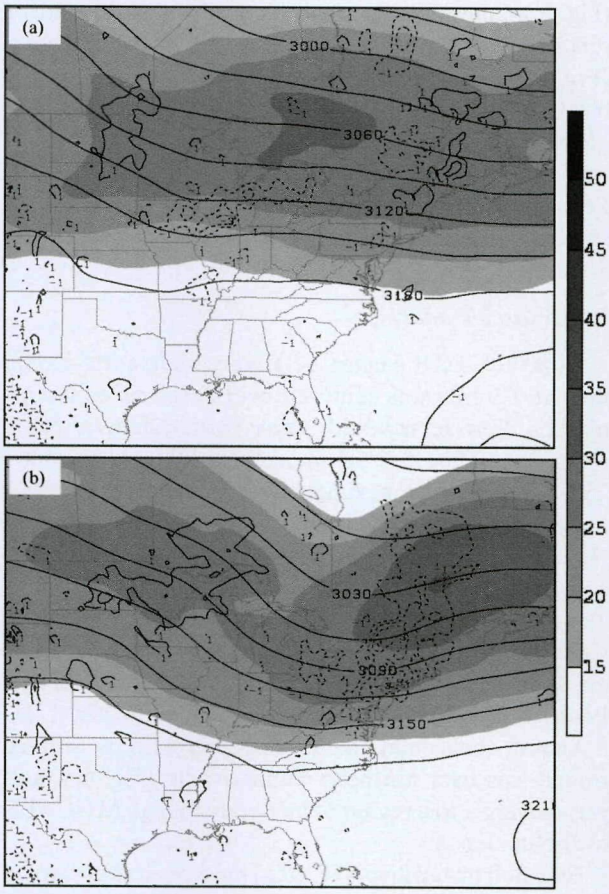

FIG. 3. Large-scale composite of NYC–LI tornado events showing 300-hPa isotachs (shaded every 5 m s^{-1} starting at 15 m s^{-1}), 700-hPa geopotential heights (solid every 30 m), and 700-hPa omega (every 1×10^{-3} s^{-1}, with positive solid and negative dashed) at (a) $T - 24$ and (b) $T - 0$ h.

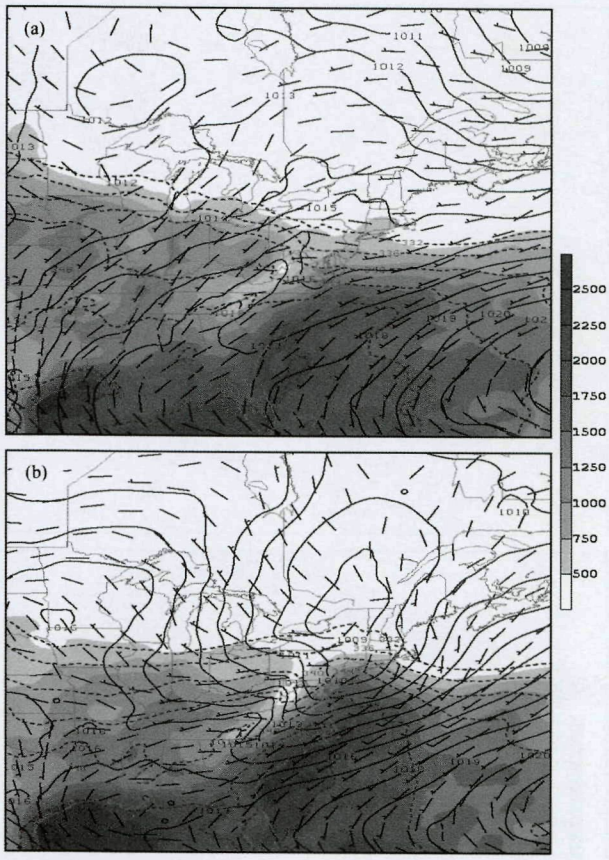

FIG. 4. Large-scale composite of NYC–LI tornado events showing sea level pressure (every 2 hPa), θ_e (dashed every 4 K), wind (full barb = 10 m s^{-1}), and MUCAPE shaded (every 250 J kg^{-1}) at (a) $T - 24$ and (b) $T - 0$ h.

4. 7–8 August 2007 event

a. Large-scale analysis

At the surface at 1200 UTC 7 August (Fig. 5a), which is about 22 h before the tornadoes affected the NYC area, there was a weak surface cyclone located over the central Great Plains (~1006 hPa) and northern Minnesota (~1004 hPa), with a warm and stationary front extending across the western and southern Great Lakes, respectively. An MCS that originated over Iowa at 0600 UTC 7 August 2007 had moved eastward to the north of the stationary front (not shown), causing severe wind and hail from southern Iowa to southern Michigan overnight and into the early morning on 7 August (Fig. 6a). Meanwhile, there was surface high pressure and weak surface winds from the northern Gulf of Mexico northeastward to New England. At 700 hPa (Fig. 5b), there was a broad ridge centered over the southeastern United States, with nearly zonal flow across the upper Midwest to the northeastern United States. One short-wave trough was located over the central Great Lakes, while another was over the upper Midwest and southern Canada. The 300-hPa jet was located well north along the Canadian border at 1200 UTC 7 August (Fig. 5b), and it remained well to the north of southern New England during this event (not shown), so it was likely not much of a factor across the NYC region.

At 1300–1600 UTC 7 August, the most active convection was located along Lake Erie as well as parts of southwest Ontario, northeast Ohio, and northwest Pennsylvania (Figs. 6b,c). The MCS continued to produce severe wind and hail across northeast Ohio around this time (Fig. 6a). Shortly after 1600 UTC (not shown), the convection weakened as the MCS moved into western Pennsylvania. This weakening occurred during peak diurnal heating, with surface temperatures averaging 32°C and the MUCAPE exceeding 2000 J kg^{-1} across western Pennsylvania (not shown). However, at 1500 UTC 7 August, western Pennsylvania was under

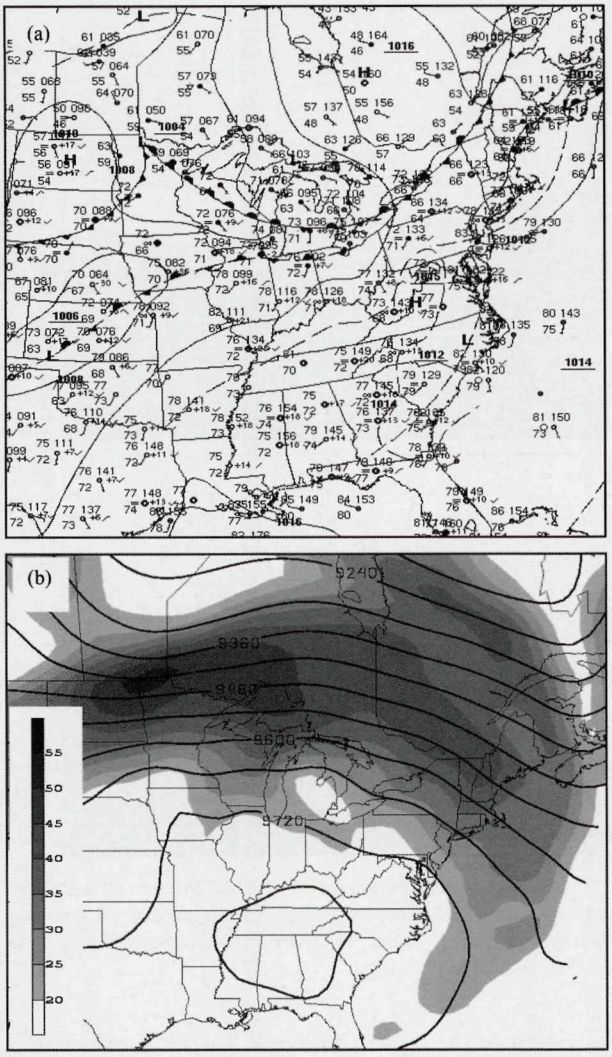

FIG. 5. (a) National Centers for Environmental Prediction (NCEP) surface analysis and conventional station models (°F) from 1200 UTC 7 Aug 2007. (b) The 700-hPa geopotential heights (solid every 60 m) and 300-hPa wind speeds (gray shaded in kt using inset scale) at 1200 UTC 7 Aug 2007.

the short-wave 700-hPa ridge (Fig. 7a), with 100 J kg^{-1} of surface-based convective inhibition (SBCIN), 0–3-km shear less than 5 m s^{-1} (not shown), and there was little Q-vector forcing and low-level frontogenesis (Figs. 7a,b). Thus, there was little shear to maintain the convection or synoptic forcing to help lift parcels to their LFC (~850 hPa). By 2300 UTC 7 August (Fig. 6d), the main region of stratiform precipitation advanced ahead of the midlevel short wave from central New York to eastern Pennsylvania.

By 0300 UTC 8 August, a 700-hPa short-wave trough with quasigeostrophic (QG) forcing for ascent (**Q**-vector convergence) was located over southern New York (Fig. 7c). Convection began to reorganize slightly south of this 700-hPa QG ascent over northwestern Pennsylvania (Fig. 6d) in a region of warm advection (Fig. 7d). At 0600 UTC, the 700-hPa QG ascent is still situated over northern New Jersey (not shown). Meanwhile, the convection became more organized and intense across eastern Pennsylvania to northern New Jersey (Fig. 6e), which subsequently moved over the NYC area by 1000 UTC 8 August (Fig. 6f).

b. Mesoscale analysis

At 0000 UTC 8 August, relatively weak anticyclonic flow at 925 hPa was centered over eastern Pennsylvania and New Jersey, with weak westerlies over NYC (Fig. 8a). The axis of moderate MUCAPE (500–2500 J kg^{-1}) and a maximum in θ_e at 925 hPa (~350 K) extended northeastward into eastern Pennsylvania and New Jersey. By 0600 UTC 8 August, 925 hPa $\theta_e > 350$ K and a MUCAPE > 2000 J kg^{-1} extended northward along the coastline as the 925-hPa winds became more southerly (Fig. 8b). At the surface, the lee trough deepened and pressure within a mesoscale low over eastern Pennsylvania fell to 1007.5 hPa (Fig. 9a). By 0600 UTC 8 August, convection had developed around the surface trough axis over northeast Pennsylvania (Fig. 6e), and was moving into a region of moderate to high MUCAPE over New Jersey.

Elevated mixed layers (EMLs) have been shown to be important for northeast U.S. severe weather events (Banacos and Ekster 2010). During this event, there was an EML between 800 and 600 hPa, with a relatively steep lapse rate in the RUC analysis (7.5°–8°C km^{-1}) in this layer over the Great Lakes at 1200 UTC 6 August (Fig. 10a). This EML had originated just east of the Rocky Mountains over the central and northern plains on 4 August 2007. The west-northwest flow at midlevels advected this EML eastward to western New York and Pennsylvania by 1200 UTC 7 August (Fig. 10b) and just off the Northeast coast by 1200 UTC 8 August (Fig. 10d). The EML is present in the Pittsburgh, Pennsylvania, sounding at 1200 UTC 7 August (Fig. 10c), with a ~8°C km^{-1} lapse rate and relatively dry air between 800 and 600 hPa. At 0305 UTC 8 August (Fig. 11a), an elevated layer of steep lapse rates was present between 800 and 600 hPa at Newark (EWR in Fig. 1). At 0300 UTC 8 August, the RUC sounding profile illustrates relatively dry air and nearly dry-adiabatic lapse rate between 750 and 650 hPa (Fig. 11c), although the lapse rate was not as steep as in earlier observations.

By 0600 UTC 8 August, the leading edge of the precipitation was approaching the NYC area (Fig. 12a). By 0830 UTC 8 August, there was mainly stratiform precipitation pattern around NYC (Fig. 12b). Meanwhile,

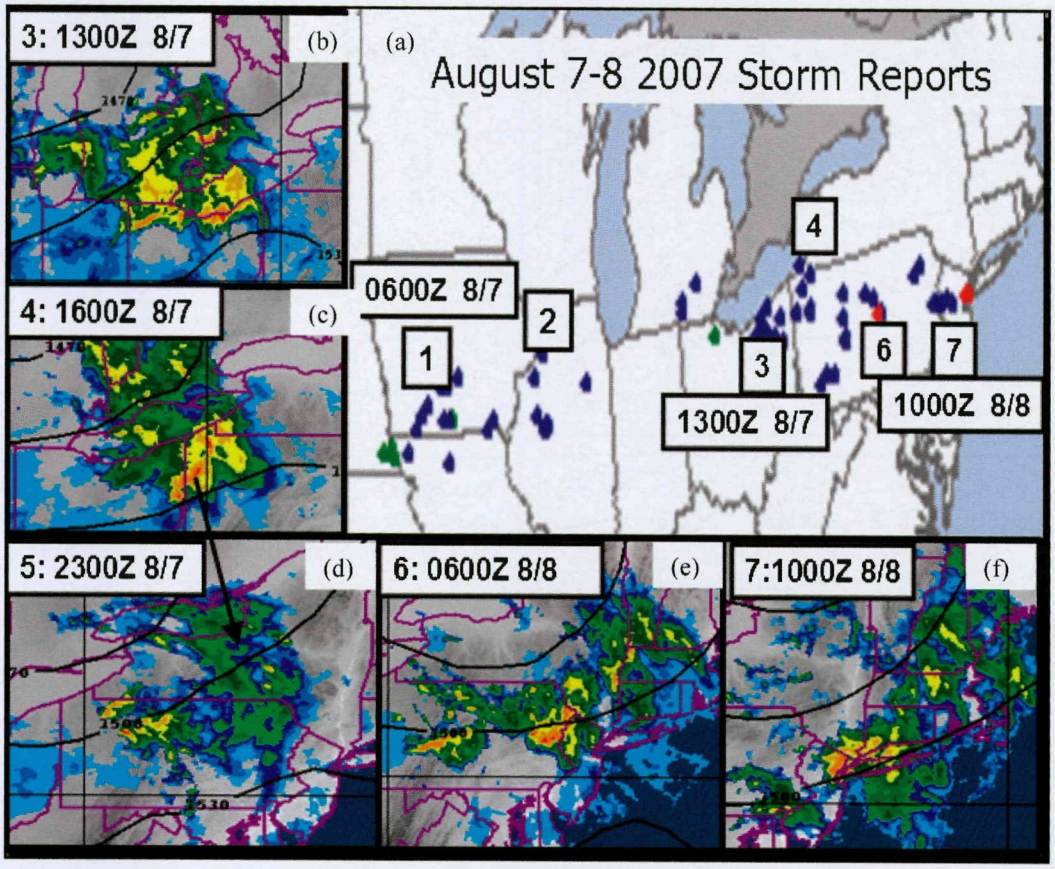

FIG. 6. (a) Severe storm reports for 7–8 Aug 2007 and composite radar reflectivity at (b) 1500 UTC 7 Aug, (c) 2300 UTC 7 Aug, (d) 0300 UTC 8 Aug, (e) 0600 UTC 8 Aug, and (f) 1000 UTC 8 Aug. The blue, green, and red dots in (a) are the severe wind, hail, and tornado reports from 0600 UTC 7 Aug to 1000 UTC 8 Aug 2007.

the 0827 UTC 8 August EWR ACARS sounding showed a 5°–10°C cooling in the 850–700-hPa layer from evaporation and adiabatic ascent since 0300 UTC (Fig. 11b). This created a nearly moist-adiabatic layer between 800 and 700 hPa, which helped steepen the lapse rate below this layer. Meanwhile, this cooling stabilized the layer immediately above, which is why the steepest lapse rates in the 800–600-hPa layer were offshore of the Northeast coast at 1200 UTC 8 August (Fig. 10d). The cooling also is seen in the RUC analyses between 0300 and 0900 UTC (Fig. 11c), although the magnitude of the cooling was less in the RUC than in the ACARS observations. Although the RUC MUCAPE values across NYC (~1500 J kg^{-1}) did not change much between 0600 and 0900 UTC (Figs. 11b,c), a comparison between the ACARS profile at EWR and the RUC profile at 0900 UTC indicates that the RUC is likely too stable between 900 and 800 hPa (Figs. 11b,c). Meanwhile, the LFC in the RUC analysis did drop from ~700 hPa at 0600 UTC 8 August to ~850 hPa at 0900 UTC 8 August. Overall, the leading edge of the MCS helped to create a more favorable thermodynamic environment for deep convection around NYC.

At the surface at 1000 UTC 8 August (Fig. 9b), the mesolow moved eastward along the surface baroclinic zone and was located in northern New Jersey. Surface pressures within the circulation fell to 1006.5 hPa. The most intense convection was focused within the north and northeast quadrants of the mesolow (Figs. 9c,d), where southerly winds and warm advection (isentropic lift) help to focus the convection (Fig. 6b).

Meanwhile, the low-level shear became more enhanced as the 925 hPa southwesterly winds increased to ~15 m s^{-1} over NYC by 0826 UTC 8 August (Fig. 11b). There are low-level veering winds at EWR, with a ~22 m s^{-1} southwesterly flow near 800 hPa and at least 17 m s^{-1} of southwesterly shear in the lowest kilometer (Fig. 11b). This 0–1-km shear is the largest of all severe convective events analyzed by Lombardo and Colle (2011) from 2002 to 2007 over the coastal northeast United States. For the east-central United States, Parker and Ahijevych (2007) highlighted the importance of

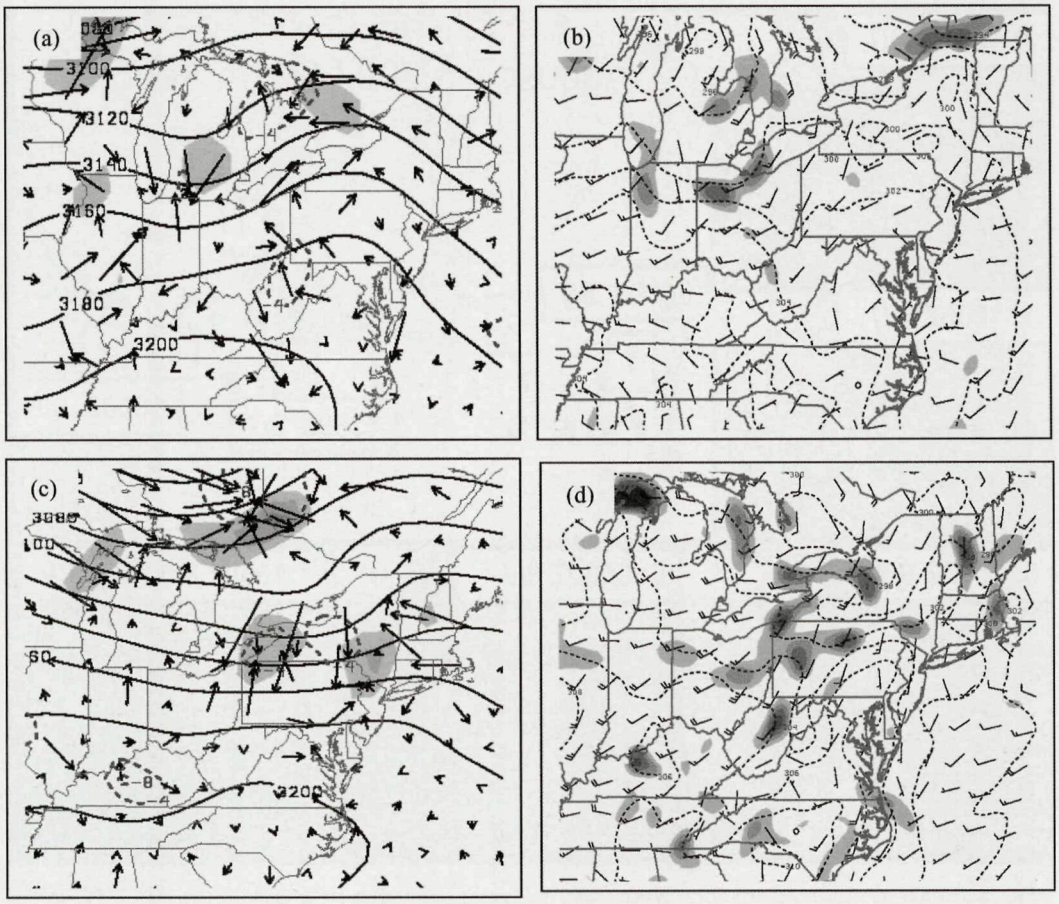

FIG. 7. (a) The 700-hPa **Q** vector and **Q**-vector convergence (shaded in K m^{-1} s^{-1} $\times 10^{-11}$), 700-hPa geopotential heights (solid, m), 700-hPa omega (dashed, m s^{-1} $\times 10^{-3}$) at 1500 UTC 7 Aug 2007. (b) The 950-hPa frontogenesis [shaded, 1 K $(100\ km)^{-1}$ $(3\ h)^{-1}$ $\times 10^{-1}$], potential temperature (dashed, every 2 K), and winds (full barb = 10 kt) at 1500 UTC 7 Aug 2007. (c) As in (a), at for 0300 UTC 8 Aug 2007. (d) As in (b), but at 0300 UTC 8 Aug 2007.

lower-tropospheric shear for the organization of mesoscale convection. Furthermore, the importance of low-level environmental shear for tornadogenesis has been well documented throughout the literature (Weisman and Klemp 1982, Brooks et al. 1994; Thompson et al. 2003; Markowski et al. 2003). For example, Thompson et al. (2003) and Markowski et al. (2003) showed that all cases with tornadoes and significant tornadoes (F2 or greater) over the central plains have mean 0–1-km vertical wind shears of 6–8 and 10–12 m s^{-1}, respectively. The enhanced environmental shear present at EWR suggests that the MCS had the potential to become tornadic when approaching the NYC area. Within this evolving environment, the leading edge of the convective area at 0930 UTC 8 August weakened as it moved into northeastern New Jersey (Fig. 12c), while a second, more linear, region of convection organized along the surface trough boundary located over north-central New Jersey. At 1013 UTC 8 August (Fig. 12d), this back edge of the MCS moved over NYC and was associated with the tornadic convection.

The intense cell approaching northern Staten Island around 1013 UTC was associated with a wind-shift line with westerly winds to the rear (Fig. 12d), which was attached to the surface trough over northern New Jersey (Fig. 9b). Meanwhile, the surface temperatures were approximately 75°–77°F (23.9°–25°C) around Staten Island and NYC, and 73°–74°F (22.8°–23.3°C) to the northwest of the trough axis and convective system (Fig. 12d), so there was no well-defined cold pool with this system. On the southern boundary of this intense convection, a mesocyclone developed along this wind-shift boundary approaching northern Staten Island. Over a period of 10 min, the gate-to-gate shear increased from 6.5 to 26 m s^{-1} (not shown). By 1017 UTC, a weak mesocyclone using the TEWR radar (radar beam height was 300–400 m) was crossing northern Staten Island (Fig. 13b).

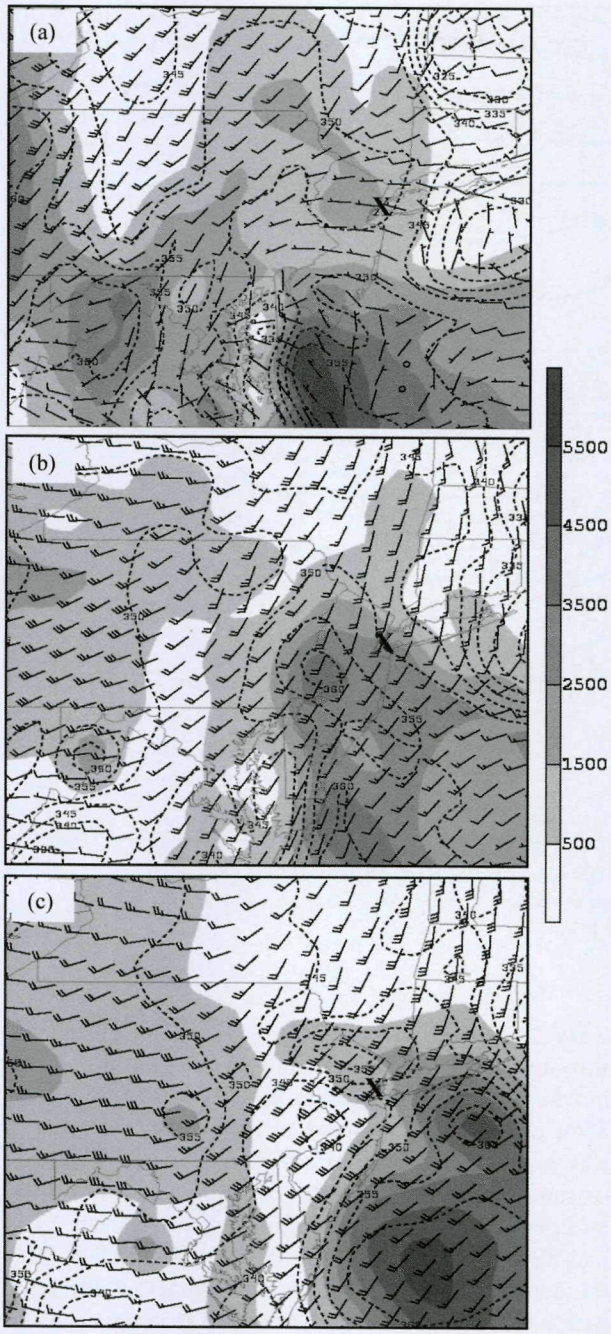

FIG. 8. MUCAPE gray shaded every 500 J kg^{-1} starting at 500 J kg^{-1} for 925-hPa θ_e (dashed, every 5 K) and 925-hPa winds (full barb = 10 kt) at (a) 0000, (b) 0600, and (c) 0900 UTC 8 Aug 2007 from the RUC analyses. NYC is shown at the location marked by X.

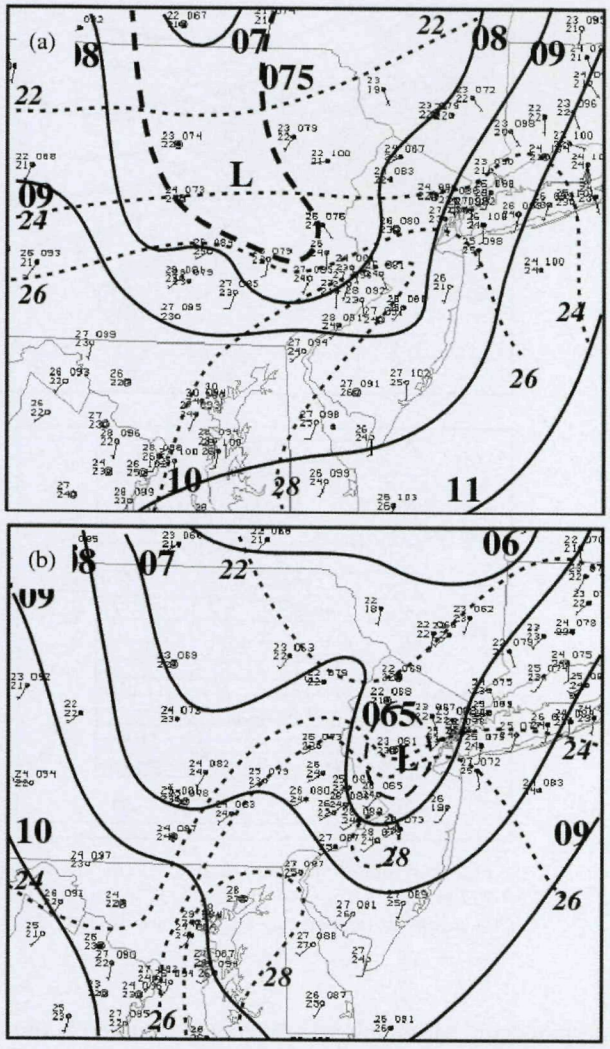

FIG. 9. Manual surface analysis showing sea level pressure (solid every 1 hPa) and surface temperature (dashed, every 2°C) at (a) 0600 and (b) 1000 UTC 8 Aug 2007.

By 1027 UTC the circulation over northern Staten Island merged with another low-level circulation that had developed to the northeast of Staten Island and drifted southeastward to the entrance of New York Harbor (Fig. 14a). Figure 13c shows the well-defined velocity couplet just east of Staten Island at 1029 UTC. This circulation was relatively shallow, with the velocity couplet much weaker by 1300 m AGL (Fig. 14b). A tornado warning was issued for NYC at 1028 UTC as this mesocyclone intensified and an EF1 tornado touched down near eastern Staten Island and the entrance to New York Harbor (Fig. 1 and dashed circle in Fig. 13c). By this time, the tornadic circulation was embedded within a shield of moderate precipitation (Fig. 13a). By 1035 UTC (Figs. 13d and 14c), the mesocyclone circulation moved into Brooklyn and produced an EF2 tornado. A cross section through this velocity couplet at this time illustrates that the circulation was mainly below 1500 m AGL (Fig. 14d). This circulation tracked eastward into central Brooklyn by 1040 UTC (not shown). By 1043 UTC, the mesocyclone

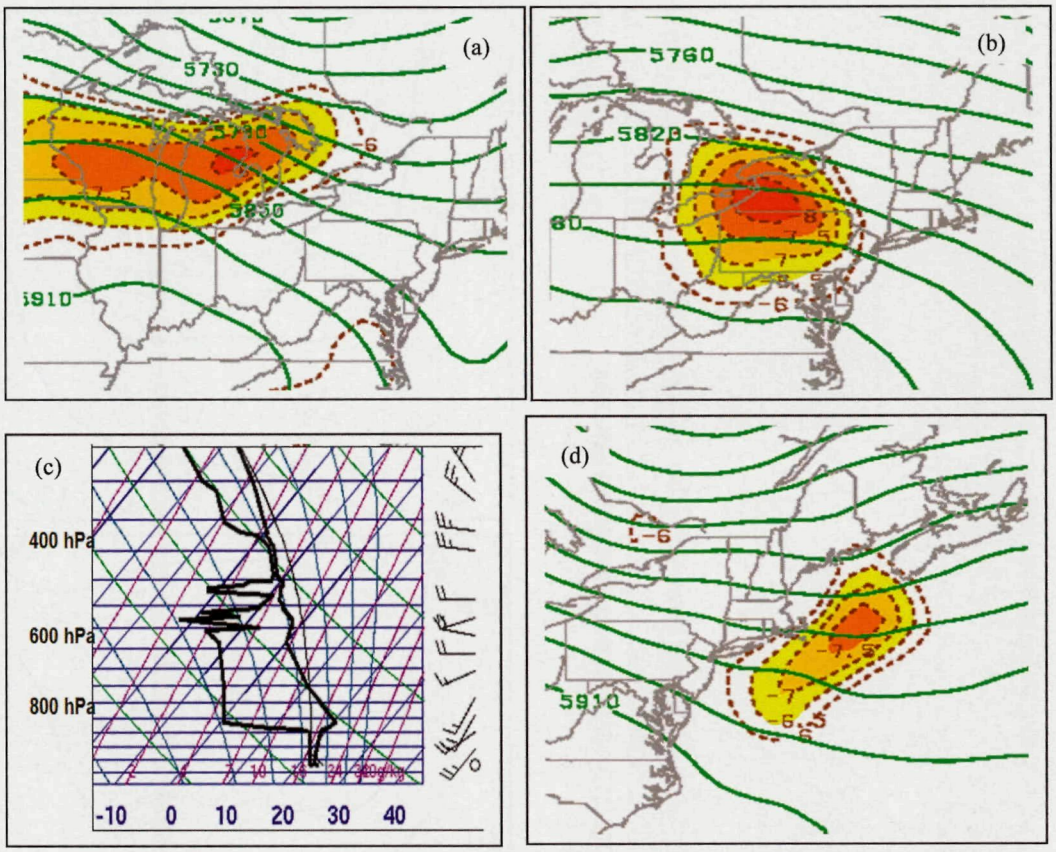

FIG. 10. Average lapse rate (shaded, °C km^{-1}) from 800 to 600 hPa using the RUC analyses at (a) 1200 UTC 6 Aug, (b) 1200 UTC 7 Aug, and (d) 1200 UTC 8 Aug. (c) Pittsburgh sounding at 1200 UTC 7 Aug showing temperature (°C), dewpoint temperature (°C), and winds (full barb = 10 kt).

occluded and within minutes dissipated completely (not shown).

5. 16 September 2010 event

a. Large-scale analysis

At 1200 UTC 16 September 2010, about 9 h before the severe weather occurred around NYC, there was a shortwave ridge over the northeastern United States at 500 hPa (Fig. 15a), while midlevel troughs were located over the western Great Lakes and western Quebec, Canada. A 300-hPa jet extended from Quebec southeastward to Nova Scotia, Canada. Given its northern location, there was likely little influence from this upper jet during this event over southern New York. There was a surface cyclone (~1002 hPa) centered over southern Lake Michigan (Fig. 15b), with a broad baroclinic zone at 950 hPa extending eastward toward the mid-Atlantic coast. Frontogenesis at 950 hPa extended from western Pennsylvania to central and southern New Jersey at this time (Fig. 15b).

By 2100 UTC 16 September, as the convective line was organizing and approaching NYC, a 700-hPa trough moved eastward to the eastern Great Lakes (Fig. 15c). Most of the 700-hPa ascent and Q-vector convergence was located over northern New York, Vermont, and extreme southeastern Ontario, while there was little midlevel ascent and forcing around southern New York and New Jersey at this level. At the surface (Fig. 15d), the cyclone (~1004 hPa) was over eastern Lake Erie, with a surface trough and associated frontogenesis extending eastward to central New York. The baroclinic zone originally over the mid-Atlantic several hours earlier was now located from eastern Pennsylvania to Long Island, with an area of 950-hPa frontogenesis around Long Island and parts of southern New England. As a result, across the NYC area there was frontogenetical forcing for ascent and a relatively strong low-level southerly jet (15–20 m s^{-1}) at 950 hPa to the south of Long Island.

The atmosphere destabilized quickly around NYC during the day on 16 September. In the ACARS sounding

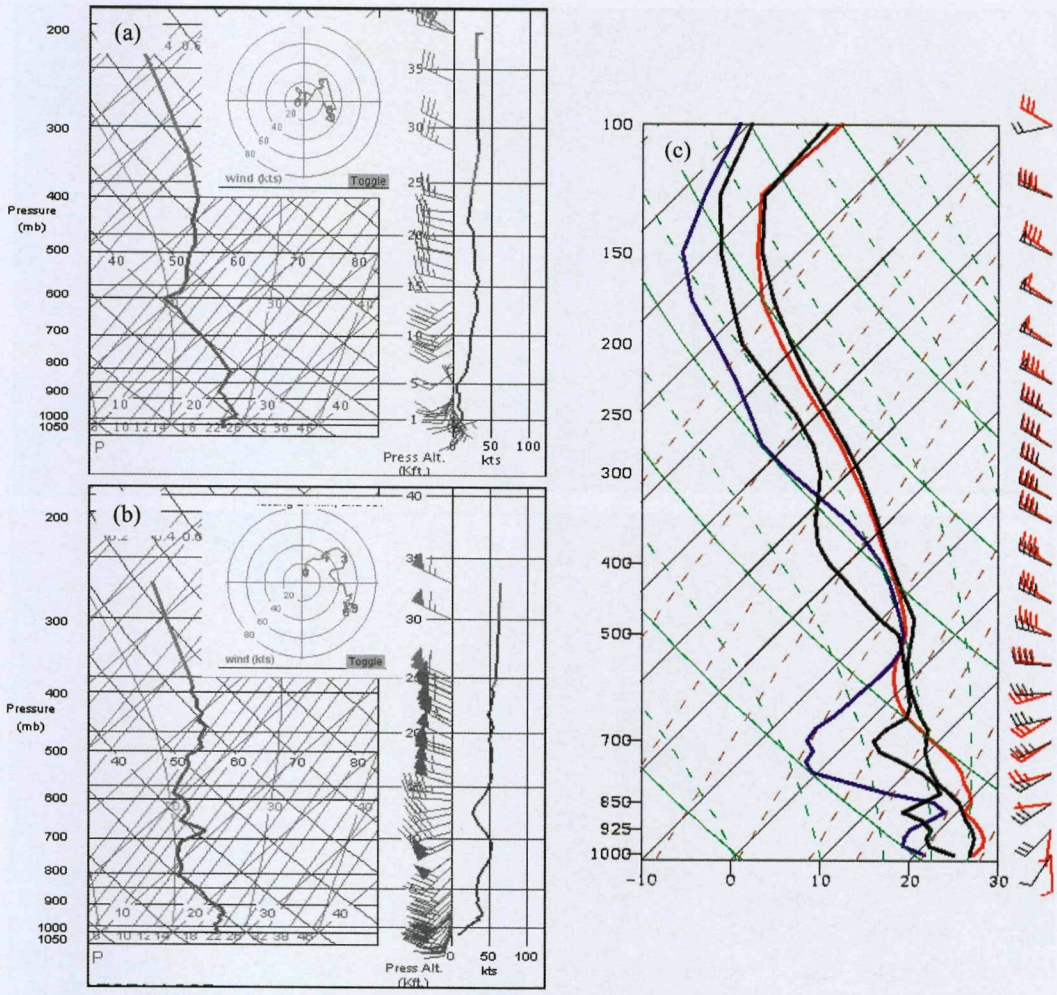

FIG. 11. ACARS temperature and wind profile at EWR at (a) 0305 and (b) 0826 UTC 8 Aug. (c) RUC analysis sounding at Newark at 0300 UTC (temperature in red and dewpoint temperature in blue) and at 0900 UTC (both temperature and dewpoint temperature in black) on 8 August.

from JFK at 1234 UTC 16 September (Fig. 16a), there was a 3°–4°C inversion between 950 and 900 hPa associated with the warm frontal zone and veering winds from south-southeasterly to southwesterly. The column was unsaturated with no MUCAPE at this time. By 1724 UTC 16 September at JFK (Fig. 16b), the low-level inversion had strengthened and deepened up to 850 hPa as a result of the warm advection near the top of this layer. Meanwhile, there was some backing in the wind profile from 750 to 650 hPa that was associated with cold advection, which led to a temperature drop of 1°–2°C around 700 hPa during the preceding 5 h. This resulted in a nearly dry-adiabatic layer between 850 and 700 hPa and ~400 J kg^{-1} of MUCAPE in the JFK sounding. It was also drier in this layer (7°–8°C dewpoint depression) as compared to the nearly saturated layer between 975 and 900 hPa. This elevated dry-adiabatic layer deepened further from 850 to 650 hPa and became drier by 2006 UTC 16 September at JFK (Fig. 16c). There were ~900 and ~1200 J kg^{-1} of MUCAPE in the JFK sounding and the 2000 UTC RUC analysis (not shown), respectively, which is about an hour before the convection moved through.

To further illustrate the origin of this nearly dry-adiabatic layer between 850 and 700 hPa at NYC, the lapse rate in this layer is plotted spatially from 0000 UTC 16 September to 2100 UTC 16 September (Fig. 17). At 0000 UTC 16 September (Fig. 17a), the greatest lapse rates between 850 and 700 hPa were over the Carolinas and Virginia westward to the Ohio River valley. There was no MUCAPE at this time across this region, since there was a subsidence inversion above 700 hPa and relatively dry air at low to midlevels (not shown). There was an anticyclone at 800 hPa over South Carolina as

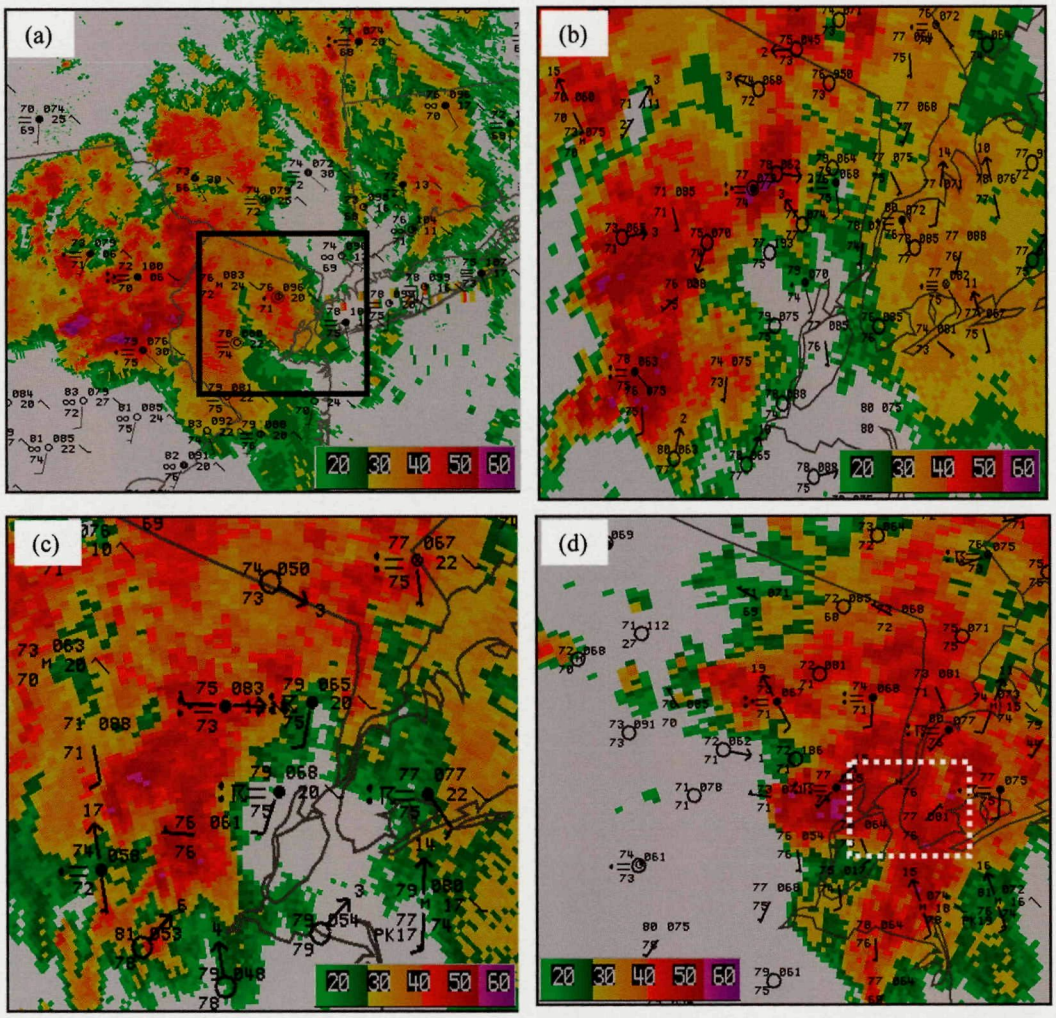

FIG. 12. (a) Regional mosaic of WSR-88D reflectivity for the 0.5° elevation scan (shaded in dBZ) combined with mesonet station models (in °F) at 0558 UTC 8 Aug 2007. (b)–(d) As in (a), but for the KDIX (see Fig. 1) radar reflectivity for the 0.5° elevation scans at (b) 0829, (c) 0930, and (d) 1013 UTC. The dashed white box in (d) is for the zoomed-in region in Fig. 13.

indicated by the clockwise flow at this level. The winds were westerly or northwesterly over the mid-Atlantic, so this air could not be advected northward along the coast at this time. By 1200 UTC 16 September, the lapse rates steepened to >9°C $(1000\ \mathrm{m})^{-1}$ in the 850–700-hPa layer around the eastern Virginia and North Carolina coast (Fig. 17b), and there was 500–1500 J kg^{-1} of MUCAPE across this region given an increase in moisture (θ_e) near the surface along the coast (not shown). Southwesterly flow at 800 hPa around Virginia advected this more unstable air northward toward New Jersey and Long Island. By 2100 UTC 16 September, the relatively steep lapse rates between 850 and 700 hPa extended northward along the mid-Atlantic coast to Long Island. This corresponds to the development of the nearly dry-adiabatic layer in the JFK ACARS profiles around this time (Fig. 16c). Meanwhile, there was an increase in low-level MUCAPE toward the coast from western New Jersey (~500 J kg^{-1}) to western Long Island (~1500 J kg^{-1}), and the southwesterly flow increased to 10–15 m s^{-1} at 800 hPa across much of the northeast United States.

b. *Mesoscale evolution*

At 1924 UTC 16 September, there was a broad area of light to moderate (35–40 dBZ) precipitation across much of eastern Pennsylvania to southeast New York, with some stronger embedded convective cells over western New Jersey (Fig. 18a). During the next hour, as this precipitation area moved eastward into a region of greater instability across central New Jersey (Fig. 18b),

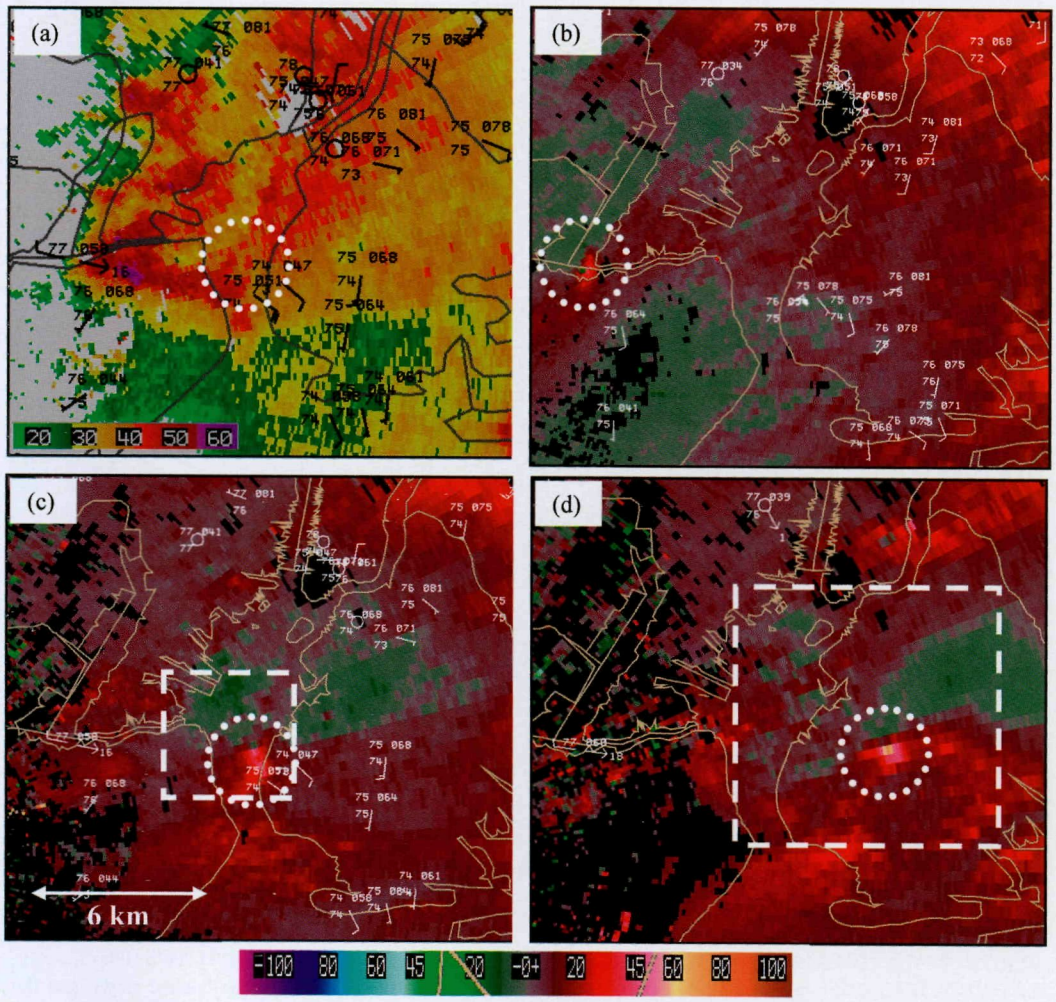

FIG. 13. (a) TEWR TDWR reflectivity (shaded in dBZ) and available surface stations (in °F) at 1029 UTC 8 Aug 2007. TJFK TDWR base velocity (shaded in kt) at (b) 1017, (c) 1029, and (d) 1035 UTC 8 Aug 2007. The circulation center is marked by the dashed circle. The dashed boxes in (c) and (d) are for the regions plotted in Figs. 14a,b and 14c, respectively.

more intense cells (to 50 dBZ) developed along its eastern side. At the surface at 2000 UTC 16 September (Fig. 19), there was a mesolow (~1010 hPa) over western New Jersey, with the warmest temperatures (27°–30°C) extending northward along the coastal plain from Maryland to central New Jersey. Meanwhile, a 2°–3°C temperature difference and wind shift (5–10 m s^{-1} southerlies to weak northerlies) existed across NYC from south to north. By 2112 UTC 16 September, the convection along the eastern side organized into a line (to ~60 dBZ) 20–30 km to the west of NYC (Fig. 18c).

The EWR and JFK TDWRs were used to investigate the detailed evolution of this convective line as it moved across NYC. At 2112 UTC 16 September using the TEWR radar (Fig. 20a), there was an eastward bulge in the line echo pattern approaching southwest Staten Island. The Doppler velocities at 2112 UTC indicate a relatively weak circulation pattern (~15 m s^{-1} gate-to-gate shear) at this location (Fig. 20b). In just 5 min (2117 UTC), this reflectivity bulge wrapped up into a well-defined spiral hook (Fig. 20c), with a well-defined velocity couplet (40 m s^{-1} gate-to-gate shear) (Fig. 20d). The TEWR radar was critical in the observation of this circulation given its rapid 1-min scans and high spatial resolution. For example, the WSR-88D radar from KDIX (~45 km to the south-southeast and a center beam height of ~800 m at this circulation location) did not have a well-defined hook or circulation center (Figs. 20e,f). There was no confirmation of tornadic damage related to this couplet; however, there were spotter reports of multiple funnel clouds approximately 5 min earlier as the vortex was near the southwest tip of Staten Island.

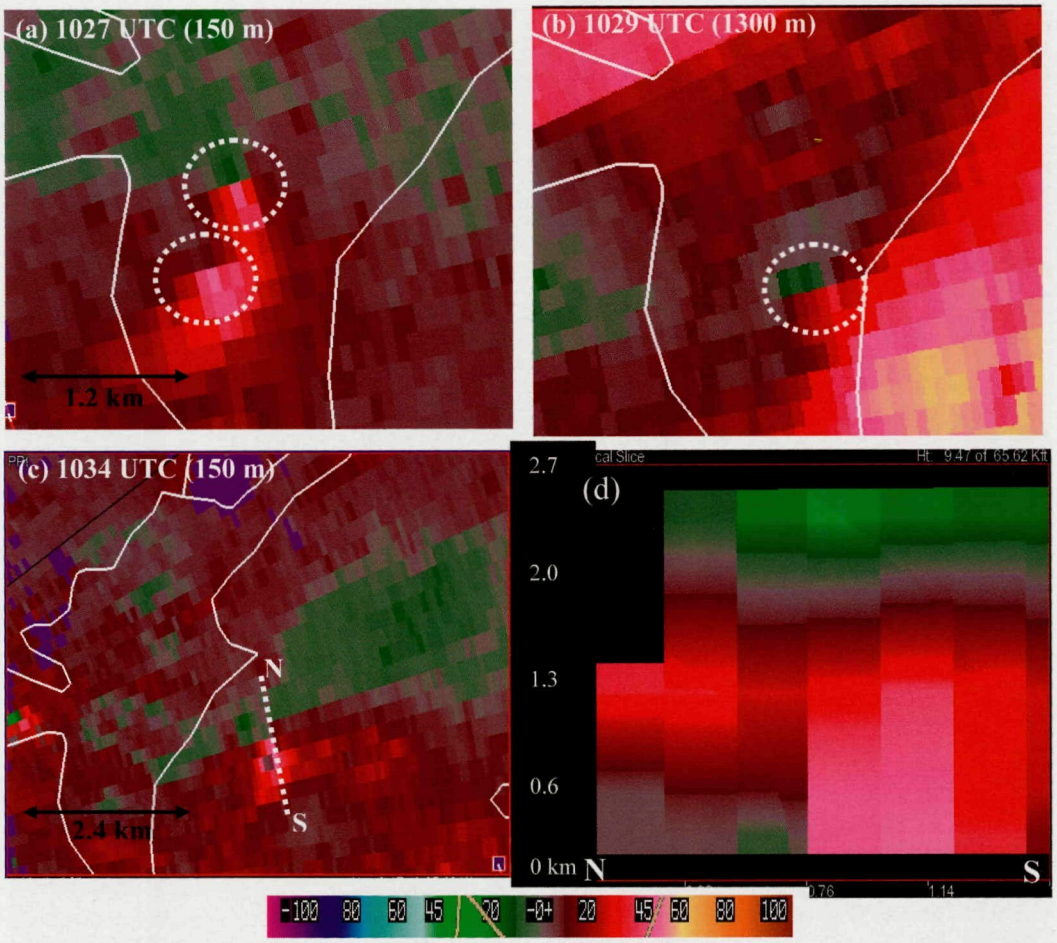

FIG. 14. TEWR TDWR base velocity (shaded in kt) at (a) 1027 UTC at 150 m AGL, (b) 1029 UTC at 1300 m AGL, and (c) 1034 UTC at 150 m AGL on 8 Aug 2007. The circulation center is located by the dashed circle, and the locations for these regions are shown in Fig. 13. (d) North–South (NS) cross-section [location shown in (c)] indicating base velocity from 0 to 2.7 km AGL.

Meanwhile, at 2117 UTC 16 September, there was another bulge in the convective line and circulation center over northwest Staten Island as seen from the TJFK radar (Figs. 21a,b). This northern Staten Island vortex wrapped up into a well-defined hook echo by 2126 UTC (Figs. 21c,d), with strong inbound (-45 m s^{-1}) and outbound (20 m s^{-1}) base radial velocities from TJFK. At this time a 107-kt wind gust (55.0 m s^{-1} where 1 kt = 0.514 m s^{-1}) was recorded at Robbins Reef (see RR in Fig. 21d), which is the highest reported wind ever recorded in the five burroughs of NYC for a convective event (NOAA/NWS, Upton). During the next 30 min this mesocyclone moved across New York Harbor and into Brooklyn and Queens, producing damaging winds and the EF0–EF1 tornadoes (Fig. 1). At 2135 UTC, there was a well-defined eastward bulge in the reflectivity pattern associated with an outbound–inbound velocity couplet (Figs. 21e,f), which was somewhat weaker than 15–20 min earlier. The horizontal scale of this mesocyclone (3–4 km) was twice as large as the 8 August event (1–2 km). The mesocyclone in the September event was also deeper than the August case (Figs. 22a,b), with a well-defined circulation pattern at 1500 m and extending up to at least 2.5 km AGL in a velocity cross section (as high as the TEWR radar can measure given its cone of silence). This mesocyclone continued northeastward and slowly weakened across the north shore of Long Island (not shown).

Overall, the tornadoes for this September 2010 event occurred along a QLCS, and were not embedded in the southwest quadrant of the MCS as in the 8 August 2007 event. Tornadoes along QLCSs have been documented over the central United States (Trapp et al. 2005). This case is consistent with the results of Lombardo and Colle (2011), which showed that just as many coastal northeast

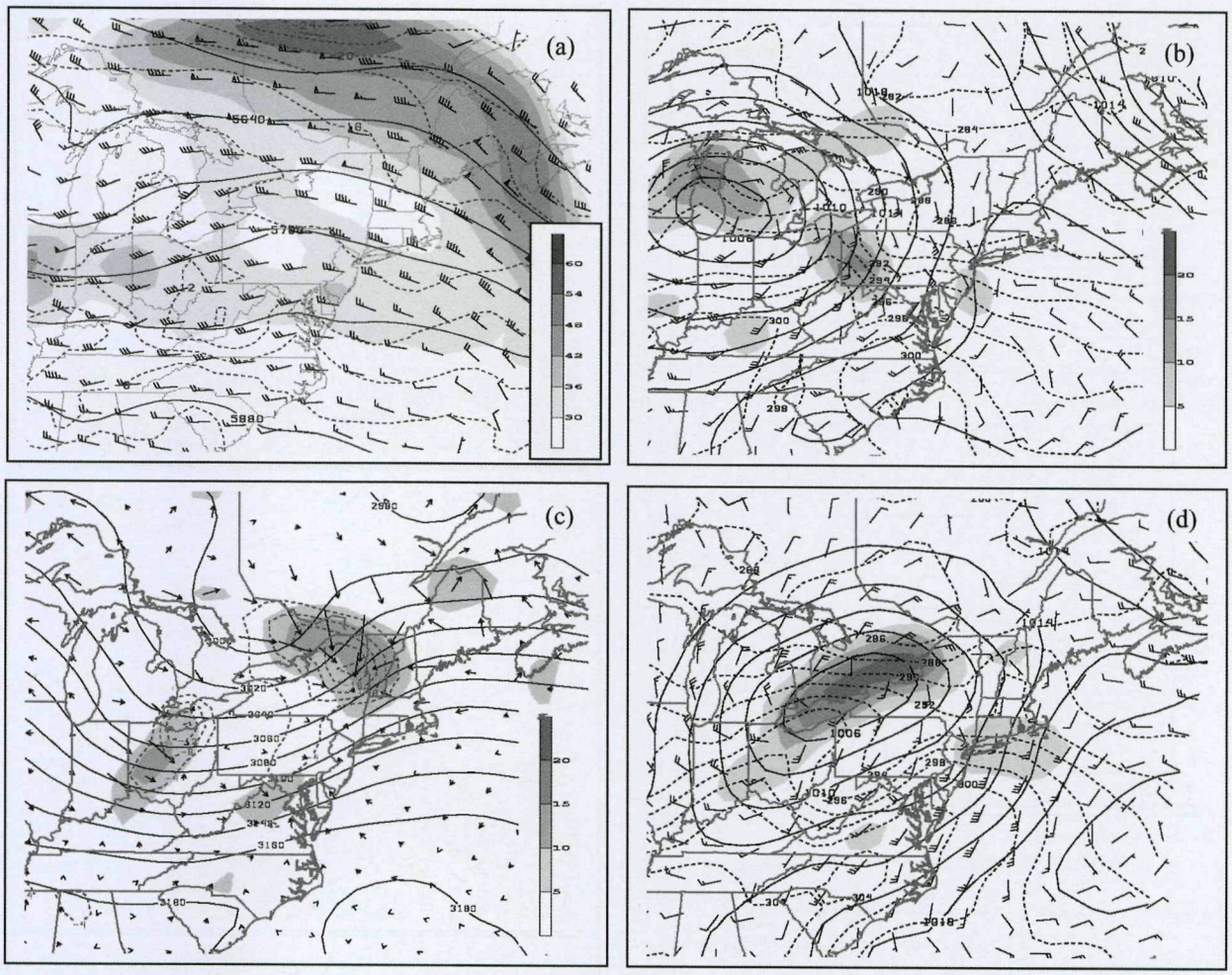

FIG. 15. (a) The 500-hPa geopotential heights (solid every 60 m), winds (full barb = 10 kt), and 300-hPa wind speed at 1200 UTC 16 Sep. (b) Sea level pressure (solid, every 2 hPa), 950-hPa potential temperature (dashed, every 2 K), frontogenesis [shaded, 1 K (100 km)$^{-1}$ (3 h)$^{-1}$ $\times$ 10^{-1}] and wind barbs at 1200 UTC 16 Sep. (c) The 700-hPa geopotential height (solid, every 20 m), **Q** vectors, **Q**-vector convergence (shaded, in K m^{-1} s^{-1} $\times$ 10^{-11}), and upward motion (dashed, every 1 m s^{-1} $\times$ 10^{-3}) at 2100 UTC 16 Sep. (d) As in (b), but at 2100 UTC 16 Sep.

U.S. tornadoes occurred for QLCSs as cellular storms (isolated supercells).

6. Discussion and summary

Although tornadoes around the New York City (NYC) metropolitan region are relatively rare as compared to the midwestern United States, they can occur and do cause significant damage. This paper describes the evolution of two tornado events on 8 August 2007 and 16 September 2010, both of which spawned two tornadoes across the NYC region and western Long Island (LI). The results from these case studies are compared to a longer-term composite evolution of 20 events across NYC–LI from 1979 to 2010, and a monthly and diurnal climatology of tornadoes was constructed for this region from 1950 to 2010.

This study also highlights the benefits of using some new observational datasets for monitoring convection in this urban coastal environment, such as the two terminal Doppler weather radars near John F. Kennedy International Airport (JFK) and Newark Liberty International Airport (EWR), aircraft soundings (ACARS), and numerous nonconventional surface observing stations. The Fort Dix (KDIX) WSR-88D is typically the first to observe rotating storms approaching NYC from central New Jersey given some of the range limitations of the TDWRs; however, the low tilt angles and rapid 1-min scans from the TDWRs allowed forecasters to issue tornado warnings with more confidence and monitor the evolving mesocyclones in the two NYC cases.

The 8 August 2007 event was associated with an MCS that formed 2 days prior in advance of a midlevel

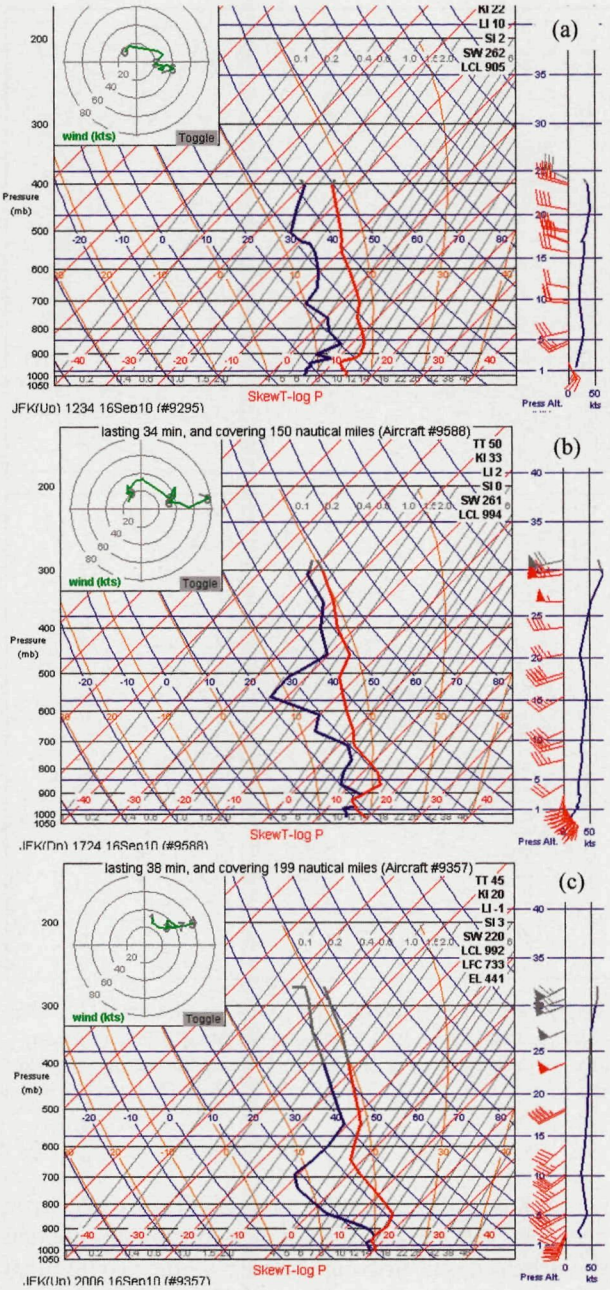

FIG. 16. ACARS temperature (red), dewpoint (blue), and wind profile at JFK at (a) 1234, (b) 1724, and (c) 2006 UTC 16 Sep.

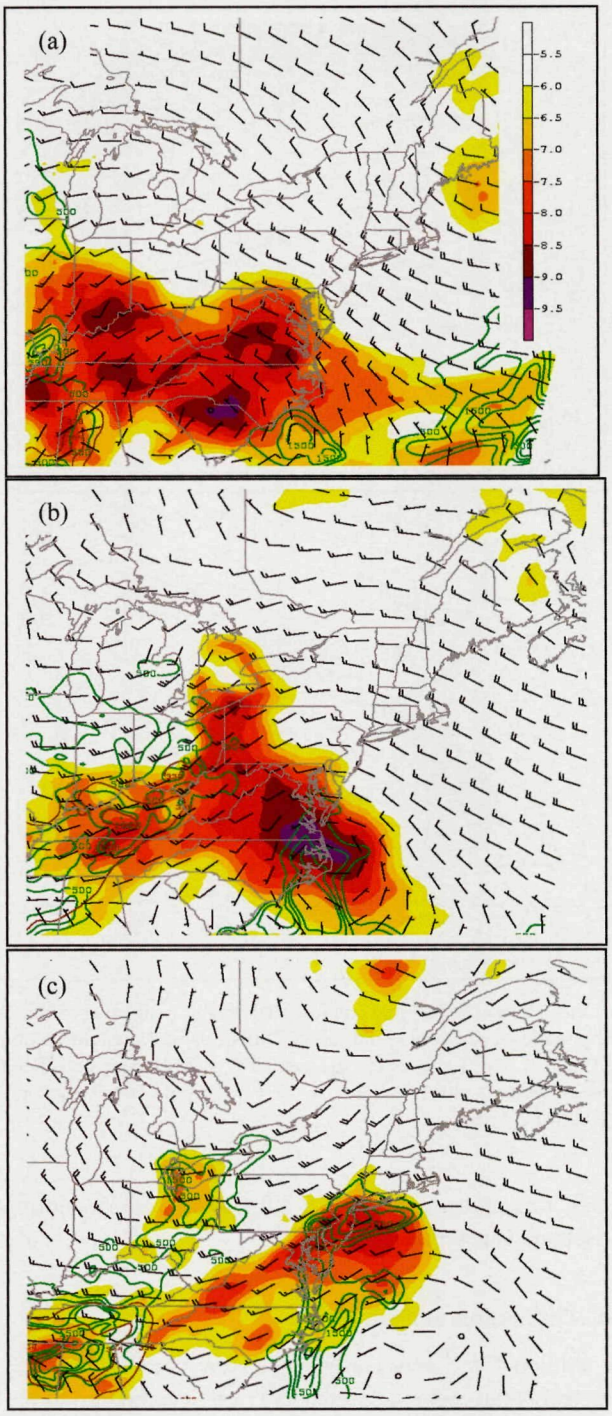

FIG. 17. Average lapse rate between 850 and 700 hPa (color shaded), MUCAPE (green, every 500 J kg^{-1}), and 800-hPa winds (full barb = 10 kt) at (a) 0000, (b) 1200, and (c) 2100 UTC 16 Sep 2010.

short-wave trough embedded in west-northwest flow. Supported by midlevel quasigeostrophic ascent and low-level frontogenesis, the MCS moved eastward along a surface baroclinic zone causing severe weather from Iowa to Ohio. Upon encountering the complex Appalachian terrain, the MCS weakened as the convection moved ahead of the midlevel forcing and vertical shear. During the overnight hours of 8 August, the MCS reintensified in the immediate lee of the terrain along a new frontogenetical boundary and associated surface mesolow development. There was enhanced near-surface warm advection, frontogenesis, and cyclonic shear near the region of tornadogenesis. At this time,

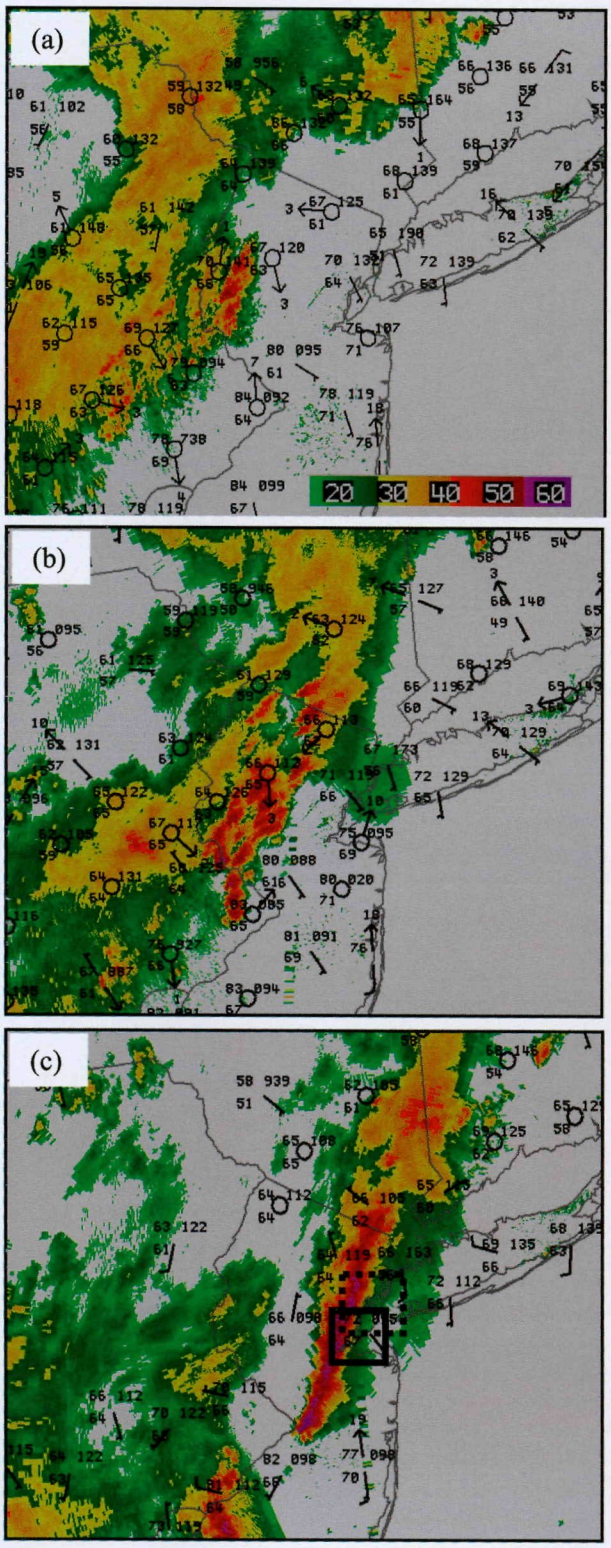

FIG. 18. Regional mosaic of WSR-88D reflectivity (shaded in dBZ) for the 0.5° scan at (a) 1924, (b) 2024, and (c) 2112 UTC 16 Sep 2010. Available surface observations are also shown. The solid and dashed boxes in (c) are the radar regions shown in Figs. 20 and 21, respectively.

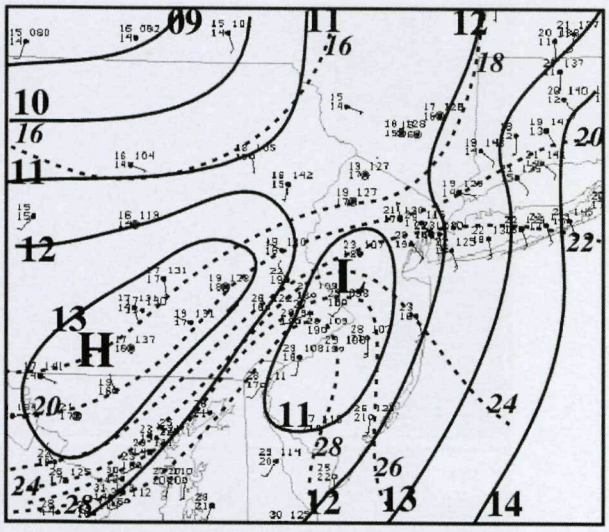

FIG. 19. Manual surface analysis showing sea level pressure (solid, every 1 hPa) and surface temperature (dashed, every 2°C) at 2000 UTC 16 Sep 2010.

there was midlevel forcing over eastern Pennsylvania with MUCAPE values approaching 1500 J kg^{-1}. A mesocyclone developed on the southwest boundary of the heavy precipitation shield, with a tornado causing EF1 and EF2 damage across eastern Staten Island and Brooklyn, respectively.

The 16 September 2010 QLCS also developed along the leading edge of the cold pool and interacted with a west–east baroclinic zone in advance of an upper-level short-wave trough and surface cyclone over the lower Great Lakes. The development of a southerly low-level jet ahead of this system led to rapid destabilization and frontogenesis over the NYC area on 16 September. The QLCS intensified rapidly as it approached NYC late in the afternoon, and low-level vortices developed as the line bowed outward in a few places.

An intriguing aspect of the August 2007 event was that the MCS redeveloped and crossed the Appalachians at night, and tornadogenesis occurred during the diurnal minimum of heating (around 0600 EDT). A climatology of all events revealed that nearly half (18 of 34 events) of NYC–LI tornadoes developed between 0900 and 1800 UTC (0500–1300 EDT). Parker and Ahijevych (2007) found that a successful MCS crossing of the Appalachians was favored during periods of higher instability around peak diurnal heating. Instability downstream of the mountains was also important in maintaining the MCS across the barrier (Letkewicz and Parker 2010). The development of a cold pool in the lee of the barrier can also help regenerate convection (Frame and Markowski 2006). For the 8 August event and several of the composite events,

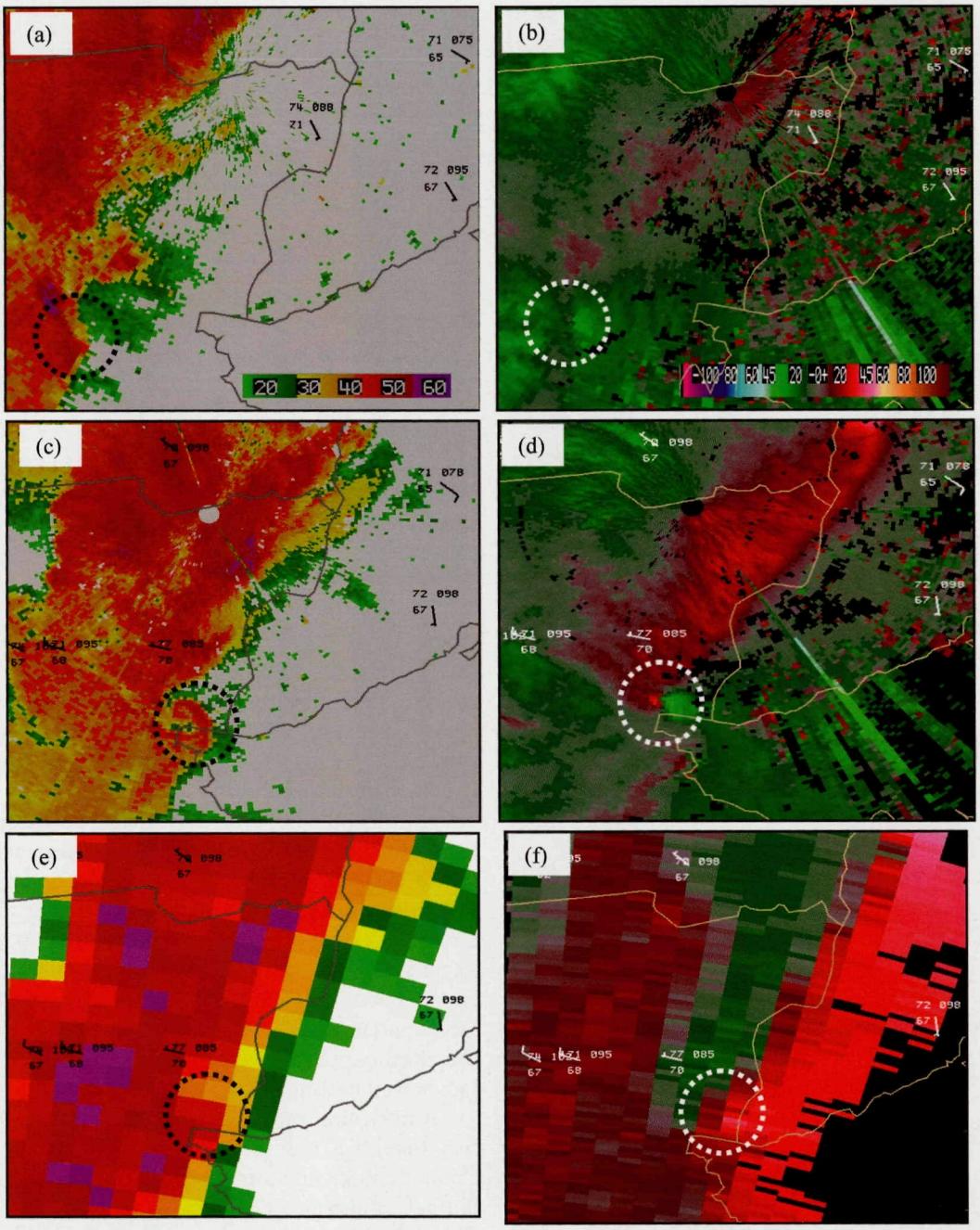

FIG. 20. (a) Radar reflectivity (shaded in dBZ) and (b) base radial velocity (shaded in kt) from TEWR at 2112 UTC 16 Sep for the 0.3° elevation scan for the small solid/dashed region in Fig. 18c. (c),(d) As in (a),(b), but at 2117 UTC 16 Sep 2010. The dashed circled region shows the circulation center developing along the convective line. The center beam height from TEWR to the circulation is 140 m ASL for (c) and (d). (e) As in (c), but for the KDIX WSR-88D. (f) As in (d), but for the KDIX WSR-88D. The center beam height from KDIX to the circulation center (shown by the white dashed circle) 55 km to the NNE is 850 m ASL.

the convection reintensified during the overnight hours so there was no diurnal heating, and there is little evidence of a well-defined cold pool. Rather, both NYC convective systems redeveloped as low-level warm advection (16 September) or evaporative cooling and adiabatic cooling aloft (7–8 August) destabilized the column in a region of favored synoptic ascent and relatively strong low-level wind shear.

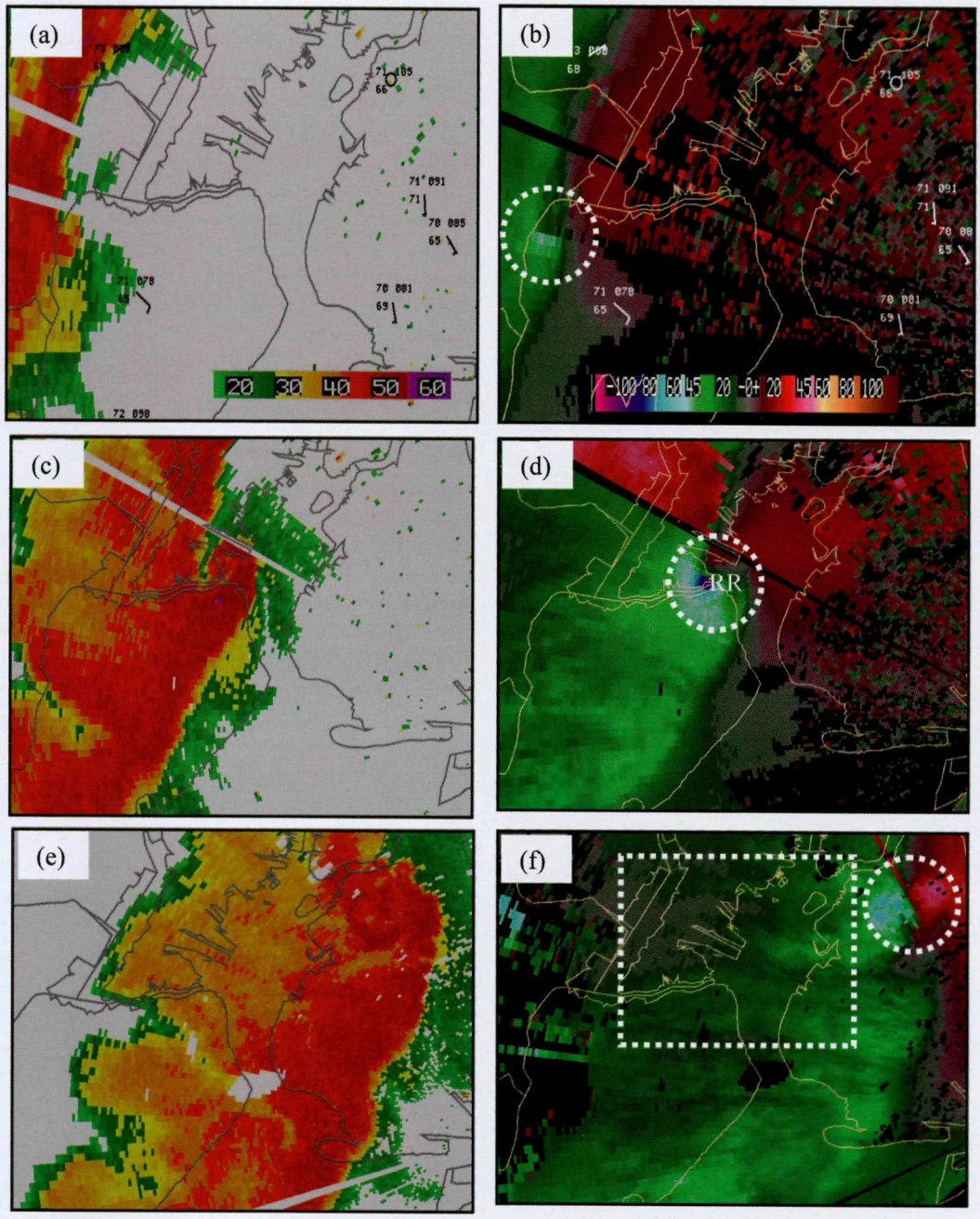

FIG. 21. As in Fig. 20, but for the TJFK terminal Doppler radar (0.5° scan) and plotted for the small dashed boxed region in Fig. 18c showing reflectivity (shaded in dBZ) at (a) 2117, (c) 2126, and (e) 2135 UTC 16 Sep and base radial velocities (shaded in kt) at (a) 2117, (c) 2126, and (e) 2135 UTC 16 Sep. The dashed circled region shows the circulation center developing along the convective line. Robbins Reef is shown by RR in (d), which recorded a 107-kt wind gust. The dashed box in (d) is for the region shown in Fig. 22a.

During the August 2007 event, there was an elevated mixed layer (EML) between 800 and 600 hPa that originated from the upper Midwest 2 days prior under west-northwest flow aloft. A steep lapse rate aloft also developed over NYC during the 16 September event between 850 and 750 hPa under southwesterly flow. EMLs have been shown to be important in destabilizing the column for northeastern U.S. severe weather (Banacos and Ekster 2010). During the 8 August event, upward motion (adiabatic cooling) and evaporation of precipitation at the leading edge of the MCS led to cooling at midlevels within the EML, thus helping to steepen the lapse rates at low levels.

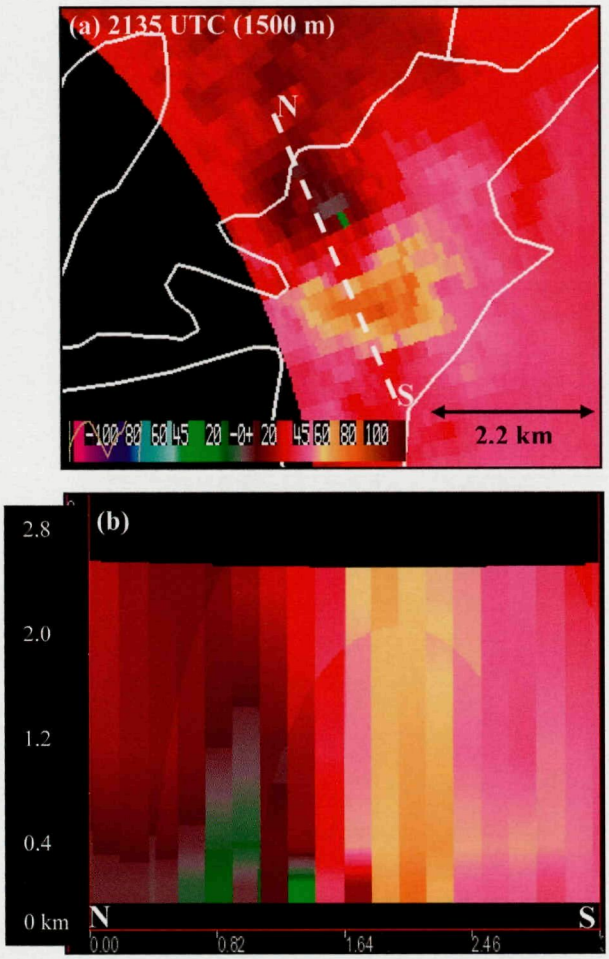

FIG. 22. TEWR TDWR base velocity (shaded in kt) for 1500 m AGL at 2135 UTC 16 Sep. The location of this region is shown in Fig. 21f. (b) NS cross-section [location shown in (a)] showing base velocity from 0 to 2.8 km AGL.

The composite analysis of all NYC–LI tornado events revealed that there is often an Appalachian lee trough and baroclinic zone that extends west–east across the region, similar to the August 2007 and September 2010 events. The low-level advection of heat and moisture ahead of this lee trough helps to destabilize the column, while frontogenesis helps trigger the convection. Several studies have shown the importance of low-level mesoscale boundaries for the development of tornadic supercells (Maddox et al. 1980; Markowski et al. 1998; Rasmussen et al. 2000). Maddox et al. (1980) diagnosed the role of low-level convergence and the generation of vertical vorticity within a thin band along the warm side of a thermal boundary. Both the 8 August and 16 September events occurred along a broad baroclinic zone; thus, this thermal and wind boundary may have played an important role in enhancing the vorticity within the convective system.

The diurnal climatology suggests that there may be different environments supporting tornadoes around NYC–LI as compared to just 40–50 km to the north over the mainland. It is hypothesized that the ambient conditions over the NYC–LI region are more maritime than coastal Connecticut, since NYC–LI experiences a more direct influence with southerly flow (sea breezes) from the ocean. First, this causes the peak in annual activity on LI to occur later in the season (August) when the surrounding Atlantic waters are at their warmest. Convective systems moving eastward often weaken upon encountering the cooler New York coastal waters during June–July (Murray and Colle 2011). In contrast, tornadoes occur in coastal Connecticut near the diurnal heating maximum, a trend more characteristic of continental region. The relatively large number of morning tornadoes over NYC–LI suggests that the diurnal heating plays a smaller role in tornadic events at the coast. Rather, unstable air can advect up the coast or an EML can move in aloft during the early morning and the midlevels can become further destabilized from evaporation and adiabatic ascent, as in the 8 August 2007 tornado event. Furthermore, Murray and Colle (2011) showed that convective systems tend to move across the Appalachians and reach the coastal waters around LI by the late night, which is similar to the 8 August event.

The composite patterns of the evolution of both the surface mass field and the instability are similar for the 8 August and 16 September events (Figs. 6b and 12d). The relatively low to moderate values of MUCAPE indicate that large instability is not necessary for tornado development in the NYC area. In both the composite and the individual events, a zonally (east–west) oriented baroclinic zone is present across northern New Jersey and NYC at the time of tornadogenesis, which suggests that baroclinic processes are important for rotating storms in the region. Similar features have been found for tornadic events in central Florida (Wasula et al. 2007), in which the east–west baroclinic zone served as low-level frontogenetical forcing for ascent as well as a potential source of cyclonic vorticity. Also, they noted the importance of surface mesolows along the front, which can locally enhance the vorticity and warm advections. The NARR composite cannot capture these small-scale features approaching NYC–LI, but surface observations in both case studies suggested that there were surface mesolows along the baroclinic zone (Figs. 6 and 16).

Lombardo and Colle (2011) showed that severe weather over the northeastern United States can occur for a wide variety of MUCAPE environments; however, the lower MUCAPE severe events (<500 J kg^{-1}) are typically associated with relatively large dynamical

forcing. The NYC tornado case studies and composites illustrate that there were moderate MUCAPE values (1000–1500 J kg^{-1}), relatively large low-level vertical shear, and synoptic forcing either from quasigeostrophic ascent at midlevels or low-level frontogenesis. The composite analysis illustrated that many tornadic events occur near the right-entrance region of an upper-level jet. With the approach of an upper-level trough and front, there is also relatively strong low-level shear for these NYC tornado events, and this shear may be enhanced by increased cyclonic flow with mesolow development near the surface, such as during the 8 August 2007 and 16 September 2010 events. The importance of dynamical forcing under relatively weak MUCAPE and vertical shear conditions is consistent with the findings of Markowski and Straka (2000), Monteverdi et al. (2003), and Hanstrum et al. (2002).

The 16 September tornado event was associated with the development of mesovortices and a bowing of the QLCS as it moved over NYC. One question is whether the coastal boundary enhanced this low-level vortex development. Wheatley and Trapp (2008) showed that the vortices with a QLCS can develop in homogeneous conditions through the tilting of crosswise horizontal baroclinic vorticity in downdrafts, as revealed by the idealized simulations of Trapp and Weisman (2003). Thus, the role of the coast may have been secondary for the NYC events, but it is interesting that for both cases the mesocyclone spun up rapidly approaching the coast near western Staten Island. One could speculate that the enhanced horizontal shear (vorticity) in the winds between water and land may enhance vortex development. Also, the relatively warm water compared to the cooler land during the late night or early morning of late summer may provide some additional low-level baroclinicity for vortex generation. Separation of the coastal effects and internal storm dynamics on rotating storms in this urban-coastal region will require more careful analysis and high-resolution model simulations.

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