Ambient Conditions Associated with the Maintenance and Decay of Quasi-Linear Convective Systems Crossing the Northeastern U.S. Coast

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ABSTRACT

Quasi-linear convective systems (QLCSs) crossing the Atlantic coastline over the northeastern United States were classified into three categories based on their evolution upon encountering the coast. Composite analyses show that convective lines that decay near the Atlantic coast or slowly decay over the coastal waters are associated with 900–800-hPa frontogenesis, with greater ambient 0–3-km vertical wind shear for the slowly decaying lines. Systems that maintain their intensity over the coastal ocean are associated with 900-hPa warm air advection, but with little low-level frontogenetical forcing. Neither sea surface temperature nor ambient instability was a clear delimiter between the three evolutions. Sustaining convective lines have the strongest environmental 0–3-km shear of the three types, and this shear increases as these systems approach the coast. In contrast, the low-level shear decreases as decaying and slowly decaying convective lines move toward the Atlantic coastline. There was also a weaker mean surface cold pool for the sustaining systems than the two types of decaying QLCSs, which may favor a more long-lived system if the horizontal vorticity from this cold pool is more balanced by low-level vertical shear.

1. Introduction

Observational evidence suggests that organized convective systems can be modified by water bodies, such as lakes (e.g., Bosart and Sanders 1981; Bosart and Galarneau 2005) as well as coastal oceans (e.g., Mapes et al. 2003; Lericos et al. 2007; Murray and Colle 2011). For example, the mesoscale convective complex responsible for the 1977 Johnstown, Pennsylvania, floods temporarily weakened while traversing the relatively cool waters of the Great Lakes, and subsequently reintensified over Pennsylvania (Bosart and Sanders 1981). During the 2003 Bow Echo and MCV experiment (BAMEX; Davis et al. 2004), the western portion of an intensively observed squall line that crossed over Lake Erie decayed more rapidly than the eastern section of the convective line that remained over Ohio (Bosart and Galarneau 2005).

Recent research has sought to identify the most influential ambient parameters on evolving organized convective storms crossing a land–water boundary. In a climatological study, Workoff (2010) analyzed the effect of Lake Erie on preexisting convective storms, evaluating a number of parameters, including the land–lake air temperature differences highlighting variations in boundary layer stability, the strength of the convectively generated cold pools, as well as vertical wind shear. No one parameter could predict the survival or decay of a quasi-linear convective system crossing over the lake, though a moderate correlation existed between weakening systems and 0–3-km ambient vertical wind shear values less than 15 m s^{-1} (Workoff 2010). Lericos et al. (2007) illustrated similar results utilizing 2D idealized simulations of squall lines moving onshore from an ocean environment to inland areas. Horizontal variations in low-level wind shear associated with the changing underlying surface were more influential on the evolving convective lines than the variations in ambient thermodynamic properties moving toward the inland. However, they noted the limitations of 2D simulations when reproducing natural convective phenomena, and stressed the need for a 3D idealized modeling study as well as comparisons to observations.

In a study of northeastern U.S. convection, Lombardo and Colle (2010) illustrated that convective development

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declines eastward from the inland areas to the Atlantic waters, with a minimum in convective initiation over the coastal ocean. Murray and Colle (2011) also highlighted a sharp decrease in the climatological convective storm frequency near the Atlantic coast, with the largest gradient during the daytime. They also showed that southerly surface winds from the relatively cool Atlantic are associated with reduced convection over much of southern New England, while coastal severe events occur for west-southwesterly low-level flow ahead of an approaching trough (Murray and Colle 2011). While this provides some insight into the flow regimes that favor coastal ocean storms, it only provides a broad synoptic perspective.

With a limited number of studies explicitly examining the impact of large water bodies on convective systems, relatively little is known about the interaction between organized storms and the marine environment. This paper seeks to elucidate the dynamic and thermodynamic properties that control the evolution of organized convective storms crossing a land–water boundary by examining northeastern U.S. quasi-linear convection interacting with the coastal Atlantic Ocean. More specifically, we aim to explain why some mesoscale convective systems (MCS) decay upon reaching a coastal boundary, while others successfully traverse the coastline and continue over the relatively cool water uninfluenced for hours. Furthermore, since linear convection is the most likely convective structure to produce severe weather over the northeastern United States (Lombardo and Colle 2011), it is important we understand the processes that lead to the decay or survival of quasi-linear convective systems (QLCSs) in this densely populated coastal area.

As illustrated by Murray and Colle (2011), warm season northeastern U.S. convective storms are less likely over the Atlantic waters in May–June than July–August, since sea surface temperatures (SSTs) in May (9°–13°C) are cooler than August (21°–25°C; not shown). Cooler SSTs promote a stronger surface-based marine layer and increased low-level stability. Contradictory to this result, Fig. 1 uses radar reflectivity to illustrate a weakening QLCS approaching the coast in midsummer on 23 July 2002 (Figs. 1a–c) and a QLCS sustaining its intensity in late spring on 31 May 2002 (Figs. 1g–i). Metz (2011) showed similar examples of mesoscale convective systems successfully crossing Lake Michigan early in the warm season, while lake waters were cooler than the surrounding land. This juxtaposition implies that SSTs and the associated atmospheric stability profile may not always be the dominant variable regulating the evolution of coastal quasi-linear storms.

2. QLCS evolution classification

To understand the evolution of quasi-linear storms over the Northeast coastal region during the warm season (May–August), a dataset was constructed of QLCSs that crossed the Atlantic coastline over the domain defined by Fig. 2. By manually examining 2-km National Operational Weather radar (NOWrad) reflectivity images at 15-min intervals over 6 warm seasons from 2002–07, 59 QLCS events were identified. By definition, a linear system must be at least 50 km in length and exhibit a continuous line of ≥50-dBZ radar reflectivity, or a line of ≥35-dBZ radar reflectivity with embedded ≥50-dBZ reflectivity cores, with a length to width ratio of 5:1 (Lombardo and Colle 2010).

All 59 QLCS events weaken near the coast or farther offshore (Fig. 2), with weakening defined as a decrease in radar reflectivity (dBZ) in the leading convective line. Some of this weakening for systems well offshore (>100 km) may be the result of increasing radar beam height, with a maximum horizontal radar beam range for the Weather Surveillance Radar-1988 Doppler (WSR-88D) of approximately 230 km. Coastal quasi-linear convective systems can weaken several different ways. The entire leading convective line can weaken uniformly below 50 dBZ. The center point of the line is defined as the decay location, and the decay time is the time the convection decreases below 50 dBZ (Fig. 3a). A QLCS may also decay nonuniformly, experiencing one of two evolutions (Figs. 3b,c). A segment of the leading convective line may weaken below 50 dBZ, with the center of the initial decaying segment as the decay location (Fig. 3b). A segment of the leading convective line may also initially break into a line of small, individual convective cells with ≥50 dBZ, and these cells subsequently decrease in intensity, visible in the 15-min reflectivity images. For this evolution, the decay time is selected as the initial time the convective line transitions from a solid line of ≥50 dBZ to individual cells ≥50 dBZ, and the location is the center point of the decaying segment (Fig. 3c).

The 59 events were classified into three categories based on their evolution. Thirty-two decaying events (Figs. 1a–c) weaken between 50 km inland of the coast and 20 km offshore. Slowly decaying events weaken over the ocean between 20–100 km of the coast (18 events; Figs. 1d–f). Nine sustaining events maintained their intensity >100 km offshore from the coast (Figs. 1g–i). An additional 14 QLCS events were not included in the analysis because they did not fit the selection criteria. They either intensified at the coast or farther offshore, transitioned to nonlinear systems, or moved perpendicular to the coastline.

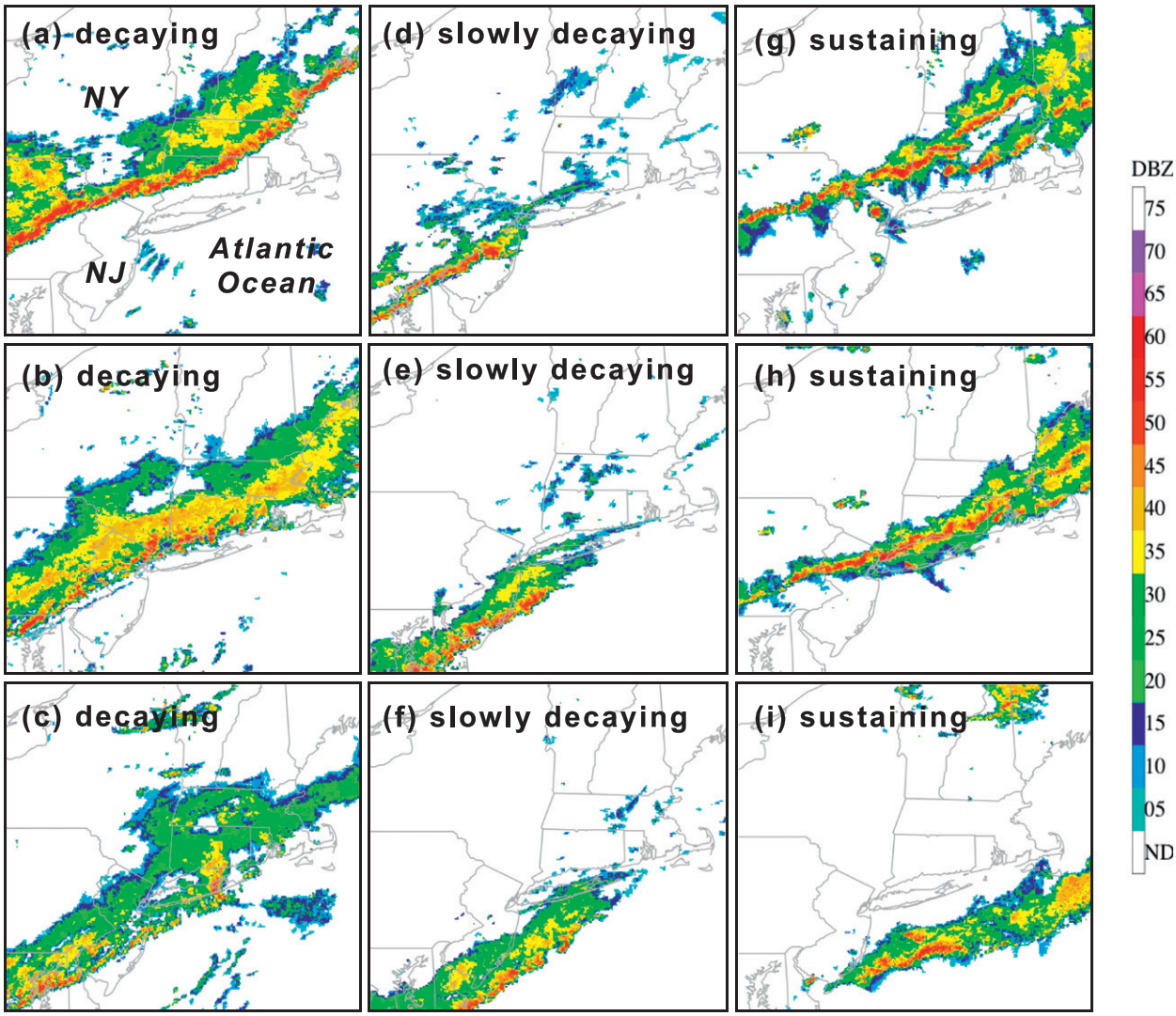

FIG. 1. 2-km NOWrad radar reflectivities (shaded every 5 dBZ) at (a) 2000 UTC 23 Jul 2002, (b) 2200 UTC 23 Jul 2002, (c) 0000 UTC 24 Jul 2002, (d) 2000 UTC 30 Aug 2003, (e) 2145 UTC 30 Aug 2003, (f) 2300 UTC 30 Aug 2003, (g) 2300 UTC 31 May 2002, (h) 0030 UTC 1 Jun 2002, and (i) 0430 UTC 1 Jun 2002.

An initial hypothesis was that the decay location for a QLCS was primarily a function of the SST, with linear systems early in the warm season having an increased chance of decay over relatively cooler ocean waters given the greater surface-based stability compared to late summer. As illustrated in Figs. 2 and 4, QLCS decay locations do not appear to be exclusively related to the warm season month (Fig. 2), with a similar number of decaying/slowly decaying systems in May–June (23 events; 89%) as July–August (27 events; 82%; Fig. 4). Since SSTs are not a clear delimiter when determining the longevity of coastal quasi-linear systems, composite analyses were created to highlight the dynamical forcing mechanisms and thermodynamic parameters associated with the three QLCS evolutions defined in this study.

3. Composite analysis

To examine the synoptic and thermodynamic differences between each of the three evolutions, feature-based composites were created using the 32-km North American Regional Reanalysis (NARR; Mesinger et al. 2006) centered on the point where the QLCS crossed the coastal boundary at the closest 3-h NARR time prior to the crossing. The point for the feature-based composite was chosen just ahead of the linear system (by 0.75°) to capture the ambient conditions

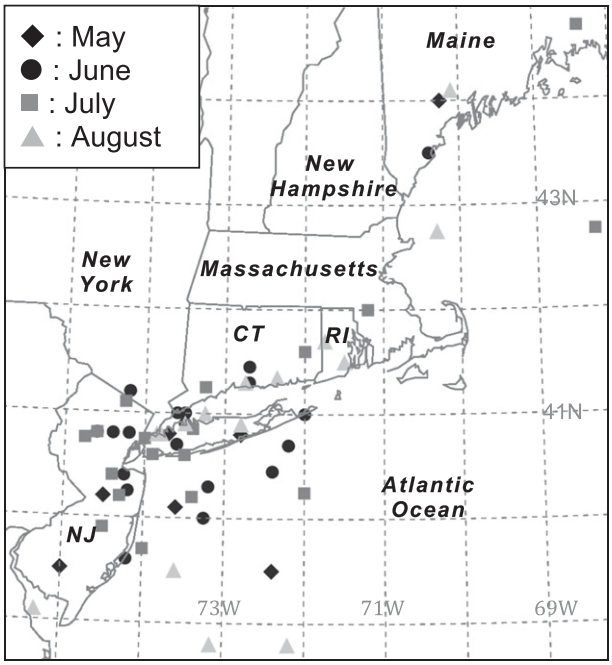

FIG. 2. Decay locations for 59 QLCSs over the northeastern United States from May–August 2002–07 categorized by month, with 7 May events, 19 June events, 18 July events, and 15 August events.

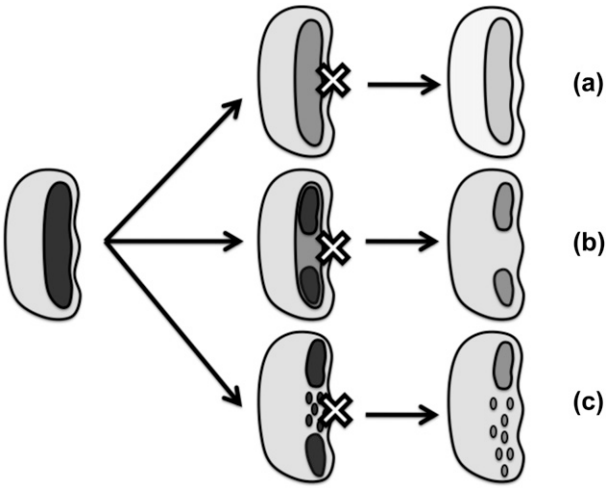

FIG. 3. Schematic depicting three different types of decay evolutions experienced by northeastern U.S. quasi-linear convective systems. The darkest gray shading represents radar reflectivity values ≥50 dBZ. The medium, light, and lightest gray shadings represent radar reflectivity values <50 dBZ, with descending reflectivity values progressing toward the lightest shading. The "X" marks the spatial and temporal point of decay ($T - 0$ h) for the quasi-linear convective system, as plotted in Fig. 2 and described in the text.

encountered by the QLCS while preventing convective contamination.

a. Decaying events (32 events)

For decaying lines (Fig. 5), the mean QLCS at $T - 0$ h (time of decay) is collocated with a surface pressure trough and a low-level thermal ridge (Fig. 5a). The mean surface trough is the result of two primary surface regimes. In ~38% of events (12 events), the decaying QLCS is collocated with a surface pressure trough, with a cold front 100–700 km to the west or north of the trough, represented as a composite surface trough northwest of the compositing point (Fig. 5a). Twenty-five percent of QLCSs (eight events) are collocated with a surface cold front. Five events (16%) have a surface cold front 200–400 km to the west or north of the convection. The remaining seven events are either collocated with a surface trough with no upstream cold front, collocated with a stationary front, associated with a surface mesolow, or under a surface high pressure system.

Decaying QLCSs are located within an axis of moderate instability [most unstable CAPE (MUCAPE) ~1200 J kg^{-1}], with 84% of decaying events crossing the coastal boundary during the warmest time of day (1800–0300 UTC; Fig. 6). This area of instability extends 200–300 km downstream of the mean QLCS location over the Atlantic waters. Average most unstable convective inhibition (MUCIN) values, while only ~−15 J kg^{-1} over the QLCS, increase steadily offshore to ~−60 J kg^{-1}, helping to suppress convection over the ocean (Fig. 5b). Surface-based CAPE (SBCAPE) and surface-based CIN (SBCIN) are comparable to MUCAPE and MUCIN, respectively, both spatially as well as their magnitudes.

The importance of low-level wind shear to the strength and longevity of quasi-linear convection has been well documented (e.g., Thorpe et al. 1982; Rotunno et al.

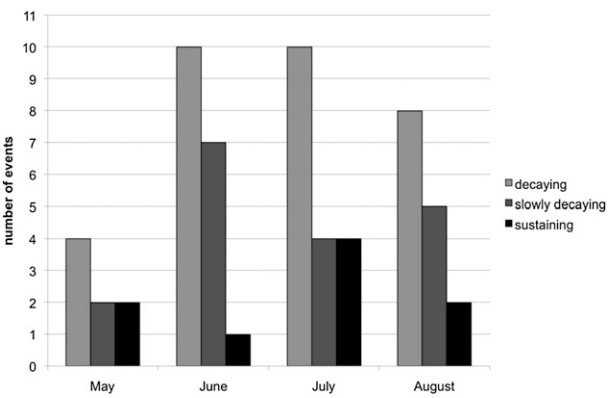

FIG. 4. Number of decaying, slowing decaying, and sustaining QLCSs categorized for May, June, July, and August.

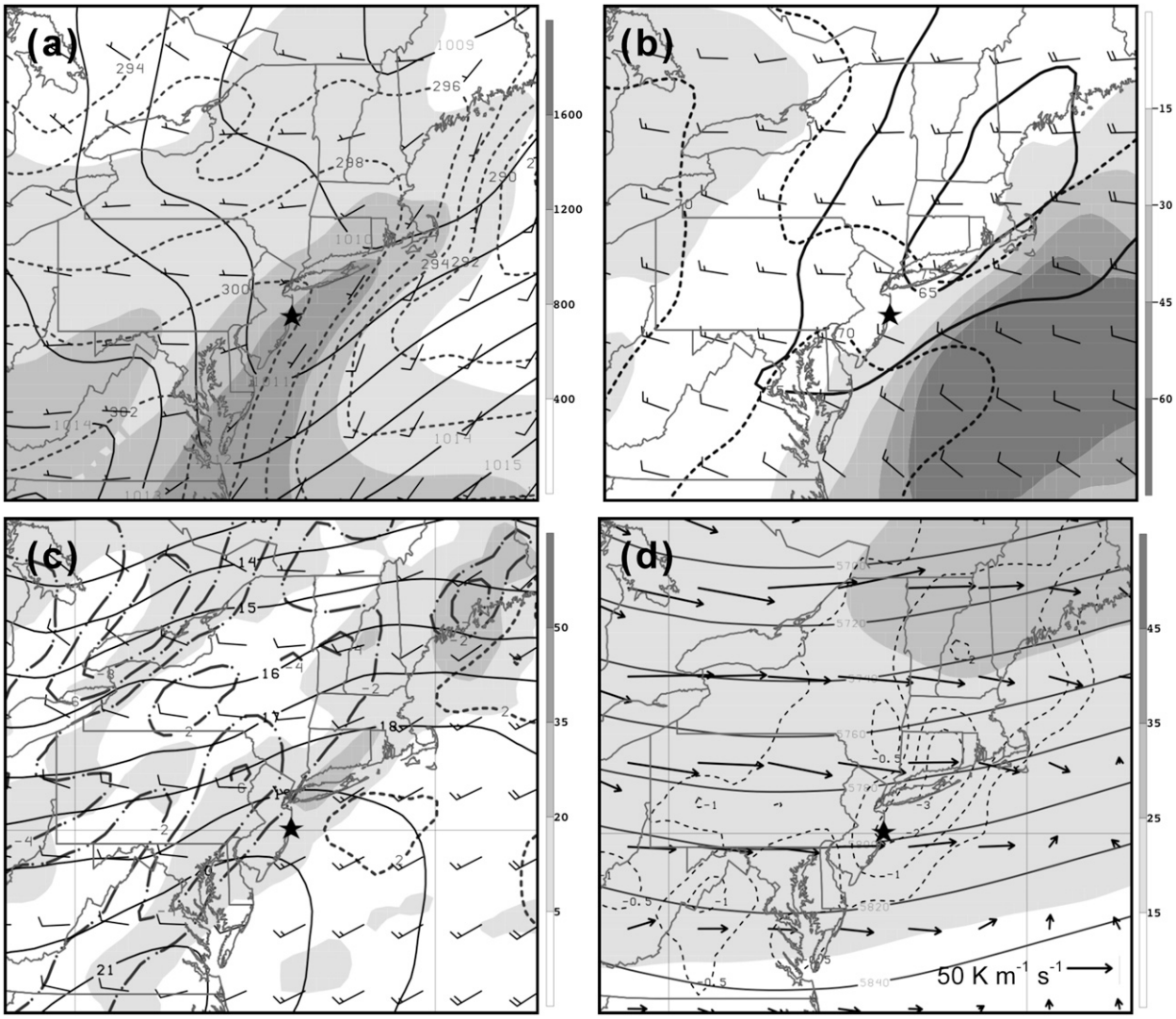

FIG. 5. Feature-based NARR composites for the decaying events showing (a) MSLP (solid every 1 hPa), 10-m potential temperature (dashed every 2 K), MUCAPE (shaded J kg^{-1}), 10-m winds (full barb = 5 m s^{-1}); (b) MUCIN (shaded J kg^{-1}), 1000–800-hPa layer-averaged relative humidity (dashed every 5%), 700–500-hPa layer-averaged relative humidity (solid every 5%), 0–3-km vertical wind shear vector (full barb = 5 m s^{-1}); (c) 900–800-hPa Miller (1948) frontogenesis [shaded every 15×10^{-2} K (100 km)$^{-1}$ (3 h)$^{-1}$, beginning at 5×10^{-2} K (100 km)$^{-1}$ (3 h)$^{-1}$], 900-hPa temperature advection (positive temperature advection dashed and negative temperature advection dash–dot every 2×10^{-5}°C s^{-1}), 900-hPa temperature (solid every 1°C), 900-hPa winds (full barb = 5 m s^{-1}); and (d) 300-hPa wind magnitude (shaded every 10 m s^{-1}), 500-hPa geopotential heights (solid every 20 dam), 800–400-hPa layer-averaged omega (dashed every 1×10^{-3} Pa s^{-1}), 500-hPa **Q** vectors (10^{-12} K m^{-1} s^{-1}). The MUCAPE and MUCIN use the maximum equivalent potential temperature (θ_e) from 0 to 180 hPa above the surface. Wind shear values represent a vector difference between two layers. A star indicates the compositing point with the geography included for scale reference.

1988; Weisman et al. 1988; Weisman and Rotunno 2004). For decaying QLCS events in this study, the mean 0–3-km vertical wind shear vector over and downstream of the convection is from the west-northwest at 7.7 m s^{-1} (Fig. 5b). This is smaller than the average 0–3-km shear that supports northeastern U.S. QLCSs (9.8 m s^{-1}; Lombardo and Colle 2010), thus promoting weakening of the system. The importance and implications of this weak vertical wind shear for decaying QLCSs will become apparent when compared to the vertical wind shear during slowly decaying and sustaining QLCS events in sections 3b and 3c.

The 1000–800-hPa and 700–500-hPa layer-averaged RH near the QLCS is 73% and 55%, respectively. The relatively dry midtroposphere suggests the potential for evaporative cooling and enhanced cold pool

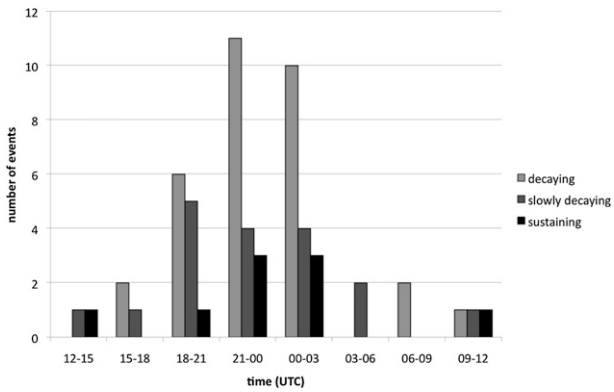

FIG. 6. Number of decaying, slowly decaying, and sustaining quasi-linear convective systems that encounter the northeastern U.S. coastline in this study, plotted as a function of time of day (3-h bins).

development (Hookings 1965; Johns and Doswell 1992; Gilmore and Wicker 1998). With only the NARR gridded data and surface aviation routine weather report (METAR) observations available for this study, the importance of the dry air aloft to cold pool development can only be speculated and will remain a topic for future exploration. Furthermore, the role of midlevel dry air in downdraft development is currently under debate. Using a 3D cloud model, James and Markowski (2010) showed that for quasi-linear systems with $\sim$1500 J kg^{-1} of CAPE, the presence of dry air aloft can be detrimental to the diabatic generation of cold pools, due to a decline in hydrometeor mass resulting from the presence of relatively dry air.

During decaying events, there is 900–800-hPa layer-averaged frontogenesis [$\sim 15 \times 10^{-2}$ K (100 km)$^{-1}$ (3 h)$^{-1}$] over the convective line (Fig. 5c), with the QLCS located on the warm side of the developing baroclinic zone within the ascending branch of the frontogenetical circulation (not shown). Limited temperature advection exists over the convective storms (Fig. 5c). At 1000 hPa, there is flow deformation collocated with a temperature gradient associated with the land–ocean boundary east of the compositing point over the marine waters (Fig. 5a), with a 1000-hPa frontogenesis value of $\sim 45 \times 10^{-2}$ K (100 km)$^{-1}$ (3 h)$^{-1}$ over the QLCS (not shown).

At midlevels, there is a broad 500-hPa trough located $\sim$600 km to the west of the QLCS (Fig. 5d), with weak **Q**-vector convergence (quasigeostrophic forcing) over the convection. Since quasigeostrophic (QG) forcing for ascent as depicted in the **Q**-vector formulation is inversely proportional to the ambient static stability, the presence of moderate instability (MUCAPE; Fig. 5a) suggests that less dynamical forcing is required to support deep-layer ascent during these events. In an environment with dry midlevels (55% RH) and more humid low levels (73%), lift can cause adiabatic cooling within the drier midlevels and less rapid diabatic cooling in the low levels, destabilizing the atmosphere. The collocation of the maximum of MUCAPE (Fig. 5a), 700–500-hPa relative humidity minimum (Fig. 5b), and the 800–400-hPa vertical motion (Fig. 5d) suggests that ascent within the region of midlevel dry air may have steepened lapse rates and enhanced instability over the northeastern U.S. coastal region. The ascending motions help to erode capping inversions through the profile as well. Furthermore, the convection is located in the right entrance region of a 25 m s^{-1} 300-hPa jet (Fig. 5d), a favorable region for ascent because of ageostrophic jet circulations. As indicated above, deep-layer ascent (800–400-hPa omega $\sim -2 \times 10^{-3}$ Pa s^{-1}) is present over the QLCS, which is collocated with both the low-level frontogenetical forcing (Fig. 5c), midlevel QG forcing for ascent, and the upper-level jet entrance region (Fig. 5d), implying their contributions to the large-scale vertical motion. Regardless of the forced ascent as well as the favorable instability, the QLCSs decay.

b. *Slowly decaying events (18 events)*

QLCS events that slowly decay upon crossing the Atlantic coast are located within the axis of a surface pressure trough, which is collocated with a thermal ridge (Fig. 7a), similar to the decaying QLCS composite. However, inspection of the individual events reveals that a majority (67%; 12 events) of slowly decaying QLCSs cross the coastal boundary while close to a surface mesolow and the associated surface boundaries, with only 2 events displaying convection farther than 200 km from the surface low. The remaining five events are either collocated with a surface trough (one event) or a cold front (two events), 50 km north of a warm front (one event), or 200 km south of a warm front with a trough 200 km to the west (one event). Therefore, slowly decaying QLCSs organize within a close proximity to robust surface features and surface boundaries, more so than decaying convective lines.

These slowly decaying systems weaken more slowly even though the average most unstable CAPE (MUCAPE $\sim$ 800 J kg^{-1}; Fig. 7a) is $\sim$25% smaller than for decaying QLCS events (Fig. 5a). The magnitude and spatial distribution of SBCAPE (not shown) are very similar to MUCAPE. A slightly smaller percentage (72%) of slowly decaying QLCSs cross the coast between 1800–0300 UTC, compared to the percent of decaying events (Fig. 6), complementing the weaker instability. The offshore gradient of MUCIN (Fig. 7b) is similar to decaying events, ranging from -10 to -15 J kg^{-1} over the

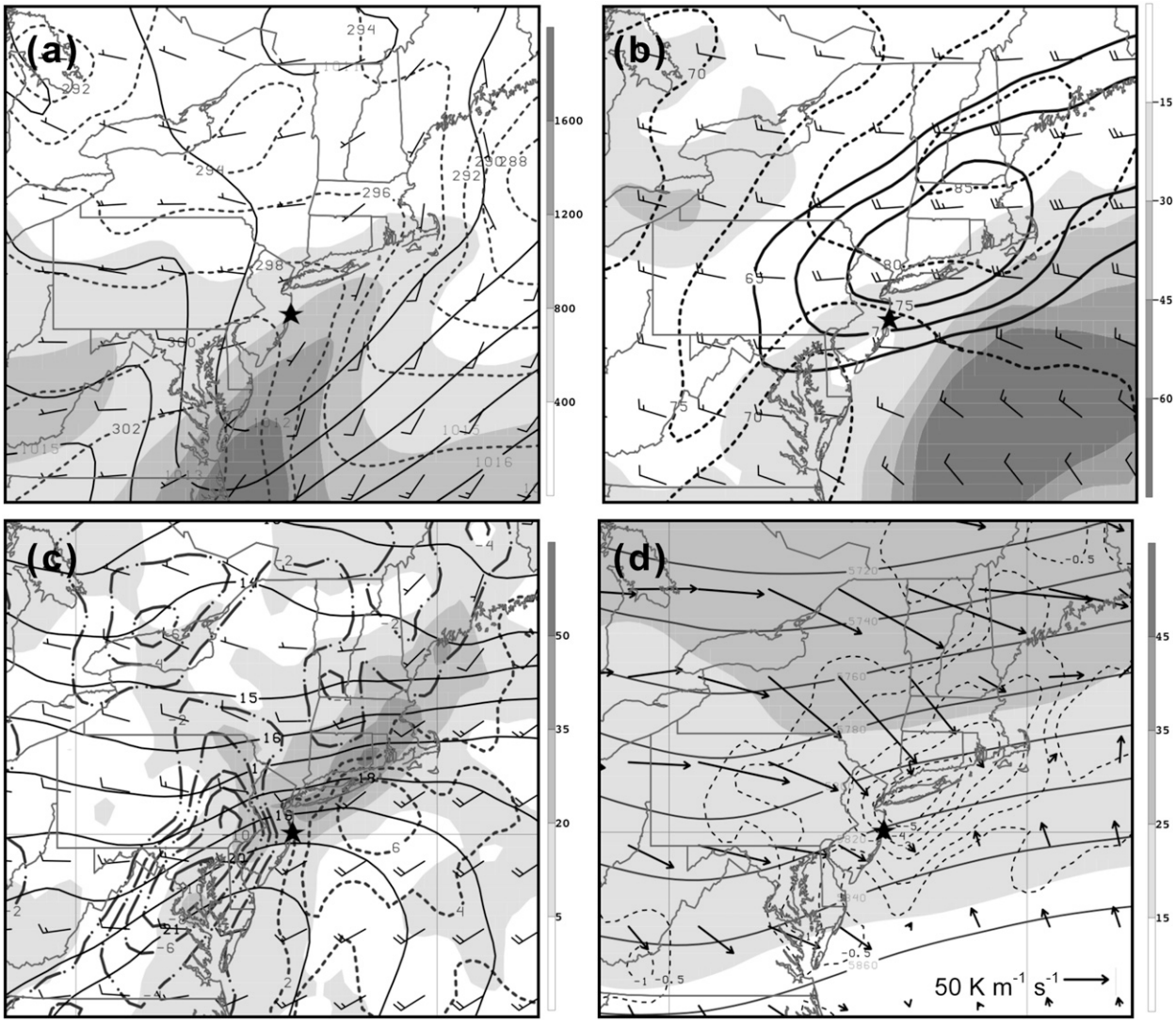

FIG. 7. As in Fig. 5, but for the slowly decaying events.

composite QLCS, with similar values of 1000–800-hPa RH (76%; Fig. 7b). As for the instability fields, SBCIN (not shown) is comparable to MUCIN. Given the less favorable horizontal distribution of instability, the longer maintenance of these systems is in part due to support from the robust dynamical surface boundaries spatially close to the QLCSs.

The mean 0–3-km vertical wind shear vector is 10.5 m s^{-1} from the west over and downstream of the QLCS (Fig. 7b), which is consistent with climatological values for quasi-linear events over the Northeast (Lombardo and Colle 2010), and significantly (95% level) greater than for decaying QLCSs. This larger low-level shear is consistent with the stronger low-level forcing associated with slowly decaying events than decaying events. The QLCS is on the warm side of a 900–800-hPa layer-average frontogenesis maximum (not shown), within the ascent region of the frontogenetical circulation, with a value of $\sim 26 \times 10^{-2}$ K (100 km)$^{-1}$ (3 h)$^{-1}$ over the compositing point (Fig. 7c). There is weak 900-hPa warm air advection (2×10^{-2}°C s^{-1}) over the QLCS, which is consistent with the location of the developing baroclinic zone in the layer (Fig. 7c). Compared to decaying QLCS events, the 1000-hPa frontogenesis maximum [$\sim 65 \times 10^{-2}$ K (100 km)$^{-1}$ (3 h)$^{-1}$; not shown] associated with a 1000-hPa baroclinic zone ~200 km east of the slow decay compositing point (Fig. 7a) is greater as well.

At midlevels, a 500-hPa relatively sharp trough axis is located ~400 km west of the QLCS, which is associated with more robust 500-hPa **Q**-vector convergence over and to the east of the convection (Fig. 7d) than for

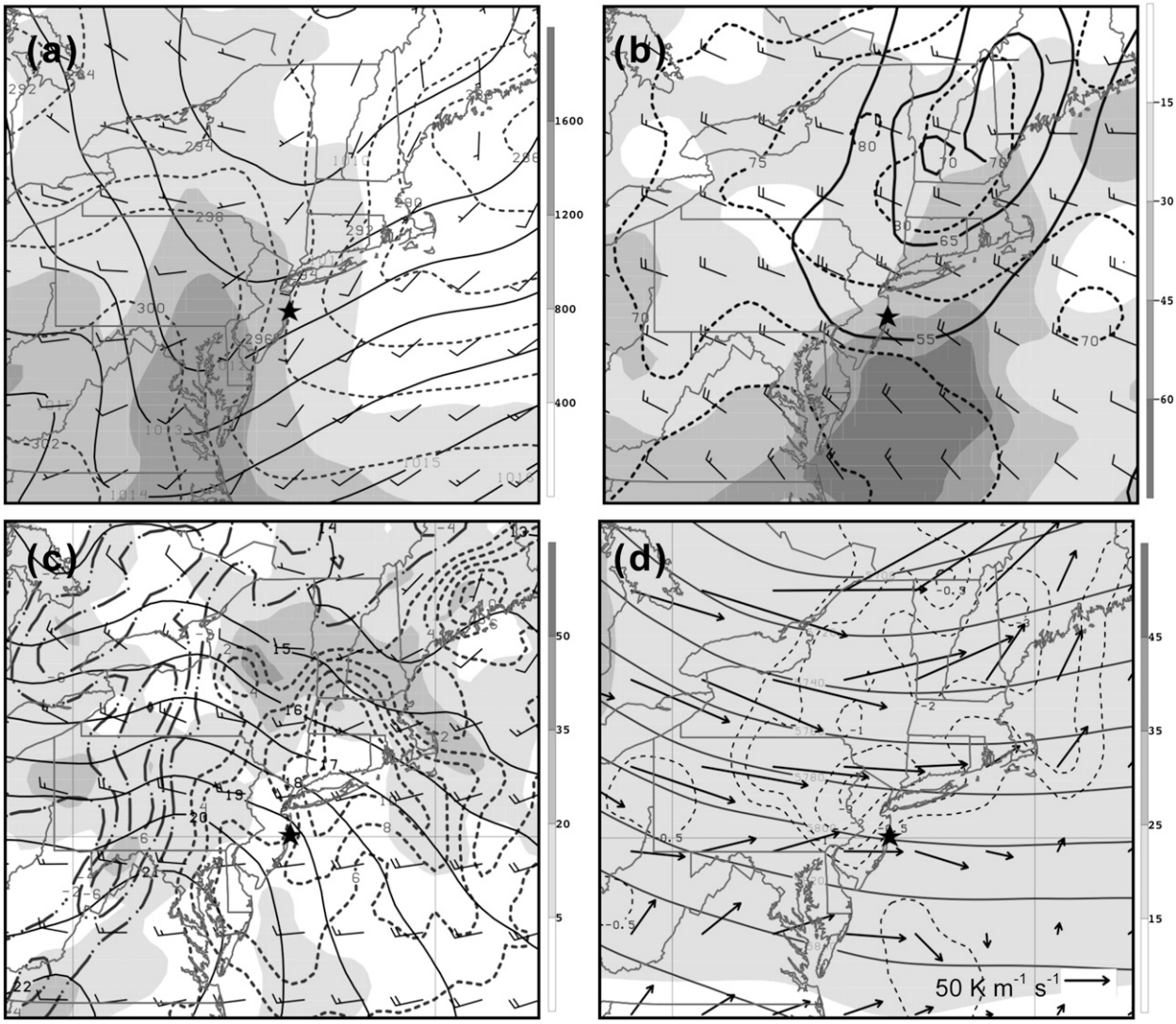

FIG. 8. As in Fig. 5, but for the sustaining events.

decaying events (Fig. 5d). The QLCS is only ~400 km southeast of the 25 m s^{-1} 300-hPa jet entrance region (Fig. 7d), a smaller distance than in the decaying QLCS composite, indicating that ageostrophic circulations may be more influential during slowly decaying events. There is more robust deep ascent (800–400 hPa; -5×10^{-3} Pa s^{-1}) over the slowly decaying QLCS than the decaying lines, which is collocated with the 900–800-hPa frontogenesis maximum (Fig. 7c), the converging 500-hPa **Q** vectors, and the 300-hPa jet entrance region (Fig. 7d). Associated with the stronger vertical motion, midlevels are significantly (95%) more saturated during slowly decaying events than decaying events, with 700–500-hPa relative humidity values of 64% (Fig. 7b). Compared to decaying QLCS events, slowly decaying systems have weaker instability, but develop with stronger dynamical forcing at the surface as well as through the troposphere.

c. *Sustaining events (nine events)*

The spatial composites of sustaining QLCS events highlight several synoptic-dynamic differences compared to the decaying and slowly decaying QLCSs. Sustaining convection is located along a surface baroclinic zone (Fig. 8a), which is associated with 45×10^{-2} K (100 km)$^{-1}$ (3 h)$^{-1}$ of 1000-hPa frontogenesis (not shown), while the surface pressure trough is 250–300 km west of the convection. Examination of the individual events reveals that no QLCSs included in this composite crossed the coast while collocated with a surface cold front. Four events showed a cold front 100–500 km (west, east, or north) from the convection.

In three events, the convective line moved perpendicular to and along a warm front and two events showed no synoptic boundaries in the eastern United States.

At $T - 0$ h, the mean MUCAPE (Fig. 8a) and MUCIN (Fig. 8b) over the convection are ~800 and ~−30 J kg^{-1}, respectively. Surface-based CAPE and surface-based CIN values as well as the spatial distributions (not shown) are comparable to MUCAPE and MUCIN, respectively. Although the thermodynamic indices are either equivalent or less favorable than for QLCSs that decay and slowly decay, the 0–3-km vertical wind shear vector over and downstream of the QLCS is 13–15.5 m s^{-1} from the northwest (Fig. 8b), ~5.5–8 m s^{-1} and ~2.5–5 m s^{-1} greater than the decaying and slowly decaying events, respectively. The 1000–800-hPa RH is 74% (Fig. 8b), a similar value seen during other evolutions (Figs. 5b and 7b), with 700–500-hPa RH values of 57% (Fig. 8c). While the mean midlevel relative humidity value is somewhat comparable to decaying events (Fig. 5b), it is not significantly (95% level) smaller than for slowly decaying events.

While there is little 900–800-hPa frontogenetical forcing over the convection, the QLCS is collocated with a maximum in warm air advection (7×10^{-2}°C s^{-1}; Fig. 8c). A manual inspection of the cases revealed that all nine maintaining events occurred with 900–800-hPa warm air advection (not shown). The spatially localized region of low-level warm advection is a forcing term for ascent following the QG omega equation. With little temperature advection above the 900–800-hPa layer (not shown), the vertically localized low-level warm air advection maximum may also be important in destabilizing the atmosphere just above the shallow marine layer.

The main region of deep midlevel synoptic-scale ascent, supported by 500-hPa converging **Q** vectors, is ~100 km west as well as 400–500 km north of the convection, with -0.5×10^{-3} Pa s^{-1} over the sustaining convective line. The localized warm air advection at this level may contribute to this weak region of ascent, with the Laplacian of thermal advection implying ascent following the QG omega equation. The composite 300-hPa jet is highly uniform over the entire domain (Fig. 8d), indicating little contribution to synoptic ascent. Overall, there is less frontogenetical support and weaker mid- to upper-level forcing for ascent for sustaining events than QLCSs that decay and slowly decay. Thus, it is likely that sustaining events are more reliant on cold-pool dynamics to propagate the QLCS over the marine waters, which will be addressed in section 4.

4. Discussion

It is important to highlight the ambient conditions that contribute to the longevity of convective lines over the northeastern United States, to allow for comparison to storms over the central United States. For example, both deep vertical wind shear and low-level frontogenesis have been shown to contribute to the longevity and maintenance of MCSs over the central United States (Coniglio et al. 2007, 2010). To underscore the distinguishing environmental variables governing the evolution of northeastern U.S. quasi-linear systems identified in this study, box-and-whisker plots are used to highlight ambient quantities averaged over a 1.5° latitude–longitude box immediately ahead of the QLCS for each individual event (Figs. 9–11). Area-average values were calculated at both $T - 3$ h and $T - 0$ h to capture the evolving environmental conditions influencing the quasi-linear convection as it moved toward the coastline. The $T - 0$ h time represents the same time used for the feature-based composites (Figs. 5, 7, and 8), while $T - 3$ h represents 3 h prior. For each event the 0.75° latitude–longitude box was moved between $T - 3$ h and $T - 0$ h to a location immediately downstream of the convective storms. In the event that no convection was present at $T - 3$ h, the center of the latitude–longitude box was placed at the location of convective initiation.

Figure 9 highlights the evolution of the ambient instability/stability associated with the linear systems as they move toward the Atlantic coastline. The mean MUCAPE during decaying QLCS events decreases from 907 ($T - 3$ h) to 831 J kg^{-1} ($T - 0$ h), though the instability increases slightly from 681 ($T - 3$ h) to 709 J kg^{-1} ($T - 0$ h) during slowly decaying events (Figs. 9a,b). For linear systems that successfully move over the water, the mean instability declines from 946 to 803 J kg^{-1} (Figs. 9a,b). However, at both $T - 3$ h and $T - 0$ h, the mean MUCAPE differences between the three storm evolutions are not statistically significant, indicating that instability is not a strong delimiter when evaluating coastal QLCS longevity. This is consistent with Coniglio et al. (2010), who found little difference in CAPE values immediately ahead of long-lived and short-lived MCSs over the central United States.

For mean MUCIN values, the stability increases between $T - 3$ h and $T - 0$ h (Figs. 9c,d) for all three convective evolutions, which is consistent with movement of the convection toward the Atlantic Ocean and the increased influence of the marine layer on the convective inhibition. Mean ambient MUCIN strengthens ~16 J kg^{-1} for decaying and slowly decaying events, yet only ~11 J kg^{-1} in sustaining cases (Figs. 9c,d), which implies that weaker MUCIN may be important to the survival of coastal QLCSs. However, the differences between the mean MUCIN values are not significant, suggesting that further investigation is needed to support this hypothesis.

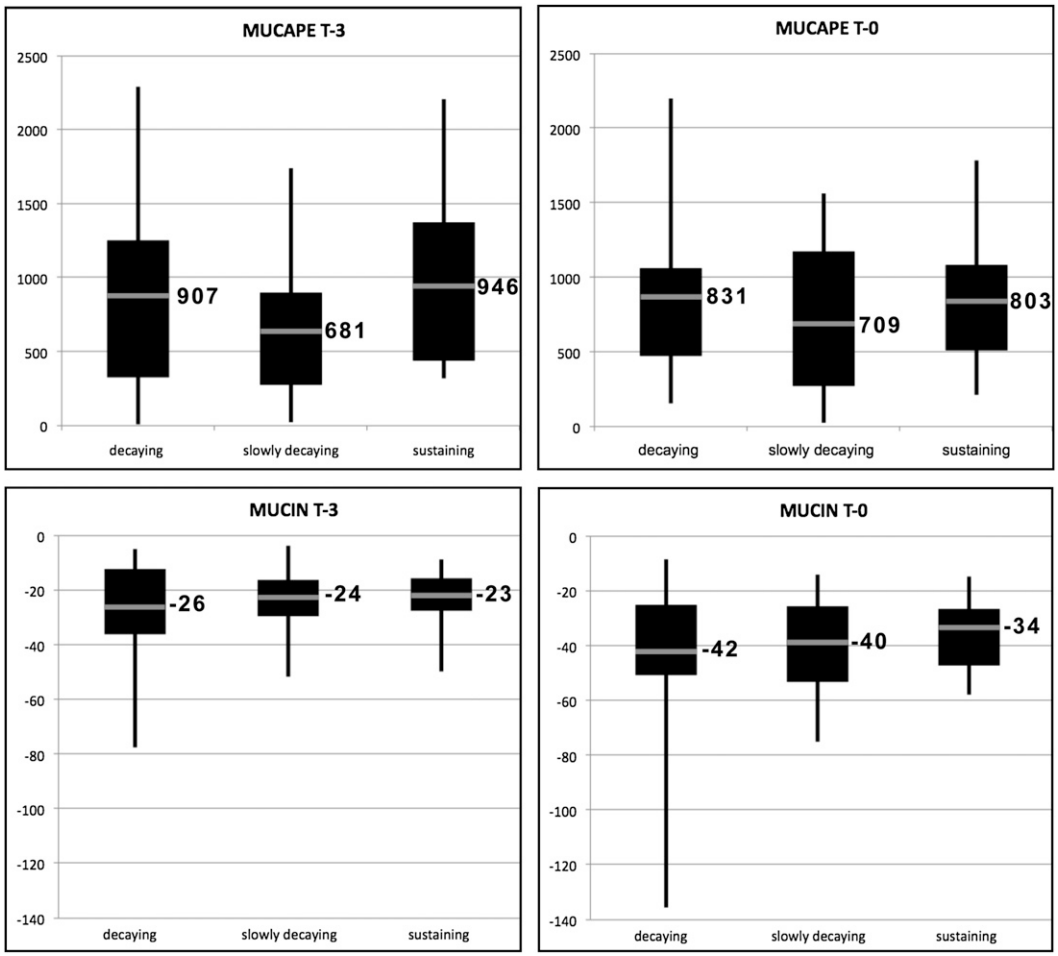

FIG. 9. Box-and-whisker plots showing 1.5° latitude–longitude area-averaged values of MUCAPE (J kg^{-1}) at (a) $T-3$ h and (b) $T-0$ h and MUCIN (J kg^{-1}) at (c) $T-3$ h and (d) $T-0$ h for the 32 decaying events, 18 slowly decaying events, and 9 sustaining events. The bottom and top of the solid black box are the 25th and 75th quartile, respectively. The mean is denoted by a gray bar with its value also noted. The maximum and minimum outliers are denoted by the vertical solid lines.

While the importance of vertical wind shear in organizing quasi-linear convective systems has been well documented in the literature (e.g., Thorpe et al. 1982; Rotunno et al. 1988; Parker and Johnson 2000; Weisman and Rotunno 2004; Cohen et al. 2007), the role of low-level shear verses deep-layer shear, as well as the physical interpretation of their significance is still being investigated (e.g., Houston and Wilhelmson 2011; Coniglio et al. 2012). For the decaying and slowly decaying events explored in this study, the mean low-level shear (0–3 km) declines ~1.4 and ~0.7 m s^{-1}, respectively, between $T-3$ h and $T-0$ h, though it increases by 4.7 m s^{-1} for sustaining coastal linear convection (Figs. 10a,b). This suggests that for sustaining events, the ambient low-level shear is becoming more favorable for linear convection, while it becomes less favorable for events that decay and slowly decay over the waters. At $T-0$ h, the mean ambient shear values for sustaining convection (13.5 m s^{-1}) is significantly (95% level) greater than for decaying lines (7.4 m s^{-1}), providing confidence to this claim, though the difference between slowly decaying (10.7 m s^{-1}) and sustaining lines is not significant. This may be in part due to the small sample size of sustaining events. The difference in low-level shear between slowly decaying (11.4 and 10.7 m s^{-1}) and decaying (8.8 and 7.4 m s^{-1}) is statistically (95%) significant. In contrast, Coniglio et al. (2010) found no significant difference in the low-level (0.5–2 km, 0.5–3 km, 1–3 km, etc.) vertical wind shear values between short-lived (≤5 h) and long-lived (≥8 h) MCSs over the central United States, within 200 km downstream of the composite system location.

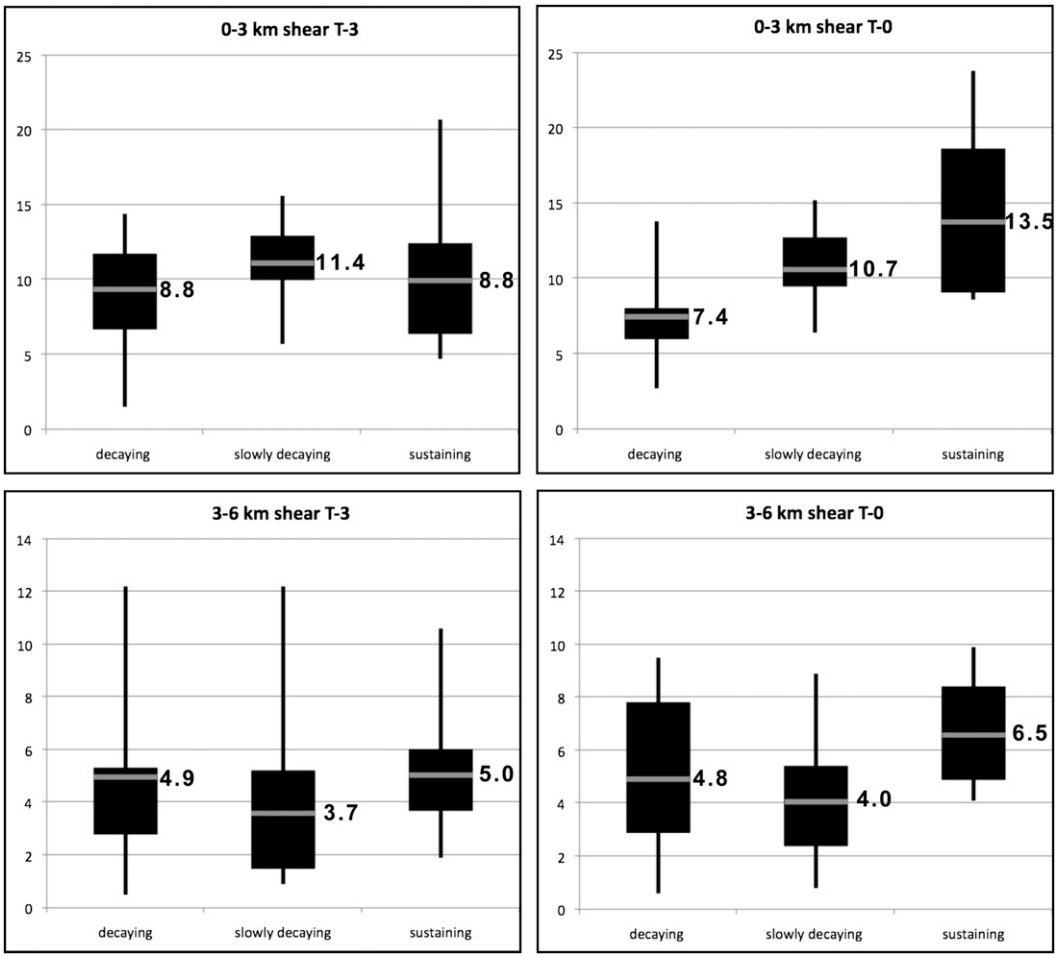

FIG. 10. As in Fig. 9, but for 0–3-km vertical wind shear (m s^{-1}) at (a) $T - 3$ h and (b) $T - 0$ h, and 3–6-km vertical wind shear (m s^{-1}) at (c) $T - 3$ h and (d) $T - 0$ h.

The vertical wind shear between 3 and 6 km has been shown to be a good discriminator between long- and short-lived MCSs over the central United States, with greater mid- and upper-level shear found for longer-lived storms (Coniglio et al. 2010). For northeastern U.S. decaying and slowly decaying quasi-linear systems examined in this study, the mean 3–6-km vertical wind shear changes little from $T - 3$ h (Fig. 10c) to $T - 0$ h (Fig. 10d), decreasing from 4.9 to 4.8 m s^{-1} for decaying events, and increasing from 3.7 to 4.0 m s^{-1} for slowly decaying events. There is a more noticeable increase in midlevel shear for sustaining events (5 to 6.5 m s^{-1}), though there is no significant difference between 3–6-km shear values at either $T - 3$ h or $T - 0$ h between the three linear evolutions.

Explicating the importance of vertical wind shear to squall-line dynamics, Rotunno et al. (1988) proposed that in the absence of synoptic forcing, optimal gust front lifting at the leading edge of surface-based squall lines occurs if the vorticity generated by the low-level (0–2.5 km) vertical wind shear is balanced by the vorticity generated by the diabatically generated cold pool [Rotunno–Klemp–Weisman (RKW) theory]. Considering the mean 0–3-km vertical wind shear for the quasi-linear systems examined in this study, the greater low-level shear during sustaining (13.8 m s^{-1}) and slowly decaying events (10.7 m s^{-1}) may have helped balance the cold pool leading to longer-lived systems. However, given that decaying and slowly decaying systems are more strongly synoptically forced than sustaining events, the application of RKW theory may be less straightforward in these types of events. To explore this hypothesis, information regarding the cold pool temperature perturbations for each quasi-linear system crossing the coast was estimated using hourly METAR surface observations. The NARR data were not used given their coarse spatial and temporal resolution, as well as the fact the NARR does not assimilate convective storm information

(radar data). Surface temperature values for 2–5 Automated Surface Observing Stations (ASOS) impacted by each quasi-linear system were averaged 1 h prior to ($T - 1$ h) and 1 h after ($T + 1$ h) the convection passed over the stations. The number and location of surface stations varied given the QLCS coastal crossing location and the density of surface observations at that location. To calculate the surface temperature perturbations, differences were taken between the $T - 1$ h and $T + 1$ h averaged surface temperature values for each event, which provided an estimate of the cold pool strength. The authors acknowledge the limitations and the uncertainties introduced through this methodology, including a lack of knowledge about the depth and vertical structure of cold pool.

The mean cold pool temperature perturbation for both decaying and slowly decaying QLCSs is 5.2 K, with a range of 0–10.4 and 0.3–10.6 K, respectively (not shown). These perturbations are within the neighborhood of the mean magnitude of cold pools associated with MCSs over Oklahoma, which range from 8.5 K during MCS initiation to 5.4 K during dissipation on average (Engerer et al. 2008). With similar estimated cold pool temperature perturbation values associated with decaying and slowly decaying events, one could hypothesize that the larger 0–3-km shear during slowly decaying events (Fig. 10b) may have contributed to the slower dissipation of the storms, following RKW theory. However, there was no correlation between the strength of the 0–3-km shear vector magnitude and the strength of the cold pool temperature perturbation when the individual decaying and slowly decaying events were examined (not shown). This may be in part due to the limitations in estimating the cold pool strength, though it may also simply highlight the importance of the stronger surface and low-level dynamical forcing during slowly decaying events.

Interestingly, the mean cold pool temperature perturbation during sustaining events is only 2.8 K with a range from 0–4.5 K (not shown), which is significantly (95% level) smaller than for decaying QLCSs. Recall that the 0–3-km wind shear for sustaining events is significantly (95%) greater than decaying events. This suggests that the horizontal vorticity associated with the weaker cold pool may be more easily balanced by the relatively strong low-level shear, thus allowing the system to maintain its intensity over the ocean. However, high-resolution analyses of individual case studies are necessary to support this assessment.

It is difficult to confidently conclude the role of the midlevel dry air in the development of cold pools during the QLCS events examined in this study. For the sustaining events, the midlevel RH was not significantly different from midlevel RH values seen during decaying and slowly decaying events; however, the magnitude of the mean cold pool for sustaining events was significantly smaller than the decaying events. Furthermore, the midlevel RH for decaying events was significantly less than for slowly decaying events, yet the magnitudes of the mean cold pools as well as the range in values were equivalent. This suggests that the midlevel dry air had little influence on the magnitude of the cold pool. These findings contradict previous studies illustrating that midlevel dry air promotes cold pool development through enhanced evaporative cooling (Hookings 1965; Johns and Doswell 1992; Gilmore and Wicker 1998), as well as studies that have shown that dry air reduces total condensate mass, hindering cold pool development in the leading portion of a QLCS (James and Markowski 2010). However, the conclusions from this current study are based on a mean analysis with a limited number of observations. It is likely that variations exist between cases, emphasizing the need for a detailed modeling study examining the diabatic processes during northeastern U.S. QLCS events.

The spatial composites illustrate noticeable differences in 900–800-hPa frontogenesis as well as 900-hPa temperature advection between sustaining convective lines (Fig. 8c) and the two modes of decaying QLCSs (Figs. 5c and 7c). At both $T - 3$ h and $T - 0$ h, the majority of slowly decaying lines is associated with larger 900–800-hPa frontogenesis values than sustaining QLCSs (Figs. 11a,b), with a significantly (95% level) greater mean value for slowly decaying lines than sustaining lines. The mean low-level frontogenesis value for slowly decaying lines is also greater than during decaying events, though the differences are not significant. Moreover, at $T - 0$ h, ~33% of sustaining events occur with 900–800-hPa frontolysis, compared to 12% of slowly decaying and 17% of decaying events (Fig. 6a). Thus, frontogenetical forcing is not required for a system to maintain its intensity over the ocean.

Low-level warm air advection is important for a QLCS to maintain its intensity over the Atlantic waters. Three-quarters of sustaining events experience warm advection larger than $\sim 2.5 \times 10^{-5}$°C s^{-1} over the QLCS at both $T - 3$ h and $T - 0$ h, while three-quarters of decaying and slowly decaying events have values less than $\sim 2.5 \times 10^{-5}$°C s^{-1} (Figs. 11c,d). Furthermore, the mean temperature advection (3.7×10^{-5}°C s^{-1} at $T - 3$ h; 7.6×10^{-5}°C s^{-1} at $T - 0$ h) for sustaining events is significantly (95% level) greater than for decaying events. In addition to its contribution to QG ascent, which is favorable for convective development (e.g., Maddox 1983), relatively strong thermal advection also implies the presence of geostrophic wind shear,

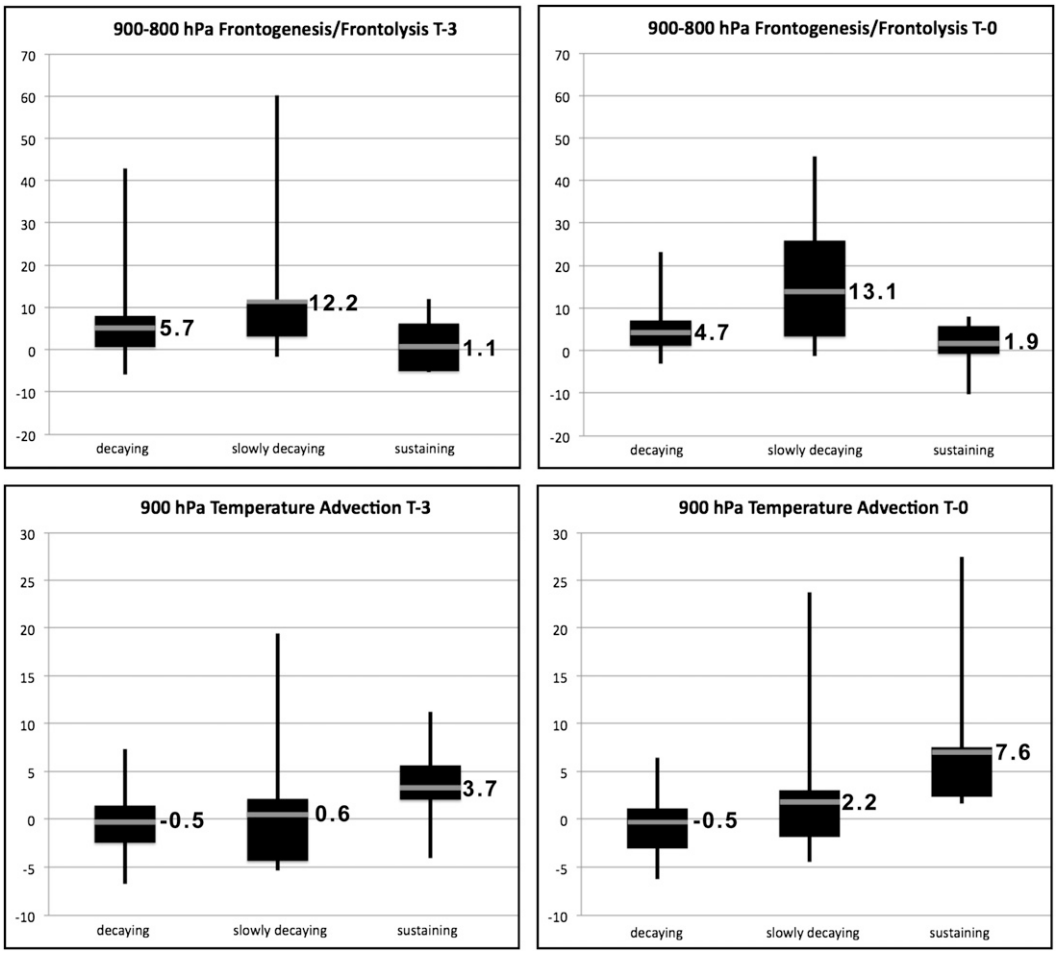

FIG. 11. As in Fig. 9, but for 900–800-hPa layer-averaged frontogenesis/frontolysis [10^{-2} K (100 km)$^{-1}$ (3 h)$^{-1}$] at (a) $T - 3$ h and (b) $T - 0$ h, and 900-hPa temperature advection (10^{-5}°C s^{-1}) at (c) $T - 3$ h and (d) $T - 0$ h.

as illustrated by the relatively strong low-level (0–3 km) wind shear.

Considering both the spatial composites as well as the statistical analyses, low- and midlevel dynamics appear to have more control on the evolution of decaying and slowly decaying quasi-linear convection than sustaining convective lines. For slowly decaying events, the stronger 900-hPa frontogenesis (Figs. 7c and 11a,b) and more robust surface boundaries help to sustain the QLCS slightly offshore (within 100 km) as it encounters the gradient of increasing convective inhibition associated with the marine layer (Fig. 7b). For the decaying events, the spatial distribution of MUCIN (Fig. 5b) is comparable to slowly decaying systems (Fig. 7b), yet there is weaker frontogenetical support (Fig. 5c) and fewer surface boundaries within close proximity. The weaker forcing may not be enough to overcome the stable marine boundary layer and support the convection over the waters. Furthermore, the larger 0–3-km vertical wind shear during slowly decaying events (Figs. 10a,b), which is a reflection of the stronger low-level dynamics, helps to balance the cold pool briefly as the QLCS moves offshore (Rotunno et al. 1988). Low-level wind shear may be most important during sustaining events, (Figs. 8c and 11c,d), with three of the sustaining QLCS events moving along a surface warm front, which provides area of even more enhanced directional shear.

5. Conclusions

Quasi-linear convective systems (QLCSs) encountering the Atlantic coastline are examined for three different evolutions: decaying near the coastline, slowly decaying after traversing the coastline, and sustaining intensity over the Atlantic waters more than 100 km from the coast.

Feature-based composites indicate that decaying and slowly decaying QLCSs are collocated with a surface

pressure trough and a 900–800-hPa frontogenesis maximum, indicating that both evolutions are associated with a developing low-level baroclinic zone. There is stronger frontogenetical forcing over slowly decaying QLCSs, and inspection of the individual cases highlights the presence of more robust surface features (e.g., mesolows) than during decaying convective events. Though there is greater instability during decaying events (MUCAPE $\sim$ 1200 J kg^{-1}) than slowly decaying events (MUCAPE $\sim$ 800 J kg^{-1}) and similar distributions of MUCIN increasing offshore, the stronger low-level forcing helps support slowly decaying convective lines for a finite period of time upon encountering the coastal waters. Furthermore, QLCS events that slowly decay offshore are associated with stronger mid- and upper-level QG forcing for ascent than decaying events, indicating that these events are associated with more robust dynamical forcing and are less dependent on instability.

The dynamic and thermodynamic environment associated with the maintenance of a QLCS over the Atlantic waters is different than for the two decaying composites. Sustaining convection is $\sim$250–300 km downstream (east) of a surface pressure trough, within a localized maximum of 900-hPa warm air advection, helping to destabilize the atmosphere as well as promote ascent. Overall there is less dynamical forcing than decaying and slowly decaying lines, with little 900–800-hPa frontogenesis, midlevel QG forcing for ascent, and ascent associated upper-level jet ageostrophic circulations. The thermodynamics associated with sustaining convective lines is equivalent to or less favorable than systems that decay and slowly decay offshore, with a mean instability similar to that over slowly decaying systems, and a larger average MUCIN of 30–35 J kg^{-1}. One important difference is that the average 0–3-km shear (13.3 m s^{-1}) is largest for QLCS events that are maintained over the Atlantic waters. There was also a weaker surface cold pool for the sustaining systems than the decaying events, which may favor more of a balance of horizontal vorticity from the cold pool and the vertical shear in the sustaining events.

Future exploration into individual case studies, using a combination of observational analyses, numerical simulations of the individual events, as well as sensitivity tests of the case-study simulations, will quantify the development and structure of the cold pool associated with individual quasi-linear convective cases, the role of vertical wind shear, and the role of the offshore stability on the QLCS evolution near the coast. Trajectories will be used to highlight the source region of air (surface or elevated) ingested into the storms as they traverse the Atlantic waters as well, to provide further insight into the processes contributing to their maintenance or decay.

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