Persistence and dissipation of Lake Michigan-crossing mesoscale convective systems

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PERSISTENCE AND DISSIPATION OF LAKE MICHIGAN-CROSSING
MESOSCALE CONVECTIVE SYSTEMS

by

Nicholas D. Metz

A Dissertation
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ABSTRACT

This thesis investigates mature mesoscale convective systems (MCSs) that traversed Lake Michigan to elucidate synoptic-scale and Lake Michigan-related features that discriminate between persistence and dissipation. Of the 110 coherent MCSs that crossed Lake Michigan during the warm seasons (April–September) of 2002–2007, 47 (43%) persisted, while 63 (57%) dissipated. Persistence was favored during July and August, when Lake Michigan was warmer and during the evening and overnight, when the low-level jet (LLJ) was most intense. However, a number of MCSs also persisted during the early warm season when the Lake Michigan water temperature was cooler than the surrounding land.

Climatological, compositing, and case study results identified common environmental ingredients for MCS persistence. On average, persisting MCSs were associated with a 5 m s\(^{-1}\) stronger 850-hPa LLJ, 500–750 J kg\(^{-1}\) of increased CAPE downstream of Lake Michigan, a 10 m s\(^{-1}\) stronger 200-hPa jet stream, 100–200 J kg\(^{-1}\) of increased downdraft CAPE, and \(1.5–2.0\times10^{-7}\) s\(^{-1}\) more moisture convergence than dissipating MCSs. Early-season MCSs persisted upon crossing the relatively cold waters of Lake Michigan because a strong, shallow near-surface-based inversion insulated the cold water from the flow above, allowing MCSs to traverse the inversion much like a “speed bump.” Climatologically, the shallow near-surface inversion strength was about 1°C stronger for persisting MCSs. Additionally, persisting MCSs typically had strong, deep convective cold pools, and associated vigorous convective ascent, which typically overcame any effects of the shallow vertical circulation generated by the temperature gradient at the Lake Michigan shore. Late warm-season MCSs generally persisted
through normal convective processes, as Lake Michigan was similar in temperature to the surrounding land environment.

Numerical simulations utilizing the WRF model were conducted for representative case studies with Lake Michigan included (control) and removed (noLM). These studies showed that the synoptic-scale environment was the main control on MCS persistence and dissipation. Within a favorable synoptic-scale environment, both persisting and dissipating noLM MCS simulations produced up to 50 mm of additional precipitation in conjunction with increased convection and instability. However, once the synoptic-scale environment became unfavorable, both control and noLM simulated MCSs dissipated simultaneously.
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degree would not have happened without her, and I look forward to the exciting future endeavors that we will share together! This thesis is dedicated to her!
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FIG. 1.1: (a) Global distributions of MCCs (black dots) and regions of widespread deep atmospheric convection as inferred by outgoing longwave radiation (shading indicates outgoing longwave radiation minima) [from Laing and Fritsch (1997)]. (b) Isolines of the average number of derechos (contoured every 3 beginning at 3) located within 2° × 2° boxes for 1980–1983 [from Johns and Hirt (1987), modified by Coniglio and Stensrud (2004)].

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FIG. 2.5: Hourly distributions of MCSs that persisted (green) and dissipated (red) upon crossing Lake Michigan during the warm seasons of 2002–2007. The purple numbers show the total number of storms in the climatology that occurred during each hour range. Source: NOWrad composite data.

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FIG. 2.14: Organizational structure of MCSs that persisted (green) and dissipated (red) upon crossing Lake Michigan during the warm seasons of 2002–2007. The black numbers show the total number of MCSs in the climatology that either persisted or dissipated within each category. The purple number shows the persistence percentage within each category. Source: NOWrad composite data.

FIG. 3.1: Idealized schematics illustrating favorable situations for the development of (a) dynamic bow echo MCSs and (b) progressive bow echo MCSs. The PJ, LJ, and SJ refer to the polar, low-level, and subtropical jets, respectively. The shading in (a) represents the most favorable area for MCS development, while the line B–M–E in (b) refers to the typical track of a MCS [from Johns (1993)].

FIG. 3.2: Synoptic-scale composites of SLP (black lines every 1 hPa) and 1000–500-hPa thickness (red dashed lines every 3 dam) for dynamic MCSs that (a) persisted and (b) dissipated upon crossing Lake Michigan. The green star in each panel reveals the location of the MCS-relative compositing point. Source: 1.0° GFS analyses.

FIG. 3.3: Synoptic-scale composites of 200-hPa wind speed (warm colors shaded according to color bar in m s$^{-1}$), 200-hPa height (black lines every 12 dam), 850-hPa wind speed (cool colors shaded according to color bar in m s$^{-1}$), and 850-hPa winds (barbs greater than 10 m s$^{-1}$, where each long barb equals 5 m s$^{-1}$) for dynamic MCSs that (a) persisted and (b) dissipated upon crossing Lake Michigan. The green star in each panel reveals the location of the MCS-relative compositing point. Source: 1.0° GFS analyses.
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FIG. 3.5: Synoptic-scale composites of 925-hPa moisture convergence (shaded according to color bar in $\times 10^{-7}$ s$^{-1}$), 925-hPa mixing ratio (red lines every 1 g kg$^{-1}$), and 925-hPa winds (barbs in m s$^{-1}$, where each long bar equals 5 m s$^{-1}$) for dynamic MCSs that (a) persisted and (b) dissipated upon crossing Lake Michigan. The black star in each panel reveals the location of the MCS-relative compositing point. Source: 1.0° GFS analyses.

FIG. 3.6: Synoptic-scale composites of MUCAPE (shaded according to color bar in J kg$^{-1}$) and 0–6-km shear (barbs in m s$^{-1}$, where one long bar equals 5 m s$^{-1}$) for dynamic MCSs that (a) persisted and (b) dissipated upon crossing Lake Michigan. The green star in each panel reveals the location of the MCS-relative compositing point. Source: 1.0° GFS analyses.

FIG. 3.7: As in Fig. 3.2 except for progressive MCSs. Source 1.0° GFS analysis.

FIG. 3.8: As in Fig. 3.3 except for progressive MCSs. Source 1.0° GFS analysis.

FIG. 3.9: As in Fig. 3.4 except for progressive MCSs. Source 1.0° GFS analysis.

FIG. 3.10: As in Fig. 3.5 except for progressive MCSs. Source 1.0° GFS analysis.

FIG. 3.11: As in Fig. 3.6 except for progressive MCSs. Source 1.0° GFS analysis.

FIG. 4.1: Isochrones every 3 h of the leading 45-dBZ line of MCS-P1 (purple) overlaid upon severe storm reports of wind (blue), hail (green), and tornadoes (red) between 12Z/07 and 06Z/08 June 2008. Source: Severe Plot v.3.0 and UAlbany radar archive.

FIG. 4.2: NEXRAD level-III composite reflectivity data (in dBZ) at (a) 1503Z/07, (b) 1701Z/07, (c) 1900Z/07, (d) 2105Z/07, (e) 2304Z/07, (f) 0104Z/08, (g) 0302Z/08, and (h) 0503Z/08 June 2008. The arrows in panels (a), (d), and (g) illustrate the location of MCS-P1. Green and orange shading begin at 20 and 45dBZ, respectfully. Source: UAlbany radar archive.

FIG. 4.3: Water-vapor imagery from (a) 1515Z/07, (b) 1815Z/07, and (c) 2115Z/07 June 2008. The green arrows in panels (a) and (b) indicate the location of a first upper-level short-wave trough, while the white arrows in panels (b) and (c) indicate the location of a second short-wave trough. Source: RAP weather.
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FIG. 4.5: Atmospheric sounding of temperature (red line in °C), dewpoint (blue line in °C), and wind (black barbs in m s$^{-1}$, where one long barb equals 5 m s$^{-1}$) for 18Z/07 June 2008 from DVN. The inset in the upper-right corner shows the location of DVN. Source: UAlbany sounding archive.

FIG. 4.6: Manual surface analysis of SLP (black lines every 2 hPa), temperature (red dashed lines every 3°C), and mixing ratio equal to 18 g kg$^{-1}$ (green line) for 20Z/07 June 2008. The “P1” illustrates the location of MCS-P1. IR satellite imagery from 2010Z/07 June 2008 is inset into the lower-left corner of the image. Source: UAlbany surface archive and UCAR MMM image archive.

FIG. 4.7: As in Fig. 4.4 except for 00Z/08 June 2008. The “P1” in each panel illustrates the location of MCS-P1. IR satellite imagery from 0010Z/08 June 2008 is inset into the lower-right corner of panel (a). Source: 20-km RUC and UCAR MMM image archive.

FIG. 4.8: 850-hPa Q-vectors (arrows greater than 5×10$^{-7}$ Pa m$^{-1}$ s$^{-1}$), Q-vector convergence (shaded according to color bar in ×10$^{-12}$ Pa m$^{-2}$ s$^{-1}$), temperature (green dashed lines every 3°C), and height (black lines every 3 dam) for 00Z/08 June 2008. A reference vector is located in the lower-right corner of the image. IR satellite imagery from 0010Z/08 June 2008 is inset into the upper-left corner of the image. Source: 1.0° GFS analysis and UCAR MMM image archive.

FIG. 4.9: (a) The 975-hPa previous 4-h potential temperature decrease (shaded according to color bar in K) and 0–3-km shear (barbs in m s$^{-1}$, where one long barb equals 5 m s$^{-1}$) for 23Z/07 June 2008, (b) cross section A–A’ of the previous 4-h potential temperature decrease (shaded according to color bar in K), potential temperature (black lines every 1 K), and wind (barbs in m s$^{-1}$, where one long barb equals 5 m s$^{-1}$) for 23Z/07 June 2008, and (c) AMDAR sounding of temperature (red line in °C), dewpoint (blue line in °C), and wind (red barbs in m s$^{-1}$, where one long barb equals 5 m s$^{-1}$) for 2317Z/07 June 2008. NEXRAD level-III composite radar imagery for 19Z/07 and 23Z/07 June 2008 are inset into the lower-left and right corners of (a), respectively, with arrows indicating MCS-P1 location. The red hexagon in (a) indicates the AMDAR sounding location. Source: 20-km RUC analyses, UAlbany radar archive, and AMDAR.
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FIG. 4.11: As in Fig. 4.1 except for between 06Z/04 and 21Z/04 June 2005 with MCS-D1. Source: Severe Plot v.3.0 and UAlbany radar archive.

FIG 4.12: NOWrad composite reflectivity data (in dBZ) at (a) 14Z/04, (b) 16Z/04, (c) 18Z/04, (d) 19Z/04, (e) 20Z/04, (f) 21Z/04, and 22Z/04 June 2005. The arrows in panels (a), (c), and (e) illustrate the location of MCS-D1. Green and orange shading begin at 20 and 45 dBZ, respectively. Source: UCAR MMM image archive.

FIG. 4.13: (a) SLP (black lines every 4 hPa), 1000–500-hPa thickness (red dashed lines every 6 dam), and 925-hPa mixing ratio (shaded according to color bar in g kg⁻¹), (b) 200-hPa wind (shaded according to color bar in m s⁻¹), 200-hPa height (black lines every 12 dam), and 850-hPa wind (barbs greater than 12.5 m s⁻¹, where one long barb equals 5 m s⁻¹), (c) 400-hPa absolute vorticity (shaded according to color bar in ×10⁻⁵ s⁻¹), 400-hPa height (black lines every 6 dam), and 400-hPa wind (barbs in m s⁻¹, where one long barb equals 5 m s⁻¹), and (d) CAPE (shaded according to color bar in J kg⁻¹) and 0–6-km shear (barbs in m s⁻¹, where one long barb equals 5 m s⁻¹) for 15Z/04 June 2005. The “D1” in each panel illustrates the location of MCS-D1. IR satellite imagery from 1510Z/07 June 2005 is inset into the lower-right corner of panel (a). Source: 20-km RUC analyses and UCAR MMM image archive.

FIG. 4.14: As in Fig. 4.5 except for a RUC sounding at 40.96°N latitude and 87.62°W longitude for 18Z/04 June 2005. The inset in the upper-right corner shows the location of the sounding. Source: 20-km RUC analysis.

FIG. 4.15: Manual surface analysis of SLP (black lines every 2 hPa), temperature (red dashed lines every 3°C), and mixing ratio equal to 16 g kg⁻¹ (green line) for 18Z/07 June 2005. The “D1” illustrates the location of MCS-D1. IR satellite imagery from 1810Z/07 June 2005 is inset into the upper-left corner of the image. Source: UAlbany surface archive and UCAR MMM image archive.

FIG. 4.16: As in Fig. 4.8 except for 18Z/04 June 2005. The “D1” illustrates the location of MCS-D1. IR satellite imagery from 1810Z/04 June 2005 is inset into the upper-left corner of the image. Source: 1.0° GFS analysis and UCAR MMM image archive.

FIG. 4.17: As in Fig. 4.13 except for 21Z/04 June 2005. The “D1” in each panel illustrates the location of MCS-D1. IR satellite imagery from 2110Z/04 June 2005 is inset in the lower-right corner of panel (a). Source: 20-km RUC and UCAR MMM image archive.
**Fig. 4.18:** (a) The 975-hPa previous 2-h potential temperature decrease (shaded according to color bar in K) and 0–3-km shear (barbs in m s$^{-1}$, where one long barb equals 5 m s$^{-1}$) for 19Z/04 June 2005, (b) cross section B–B’ of the previous 2-h potential temperature decrease (shaded according to color bar in K), potential temperature (black lines every 1 K), and wind (barbs in m s$^{-1}$, where one long barb equals 5 m s$^{-1}$) for 19Z/04 June 2005, and (c) AMDAR sounding of temperature (red line in °C), dewpoint (blue line in °C), and wind (red barbs in m s$^{-1}$, where one long barb equals 5 m s$^{-1}$) for 1955Z/04 June 2005. NOWrad composite radar imagery for 17Z/04 and 19Z/04 June 2005 are inset into the lower-left and right corners of (a), respectively, with arrows indicating MCS-D1 location. The red hexagon in (a) indicates the AMDAR sounding location. Source: 20-km RUC analyses, UAlbany radar archive, and AMDAR.

**FIG. 4.19:** (a) As in Fig. 4.10a except at ARR from 12Z/04 to 23Z/04 June 2005, (b) as in Fig. 4.10b except for from 12Z/04 to 23Z/04 June 2005, and (c) the locations of the meteograms in (a) and (b). Source: UAlbany surface archive and NDBC.

**FIG. 4.20:** As in Fig. 4.1 except for between 10Z/18 and 03Z/19 June 2010 with MCS-P2. Source: Severe Plot v.3.0 and UAlbany radar archive.

**FIG. 4.21:** As in Fig. 4.2 except for (a) 14Z/18, (b) 16Z/18, (c) 18Z/18, (d) 20Z/18, (e) 22Z/18, (f) 00Z/19, and (g) 02Z/19 June 2010. The arrows in panels (a), (c), and (e) illustrate the location of MCS-P2. Source: UAlbany radar archive.

**FIG. 4.22:** As in Fig. 4.13 except for 16Z/18 June 2010. The “P2” in each panel illustrates the location of MCS-P2. IR satellite from 1610Z/18 June 2010 is inset in the lower-right corner of panel (a). Source: 20-km RUC and UCAR MMM image archive.

**FIG. 4.23:** As in Fig. 4.5 except for a sounding from GRB at 18Z/18 June 2010. The inset in the upper-right corner shows the location of the sounding. Source: UAlbany sounding archive.

**FIG. 4.24:** As in Fig. 4.6 except for 20Z/18 June 2010. The “P2” illustrates the location of MCS-P2. IR satellite imagery from 2010Z/18 June 2010 is inset into the upper-left corner of the image. Source: UAlbany surface archive and UCAR MMM image archive.

**FIG. 4.25:** As in Fig. 4.8 except for 18Z/18 June 2010. The “P2” illustrates the location of MCS-P2. IR satellite imagery from 1810Z/18 June 2010 is inset into the upper-left corner of the image. Source: 1.0° GFS analysis and UCAR MMM image archive.

**FIG. 4.26:** As in Fig. 4.13 except for 22Z/18 June 2010. The “P2” in each panel illustrates the location of MCS-P2. IR satellite from 2210Z/18 June 2010 is inset in the lower-right corner of panel (a). Source: 20-km RUC and UCAR MMM image archive.
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FIG. 4.30: As in Fig. 4.1 except for between 00Z/24 and 16Z/24 June 2003 with MCS-D2. Source: Severe Plot v.3.0 and UAlbany radar archive.

FIG. 4.31: As in Fig. 4.12 except for (a) 02Z/24, (b) 04Z/24, (c) 06Z/24, (d) 08Z/24, (e) 10Z/24, (f) 12Z/24, (g) 14Z/24, and (h) 16Z/24 June 2003. The arrows in panels (a), (d), and (g) illustrate the location of MCS-D2. Source: UCAR MMM image archive.

FIG. 4.32: As in Fig. 4.13 except for 08Z/24 June 2003. The “D2” in each panel illustrates the location of MCS-D2. IR satellite imagery from 0808Z/24 June 2003 is inset in the lower-right corner of panel (a). Source: 20-km RUC and BAMEX field catalog.

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FIG. 4.34: As in Fig. 4.6 except for 12Z/24 June 2003. The “D2” illustrates the location of MCS-D2. IR satellite imagery from 1209Z/24 June 2003 is inset into the upper-left corner of the image. Source: UAlbany surface archive and BAMEX field catalog.
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FIG. 4.41: As in Fig. 4.13 except for 04Z/22 August 2007. The “P3” in each panel illustrates the location of MCS-P3. IR satellite imagery from 0410Z/22 August 2007 is inset in the lower-right corner of panel (a). Source: 20-km RUC and UCAR MMM image archive.

FIG. 4.42: As in Fig. 4.5 except for a RUC sounding at 42.23°N latitude and 89.03°W longitude for 06Z/22 August 2007. The inset in the upper-right corner shows the location of the sounding. Source: 20-km RUC analysis.

FIG. 4.43: As in Fig. 4.6 except for 06Z/22 August 2007. The “P3” illustrates the location of MCS-P3. IR satellite imagery from 0610Z/22 August 2007 is inset into the lower-right corner of the image. Source: UAlbany surface archive and UCAR MMM image archive.
FIG. 4.44: As in Fig. 4.8 except for 06Z/22 August 2007. The “P3” illustrates the location of MCS-P3. IR satellite imagery from 0610Z/22 August 2007 is inset into the lower-left corner of the image. Source: 1.0° GFS analysis and UCAR MMM image archive.

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FIG. 4.47: (a) As in Fig. 4.10a except at RFD from 06Z/22 to 13Z/22 August 2007, (b) as in Fig. 4.10b except for from 06Z/22 to 13Z/22 August 2007, and (c) the locations of the meteograms in (a) and (b). Source: UAlbany surface archive and NDBC.

FIG. 4.48: As in Fig. 4.1 except for between 00Z/06 and 16Z/06 July 2003 with MCS-D3. Source: Severe Plot v.3.0 and UAlbany radar archive.

FIG. 4.49: As in Fig. 4.12 except for (a) 04Z/06, (b) 06Z/06, (c) 08Z/06, (d) 10Z/06, (e) 12Z/06, (f) 14Z/06, (g) 16Z/06, and (h) 18Z/06 July 2003. The arrows in panels (a), (d), and (g) illustrate the location of MCS-D3. Source: UCAR MMM image archive.

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FIG. 4.51: As in Fig. 4.5 except for a sounding from GRB at 12Z/12 July 2003. The inset in the upper-right corner shows the location of the sounding. Source: UAlbany sounding archive.

FIG. 4.52: As in Fig. 4.15 except for 12Z/06 July 2003. The “D3” illustrates the location of MCS-D3. IR satellite imagery from 1209Z/06 July 2003 is inset into the lower-left corner of the image. Source: UAlbany surface archive and BAMEX field catalog.
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FIG. 4.55: a) The 975-hPa previous 2-h potential temperature decrease (shaded according to color bar in K) and 0–3-km shear (barbs in m s$^{-1}$, where one long barb equals 5 m s$^{-1}$) for 14Z/06 July 2003 and (b) cross section F–F’ of the previous 2-h potential temperature decrease (shaded according to color bar in K), potential temperature (black lines every 1 K), and wind (barbs in m s$^{-1}$, where one long barb equals 5 m s$^{-1}$) for 14Z/06 July 2003. NOWrad composite radar imagery for 12Z/06 and 14Z/06 July 2003 are inset into the lower-left and right corners of (a), respectively, with arrows indicating MCS-D3 location. Source: 20-km RUC analyses and UAlbany radar archive.

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FIG. 6.1: Idealized schematic of common features associated with persisting Lake Michigan-crossing MCSs. A region of MUCAPE (purple shading), DCAPE (magenta shading), moisture convergence (green shading), as well as an upper-level jet stream (orange shading), a LLJ (blue barbs), and a cold-pool-induced boundary (red dashed line) are indicated in this image.

FIG. 6.2: Idealized west–east cross section schematic of an early warm-season persisting MCS. Circulations are those associated with the convective cold pool, vertical shear, and Lake Michigan marine layer at (a) a time prior to intersecting Lake Michigan and (b) a time after intersecting Lake Michigan. The green arrow indicates convective ascent.
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**TABLE 2.1**: Breakdown of the number and percentage of MCSs that persisted and dissipated during the warm seasons of 2002–2007.
1 Introduction

1.1 Motivation and Purpose

The purpose of this dissertation is to gain insight into mature mesoscale convective systems (MCSs) that traverse the Great Lakes. More specifically, this investigation probes the important features and processes that differentiate between MCSs that persist and dissipate upon encountering a seasonally and diurnally varying Lake Michigan marine environment. This research employs a multiscale approach, to elucidate the roles of the synoptic-meso-, and storm-scale processes that affect MCS persistence and dissipation throughout the warm season.

MCS activity is favored in the Great Lakes region (e.g., Johns and Hirt 1987; Laing and Fritsch 1997). Dynamically, MCSs persist through a balance between circulations that are induced by the low-level vertical wind shear and the horizontal buoyancy gradient imparted by a convective cold pool [e.g., Rotunno et al. 1988 (hereafter RKW); Weisman and Rotunno 2004]. As an MCS approaches the marine air associated with the Great Lakes, the environmental conditions upon which the MCS is dependent on for persistence change. The summertime marine layer is cooler than the nearby land, containing reduced amounts of Convective Available Potential Energy (CAPE). As an MCS progresses into reduced-CAPE air, the likelihood of persistence typically decreases (e.g., Coniglio et al. 2004). However, a previous cursory subject analysis reveals that 68% of intense Lake Michigan-crossing MCSs maintain their radar-indicated base-reflectivity structure and produce a similar number of severe reports both

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1 The authors limited their focus to intense MCSs and generally only considered those MCSs that affected the Grand Rapids (GRR) forecast area (R. Graham 2009, personal communication)
upstream and downstream of the lake (Graham et al. 2004). Additionally, forecasters often struggle to determine whether an MCS will persist upon crossing the Great Lakes (R. Graham 2009, personal communication).

This dissertation will utilize the hourly gridded National Centers for Environmental Prediction (NCEP) 20-km Rapid Update Cycle (RUC) grids (Benjamin et al. 2004) for the analysis of the environmental conditions associated with MCSs that cross Lake Michigan. The north–south orientation and relatively straight coastline of Lake Michigan coupled with generally west to east progressing MCSs results in most MCSs intersecting Lake Michigan at a nearly perpendicular angle of incidence with the coastline. Additionally, the following data will be used extensively: (i) lake buoys from the National Data Buoy Center (NDBC), (ii) composite National Operational Weather radar (NOWrad) imagery [obtained from the University Corporation for Atmospheric Research (UCAR) and the WSI Corporation], (iii) atmospheric soundings from the National Weather Service (NWS) and Aircraft Meteorological Data Reports (AMDar), (iv) 1.0° GFS analysis grids, and (v) the 32-km North American Regional Reanalysis (NARR), to investigate the MCSs and their associated environments surrounding Lake Michigan. This data, along with satellite imagery and surface observations, augment the RUC-analysis grids and will be utilized to identify a climatology of warm-season MCSs that crossed Lake Michigan between 2002 and 2007. Further investigation into the persistence/dissipation of Lake Michigan-crossing MCSs will be performed through case study, composite, and Weather Research and Forecasting (WRF; Skamarock et al. 2005, 2008) modeling analyses.

The above analysis techniques will be employed in an attempt to understand the
following questions:

- What percentage of mature warm-season MCSs that cross Lake Michigan persist versus dissipate?
- What are the environmental and lake conditions associated with MCSs that persist versus those that dissipate?
- What are the implications in terms of RKW theory as a mature MCS interacts with the Lake Michigan shallow surface-based cold dome?
- What determines whether an MCS approaching Lake Michigan will ingest marine-layer air or become elevated over a lake surface-based cold dome, and if MCSs do become elevated over the Lake Michigan cold dome do their associated characteristics change (e.g., speed, organization)?
- What are the relative roles of the synoptic scale forcing and mesoscale lake interactions on MCS persistence and dissipation?

1.2 Thesis Organization

The remainder of this first chapter will provide an overview of related previous research that will serve as background scientific information, and allow this research to be placed into the proper perspective. Chapter 2 will provide results from a 6-yr subjective climatology of MCSs that crossed Lake Michigan spanning the warm seasons of 2002 through 2007. Chapter 3 will present compositing analyses of persisting and dissipating MCS environments from the same years. Chapter 4 will examine selected real data case studies of persisting and dissipating MCSs that occurred during a variety of different times of the day and year, while chapter 5 will discuss WRF-simulated case
studies to provide a clearer view on the role of Lake Michigan in MCS persistence and dissipation. Chapter 6 will summarize important results and place them into the context of previous research.

1.3 Background

1.3.1 MCS Overview

The *Glossary of Meteorology* defines an MCS as “a cloud system that occurs in connection with an ensemble of thunderstorms and produces a contiguous precipitation area on the order of 100 km or more in horizontal scale in at least one direction. An MCS exhibits deep, moist convective overturning contiguous with or embedded within a mesoscale vertical circulation that is at least partially driven by convective overturning” (Glickman 2000, p. 486). Thus, MCSs are at least meso-β (20–200 km) in size (Orlanski 1975). MCSs can cause significant wind damage (e.g., Johns and Hirt 1987) as well as extreme amounts of precipitation (e.g., Schumacher and Johnson 2005). The two main categories of MCSs that have been studied extensively are the mesoscale convective complex (MCC) and the derecho.

The MCC definition was formulated by Maddox (1980) and relates to the physical characteristics of midlatitude convective weather systems that can be observed from enhanced infrared (IR) satellite imagery. The characteristics of an MCC are as follows:

(a) The cloud shield must have continuously low IR temperatures ≤ −32°C over an area ≥ 10^5 km^2.
(b) An interior portion of the cold cloud shield must have IR temperatures ≤ −52°C and cover an area ≥ 0.5 × 10^5 km^2.
(c) Size definitions (a) and (b) must be met for a minimum of 6 h.
(d) The eccentricity between the major and minor axis must be ≥ 0.7 at the time of maximum extent.
MCCs can occur across the world, but they have a local maximum of occurrence across the Midwest and western Great Lakes region (Laing and Fritsch 1997; Fig. 1.1a).

A second major subset of MCSs is derechos. In the U.S., rapidly moving MCSs that produce significant severe wind damage during the warm season are common. These wind-producing MCSs were first observed during the late 19th century and were referred to as derechos, a Spanish word meaning “straight ahead” (Hinrich 1888). Later, Fujita and Wakimoto (1981) referred to wind damage on a similar scale as a derecho as a family of downburst clusters, where the major axis of damaging wind was at least 400-km long. Johns and Hirt (1987) adapted the term derecho to include any family of downburst clusters produced by MCSs and set forth the following criteria for a derecho:

(a) A concentrated area of wind damage or convective gusts > 26 m s\(^{-1}\) must be present covering at least 400 km.
(b) The reports must show a nonrandom chronological pattern of progression.
(c) There must be at least three reports, separated by 64 km or more, of either F1 damage (Fujita 1981; Fujita and Wakimoto 1981) and/or convective gusts of 33 m s\(^{-1}\) or greater within the path of strong wind.\(^3\)
(d) No more than 3 h can elapse between successive wind damage (gust) events.
(e) Each convectively induced system must have temporal and spatial continuity on NWS radar charts, but the radar echoes need not be continuous.
(f) Multiple swaths of wind damage must be associated with the same MCS.

The main corridor for derecho activity stretches from the Upper Midwest across the southern Great Lakes (Johns and Hirt 1987; Fig.1.1b). Derechos appear to account for about 40% of the causalities associated with convectively induced nontornadic winds

\(^2\) In some subsequent studies (e.g., Bentley and Mote 1998), the derecho criteria have been modified slightly from the original Johns and Hirt (1987) definition [e.g., an elimination of (c) above].
\(^3\) The Fujita scale for damaging wind is as follows: F0 (18–33 m s\(^{-1}\)); F1 (33–50 m s\(^{-1}\)); F2 (50–70 m s\(^{-1}\)); F3 (70–92 m s\(^{-1}\)); F4 (92–117 m s\(^{-1}\)); F5 (117–143 m s\(^{-1}\)).
(Ashley and Mote 2005). During 1986–2003, derechos caused 153 deaths and led to 2605 injuries across the U.S. (Ashley and Mote 2005). In comparison, during the same period, 254 people perished because of hurricanes striking the U.S.

Many of the MCSs that meet the derecho criteria are associated with a bow echo (e.g., Fujita 1978; Weisman 2001) or are linear in organization. The term bow echo reflects the radar representation of an MCS where an intense line of convective cells evolves into a convective bow shape with a trailing stratiform precipitation region. As an MCS bows out, cyclonic and anticyclonic bookend vortices often form on the northern and southern flanks of the bow, respectively, and a rear inflow jet develops along the apex of the bow. In time, the cyclonic vortex dominates and the system weakens.

1.3.2 Great Lakes Structure and Atmospheric Effects

Relatively few studies have focused on warm-season convection in the Great Lakes region. This is surprising given that the Great Lakes region is an area of frequent MCS activity (e.g., Johns and Hirt 1987; Laing and Fritsch 1997; Figs. 1a,b). Changnon (1968) found a 6% annual reduction in the amount of precipitation that fell over Lake Michigan when compared to the surrounding land. This percentage increased to 14% when only summer months were considered. Bolsenga (1977) utilized precipitation gauges across northern Lake Michigan to confirm the decreased over-lake precipitation when Lake Michigan was colder than the surrounding land from May through August (Fig. 1.2). However, when Lake Michigan was warmer than the surrounding land in September, precipitation totals over the lake were actually larger than over land (Fig. 1.2). During the international field year for the Great Lakes (April 1972–May 1973), the
amount of precipitation over Lake Ontario and the surrounding region was measured (Wilson 1977). In general, the cooler lake waters over the summer suppressed precipitation by about 10% when compared to the onshore regions within 30 km of Lake Ontario (Wilson 1977). This convective suppression was most favored when convection was not tied to large-scale, well-organized weather systems. Conversely, Chandik and Lyons (1973) found no systematic suppression of over-lake, warm-season rain based on WSR–57 radar data from Chicago.

The Lake Michigan near-surface environment is typically much cooler than the surrounding land throughout much of spring and summer due to the higher specific heat capacity of water as compared with land. Thus, Lake Michigan can cause downwind cumulus cloud clearing throughout much of the early and mid warm season as cooler marine impacted air advects onshore, thermally stabilizing the environment (e.g., Purdom 1990; Segal et al. 1997). Additionally, the cool water temperature of Lake Michigan, especially during the early warm season, promotes the development of an intense near-surface lake inversion (Bellaire 1965). These inversions develop as warm air is advected over Lake Michigan. The cool lake surface water modifies the air temperature of the environment immediately above the water. The temperature then increases upward until reaching the free atmosphere.

In a field experiment during early May 1976, Wylie and Young (1979) steamed a ship from southwest to northeast across central Lake Michigan (Fig. 1.3a) to measure the depth of the Lake Michigan marine layer. On the day of this experiment, there was ambient southwesterly flow at Milwaukee, Wisconsin (MKE). Wylie and Young (1979) took a vertical profile of temperature at both 43 and 150 km distance from MKE along
the ship transit line (Fig. 1.3a) and found that the inversion associated with this lake-induced cold dome extended upward ~140 m from the surface, and the temperature increased 6–7°C through this layer in both locations (Fig. 1.3b). The 150-km profile showed an isothermal layer that extended another ~100 m on top of this inversion, indicating that stable conditions were somewhat deeper as air crossed a longer fetch of Lake Michigan. Substantial momentum decoupling occurred between the inversion and the overlying atmosphere. Near surface wind speeds of 5–10 m s$^{-1}$ nearly doubled over the lowest ~100 m of the over-lake atmosphere. Average temperature increase rates at a point above the Lake Michigan inversion, were similar to the surface rates at nearby MKE, indicating that the surface environment at MKE was likely similar to the “surface” environment immediately above the Lake Michigan cold dome.

The difference in temperature between the cool marine layer associated with Lake Michigan and the onshore land temperature can induce a lake- (land)-breeze circulation, with onshore (offshore) flow during the daytime (nighttime) hours (e.g. Lyons 1972; Ryznar and Touma 1981; Laird et al 2001). Laird et al. (2001) examined a 15-yr climatology of Lake Michigan lake-breeze events and found their maximum occurrence to be in August. Typically, Lake Michigan lake breezes are associated with surface high pressure and surface winds of less than 5 m s$^{-1}$. Thus, even though the land–lake temperature gradient is largest in May and June, there are not as many lake-breeze events during these months because the synoptic-scale flow is too strong. This result agrees with Arritt (1993) who found that onshore synoptic-scale flow of only a few m s$^{-1}$ was sufficient to suppress lake breeze development. The presence of a lake breeze can cool the summer mean temperature by up to 2°C within 80 km of Lake Michigan (Scott and
Additionally, summertime convection can often form along the boundary interface between these mesoscale breezes and the ambient environment (e.g., Laird et al 1995; Smith et al. 2005).

1.3.3 MCSs near the Great Lakes

Derechos (e.g., Johns and Hirt 1987; Bentley and Sparks 2003) and MCCs (e.g., Maddox 1980) tend to progress across the upper Midwest and Great Lakes, especially during the warm season. Thus, it is necessary to understand evolutionary and lifecycle changes to these MCSs upon crossing Lake Michigan and the associated lake-induced cold dome. Augustine and Howard (1991) show several MCC tracks during the 1986 and 1987 warm seasons (their Figs. 2 and 3) that crossed Lake Michigan and moved into Lower Michigan, while others terminated over the water. In general, the amount of lightning and associated convective activity increases near Lake Michigan and the Great Lakes in general, as the warm-season progresses from spring to summer, reflecting the climatologically favored time for MCSs in this region (Johns and Hirt 1987; Changnon 1988).

Many of the severe MCSs that eventually cross the Great Lakes initially form to the lee of the Rocky Mountains (e.g., Carbone et al. 2002; Ahijevych et al. 2004; Trier et al. 2006, 2010). These long-lived convective episodes, or time–space clusters of precipitation that result from sequences of MCSs, progress eastward after forming and can cover up to 2500 km before dissipating (Fig. 1.4). Ahijevych et al. (2004) showed that the frequency of these convective episodes is strongly associated with the time of day and the distance east of the Rocky Mountains. These convective episodes often occur
along preferred corridors that last for 2–7 days (Tuttle and Davis 2006). When the preferred corridor is far enough north, these convective episodes can repeatedly move into close proximity with and even cross Lake Michigan and subsequently the other Great Lakes (e.g., Metz and Bosart 2010). However, in order for a convective episode to progress as far eastward as Lake Michigan, a process such as coherent regeneration often needs to occur in order to reinvigorate the MCS and allow the system to cross such a large distance (e.g., Tuttle and Carbone 2004; Metz and Bosart 2010).

Graham et al. (2004) investigated a subset of MCSs that crossed Lake Michigan to determine the result of MCS–lake interaction. While the study was somewhat limited in scope, the authors did find that 68% of intense, severe-weather producing MCSs identified from 1996 to 2001 persisted upon crossing Lake Michigan. Although the authors did not give precise numbers, they found that when the lake water temperature was at least 2.5°C colder than the 5-m buoy air temperature, MCSs tended to persist upon crossing Lake Michigan, likely becoming elevated over the Lake Michigan cold dome. Workhoff (2010) examined a climatology of convective storms that crossed Lake Erie and found that as the organizational structure of the storms increased (e.g., isolated/cluster storms, to linear/complex MCSs), the storms weakened/dissipated at a reduced rate.

1.3.4 MCS Convective Ingredients, Maintenance, and Evolution

It has long been understood that thunderstorms have a nocturnal maximum over the central U.S. (e.g., Maddox 1980). Thunderstorms, and more specifically MCSs that occur at night or on the cool side of a surface boundary, can be elevated and decoupled
from the boundary layer (Colman 1990a,b). Lower tropospheric warm advection over a surface boundary in the presence of a low-level jet (LLJ) can lift parcels into these MCSs (e.g., Fritsch and Forbes 2001). These elevated MCSs often behave as bore-like systems (e.g., Clarke et al. 1981; Klemp et al. 1997) as a well-developed density current associated with an MCS progresses into a nocturnal, near-surface stable region. Thus, the classical cold pool viewpoint on convective evolution, where MCSs persist through a balance between circulations induced by the low-level vertical wind shear and the horizontal buoyancy gradient caused by the convective cold pool, may not apply (e.g. RKW; Carbone et al. 1990; Weisman and Rotunno 2004). However, some MCSs that occur on the cool sides of boundaries do still produce severe weather (e.g., Johns and Hirt 1987; Coniglio and Stensrud 2001). Parker (2008) allowed simulated storms to mature while gradually cooling the lowest model levels. He found that even with low-level cooling of up to 10°C, boundary-layer parcels still could be lifted to their level of free convection (LFC) by the convective cold pool. Thus, many of the MCSs once thought to be elevated, perhaps instead were “surface” based even if they occurred over nocturnal cold pools. Provided an elevated source of CAPE was available, the simulated MCSs did eventually become elevated when even more cooling occurred in the model above the aforementioned 10°C, and the MCSs then propagated at a speed similar to a bore.

The aforementioned cold-pool viewpoint on convective evolution is described by RKW theory, which addresses how typical thunderstorms are regenerated along a line, resulting in an MCS. RKW found that the ability of an old cell to be lifted by cold surface outflow to its LFC is enhanced by the presence of low-level vertical shear. The authors concluded that cold-pool–shear interactions are crucial to understanding MCS
maintenance (Weisman and Rotunno 2004). For a cold pool spreading at the surface in a
nonsheared environment, the circulation at the leading edge can be described by density
current dynamics (e.g., Simpson 1997). When environmental vertical shear is present, the
shear-induced horizontal circulation will partially counteract the horizontal circulation
associated with the convective cold pool on the downshear side. The interaction between
these two circulations, which are manifest as vorticity about a horizontal axis, results in
deep, upright lifting on the downshear side of the cold pool (Weisman and Rotunno
2004).

In order to visualize RKW theory, consider the life cycle of a squall line where C
is the theoretical speed of propagation of a cold pool (e.g., Benjamin 1968) that can be
related to the strength of the horizontal circulation associated with a convective cold pool
and ΔU is the strength of the horizontal circulation associated with low-level ambient
vertical wind shear. Before cold pool development, the convective cells within squall line
tilt downshear due to the ambient vertical shear (Weisman and Rotunno; Fig. 1.5a). Once
a cold pool develops, its circulation about a horizontal axis counteracts that associated
with the ambient vertical shear (Fig. 1.5b). As a result, deep ascent occurs at low levels
and the MCS becomes upright. Finally, the cold pool becomes stronger and the
circulation associated with the cold pool begins to dominate over that of the shear (Fig.
1.5c). This final stage leads the convective cells and associated heating/cooling to shift
rearward, inducing a rear inflow jet. These rear-inflow jets are generated by a buoyancy
gradient associated with warm air aloft in the ascending front-to-rear flow and the surface
convective cold pool (e.g., Weisman 1992; Fritsch and Forbes 2001; Fig. 1.5c).

The optimal state in RKW theory occurs when C/ΔU is one, resulting in vertically
erect updrafts at the leading edge of the MCS. However, in many MCSs, \( C/\Delta U \) can become greater than one as the cold pool evolves and strengthens, resulting in bow echo development. In idealized simulations, optimal squall lines occur with moderate-to-strong vertical wind shear oriented perpendicular to the squall line in the lowest 2.5 km of the atmosphere (e.g., Weisman et al. 1988; Weisman 1992). RKW theory only considers the balance between the strength of the convective cold pool and the environmental shear and does not take into account buoyancy heterogeneities in the environment. Lericos et al. (2007) found that in simulated squall lines that passed from water to land, the tilt of the updraft circulation did change depending upon the strength and depth of the shear profile.

Since the convective cold pool is so important to MCS maintenance, the potential for strong, evaporatively cooled downdrafts must be considered when forecasting MCSs. Thunderstorms that form in environments with dry midtropospheric air, often associated with an elevated mixed layer advected from off the Mexican Plateau, have the potential for stronger downdrafts (e.g., Browning and Ludlam 1962). Emanuel (1994) derived downdraft CAPE (DCAPE) as the maximum increase in kinetic energy that could result from the evaporative cooling of air. Cohen et al. (2007) found that DCAPE could be used to identify MCS intensity as MCS organization and severity increased as the amount of DCAPE in the environment increased.

Through idealized numerical simulations, Weisman (1993) determined the CAPE and shear environment associated with different convective modes. Based on the results of the experiments, Weisman (1993) created a CAPE–shear phase space for 0–2.5 km (Fig. 1.6a) and 0–5 km shear (Fig. 1.6b). Four main types of convective organization were identified. The first type, “W,” consists of weak, upshear-tilted systems where the
cold pool circulation overwhelms the vertical shear. These systems tend to dissipate quickly (gust out) as an upshear-tilt results in rain behind the gust front. These weak, upshear-tilted systems occur in association with minimal low-level shear regardless of the amount of CAPE present. The second type of convective organization, “B,” is the long-lived bow echo. In the idealized experiments, bow echoes occur with 20–30 m s\(^{-1}\) of shear and CAPE values above 2000–3000 J kg\(^{-1}\). Isolated long-lived splitting supercells, “S,” form with the highest shear values in the numerical experiments. The “I” designation refers to a hybrid organization between weaker convective systems and bow echoes. Weisman (1993) conceded that observations show that bow echoes can develop with \(\sim 5\) m s\(^{-1}\) weaker shear than in the idealized simulations. Studies by Evans and Doswell (2001) and Coniglio et al. (2004) confirmed that observations reveal that bow echoes tend to form in slightly weaker shear than was identified in Weisman (1993; Figs. 1.6a,b).

The low-level jet (LLJ) has long been recognized as another important contributor to MCS maintenance. An LLJ occurs in many locations throughout the world, but perhaps none more frequently than the central U.S. (e.g., Bonner 1968; Douglas 1993; Stensrud 1996). The maximum speed associated with the LLJ typically occurs at night, near the top of the nocturnal boundary layer (e.g., Means 1952; Blackadar 1957; Bonner 1968). These LLJs have been observed in a majority of MCC (e.g., Maddox 1983; Laing and Fritsch 1997) and derecho (e.g., Johns and Hirt 1987; Coniglio et al. 2004; Metz and Bosart 2010) environments. Laing and Fritsch (2000) showed that in a composite of 12 summertime MCCs, a robust LLJ impinged upon the MCC-genesis region (Fig. 1.7). The LLJ is favorable for convective development, enhancement, and maintenance for multiple reasons. First, an LLJ can come into a favorable superposition with an upper-
level jet stream, enhancing synoptic scale upward motion favorable for convective development (Uccellini and Johnson 1979). Perhaps more importantly though, when an LLJ is present, sensible heat and moisture transport (Fig. 1.7) increase two and three times, respectively, over situations without an LLJ (Uccellini and Johnson 1979). When these LLJs intersect surface fronts at a perpendicular angle, extreme rainfall-producing MCSs can result in conjunction with enhanced near-surface convergence and echo training (e.g., Maddox et al., 1979; Schumacher and Johnson 2005, 2006, 2009; Galarneau et al. 2010; Schumacher et al. 2011). Thus, the LLJ allows for the maintenance of convective episodes that often develop in the lee of the Rockies, and prolong the longevity of individual MCSs (Carbone et al. 1990).

Letkewicz and Parker (2010) examined Appalachian Mountain-crossing MCSs to deduce how local effects might alter MCS persistence and dissipation. They found that MCSs that crossed the Appalachians actually persisted with ~7 m s\(^{-1}\) of less shear than those that dissipated. The authors formulated a couple of possible hypotheses for this apparent contradiction. First, the associated weaker downslope flow to the lee of the mountains in the weaker shear cases might result in less convective suppression. Second, since the terrain partially blocks the MCS cold pool, weaker shear to the lee of the Appalachians may be beneficial to balance a weaker cold pool. However, even when considering these local effects, the amount of CAPE that the MCSs moved into downstream of the mountains was the main differentiating factor between the MCSs that persisted and dissipated. The MCSs that persisted crossed into an environment with most-unstable CAPE (MUCAPE) values that were ~1600 J kg\(^{-1}\) greater than those that dissipated.
1.4 Chapter 1 Figures

FIG 1.1: (a) Global distributions of MCCs (black dots) and regions of widespread deep atmospheric convection as inferred by outgoing longwave radiation (shading indicates outgoing longwave radiation minima) [from Laing and Fritsch (1997)]. (b) Isolines of the average number of derechos (contoured every 3 beginning at 3) located within $2^\circ \times 2^\circ$ boxes for 1980–1983 [from Johns and Hirt (1987), modified by Coniglio and Stensrud (2004)].
FIG. 1.2: Average monthly lake (solid line) and land (dashed line) precipitation for 1963–1968 measured from stations surrounding northern Lake Michigan. The lake stations were situated on islands in the middle of northern Lake Michigan, while land stations were in either Wisconsin or Michigan [from Bolsenga (1977)].
FIG. 1.3: (a) Surface analysis near Lake Michigan for 1200 UTC 4 May 1976 of sea-level pressure (every 4 hPa) and selected surface wind observations (m s$^{-1}$). The track of the ship Neeskay is illustrated by the dashed line. (b) Vertical profiles of temperature made at 43 and 150 km along the dashed line in (a). The dashed line next to the 43-km profile represents a dry adiabat [from Wylie and Young (1979)].
FIG. 1.4: Time–longitude Hovmöller diagram of radar-estimated rainfall rate averaged between 30° and 48°N for 18–21 July 2000. Three long-lived precipitation episodes are evident, all beginning at approximately 2200 UTC, 105°W, on 17, 19, and 20 July [from Ahijevych et al. (2004)].
FIG. 1.5: Stages in the development of an MCS where (a) the circulation caused by the ambient vertical shear dominates over the cold pool circulation, (b) the circulation caused by the ambient vertical shear is nearly equal to the cold pool circulation, and (c) the cold pool circulation dominates the circulation caused by the ambient vertical shear and a rear-inflow jet is induced [from Weisman and Rotunno (2004)].
FIG. 1.6: Modes of convective organization associated with CAPE and (a) 0–2.5-km shear and (b) 0–5-km shear. The letters refer to idealized model output where “W” represents weak upshear-tilted MCSs, “I” represents an intermediate form, “B” represents a bow echoes, and “S” represents split supercells. The shapes on each phase space represent observed MCSs according to the legend at the top of the figure [from Coniglio et al. (2004), adapted from Weisman (1993)].
FIG. 1.7: A synoptic-scale composite of the mean MCC-genesis environment showing 850-hPa geopotential height (solid lines in m), 850-hPa mixing ratio (dashed lines in g kg\(^{-1}\)), and 850-hPa wind vectors (arrows in m s\(^{-1}\)). The composite consists of 12 individual MCC cases spanning 1986–1987. The quadrilateral in the center of the image indicates the composite MCC genesis region. [from Laing and Fritsch (2000)].
2. Climatology of Lake Michigan-Crossing MCSs

2.1 Purpose

The purpose of this chapter is to examine a climatology of recent Lake Michigan-crossing MCSs. Graham et al. (2004) provided a brief examination of MCSs that crossed Lake Michigan. However, this work only focused on a random subset of these lake-crossing MCSs that were intense and entered the GRR forecast area in Lower Michigan. The goal of this chapter is to provide a much more comprehensive examination of all MCSs that crossed Lake Michigan during the warm-seasons (April–September) of 2002 through 2007. Not only will the persistence and dissipation of these MCSs be recorded, but the environmental and lake characteristics associated with each lake-crossing MCS will also be categorized. Additionally, the results from the climatology will serve as a basis for compositing and case study analyses in subsequent chapters.

2.2 Data and Methods

NOWrad⁴ composite radar imagery was utilized to compile the climatology of MCSs that crossed Lake Michigan during the warm-seasons of 2002–2007. April through September was chosen for the warm-season definition because these months typically feature MCSs progressing through the Great Lakes region (e.g., Johns 1982). The years 2002 through 2007 were chosen for the length of the climatology because the NOWrad radar imagery dataset was available in a consistent format from UCAR for this entire period. The NOWrad radar imagery was produced by the WSI Corporation and utilizes all radars in the WSR-88D network to produce a reflectivity composite image every 15

⁴ Acquired from: http://www.mmm.ucar.edu/imagearchive/WSI/
min (e.g., Ahijevych et al. 2004). The details of the NOWrad dataset are proprietary to the WSI Corporation.

This climatology was constructed subjectively by examining the NOWrad radar imagery every 30 min for all six warm seasons. In order for an MCS to be added to the climatology, it needed to be at least $100 \times 50$ km$^2$ on the NOWrad radar imagery (e.g., Glickman 2000, p. 486) and contain a continuous region at least 100 km long in any direction of 45-dBZ or greater radar returns. This second criterion ensured that each MCS was in fact convective. These two reflectivity criteria needed to be met on consecutive 30-min NOWrad radar images for at least three hours (seven images) leading up to MCS intersection with Lake Michigan to ensure that the MCSs had continuity. MCSs persisted upon crossing Lake Michigan if they still satisfied the two aforementioned reflectivity criteria and produced at least one severe weather report after traversing the water.

In addition to utilizing the NOWrad composite radar imagery to construct the climatology of Lake Michigan-crossing MCSs, other datasets were also employed. The MUCAPE and 0–6-km shear associated these MCSs were calculated using proximity soundings both upstream and downstream from Lake Michigan at Green Bay, WI (GRB), Davenport, IA (DVN), Lincoln, IL (ILX), and White Lake, MI (DTX). In order for a proximity sounding to be valid on the upstream side of Lake Michigan, the sounding needed to occur within 150 km and 2 h of MCS passage and not sample environmental air altered by convection (e.g., Evans and Doswell 2001). Proximity soundings from DTX, downstream of Lake Michigan, were taken within 2 h of the MCS being over Lake Michigan and could not sample environmental air altered by convection. Thus, not every MCS had a valid proximity sounding. The MUCAPE was calculated by utilizing the
parcel with the highest equivalent potential temperature in the lowest 300 hPa of the sounding. MUCAPE was chosen over other CAPE varieties, because it enabled sounding comparisons from all times of the day. For example, surface-based CAPE is nearly zero in many of the 1200 UTC soundings, even though there is ample CAPE present immediately above the shallow nocturnal inversion.

DCAPE was also calculated from upstream proximity soundings at GRB, DVN, and ILX. These were the same upstream soundings utilized for MUCAPE and 0–6-km shear. The value of DCAPE in each proximity sounding was calculated by following the moist adiabat to the ground from the minimum equivalent potential temperature value in the lowest 400 hPa of the sounding. This technique for calculating DCAPE is similar to the technique employed by the Storm Prediction Center5.

Lake surface temperatures were calculated by utilizing two buoys located in the center of the northern (45002) and southern (45007) portions of Lake Michigan. Meteorological data from the buoy that was located closest to each Lake Michigan-crossing MCS was employed in creating this climatology. The intensity of the near-surface inversion over Lake Michigan was calculated by taking the difference in the 5-m and lake surface temperatures from the closest Lake Michigan buoy at the time of MCS passage. The strength of the horizontal land–lake temperature difference was calculated by subtracting the lake water temperature at the nearest buoy from the average temperature calculated from all available land stations immediately ahead of each MCS in the hour prior to intersection with Lake Michigan.

The strength of the 850-hPa LLJ impinging upon lake-crossing MCSs was

5 http://www.spc.noaa.gov/exper/soundings/help/index.html
calculated from the NARR\(^6\) by examining the maximum magnitude of the 850-hPa winds that intersected each MCS at the time of lake crossing. The NARR dataset was chosen because it was available at 32-km resolution over the entire length of the 6-yr climatology. The RUC\(^7\) dataset was only available at 40-km resolution during 2002 before transitioning to 20-km in 2003. The NARR is output every 3 h, and the closest time to lake crossing was chosen for analysis.

The magnitude of the 850-hPa Q-vector convergence/divergence associated with each Lake Michigan-crossing MCS was calculated using the 1.0° resolution GFS-analysis grids. The GFS was utilized for this analysis because the Q-vector convergence (divergence) is a representation of the synoptic-scale forcing for ascent (descent), and thus, the resolution of the GFS was appropriate to identify these synoptic-scale signatures. The quasi-geostrophic omega equation can be approximated as

\[
\left(\nabla^2_p + \frac{f_0^2}{\sigma} \frac{\partial^2}{\partial p^2}\right) \omega \approx -2\nabla_p \cdot \vec{Q}
\]

where the Q-vector is represented as

\[
\vec{Q} = -\frac{R}{\sigma} \begin{pmatrix} \frac{\partial \nabla_x}{\partial x} \cdot \nabla_p T \\ \frac{\partial \nabla_y}{\partial y} \cdot \nabla_p T \end{pmatrix} = \begin{pmatrix} Q_1 \\ Q_2 \end{pmatrix}
\]

with R and \( \sigma \) representing the dry gas constant \((287 \text{ J K}^{-1} \text{ kg}^{-1})\) and the static stability parameter \((2 \times 10^{-6} \text{ m}^2 \text{ Pa}^{-2} \text{ s}^{-2})\), respectively (Bluestein 1992, p. 352–353). This form of the quasi-geostrophic omega equation neglects a second right hand side “Beta” term, which is typically much smaller than the first right hand side term listed that represents

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\(^6\) Acquired from: http://nomads.ncdc.noaa.gov/#narr_datasets

\(^7\) http://ruc.noaa.gov/
the convergence/divergence of the Q-vector. Q-vector convergence (forcing for ascent) is associated with a positive right hand side of the aforementioned quasi-geostrophic omega equation. The GFS is output every 6 h, so the analysis grid closest to the time each MCS initially intersected Lake Michigan was chosen for analysis. If the MCS intersected Lake Michigan directly between two analysis times, the earlier time was chosen for consistency\(^8\). In each case, the magnitude of the 850-hPa Q-vector convergence was calculated by identifying the location of the MCS centroid at the analysis time and averaging in a 200-km circle around this location\(^9\).

This chapter presents many meteorological variables by comparing values associated with MCSs that persist and dissipate. Thus, in order to test the robustness of any differences, statistical significance testing was performed using a Student’s T-test for paired samples (e.g., Wilks 2006). The Mann–Whitney test was used to confirm results found with the Student’s t-test, because the Mann–Whitney test does not require normal distributions of variables (e.g., Wilks 2006).

2.3 General Characteristics

In order to begin to understand the effects of Lake Michigan on the distribution of convection in the Great Lakes region, daily time series were created representing the average number of square kilometers covered by 45-dBZ or greater convection at every hour of the day within a box. Three time series are color coded to match three boxes, one to the west, one directly over, and one to the east of Lake Michigan (Fig. 2.1a). These averages spanned the warm-seasons of 2002–2007, just like the climatology of MCSs.

\(^8\) The results do not change if the later time was chosen instead.
\(^9\) Other averaging radii were tested and did not produce significantly different results.
Additionally, the plot is divided into three regimes, regime I that spans the late night and early morning hours, regime II that spans the late morning and afternoon hours, and regime III that spans the evening and most of the overnight hours. In regime I, the average areal coverage of 45-dBZ or greater convection decreases in each box, as would be expected during these hours (Fig. 2.1a). In regime II, daytime heating occurs and convective coverage increases, resulting in the increase of 45-dBZ or greater convection in all three boxes. More convective development occurred in the overland boxes than over Lake Michigan, likely due to the thermal stability of the lake. During regime III, west–east moving MCSs often enter the three boxes (e.g., Johns and Hirt 1987; Augustine and Howard 1991), and there is a ~40% decrease in 45-dBZ or greater convective coverage from west to east across Lake Michigan. An examination of three similar time series associated with boxes to the south of Lake Michigan (Fig. 2.1b), does not show the same 40% reduction in convective coverage in regime III. In fact, the three time series look quite similar\textsuperscript{10}. The precise role of Lake Michigan in this west to east reduction in convective coverage during regime III will be investigated in this dissertation (Fig. 2.1a).

Using the criteria discussed in section 2.2, 110 total MCSs were found to intersect Lake Michigan during the warm seasons of 2002–2007. Of these 110, 47 (43%) persisted, while 63 (57%) dissipated (Table 2.1). These persistence/dissipation numbers are somewhat different from the results of Graham et al. (2004) where they found that 68% of MCSs persisted, 8% weakened, and 24% dissipated. Recall, that the Graham et al. (2004) climatology covered 1996–2001 and tended to focus on intense MCSs, which

\textsuperscript{10} It is important to note that the convective environments at the latitudes of Lake Michigan (Fig. 2.1a) and further to the south (Fig. 2.1b) are not always the same.
could explain these differences. Additionally, this thesis does not separate out MCSs that weakened and dissipated. Of the 63 MCSs that crossed Lake Michigan and were included in the dissipation category, a few did still have isolated regions of 45 dBZ or greater convection. However, these regions of convection no longer met the 100-km long criterion, and none of these MCSs produced any severe weather after crossing Lake Michigan. Most of this isolated remaining convection typically dissipated soon after crossing Lake Michigan. Therefore, these MCSs were included in the dissipation category.

Daily time series similar to those in Fig. 2.1a were created, but this time only encompassing warm-season days from 2002–2007 on which one of the 110 Lake Michigan-crossing MCSs was in existence\(^\text{11}\) (Fig. 2.2). During regime I, the amount of 45 dBZ or greater convection decreases in all boxes, in a very similar manner to the time series that encompassed all warm-season days (Figs. 2.1a and 2.2). In regime II, the amount of convection in all three boxes once again increases, coincident with an increase in daytime heating (Fig. 2.2). In regime III, there is a 50% or greater decrease in the amount of convection from west to east across Lake Michigan (Fig. 2.2). Recall, that during the evening and overnight hours, there are typically west to east moving MCSs progressing across the Great Lakes region (e.g., Johns and Hirt 1987; Augustine and Howard 1991), many of which appear in the climatology (Table 1). The west-to-east decrease in 45 dBZ or greater convection (Fig. 2.2) is consistent with the 57% of warm-season MCSs in the climatology that dissipated upon crossing Lake Michigan (Table 1)

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\(^{11}\) Note that the number of square kilometers covered by 45 dBZ or greater convection is much higher in Fig. 2.2 than in Fig. 2.1a. The average in Fig. 2.1a covers the entire warm seasons of 2002–2007, including many days when there was no convection in any of the boxes.
and objectively supports the subjective climatology.

The 110 MCSs in the climatology were not equally distributed by year, as there was interannual variability in the number of Lake Michigan-crossing MCSs. The maximum number of MCSs (30) occurred in 2002, while the minimum number (11) occurred in 2006 (Fig. 2.3). Generally, there were more dissipating MCSs than persisting MCSs during each year. However, 2003 and 2004 contained more persisting MCSs than dissipating MCSs (Fig. 2.3). These two years were rife with long-track, intense MCSs, many of which were part of convective episodes that progressed eastward from the lee of the Rockies (e.g., Metz and Bosart 2010).

The MCSs that made up the climatology occurred most often during the summer months, as this time of the year is the most favored period for MCSs to occur in the Great Lakes region (e.g., Johns and Hirt 1987; Fig. 2.4). Additionally, the greatest number of MCSs persisted during July and August when the Lake Michigan water temperature was the warmest. During these two months, the average Lake Michigan water temperature is near or in excess of 20°C. However, a number of MCSs also persisted upon crossing Lake Michigan during the earlier months of the warm season when the lake water temperature was much cooler (3–10°C; Fig. 2.4).

Additionally, MCSs in the climatology were most favored to occur and persist during the early evening and overnight hours (Fig. 2.5) as these hours coincide with the climatological maximum of the LLJ (e.g., Stensrud 1996). Furthermore, during the overnight hours, over-lake convergence is possible, especially during the later warm season as the water temperature of Lake Michigan is often warmer than the nearby land temperature. This warmer water temperature can result in offshore flow (e.g., Case et al.
2005) on both sides of Lake Michigan, possible mid-lake convergence, and the potential enhanced upward vertical motion. Furthermore, due to nocturnal cooling, many of the overnight MCSs are elevated in nature, feeding off unstable air located just above the nocturnal surface inversion (e.g., Colman 1990a,b). Given the shallow nature of the Lake Michigan cold dome, many of these elevated MCSs may never “see” Lake Michigan below (more discussion of elevated MCSs will occur in chapter 4). The evening and overnight also feature many dissipating MCSs and this time period corresponds to regime III in Figs. 2.1a and 2.2, when a notable decrease in 45-dBZ or greater convection occurs from west to east over Lake Michigan. Much of this decrease is likely associated with the dissipating MCSs in the climatology during the same hours (Fig. 2.5).

Previous research has found that the average MCS duration is ~9 h (e.g., Johns and Hirt 1987; Geerts 1998). Thus, an examination of the number of hours after formation that each MCS encountered Lake Michigan is warranted to see if the MCSs that dissipated were generally near the end of their life cycle. Sixty-three of the 110 MCSs in the climatology existed for 3–6.5 hours before crossing Lake Michigan, and these MCSs persisted at a 46% rate (Fig. 2.6), just slightly higher than the 43% persistence rate for the entire climatology (Table 1). Another 29 MCSs existed for 7–10.5 hours, and these persisted at a 41% rate. Finally, 18 MCSs existed for 11 hours or more prior to intersecting Lake Michigan. These MCSs persisted at a 33% rate. Even though MCSs that were around for a longer period of time, persisted at a slightly lower rate, there is no statistical significance in this data, and the number of hours an MCS was present prior to lake crossing, is not a good predictor of MCS persistence/dissipation upon crossing Lake Michigan.
2.4 Meteorological Characteristics

A favorable CAPE/shear environment is vital to convective persistence (e.g., Weisman 1993). Proximity soundings from GRB, DVN, and ILX were examined to gain insight into the amount of CAPE and shear in the environment upstream of Lake Michigan as the MCSs crossed the lake. The MUCAPE/0–6-km shear\(^\text{12}\) phase space associated with the 54 MCSs that had an upstream proximity sounding reveals that MCSs persisted and dissipated in a wide variety of upstream MUCAPE and 0–6-shear environments (Fig. 2.7a). There was only a small bias to higher MUCAPE and 0–6-km shear for the MCSs that persisted over those that dissipated. MCSs persisted (dissipated) with an average MUCAPE of 2882 J kg\(^{-1}\) (2545 J kg\(^{-1}\)) and 0–6-km shear of 16.3 m s\(^{-1}\) (14.3 m s\(^{-1}\)). In general, the phase space shows no distinguishable relationship between MCSs that persisted and dissipated upon crossing Lake Michigan. Therefore, there are likely other meteorological parameters besides the amount of upstream MUCAPE and 0–6-km shear that determine whether an MCS will persist upon crossing Lake Michigan. This conclusion is perhaps not surprising given that the MCSs that dissipated over Lake Michigan still required favorable conditions to last long enough to be considered in the climatology.

Proximity soundings from DTX were examined to identify the amount of MUCAPE and 0–6-km shear in the environment downstream of Lake Michigan as the MCSs crossed the lake\(^\text{13}\) (Fig. 2.7b). The MUCAPE/0–6-km shear phase space is

\(^\text{12}\) Other shear depths were also tested and did not produce different results.
\(^\text{13}\) Note that although DTX is on the opposite side of Lower Michigan from Lake Michigan, it is the closest actual downstream sounding station to the lake.
noticeably different than for the upstream proximity soundings (Fig. 2.7a). There is a
general differentiation between persisting and dissipating MCSs based on the amount of
MUCAPE in the DTX proximity soundings, with persisting MCSs having generally more
MUCAPE (Fig. 2.7b). Additionally, there were a few MCSs that persisted in a low
MUCAPE, high 0–6-km shear environment. Similar to the upstream environment, the
amount of 0–6-km shear in the downstream environment does not appear to differentiate
between MCSs that persisted and dissipated (Figs. 2.7a,b). On average, MCSs persisted
(dissipated) with an average MUCAPE of 1323 J kg\(^{-1}\) (778 J kg\(^{-1}\)) and 0–6-km shear of
15.3 m s\(^{-1}\) (14.1 m s\(^{-1}\)). These differences in CAPE are statistically significant to the
95th percentile and are very similar to what Letkewicz and Parker (2010) found with
Appalachian Mountain-crossing MCSs, where these MCSs also tended to persist with
increased amounts of downstream CAPE.

The LLJ has been found to be an important feature for MCS persistence in the
central United States (e.g., Johns and Hirt 1987; Laing and Fritsch 1997; Coniglio et al.
2004; Metz and Bosart 2010). Box and whisker plots of the maximum magnitude of the
850-hPa wind incident upon the 110 MCSs at the time of lake crossing reveals that in
general, MCSs tend to persist with a more intense LLJ (Fig. 2.8). In fact, there is very
little overlap between the inter-quartile ranges associated with persisting and dissipating
MCSs. The mean maximum 850-hPa wind speed associated with persisting (dissipating)
Lake Michigan-crossing MCSs is 17.2 m s\(^{-1}\) (13.4 m s\(^{-1}\)). These differences in the
maximum strength of the LLJ are statistically significant to the 99.9th percentile.

The amount of DCAPE in the environment upstream from Lake Michigan was
also calculated from proximity soundings at GRB, DVN, and ILX (Fig. 2.9). The amount
of DCAPE can be a good proxy for the strength of convectively induced downdrafts, the resulting strength of convective cold pools, and MCSs. Persisting Lake Michigan-crossing MCSs in this climatology were found to generally be associated with greater amounts of DCAPE than those that dissipated. While the average amount of DCAPE associated with persisting and dissipating MCSs only varied by 114 J kg\(^{-1}\) (989 J kg\(^{-1}\) for persisting MCSs versus 875 J kg\(^{-1}\) for dissipating MCSs), the inter-quartile ranges of the two box and whisker plots do have some separation (Fig. 2.9). This separation resulted in a statistically significant difference in upstream environmental DCAPE at the 90th percentile.

2.5 Lake-Related Characteristics

Conventional wisdom suggests that as an MCS approaches a region with a decreased surface temperature, convection may weaken as the amount of surface-based CAPE is typically reduced (e.g., Coniglio et al. 2004). However, box and whisker plots of the Lake Michigan water temperature at the time of MCS crossing shows no differentiation between MCSs that persisted and dissipated (Fig. 2.10). In fact, both box and whisker plots look nearly identical and the inter-quartile ranges are quite “long”, a result of the seasonally varying nature of the Lake Michigan water temperature. Thus, the Lake Michigan water temperature is not a good predictor of MCS persistence/dissipation.

Since previous research has shown the depth of the Lake Michigan cold dome to be shallow (e.g., Wylie and Young 1979), an investigation into the thermal structure of this layer is warranted. Unfortunately, very little observational data is available aside
from that gathered by two central Lake Michigan buoys. This buoy data is enough to measure the strength of the near-surface inversion associated with the MCSs in the climatology (temperature at 5 m minus lake water temperature; Fig. 2.11a). The mean inversion strength for MCSs that persisted is 0.85°C more intense than for those that dissipated (2.33°C for persisting MCSs versus 1.48°C for dissipating MCSs). These strong near-surface inversion values suggest that the unfavorable Lake Michigan-altered air is never able to be ingested into the MCS updrafts, especially in the case of persisting MCSs. The persisting MCSs likely ride up and over a shallow, stable near-surface inversion (these inversions will be discussed in more detail in chapters 4 and 5). This “speed bump effect” explains why the Lake Michigan surface temperature is not a good predictor of MCS persistence and dissipation. The weaker inversions associated with dissipating MCSs may allow some of the unfavorable, near-lake air to be ingested into the MCS updrafts. The inter-quartile ranges between persisting and dissipating MCSs do have some overlap (Fig. 2.11a) because many of the persisting MCSs that occurred during the late warm season were associated with a weaker inversion since the temperature of Lake Michigan was much warmer during this time (Fig. 2.4). Still, the inversion strength differences between the MCSs that persisted and dissipated are statistically significant to the 95th percentile. Graham et al. (2004) also found an increased tendency for MCS persistence when the strength of this inversion was at least 2.5°C.

Since the inversion strength is highly dependent on the warm season month, an examination of the early warm-season (April–June) inversion strength (Fig. 2.11b) and the late warm season (July–September) inversion strength (Fig. 2.11c) is warranted. The
early warm-season box and whisker plots show an even greater separation between persisting and dissipating MCSs than the overall climatology, with the top whisker associated with persisting MCSs extending to an inversion strength of almost 6.5°C (Fig. 2.11b). The mean inversion strength for persisting (dissipating) early warm-season MCSs is 3.80°C (2.56°C), a difference of 1.24°C. These differences are statistically significant to the 99th percentile. In the late warm season, there is also some separation between the persisting and dissipating inter-quartile ranges (Fig. 2.11c). However, the persisting MCSs only have a mean inversion strength that is 0.64°C stronger than dissipating MCSs (1.25°C for persisting MCSs versus 0.61°C for dissipating MCSs), less than the overall climatology. Additionally, note how much weaker the inversion strength is during the late warm season as the Lake Michigan water temperature warms closer to the ambient land temperature. There can even be cases where the temperature decreases with height over the lowest 5 m (negative inversion strengths). Still, there is a statistically significant difference at the 90th percentile between the inversion strength for persisting and dissipating late warm season MCSs. These results illustrate that although MCSs throughout the warm season have a tendency to persist when the strength of the Lake Michigan near surface inversion is stronger, this relationship is more robust during the early warm season.

Another parameter that can be examined in association with persisting and dissipating Lake Michigan-crossing MCSs is the strength of the horizontal land–lake temperature difference just before each MCS intersected Lake Michigan (Fig. 2.12). Recall that the stronger this temperature difference, the greater the potential for a lake breeze, assuming that synoptic-scale flow is not too strong (e.g., Laird et al. 2001). An
examination of box and whisker plots of the horizontal temperature difference shows no differentiation between MCS cases that persisted and dissipated. The simple average for persisting (dissipating) MCSs is 7.52°C (7.11°C). The results presented herein compare the 2-m land temperature to the Lake Michigan water temperature. If instead the 2-m land temperature was compared to the 5-m buoy air temperature over Lake Michigan, the results do not change (not shown).

2.6 Forcing and Organization

While MCSs are inherently mesoscale phenomena, many previous authors have illustrated the importance of the synoptic-scale on convective development and evolution (e.g., Doswell 1987). Thus, the 850-hPa Q-vector forcing for ascent was calculated in association with each MCS at the time of lake crossing (see section 2.2 for details). Q-vector convergence (positive values), coincident with forcing for ascent, was associated with 102 of the 110 MCSs (Fig. 2.13). This result is not surprising because the synoptic-scale environment needed to be favorable enough for the dissipating MCSs to last long enough to intersect Lake Michigan. In general, there is no obvious pattern differentiating MCSs that persist and dissipated in relation to the magnitude of the 850-hPa Q-vector convergence. Twenty MCSs were associated with Q-vector convergence values of at least $1 \times 10^{-12}$ Pa m$^{-2}$ s$^{-1}$. If these twenty MCSs are considered to be associated with strong forcing and are removed from the climatology, there is very little overall effect on the rest of the climatology, other than to remove cases (i.e., the previous climatological results presented in this chapter do not significantly change). These cases would also be removed at a nearly even rate (percentage wise) from the persisting and dissipating
climatologies (9 out of 47 persisting MCSs for 19.1% and 11 out of 63 dissipating MCSs for 17.5%). Therefore, while Q-vector convergence appears to be necessary for both persisting and dissipating MCSs to occur, the strength of this Q-vector synoptic-scale forcing does not appear to play a role in whether MCSs persist or dissipate upon crossing Lake Michigan. Thus, no attempt will be made to show climatological plots with the most strongly forced cases removed, since doing so would result in very little change to the results already presented. The same conclusion is reached if 700-hPa or 850–700-hPa layer-averaged Q-vector convergence is considered (not shown).

In the 2002–2007 climatology of Lake Michigan-crossing MCSs, 43% of all MCSs persisted upon crossing the lake (Table 2.1). This persistence percentage is highly dependent upon the organizational structure of the MCSs (Fig. 2.14). For example, only 33% of nonlinear (e.g., Gallus et al. 2008), relatively disorganized MCSs persisted upon crossing Lake Michigan. This percentage increases to 46% for linear MCSs and 69% for bow echoes, although there are only 13 total bow echo cases in the climatology (Fig. 2.14). Weisman (1993) found that increasing the amount of CAPE generally increases the organizational structure. Thus, the bow echo cases are generally associated with very favorable synoptic- and mesoscale environments (these environments will be discussed further in chapter 4). Additionally, as MCSs bow out, the associated cold pool strength increases (e.g., RKW; Weisman and Rotunno 2004). This increased persistence rate for bow echoes with strong cold pools (Fig. 2.14), agrees with the box and whisker plots of DCAPE, which revealed that Lake Michigan-crossing MCSs persisted in association with generally higher amounts of upstream DCAPE (Fig. 2.9), and thus, potentially stronger cold pools. MCSs likely persist at a higher rate with increasing structural organization.
for a couple of reasons. First, as just mentioned, the associated synoptic-scale environment that these MCSs “feed off” is very favorable. Additionally, bow echoes tend to move faster than less well-organized systems (e.g., Johns and Hirt 1987, Weisman 1992). Therefore, bow echoes spend less time over Lake Michigan, and have a smaller chance of ingesting any unfavorable air into the MCS updrafts.\footnote{The NOWrad data is only available in 15-min increments. Thus, it was difficult to calculate the exact amount of time each MCS resided over Lake Michigan. However, bow echo MCSs generally tended to reside over the lake for \sim15–30 minutes less than other climatological MCSs.}
2.7 Chapter 2 Figures

FIG 2.1: Daily average time series (in UTC), color coded to match 4° × 2° latitude-longitude boxes of the average square kilometers of ≥45-dBZ convection. The average was constituted by using warm-season (April–September) NOWrad composite data from 2002 to 2007 for (a) three boxes, one to the west, over, and east of Lake Michigan and (b) three boxes at the same longitudes as in (a) but 4° latitude farther south. Three different regimes are indicated in both panels. Source: NOWrad composite data.
FIG. 2.2: As in Fig. 2.1a, except only for days on which the 110 MCSs in the climatology were present. Source: NOWrad composite data.
FIG. 2.3: Yearly distributions of MCSs that persisted (green) and dissipated (red) upon crossing Lake Michigan during the warm seasons of 2002–2007. The purple numbers show the total number of storms in the climatology that occurred during each year. Source: NOWrad composite data.
FIG. 2.4: Monthly distributions of MCSs that persisted (green) and dissipated (red) upon crossing Lake Michigan during the warm seasons of 2002–2007. The average monthly lake-surface temperature (in °C) at buoy 45007 in the southern portion of Lake Michigan is noted for each month. The purple numbers show the total number of storms in the climatology that occurred during each month. Source: NOWrad composite data.
FIG. 2.5: Hourly distributions of MCSs that persisted (green) and dissipated (red) upon crossing Lake Michigan during the warm seasons of 2002–2007. The purple numbers show the total number of storms in the climatology that occurred during each hour range. Source: NOWrad composite data.
FIG. 2.6: Number of hours after formation that persisting (green) and dissipating (red) MCSs crossed Lake Michigan during the warm seasons of 2002–2007. The purple numbers show the total number of storms in the climatology that occurred during each hour range. Source: NOWrad composite data.
FIG 2.7: (a) MUCAPE (in J kg$^{-1}$)/0–6-km shear (in m s$^{-1}$) phase space for MCSs that persisted (green diamonds) and dissipated (red squares) from proximity soundings at GRB, DVN, and ILX. (b) As in (a) except for proximity soundings from DTX. The insets in the corner of (a) and (b) reveal the locations of the proximity soundings. Green and red boxes in (b) illustrate clusters of persisting and dissipating MCSs on the phase space. Source: NWS rawinsonde archive.
FIG. 2.8: Box and whisker plots of the maximum magnitude of the 850-hPa wind (in m s$^{-1}$) incident on MCSs that crossed Lake Michigan during the warm seasons of 2002–2007. The green box represents persisting MCSs, while the red box represents dissipating MCSs. The colored boxes, blue line, light blue hexagon, and whiskers represent the inter-quartile range, the median, the mean, and the 10th and 90th percentiles, respectively. Source: 32-km NARR.
FIG. 2.9: As in Fig. 2.8a except for DCAPE (in J kg$^{-1}$). Source: NWS rawinsonde archive.
FIG. 2.10: As in Fig. 2.8 except for Lake Michigan water temperature (in °C). Source: The NDBC.
FIG. 2.11: As in Fig. 2.8 except for the strength of the Lake Michigan near surface inversion (in °C) for (a) the entire warm season, (b) April–June, and (c) July–September. Source: The NDBC.
FIG. 2.11: continued
FIG. 2.12: As in Fig. 2.8 except for the strength of the horizontal land–lake temperature difference (in °C) on the western side of Lake Michigan. Source: UAlbany surface archive and the NDBC.
FIG. 2.13: The magnitude of the 850-hPa Q-vector convergence (in \(10^{-12} \text{ Pa m}^{-2} \text{ s}^{-1}\)) associated with MCSs that persisted (green) and dissipated (red) upon crossing Lake Michigan during the warm seasons of 2002–2007. The Q-vector convergence values were calculated by averaging the Q-vector convergence in a 200-km radius over each MCS. Source: 1.0° GFS.
FIG. 2.14: Organizational structure of MCSs that persisted (green) and dissipated (red) upon crossing Lake Michigan during the warm seasons of 2002–2007. The black numbers show the total number of MCSs in the climatology that either persisted or dissipated within each category. The purple number shows the persistence percentage within each category. Source: NOWrad composite data.
2.8 Chapter 2 Tables

TABLE 2.1: Breakdown of the number and percentage of MCSs that persisted and dissipated during the warm seasons of 2002–2007.

<table>
<thead>
<tr>
<th></th>
<th>Persist</th>
<th>Dissipate</th>
</tr>
</thead>
<tbody>
<tr>
<td>Number of MCSs</td>
<td>47</td>
<td>63</td>
</tr>
<tr>
<td>Percentage</td>
<td>43%</td>
<td>57%</td>
</tr>
</tbody>
</table>
3. Synoptic-Scale Composites of Lake Michigan-Crossing MCS Environments

3.1 Purpose

The purpose of this chapter is to examine the synoptic-scale environment associated with persisting and dissipating Lake Michigan-crossing MCSs. Since there are 110 total MCS cases in the climatology presented in chapter 2, synoptic-scale composites will be employed to examine key environmental similarities and differences between persisting and dissipating MCSs. Previous authors have utilized synoptic and mesoscale composites to elucidate common environmental features associated with convective weather events (e.g., Schumacher and Johnson 2009). This approach will be adopted here and the results presented in this chapter will attempt to link to climatological results presented in chapter 2 and serve as a bridge to real and simulated case study results in chapters 4 and 5, respectively.

3.2 Data and Methods

The aforementioned synoptic-scale composites were created utilizing all 110 MCS cases from the Lake Michigan-crossing MCS climatology presented in chapter 2. The 1.0° GFS dataset was used to make these composites since the resolution of the GFS is appropriate to examine the synoptic-scale environment\(^\text{15}\). The GFS is output every 6 h, so the analysis grid that was closest to the time each MCS initially intersected Lake Michigan was chosen for the composites. If the MCS intersected Lake Michigan directly between two analysis times, the earlier time was chosen for consistency. Since each MCS was located in a different position relative to the lake at the analysis time, each GFS

\(^{15}\) Acquired from the UCAR Computational and Information Systems Laboratory archive at http://dss.ucar.edu/datasets/ds083.2/matrix.html
analysis grid was shifted relative to the MCS location in order to create MCS-relative composites. This shift was accomplished by finding the centroid of each MCS at the analysis time and then shifting each grid to the average position of all the analyzed MCSs in each composite. In all composites presented in chapter 3, this average position is located in extreme southeastern Wisconsin.

Originally, all of the persisting MCSs were placed into a single composite of 47 cases, while all of the dissipating MCSs were placed into a single composite of 63 cases. However, since the MCSs are associated with differing synoptic-scale patterns, the resulting composites did not resemble many of the individual MCS cases. Therefore, in order to arrive at synoptically representative composites, each MCS environment was categorized by and composited based on its associated 200-hPa synoptic-scale pattern. Two main patterns resulted. The first pattern is very similar to the Johns (1993) dynamic pattern that is typically associated with bow echo MCSs (Fig. 3.1a). This dynamic pattern for lake-crossing MCSs features an upper-level trough located to the west of the favorable MCS location and a southerly component to the flow in the upper levels. Additionally, a low-pressure system is located just to the west of the MCS location, with MCS development within the warm sector. The LLJ and the upper-level jet streams cross each other at about a 60° angle. Forty-eight of the 110 climatological MCSs are associated with the dynamic pattern, 17 that persisted and 31 that dissipated\textsuperscript{16}.

\textsuperscript{16} Note that the results from the dynamic composites compare two groups that contain differing numbers of MCSs. The dissipating dynamic composite results that appear in section 3.3 were redone by choosing only 17 MCS cases at random, to match the number of persisting dynamic MCSs. This smaller dissipating composite gave very similar results to those presented herein. Thus, the differences that will be shown between the persisting and dissipating dynamic composites are not a function of composite size.
The second pattern is a warm-season pattern, often associated with progressive bow echo MCSs (Fig. 3.1b). This set-up features MCSs moving around the periphery of a mid-tropospheric ridge, with generally zonal or a northerly component to the upper-level flow. There is typically a west–east oriented surface boundary with an upper-level jet stream located immediately to the north of this surface boundary. The LLJ curves anticyclonically into the progressive MCS environment. These MCSs often progress into increasingly warm, unstable air at a small angle relative to the surface boundary (e.g., Johns and Hirt 1987; Metz and Bosart 2010) and can be associated with northwesterly flow aloft (e.g., Johns 1982). Sixty-two of the 110 climatological MCSs were associated with this pattern, 30 that persisted and 32 that dissipated.

Q-vectors and Q-vector convergence were calculated from the GFS analysis grids utilizing the Q-vector and quasi-geostrophic omega equations presented in section 2.2. The MUCAPE in the GFS is calculated by lifting the 30-hPa layer that is the most unstable within the lowest 180-hPa layer of the GFS. The 925-hPa moisture convergence is calculated by taking the 925-hPa horizontal flux divergence of the mixing ratio. The resulting moisture convergence output is in units of s$^{-1}$ as the moisture variable is left in its unitless form (e.g., g g$^{-1}$ or kg kg$^{-1}$).

3.3 Dynamic Synoptic-Scale Composite

The persisting dynamic composite of 17 Lake Michigan-crossing MCSs has a 1006-hPa surface low pressure located to the west-northwest of the MCS location (Fig. 3.2a). The composite MCS is located in inferred surface geostrophic southwesterly flow, and on the western periphery of a 570–573 dam 1000–500-hPa thickness ridge. A
thermal trough is located well to the west of the persisting MCS composite location. The dissipating composite of 31 MCSs has a slightly weaker, broader 1007-hPa low that is positioned to the northwest of the MCS location (Fig. 3.2b). The composite MCS is located under a 567–570 dam 1000–500-hPa thickness ridge in the dissipating composite. Additionally, the thickness gradient along and to the north of the dissipating MCS composite location (Fig. 3.2b) is not as large as along and to the north of the persisting MCS composite location (Fig. 3.2a). In both composites, a region of relatively low pressure extends southwestward from the surface cyclone (Figs. 3.2a,b). This region of lower pressure is collocated with the surface cold front in these cases. The surface cyclone position is very similar to that in the Johns (1993) dynamic schematic (Fig. 3.1a), which is to be expected given that the composites were grouped by synoptic-scale patterns that matched the Johns (1993) upper-level synoptic-scale patterns.

The persisting dynamic composite of Lake Michigan-crossing MCSs is concomitant with a 200-hPa trough located to the west of the MCS location (Fig. 3.3a). Again, this is to be expected given that these composites were created by examining the 200-hPa pattern associated with each MCS and subsetting in a manner that matched the Johns (1993) schematics (Fig. 3.1). The composite persisting MCS is located in the equatorward-entrance region of a 42 m s⁻¹ 200-hPa jet stream (Fig. 3.3a). The equatorward entrance region is a favorable area for synoptic-scale ascent. Additionally, the persisting composite MCS is located near the terminus of a 17 m s⁻¹ LLJ. The dissipating dynamic composite MCS is also located in the equatorward exit region of the 200-hPa jet stream (Fig. 3.3b). However, the magnitude of this 200-hPa jet stream associated with the dissipating composite MCS is only slightly greater than 33 m s⁻¹.
about 10 m s\(^{-1}\) weaker than the 200-hPa jet stream present in the persisting composite (Figs. 3.3a,b). The dissipating dynamic composite MCS is also located near the terminus of a 14 m s\(^{-1}\) LLJ (Fig. 3.3b), which is about 3 m s\(^{-1}\) weaker than in the persisting composite (Fig. 3.3a). This difference in the strength of the LLJ mirrors the climatological results presented in chapter 2 (Fig. 2.8). Note the dynamic patterns associated with both persisting and dissipating composite MCSs look qualitatively very similar, and generally differ only in the magnitudes of the associated LLJ and upper-level jet streams (Figs. 3.3a,b).

The dynamic composite associated with persisting Lake Michigan-crossing MCSs reveals a maximum in Q-vector convergence that exceeds 1.5 \times 10^{-12} \text{ Pa m}^{-2} \text{s}^{-1} to the north of the composite location (Fig. 3.4a). The value of Q-vector convergence directly over the persisting MCS composite location is slightly above 1.0 \times 10^{-12} \text{ Pa m}^{-2} \text{s}^{-1}. Much of this Q-vector forcing for ascent is associated with vorticity advection ahead of an advancing trough that is located over Minnesota. Another contribution to the forcing for ascent and Q-vector convergence results from the 850-hPa warm advection over the MCS composite region. The dynamic composite associated with dissipating Lake Michigan-crossing MCSs is located to the south of a 1.0 \times 10^{-12} \text{ Pa m}^{-2} \text{s}^{-1} maximum in Q-vector convergence (Fig. 3.4b). The value of Q-vector convergence directly over the dissipating composite MCS is only slightly less than the aforementioned 1.0 \times 10^{-12} \text{ Pa m}^{-2} \text{s}^{-1} (Fig. 3.4b). The 850-hPa trough in the dissipating composite is somewhat broader than in the persisting composite (Figs. 3.4a,b), and the 850-hPa warm advection near the MCS composite location is slightly less intense (not shown). Even though the Q-vector convergence maximum is larger in the persisting composite, this maximum is displaced
to the north of the MCS composite location (Fig. 3.4a), and the amount of forcing for ascent, associated with Q-vector convergence, is only slightly larger over the persisting composite MCS than the dissipating composite MCS (Figs. 3.4a,b). This result agrees with the climatological results from chapter 2 where the amount of Q-vector convergence (averaged in a 200-km radius around each MCS) did not significantly differentiate between MCSs that persisted and dissipated (Fig. 2.13). Additionally, many of the persisting and dissipating cases that went into the dynamic composites were associated with climatological Q-vector convergence values on the higher end of the climatological distributions (Fig. 2.13). These larger values of Q-vector convergence are to be expected since the dynamic pattern generally features well-developed troughs approaching from the west and concomitant cyclonic vorticity advection in addition to robust warm advection.

Recall that in a convective environment, the LLJ can transport large amounts of moisture and result in enhanced convergence and heavy rainfall (e.g., Uccelini and Johnson 1979; Schumacher and Johnson 2009). The persisting dynamic composite MCS, which was situated near the terminus of a 17 m s$^{-1}$ 850-hPa LLJ, (Fig. 3.3a) is also located near the terminus of the 925-hPa LLJ (Fig. 3.5a). This 925-hPa LLJ contains winds in excess of 15 m s$^{-1}$ and terminates along a mixing ratio gradient where values range from 13 g kg$^{-1}$ over the MCS composite location to 9 g kg$^{-1}$ immediately to the north of Lake Michigan. The combination of the strong 925-hPa LLJ and the high mixing ratio air leads to large values of moisture convergence along and to the north of the MCS composite location that exceed $-4 \times 10^{-7}$ s$^{-1}$ (equivalent to $-1.44$ g kg$^{-1}$ h$^{-1}$). Heavy rain producing MCSs have been identified to favor regions of strong 925-hPa
moisture convergence of around $-1.9 \times 10^{-7} \text{s}^{-1}$ ($-0.7 \text{ g kg}^{-1} \text{h}^{-1}$; e.g., Moore et al. 2003).

On the other hand, the dissipating dynamic composite MCS is associated with a 925-hPa LLJ that is only slightly stronger than 12.5 m s$^{-1}$ (Fig. 3.5b). The 925-hPa mixing ratio values maximize over the dissipating MCS composite location at 12 g kg$^{-1}$ and the mixing ratio gradient to the north of this location is not as strong as in the persisting dynamic composite (Figs. 3.5a,b). The weaker 925-LLJ and smaller mixing ratio values in the dynamic composite result in 925-hPa moisture convergence values that are only around $-2.5 \times 10^{-7} \text{s}^{-1}$ (equivalent to $-0.96 \text{ g kg}^{-1} \text{h}^{-1}$; Fig. 3.5a), much less than the moisture convergence values near the persisting composite MCS (Fig. 3.5a).

The persisting dynamic composite associated with Lake Michigan-crossing MCSs reveals that the MCS is located in a region of 1000–1250 J kg$^{-1}$ of MUCAPE and 15 m s$^{-1}$ of 0–6 km shear (Fig. 3.6a). The dissipating dynamic composite MCS is situated in a region of 750–1000 J kg$^{-1}$ of MUCAPE and 15 m s$^{-1}$ of 0–6-km shear (Fig. 3.6b). The amount of 0–6-km shear in both composites is very similar upstream and downstream of the MCS locations. There is 2.5–5 m s$^{-1}$ more 0–6-km shear to the north of the persisting composite MCS location along the U.S./Canadian border when compared to the same region in the dissipating composite (Figs. 3.6a,b). This increased 0–6-km shear is in the same location where the persisting composite is associated with a 200-hPa jet stream that is almost 10 m s$^{-1}$ more intense than in the dissipating composite (Figs. 3.3a,b). These composites confirm climatological results that revealed similar amounts of 0–6-km shear both upstream and downstream of Lake Michigan along the paths of persisting and

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17 The values of 925-hPa moisture convergence in Moore et al. (2003) were calculated by interpolating rawinsonde data to a 190-km grid. Thus, their moisture convergence values associated with a heavy rain-producing MCS are somewhat less than those found in this section.
dissipating MCSs (Figs. 2.7a,b). The amount of MUCAPE upstream of both MCS composite locations is approximately the same, maximizing between 1000 and 1250 J kg$^{-1}$ (Figs. 3.6a,b). However, the amount of MUCAPE along and downstream of the persisting composite MCS location is 500–750 J kg$^{-1}$ larger than in the dissipating composite. This result also mirrors the previous climatological investigation from proximity soundings that showed a similar amount of MUCAPE upstream of Lake Michigan with persisting and dissipating MCSs, and a larger amount of MUCAPE downstream of Lake Michigan with persisting MCS cases (Figs. 2.7a,b).

In general, the persisting and dissipating dynamic composite patterns look qualitatively similar. However, there are differences between the two composites as the persisting composite MCS generally is associated with a more dynamically favorable environment than the dissipating composite MCS. The persisting composite MCS is located near a more robust 200-hPa jet stream, a stronger LLJ, a slightly stronger surface low, and slightly more 850-hPa Q-vector convergence, while the amount of 0–6-km shear is similar between the two composites. Furthermore, the thermodynamic environment associated with the persisting MCS composite is more favorable for convective maintenance than the thermodynamic environment associated with the dissipating composite MCS, with almost twice as much moisture convergence and twice as much MUCAPE along and downstream of the MCS composite location.

3.4 Progressive Synoptic-Scale Composites

Persisting and dissipating MCSs that comprised the progressive pattern composites are associated with vastly different synoptic-scale patterns than the dynamic
pattern composites. The persisting progressive composite of 30 MCSs is located to the east-northeast of a broad 1011-hPa lee cyclone (Fig. 3.7a). The 1011-hPa SLP contour extends eastward from the cyclone center in conjunction with a region where a warm front (or stationary front) is typically located with these progressive persisting MCSs. Geostrophic surface southwesterly flow impinges on the persisting MCS composite location. This SLP pattern and inferred frontal position agree nicely with the Johns (1993) schematic. Additionally, the persisting composite MCS is located near the downstream side of a 573 dam 1000–500-hPa thickness ridge. This is a typical location for Great Lakes progressive MCSs relative to a 1000–500-hPa thickness ridge (e.g., Metz and Bosart 2010). The dissipating progressive composite of 32 MCSs is located to the east of a slightly stronger 1009-hPa lee cyclone when compared to the persisting composite (Figs. 3.7a,b). However, in general, the SLP pattern near the dissipating composite MCS is very similar to that near the persisting composite MCS, with geostrophic southwesterly flow into the MCS region (Figs. 3.7a,b). The thickness gradient to the northwest of the dissipating composite MCS is not as strong as the thickness gradient associated with the persisting MCS. Additionally, the dissipating composite MCS is on the downstream side of a 570 dam 1000–500-hPa thickness ridge (Fig. 3.7b).

The persisting progressive composite MCS is associated with relatively zonal flow and a flat 200-hPa ridge (Fig. 3.8a). The composite MCS is located on the downstream portion of this 200-hPa ridge in slightly north-of-west flow. Immediately to the north of the persisting composite location is a 42 m s$^{-1}$ 200-hPa jet stream, and the

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18 The synoptic-scale SLP pattern may be slightly altered near the MCS composite location due to the surface effects of convection in the GFS.
composite MCS is located on the anticyclonic shear side of this jet stream near the
equatorward-entrance region (Fig. 3.8a). A 15 m s\(^{-1}\) 850-hPa LLJ curves
anticyclonically and terminates near the persisting composite MCS location. The
dissipating progressive composite MCS is also located on the anticyclonic shear side of a
200-hPa jet stream, near the equatorward jet-entrance region (Fig. 3.8b). However, this
200-hPa jet stream maximizes at about 36 m s\(^{-1}\), nearly 6 m s\(^{-1}\) weaker than the 200-hPa
jet stream concomitant with the persisting progressive composite MCS (Figs. 3.8a,b). A
12 m s\(^{-1}\) LLJ curves anticyclonically and terminates near the dissipating MCS composite
location (Fig. 3.8b). In general, the upper-level pattern associated with both the
persisting and dissipating progressive MCS composites look qualitatively similar to each
other and the Johns (1993) schematic with the composite MCSs located immediately
downstream of an upper-level ridge axis (Fig. 3.1b).

The 850-hPa Q-vector convergence directly over the persisting progressive
composite MCS location is between 0.5 and 0.75\(\times\)10\(^{-12}\) Pa m\(^{-2}\) s\(^{-1}\) (Fig. 3.9a). Much of
this Q-vector convergence over the MCS is driven by low-level warm-advection as
implied 850-hPa geostrophic winds in the MCS composite region blow across the
temperature gradient from higher to lower temperatures. The dissipating composite
MCS is situated in a region of Q-vector convergence of around 0.75\(\times\)10\(^{-12}\) Pa m\(^{-2}\) s\(^{-1}\),
slightly higher than the Q-vector convergence magnitude found over the persisting
composite MCS (Fig. 3.9b). Again, much of this Q-vector convergence is driven by 850-
hPa warm advection, similar to the persisting MCS composite discussed previously. The
maximum in Q-vector convergence to the north of the dissipating MCS composite
location is also slightly larger than in the persisting MCS composite (Figs. 3.9a,b). Still
though, the amount of Q-vector convergence directly over both the persisting and dissipating composite MCSs is quite similar in magnitude, in agreement with the climatological results shown in chapter 2 (Fig. 2.13). Furthermore, the total 850-hPa Q-vector forcing for ascent in both the persisting and dissipating progressive composites is smaller than the forcing for ascent in the dynamic composites because the progressive composites do not contain a well-developed trough approaching from the west (Figs. 3.4a,b).

The persisting progressive composite MCS is positioned in a region of 925-hPa mixing ratios that range from 13 to 14 g kg\(^{-1}\) (Fig. 3.10a). Additionally, the 925-hPa winds impinging on this composite location are between 10 and 12.5 m s\(^{-1}\). The combination of these mixing ratio and wind values result in composite 925-hPa moisture convergence values of around \(-2.5\times10^{-7}\) s\(^{-1}\) (equivalent to \(-0.90\) g kg\(^{-1}\) h\(^{-1}\)) near the persisting MCS. The dissipating composite MCS is located in a region of 925-hPa mixing ratio values near 11 g kg\(^{-1}\) (Fig. 3.10b). Furthermore, the mixing ratio gradient along and to the north of the dissipating MCS composite location is not as strong as to the north of the persisting MCS composite location (Figs. 3.10a,b). The 925-hPa wind values near the dissipating MCS composite are around 10 m s\(^{-1}\), slightly weaker than those near the persisting MCS composite (Fig. 3.10b). These mixing ratio and wind magnitudes result in a 925-hPa moisture convergence maximum that is located to the west of the dissipating MCS composite location. The composite MCS is coincident with moisture convergence values of only \(-0.5\) to \(-1.0\times10^{-7}\) s\(^{-1}\) (equivalent to \(-0.18--0.36\) g kg\(^{-1}\) h\(^{-1}\)).
The persisting progressive composite associated with Lake Michigan-crossing MCSs reveals that the MCS is located in a region of 1000–1250 J kg⁻¹ of MUCAPE and 15–17.5 m s⁻¹ of 0–6 km shear (Fig. 3.11a). The dissipating progressive composite MCS is located in a region of 500–750 J kg⁻¹ of MUCAPE and 15 m s⁻¹ of 0–6-km shear (Fig. 3.11b). The amount of 0–6-km shear near the persisting MCS composite location is slightly higher but not significantly different from the amount of 0–6-km shear near the dissipating MCS composite. Additionally, the amount of 0–6-km shear downstream of Lake Michigan is nearly identical in both progressive composites (Figs. 3.11a,b). The magnitude of 0–6-km shear to the north of the persisting composite MCS location is 2.5–5 m s⁻¹ greater than in the dissipating composite. Recall that this is the same location where the persisting composite has a 6 m s⁻¹ stronger 200-hPa jet stream than the dissipating composite (Figs. 3.8a,b). The persisting composite has about 250 J kg⁻¹ more upstream MUCAPE than the dissipating composite (Figs. 3.11a,b). However, the largest MUCAPE differences between the two composites occur along and downstream from the MCS locations, where the persisting MCS composite has 500–750 J kg⁻¹ more MUCAPE than the dissipating MCS composite. Recall that proximity soundings from DTX also revealed that MCSs tended to persist in conjunction with higher amounts of downstream MUCAPE (Fig. 2.7b). These downstream MUCAPE results agree with another regional MCS study by Letkewicz and Parker (2010), where the authors found that Appalachian Mountain-crossing MCSs had a higher frequency of persistence as the amount of downstream CAPE on the eastern side of the Appalachians increased.

In general, the persisting and dissipating progressive composite patterns look qualitatively similar. However, there are differences between the two composites as the
persisting composite MCS generally has a slightly more dynamically favorable environment than the dissipating composite MCS. The persisting composite MCS is associated with a more robust 200-hPa jet stream and a stronger LLJ, while the amounts of Q-vector convergence and 0–6-km shear are similar between the two composites. These dynamical differences between the progressive composites are not as stark as the dynamical differences between the dynamic composites. The thermodynamic environment associated with the persisting composite MCS is much more favorable for convective maintenance than the dissipating composite MCS, with twice as much moisture convergence and twice as much MUCAPE along and downstream of the MCS composite locations. These thermodynamic differences are very similar to those observed in conjunction with the dynamic MCS composites.
FIG. 3.1: Idealized schematics illustrating favorable situations for the development of (a) dynamic bow echo MCSs and (b) progressive bow echo MCSs. The PJ, LJ, and SJ refer to the polar, low-level, and subtropical jets, respectively. The shading in (a) represents the most favorable area for MCS development, while the line B–M–E in (b) refers to the typical track of a MCS [from Johns (1993)]
FIG. 3.2: Synoptic-scale composites of SLP (black lines every 1 hPa) and 1000–500-hPa thickness (red dashed lines every 3 dam) for dynamic MCSs that (a) persisted and (b) dissipated upon crossing Lake Michigan. The green star in each panel reveals the location of the MCS-relative compositing point. Source: 1.0° GFS analyses.
FIG. 3.3: Synoptic-scale composites of 200-hPa wind speed (warm colors shaded according to color bar in m s$^{-1}$), 200-hPa height (black lines every 12 dam), 850-hPa wind speed (cool colors shaded according to color bar in m s$^{-1}$), and 850-hPa winds (barbs greater than 10 m s$^{-1}$, where each long barb equals 5 m s$^{-1}$) for dynamic MCSs that (a) persisted and (b) dissipated upon crossing Lake Michigan. The green star in each panel reveals the location of the MCS-relative compositing point. Source: 1.0° GFS analyses.
FIG. 3.4: Synoptic-scale composites of 850-hPa Q-vectors (arrows greater than $5 \times 10^{-7}$ Pa m$^{-1}$ s$^{-1}$), Q-vector convergence (shaded according to color bar in $\times 10^{-12}$ Pa m$^{-2}$ s$^{-1}$), temperature (green dashed lines every 3°C), and height (black lines every 3 dam) for dynamic MCSs that (a) persisted and (b) dissipated upon crossing Lake Michigan. Reference vectors are located in the lower-right corner of each panel. The green star in each panel reveals the location of the MCS-relative compositing point. Source: 1.0° GFS analyses.
FIG. 3.5: Synoptic-scale composites of 925-hPa moisture convergence (shaded according to color bar in $\times 10^{-7} \text{ s}^{-1}$), 925-hPa mixing ratio (red lines every 1 g kg$^{-1}$), and 925-hPa winds (barbs in m s$^{-1}$, where each long barb equals 5 m s$^{-1}$) for dynamic MCSs that (a) persisted and (b) dissipated upon crossing Lake Michigan. The black star in each panel reveals the location of the MCS-relative compositing point. Source: 1.0° GFS analyses.
FIG. 3.6: Synoptic-scale composites of MUCAPE (shaded according to color bar in J kg$^{-1}$) and 0–6-km shear (barbs in m s$^{-1}$, where one long bar equals 5 m s$^{-1}$) for dynamic MCSs that (a) persisted and (b) dissipated upon crossing Lake Michigan. The green star in each panel reveals the location of the MCS-relative compositing point. Source: 1.0° GFS analyses.
FIG. 3.7: As in Fig. 3.2 except for progressive MCSs. Source 1.0° GFS analysis.
FIG. 3.8: As in Fig. 3.3 except for progressive MCSs. Source 1.0° GFS analysis.
FIG. 3.9: As in Fig. 3.4 except for progressive MCSs. Source 1.0° GFS analysis.
FIG. 3.10: As in Fig. 3.5 except for progressive MCSs. Source 1.0° GFS analysis.
FIG. 3.11: As in Fig. 3.6 except for progressive MCSs. Source 1.0° GFS analysis.
4. Case Study Analyses of Lake Michigan-Crossing MCSs

4.1 Purpose

The purpose of this chapter is to utilize individual case study examples to elucidate important environmental and lake characteristics associated with Lake Michigan-crossing MCSs. While the two previous chapters have revealed many common features concomitant with these MCSs, climatological and compositing analyses cannot reveal as much specific detail as case studies. These case studies will draw from various times of the year and different synoptic patterns in an attempt to show the spectrum of environmental and lake conditions associated with Lake Michigan-crossing MCSs. The results presented in this chapter will serve as motivation for the simulated MCS case studies in chapter 5.

4.2 Data and Methods

The case studies in this chapter were chosen to be representative of the Lake Michigan-crossing MCS climatology and composites, in order to delve into detail about different environmental facets unveiled in chapters 2 and 3. Three different comparisons of persisting and dissipating MCSs (six total MCS case studies) will be presented in this chapter. The first comparison features the 7–8 July 2008\textsuperscript{19} MCS that persisted and the 4 June 2005 MCS that dissipated upon crossing Lake Michigan. These two cases were chosen for comparison because they both occurred at around the same time of the year.

\textsuperscript{19} Although the 7 June 2008 MCS case is not in the MCS climatology presented in chapter 2, it does follow all of the climatological criteria and would be included if the climatology was extended another year. This case was utilized because it occurred in real time as the author was beginning to study Lake Michigan-crossing MCSs and initially piqued his interest.
and both crossed Lake Michigan at around the same time of the day. Additionally, both of these cases were concomitant with the dynamic synoptic-scale pattern (Johns 1993), and were relatively small MCSs with a leading convective line at the time of lake crossing.

The second pair of Lake Michigan-crossing MCS are the 18 June 2010\textsuperscript{20} MCS that persisted and the 24 June 2003 MCS that dissipated upon crossing Lake Michigan. These two MCSs were chosen for comparison because both crossed Lake Michigan at about the same time of year and both were categorized according to the dynamic synoptic-scale pattern (Johns 1993). Additionally, both MCSs were well-organized bow-echo MCSs. Recall that bow echo MCSs persisted at a much higher rate than other MCSs in the climatology (Fig. 2.14).

The final set of Lake Michigan-crossing MCSs considered in this chapter are the 22 August 2007 MCS that persisted and the 6 July 2003 MCS that dissipated. These two cases were chosen for further analysis because both crossed Lake Michigan during the overnight/early morning hours and likely ingested air located above a nocturnal surface inversion (e.g., Colman 1990a,b). Additionally, both MCSs occurred during the later half of the warm season when Lake Michigan is typically warmer (Fig. 2.4), and both MCS cases fell under the progressive MCS compositing pattern (Johns 1993) that was identified in chapter 3.

\textsuperscript{20} Again, this case is not in the climatology presented in chapter 2 but would be included if the climatology was extended through 2010.
All six case studies were investigated by utilizing the 20-km RUC analysis grids (e.g., Benjamin et al. 2002, 2004)\textsuperscript{21}. The RUC dataset has 50 vertical levels and employs a hybrid isentropic-sigma vertical coordinate. The domain of the RUC covers the continental U.S. and extends a small distance beyond in all directions. These characteristics make the RUC an excellent dataset to analyze convective weather events in high resolution over the U.S. (e.g., Markowski et al. 2003; Thompson et al. 2003; Metz and Bosart 2010). The MUCAPE in the RUC was calculated by lifting a parcel from the level of maximum buoyancy within 300 hPa of the surface. Before finding the most buoyant level, an averaging of potential temperature and water vapor mixing ratio was performed over the lowest seven RUC native levels (~40 hPa).

The composite radar imagery utilized in this chapter was obtained from the NOWrad archive at UCAR\textsuperscript{22} and the Next Generation Weather Radar (NEXRAD) level-3 archive at the University at Albany. The MCS case studies from 2007 and earlier employ the NOWrad composite dataset and those after utilize the University at Albany archive. UCAR stopped receiving the NOWrad dataset after September 2007 necessitating the utilization of the NEXRAD level-3 data for the 2008 and 2010 case studies. The infrared (IR) satellite imagery in this chapter was acquired from the UCAR Mesoscale and Microscale Meteorology (MMM) Division archive\textsuperscript{23} and the Bow Echo and Mesoscale Convective Vortex Experiment (BAMEX) online catalog\textsuperscript{24}, while the water-vapor

\textsuperscript{22} Acquired from: http://www.mmm.ucar.edu/imagearchive/WSI/
\textsuperscript{23} Acquired from: http://www.mmm.ucar.edu/imagearchive/
\textsuperscript{24} Acquired from: http://catalog.eol.ucar.edu/bamex/
satellite imagery was obtained the NCAR Research Applications Laboratory (RAP)\textsuperscript{25}. Severe reports were plotted with the Severe Plot v.3.0 program\textsuperscript{26}. NWS rawinsonde and surface data were acquired from the University at Albany archive, while buoy data was downloaded from the NDBC\textsuperscript{27}. Finally, AMDAR sounding data was acquired from the Earth System Research Laboratory Global Systems Division\textsuperscript{28}. This AMDAR data was collected by aircraft upon take off and landing.

In addition to this litany of datasets, the 1.0° GFS (instead of the 20-km RUC) was utilized to create Q-vector plots. Given that the Q-vector is a representation of the forcing for synoptic-scale vertical motion, the GFS is an appropriate dataset to calculate the Q-vector and Q-vector convergence. RUC plots of Q-vectors and Q-vector convergence were too unintelligible to be employed in this thesis. Q-vectors and Q-vector convergence were calculated from the GFS analysis grids utilizing the Q-vector and quasi-geostrophic omega equations presented in section 2.2.

4.3 Persisting MCS of 7–8 June 2008

The 7–8 June 2008 persisting MCS (hereafter MCS-P1), initially formed along the Minnesota/Iowa border around 1200 UTC 7 June (hereafter 12Z/07), progressed eastward and crossed Lake Michigan beginning at around 21Z/07, before finally weakening after 03Z/08 over lower Michigan (Fig. 4.1). In the wake of MCS-P1, many high wind, hail, and tornado reports were noted, and at least nine severe wind reports occurred between 00Z/08 and 05Z/08 in lower Michigan subsequent to MCS-P1 crossing

\textsuperscript{25} Acquired from: http://www.ral.ucar.edu/weather/
\textsuperscript{26} Acquired from: http://www.spc.noaa.gov/climo/online/sp3/plot.php
\textsuperscript{27} Acquired from: http://www.ndbc.noaa.gov
\textsuperscript{28} Acquired from: http://amdar.noaa.gov/demo_java/
Lake Michigan. In addition, upstream of MCS-P1, more convection fired contributing to the large number of severe reports in Iowa, Minnesota and Wisconsin between 12Z/07 and 06Z/08 (Fig. 4.1).

On 1503Z/07, MCS-P1 was positioned on the Minnesota/Wisconsin border (Fig. 4.2a). About 2 h later, MCS-P1 had progressed to the east-southeast (Fig. 4.2b), and by 1900Z/07, had increased in organization and was located in south-central Wisconsin (Fig. 4.2c). By 2105Z/07, MCS-P1 was about to intersect Lake Michigan (Fig. 4.2d), and at 2304Z/07, had begun to cross the lake (Fig. 4.2e). To the west of MCS-P1 at 2304Z/07, a west–east line of convection increased in areal coverage. Two hours later at 0104Z/08, MCS-P1 had reorganized slightly with the strongest echoes on the southern flank, and had emerged over lower Michigan (Fig. 4.2f), still meeting the MCS criteria defined in section 2.2. At 0302Z/08, MCS-P1 continued to progress eastward (Fig. 4.2g), and was still causing severe wind reports (Fig. 4.1). Finally, by 0503Z/08, the leading line of the MCS began to weaken a bit (Fig. 4.2h), and would weaken more over the next two hours. The west–east line of convection that formed behind MCS-P1 about 6 h earlier had grown upscale, traversed the same path as, and merged with MCS-P1.

MCS-P1 organized as an upper-level short-wave trough approached from the west. At 1515Z/07, this short-wave trough is identifiable on water-vapor imagery, and was present on the Nebraska/Iowa border (Fig. 4.3a). By 1815Z/07, the short-wave trough was located immediately to the west of MCS-P1 in Wisconsin (Fig. 4.3b), and at 2115Z/07, was positioned under the cloud shield of MCS-P1 (Fig. 4.3c). Note how the cloud-shield extent increased as the short-wave trough intersected the convection

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29 Water-vapor imagery was used to pinpoint this upper-level short-wave trough, because the short-wave trough was not clearly apparent on upper-level RUC vorticity analyses.
associated with MCS-P1 (Figs. 4.3b,c). Behind this first short-wave trough, a second, subtle short-wave trough was located over Nebraska at 1815Z/07 (Fig. 4.3b). The west–east line of convection behind MCS-P1 (Figs. 4.2d,e) initially fired as this second short-wave trough moved along the eastern Iowa/Minnesota border.

On 18Z/07, as MCS-P1 was becoming better organized along the Minnesota/Wisconsin border (Figs. 4.2b,c), the MCS was located to the east of a closed surface low-pressure in northwest Iowa (Fig. 4.4a). Although the 4-hPa SLP contours do not capture this low-pressure system, a cyclonic-wind signature was present in this region (not shown). Additionally, MCS-P1 was positioned on the western side of a 573 dam 1000–500-hPa thickness ridge and was located under 16–18 g kg⁻¹ 925-hPa mixing ratios (Fig. 4.4a). Aloft, MCS-P1 was located downstream of a 200-hPa trough and positioned under southwesterly flow aloft (Fig. 4.4b), reminiscent of the dynamic-composite MCS pattern presented in chapter 3. MCS-P1 was not located in a well-defined equatorward-entrance region concomitant with the 200-hPa jet stream, but was still positioned on the anticyclonic shear-side of a robust 50–60 m s⁻¹ 200-hPa jet stream (Fig. 4.4b).

Additionally, MCS-P1 was situated near the terminus of a strong 17.5–20 m s⁻¹ 850-hPa LLJ (Fig. 4.4b), and progressed eastward in an environment of 1500–2500 J kg⁻¹ of MUCAPE (Fig. 4.4c). MCS-P1 was located on a MUCAPE gradient where values quickly increased to in excess of 4500 J kg⁻¹ about 200 km south of MCS-P1. At 18Z/07, 27.5–30 m s⁻¹ of 0–6-km shear was located directly over MCS-P1.

A rawinsonde taken at DVN on 18Z/07, immediately to the south of the path of MCS-P1 reveals a robust convective environment. The sounding contains a veering wind profile, indicative of warm advection, where winds increased from 7.5 m s⁻¹ at the
surface to 25 m s\(^{-1}\) at 600 hPa (Fig. 4.5). A dry adiabatic layer extended up to almost 850 hPa with a capping inversion immediately above this. The surface dewpoint was around 24°C and an elevated mixed layer was present in the sounding between 725 and 600 hPa. When elevated mixed layers are present in the environment, the potential for evaporatively cooled downdrafts is enhanced (e.g., Browning and Ludlam 1962). The amounts of MUCAPE and 0–6-km shear in this sounding were about 4000 J kg\(^{-1}\) and 24 m s\(^{-1}\), respectively, in agreement with the RUC analysis at the same time (Fig. 4.4c), while the amount of DCAPE in the sounding was almost 1200 J kg\(^{-1}\) indicating the potential for strong downdrafts.

At 18Z/07, MCS-P1 was located in southern Wisconsin. The convective downdrafts associated with MCS-P1 had resulted in an intense convective cold pool across southern Wisconsin. Surface temperatures increased almost 12°C from southern Wisconsin to northern Illinois across the southern edge of the convective cold pool (Fig. 4.6). Surface mixing ratio air with values above 18 g kg\(^{-1}\) pooled immediately to the south of the convectively induced boundary. Recall that additional convection fired behind MCS-P1 (Figs. 4.2e–h). Although this subsequent convection is not the focus of this section, it should be noted that this convection fired as the LLJ impinged on this convectively induced boundary, helping parcels reach their level of free convection (not shown). Additionally at 18Z/07, the air temperature at the buoy in the southern portion of Lake Michigan was only 12°C, about 15–19°C cooler than the surrounding land stations (Fig. 4.6). However, also notice that the SLP reading at the buoy was 1012.1 hPa, which is generally situated in the synoptic-scale SLP gradient. Thus, since the pressure was not notably higher than the surrounding land stations, the depth of the Lake
Michigan cold dome can be inferred to have been quite shallow, in agreement with the findings of Wylie and Young (1979).

At 00Z/08, MCS-P1 was located over southern Lake Michigan and was crossing the shallow Lake Michigan cold dome. At this time, MCS-P1 continued to be positioned on the western periphery of a 576 dam 1000–500-hPa thickness ridge (Fig. 4.7a). Additionally, a low is identifiable in the surface wind field to the west of MCS-P1 over Western Iowa, and MCS-P1 was located near a 18 g kg⁻¹ maximum in 925-hPa mixing ratio (Fig. 4.7a). Aloft at 200-hPa, MCS-P1 continued to be positioned downstream of a 200-hPa trough, and was located in the equatorward-entrance region of a 60 m s⁻¹ 200-hPa jet stream (Fig. 4.7b). The strength of the 200-hPa jet stream maximum increased from 6-h previous, and some of this increase may be attributable to diabatic outflow from MCS-P1 and the other convection on this day (not shown). Furthermore, MCS-P1 was collocated with a 22.5 m s⁻¹ 850-hPa LLJ, which was helping to advect warm-moist air into the MCS region as MCS-P1 crossed the Lake Michigan shallow cold dome. The amount of MUCAPE over MCS-P1 at 00Z/08 was generally between 500 and 1500 J kg⁻¹, but note the MUCAPE between 500 and 1500 J kg⁻¹ also present in the downstream environment over lower Michigan (Fig. 4.7c). Recall that climatological and compositing results indicated that MCSs persisted upon crossing Lake Michigan when there was large amounts of MUCAPE in lower Michigan (Figs. 2.7b and 3.6), and the persistence of MCS-P1 agrees with these findings. Finally, at 00Z/08, MCS-P1 was located under 17.5–20 m s⁻¹ of 0–6-km shear (Fig. 4.7c).

As MCS-P1 persisted upon crossing Lake Michigan at 00Z/08, there was between 0.5 and 0.75×10⁻¹² Pa m⁻² s⁻¹ of 850-hPa Q-vector convergence located immediately
over the convective region (Fig. 4.8). Recall that Q-vector convergence is associated with forcing for ascent. The Q-vector convergence concomitant with MCS-P1 resulted from cyclonic-vorticity advection ahead of an advancing trough from the west, and warm advection over the MCS region (note the inferred 850-hPa southwesterly flow blowing across the 850-hPa isotherms from warm to cold especially across Iowa). Farther to the west over Iowa, Wisconsin and Minnesota, Q-vectors point toward warm air indicative of frontogenetical forcing over the cold-pool induced boundary from MCS-P1 (Figs. 4.6 and 4.8). Coincident with this frontogenetical forcing, intense convection increased in coverage in this region around 00Z/08 (Figs. 4.2d–f).

As MCS-P1 approached Lake Michigan, an intense convective cold pool was created at 975-hPa (Fig. 4.9a). The 975-hPa potential temperature dropped more than 5 K between 19Z/07 and 23Z/07 as MCS-P1 crossed southern Wisconsin (Fig. 4.9a). This potential temperature decrease was present in conjunction with 17.5–20 m s\(^{-1}\) of 0–3-km shear (Fig. 4.9a). Additionally, between 19Z/07 and 23Z/07, MCS-P1 progressed through cross-section line A–A’ (Fig. 4.9a). When a similar 4-h potential temperature difference is taken along cross-section A–A’, the convective cooling can be identified up to almost 850-hPa (Fig. 4.9b). Recall that the DVN sounding at 18Z/07 revealed an elevated mixed layer with high values of DCAPE (Fig. 4.5), and a strong, deep convective cold pool is consistent with these findings from the DVN sounding\(^{30}\). Climatological results revealed that MCSs tended to persist with larger values of DCAPE (Fig. 2.9) and implied stronger cold pools, and the cold pool associated with MCS-P1 corroborates these results.

\(^{30}\) Thus, the cold pool strength and depth can likely be related to the larger-scale environment.
RKW theory states that strong ascent will occur in the presence of an intense cold pool and strong vertical shear, and both of these were collocated with MCS-P1. Recall that RKW theory discusses MCS maintenance in terms of the balance between the vertical shear ($\Delta U$) and the theoretical speed of a cold pool ($C$), where $C$ can be approximated as

$$C = \sqrt{\frac{\bar{\theta}}{g}} H.$$ 

In this equation, $\bar{\theta}$ is the average environmental potential temperature, $\theta^*$ is the potential temperature perturbation in the cold pool, and $H$ is the depth of the cold pool. If only the surface potential temperature perturbation is known, the above equation can be rewritten as

$$C = \sqrt{g \frac{\theta_{sfc}}{\bar{\theta}} H}.$$ 

RKW theory shows that when the convective cold pool associated with an MCS strengthens, $C$ increases in magnitude (e.g., Weisman and Rotunno 2004). As MCS-P1 approached Lake Michigan, $C$ was estimated to be around 18 m s$^{-1}$ (using $\bar{\theta}=303$ K, $\theta^*_{sfc}=7$ K, and $H=1.4$ km estimated from Figs. 4.6 and 4.9a,b). This cold pool velocity coupled with about 17.5 m s$^{-1}$ of 0–3-km line-perpendicular shear results in a ratio of $C/\Delta U$ that is nearly optimal (e.g., Rotunno and Weisman 2004). Thus, vigorous convective ascent occurred at the leading edge of MCS-P1, and as this MCS encountered the shallow cold dome of Lake Michigan, there was no impact by the unfavorable Lake Michigan-modified air below. An AMDAR sounding along the northern periphery of the convective cold pool shows the convective inversion extending up to almost 900 hPa, in good agreement with the RUC analyses at the same location (Figs. 4.9b,c).
A surface meteogram from Madison, WI (MSN) also confirms the strength of the surface cooling (Fig. 4.10a). As MCS-P1 passed over MSN at 20Z/07, the temperature dropped about 9°C and the dewpoint dropped about 3°C. Between 19Z/07 and 22Z/07, the SLP rose about 2.5 hPa, consistent with a strong, relatively deep convective cold pool. As MCS-P1 passed over buoy 45007, the temperature only dropped about 5°C between 00Z/08 and 02Z/08 (Fig. 4.10b). The magnitude of the temperature drop at buoy 45007 was less than at MSN because the temperature at the buoy was already significantly lower due to cooling from Lake Michigan. However, the SLP still rose about 2.5 hPa between 00Z/08 and 02Z/08. Additionally, note the strength of the near-surface inversion as MCS-P1 passed over Lake Michigan. At 23Z/07, the 5-m air temperature was 6.2°C warmer than the water temperature, indicating an intense near-surface inversion. Just 1-h later at 00Z/08, this inversion strength increased to almost 8°C immediately before MCS-P1 passage. Given this strong near-surface inversion, MCS-P1 likely never “saw” the cool unfavorable air associated with Lake Michigan below, and went right up and over this shallow dome of cold air, much like a speed bump. MCS-P1 persisted above this strong inversion over Lake Michigan, as the 850-hPa LLJ fed warm-moist air into an already high MUCAPE environment (Figs. 4.6a–c).

4.4 Dissipating MCS of 4 June 2005

The 4 June 2005 dissipating MCS (hereafter MCS-D1) initially formed from a relatively disorganized west–east-oriented convective line along the Kansas/Oklahoma border (not shown). By 11Z/04, MCS-D1 had produced severe storm reports and organized upon moving into western Missouri (Fig. 4.11). Over the next 5 h, MCS-D1
continued to have a coherent leading 45-dBZ line, but did not produce any additional severe reports until about 16Z/04. Following 16Z/04, multiple severe reports occurred until around 1930Z/04 when MCS-D1 intersected Lake Michigan and began immediate weakening.

NOWrad composite radar imagery shows MCS-D1 over northeastern Missouri at 14Z/04, to the south of another weakening MCS at the same time (Fig. 4.12a). Two hours later at 16Z/04, MCS-D1 featured a broad trailing stratiform precipitation region and was located in western Illinois (Fig. 4.12b). By 18Z/04, the leading convective line had become more robust and better organized in north-central Illinois (Fig. 4.12c). At 19Z/04, MCS-D1 was near the western Lake Michigan shore (Fig. 4.12d), and by 20Z/04 had crossed out over the water and immediately began to weaken to below 45 dBZ (Fig. 4.12e). One hour later, further weakening had occurred (Fig. 4.12f), and finally by 22Z/04, almost the entirety of MCS-D1 that was over Lake Michigan had dissipated (Fig. 4.12g). A small intense convective cell formed along the southern edge of the gust front from MCS-D1 around 21Z/04 (Fig. 4.12f), and by 22Z/04 had moved southward through Chicago (Fig. 4.12g) before dissipating.

At 15Z/04, MCS-D1 was located in extreme northern Missouri, to the east of a 1000-hPa surface low (Fig. 4.13a). Additionally, MCS-D1 was located on the upstream side of a 570-hPa 1000–500-hPa thickness ridge, and under 14 g kg⁻¹ 925-hPa mixing ratio values. However, the 925-hPa mixing ratio values associated with MCS-D1 6-h prior to being over Lake Michigan were almost 4 g kg⁻¹ less than those associated with MCS-P1 6 h prior to crossing Lake Michigan (Figs. 4.4a and 4.13a). Just as with MCS-P1, MCS-D1 was located downstream of a 200-hPa trough (Fig. 4.13b), and was
associated with the dynamic pattern identified in chapter 3. However, the trough to the west of MCS-D1 appears somewhat deeper and closer than the trough associated with MCS-P1 (Figs. 4.4b and 4.13b). MCS-D1 was located in a coupled jet region of a 30–40 m s\(^{-1}\) 200-hPa jet stream, near the core of a 20 m s\(^{-1}\) 850-hPa LLJ (Fig. 4.13b). The 200-hPa jet stream associated with MCS-D1 was 10–15 m s\(^{-1}\) weaker than the one concomitant with MCS-P1 (Figs. 4.4b and 4.13b). At 400 hPa, MCS-D1 was located to the south of a vorticity maximum associated with a short-wave trough that contained 24–28×10\(^{-5}\) s\(^{-1}\) of absolute vorticity (Fig. 4.13c). This trough extended from northwestern Iowa to southern Illinois. At 15Z/04, MCS-D1 was positioned on a MUCAPE gradient, and under 500–1500 J kg\(^{-1}\) of MUCAPE and 20–22.5 m s\(^{-1}\) of 0–6-km shear (Fig. 4.13d). These MUCAPE and 0–6-km shear values are smaller than those associated with MCS-P1 6 h before crossing Lake Michigan (Fig. 4.4c).

Unfortunately, no rawinsondes were launched to sample the environmental air in the path of MCS–D1, so a representative RUC sounding from eastern Illinois is presented in an attempt to get a sense of the vertical profile of the atmosphere at 18Z/04 when MCS-D1 was located in north-central Illinois, just before intersecting Lake Michigan (Fig. 4.14). This RUC sounding contains winds that veered almost 60° between the surface and 850 hPa, with winds that increased from 5 to 15 m s\(^{-1}\) between the surface and 700 hPa. The surface temperature at 18Z/04 at the sounding location was 28°C with a dry adiabatic layer extending to about 875 hPa and a surface dewpoint of 20°C. The resulting MUCAPE and 0–6-km shear values from this RUC sounding were approximately 1100 J kg\(^{-1}\) and 14 m s\(^{-1}\), respectively, with DCAPE of around 800 J kg\(^{-1}\).
These values are significantly smaller than those associated with MCS-P1 from the DVN proximity sounding (Fig. 4.5)

At 18Z/04, MCS-D1 was located in northern Illinois, approaching Lake Michigan (Fig. 4.15). Surface temperatures only cooled about 3–5°C upon the passage of MCS-D1, and a much weaker cold-pool boundary set-up to the south of MCS-D1, when compared to the cold-pool boundary to the south of MCS-P1 (Figs. 4.6 and 4.15). Mixing ratios in excess of 16 g kg\(^{-1}\) pooled to the south of this boundary on 18Z/04 (Fig. 4.15). Note that the buoy at the southern end of Lake Michigan recorded a temperature of 16°C, about 12–14°C cooler than surrounding land stations. However, the 1010.9-hPa SLP reading at this buoy is consistent with the general synoptic-scale SLP gradient, again indicating that the Lake Michigan cold dome was quite shallow.

Furthermore, at 18Z/04, MCS-D1 was located under a region of \(0.75 \times 10^{-12}\) Pa m\(^{-2}\) s\(^{-1}\) of 850-hPa Q-vector convergence (Fig. 4.16), slightly higher than the magnitude associated with MCS-P1 upon crossing Lake Michigan (Fig. 4.8). However, the trailing frontogenetical forcing to the west of MCS-P1 was not present upstream of MCS-D1 (Figs. 4.8 and 4.16). The Q-vector convergence over MCS-D1 was again the result of a combination of cyclonic vorticity advection and warm advection. In general, the magnitude of 850-hPa Q-vector convergence over both MCS-P1 and MCS D1 was quite similar at the time of Lake Michigan crossing (Figs. 4.8 and 4.16) corroborating climatological results illustrating that the magnitude of low-level Q-vector convergence

\[31\] It is important to note that the surface analysis associated with MCS-D1 is 2 h earlier in the day than the surface analysis associated with MCS-P1. Thus, surface temperatures did not have as many hours to warm before MCS-D1 convectively cooled the surface, possibly contributing to the weaker cold-pool boundary.
does not discriminate between persisting and dissipating Lake Michigan-crossing MCSs (e.g., Fig. 2.13).

As MCS-D1 was dissipating directly over Lake Michigan on 21Z/04, there was a broad surface low-pressure to the west located in western Minnesota (Fig. 4.17a). Additionally MCS-D1 was located on the immediate upstream side of a 573-hPa 1000–500-hPa ridge, in a similar location to MCS-P1 upon crossing Lake Michigan (Figs. 4.7a and 4.17a). However, the 925-hPa mixing ratio values directly above Lake Michigan at 21Z/04 were below 12 g kg\(^{-1}\), much less than the 18 g kg\(^{-1}\) over Lake Michigan as MCS-P1 persisted. Additionally at 21Z/04, MCS-D1 remained in the broad equatorward entrance region of a 30–40 m s\(^{-1}\) 200-hPa jet stream (Fig. 4.17b). However, MCS-D1 was beginning to move away from the core of the 850-hPa LLJ, and no longer had as much LLJ support upon crossing Lake Michigan as was present with MCS-P1 (Figs. 4.7b and 4.17b). The small convective cell behind MCS-D1 that moved through Chicago between 21Z/04 and 22Z/04 (Figs. 4.12f,g), formed as the eastern edge of this LLJ impinged upon the convective outflow boundary left by MCS-D1 (not shown). At 400-hPa, MCS-D1 was still collocated with a short-wave trough (Fig. 4.17c). However, the 24–28×10\(^{-5}\) s\(^{-1}\) vorticity maximum was even farther removed from MCS-D1 at 21Z/04 compared to 6 h earlier (Figs. 4.13b and 4.17b). At 21Z/04, MCS-D1 was emerging from a region of MUCAPE of around 2500 J kg\(^{-1}\), and the amount of MUCAPE downstream in lower Michigan was generally less than 500 J kg\(^{-1}\) (Fig. 4.17d). Recall that climatological and compositing results indicated that MCSs persisted upon crossing Lake Michigan when large amounts of MUCAPE were present in lower Michigan (Figs. 2.7b and 3.6). MCS-D1 was moving into a region of almost non-existent downstream
MUCAPE and subsequently dissipated. Finally, the amount of 0–6-km shear over MCS-D1 at 21Z/04 was around 15–20 m s\(^{-1}\), very similar to the amount over MCS-P1 upon crossing Lake Michigan (Figs. 4.7c and 4.17d). Previous climatological results showed that the amount of 0–6-km shear did not differentiate between MCSs that persisted and dissipated upon crossing Lake Michigan (Figs. 2.7a,b).

As MCS-D1 progressed across northern Illinois, the 975-hPa potential temperature only decreased 3–4 K between 17Z/04 and 19Z/04 (Fig. 4.18a)\(^{32}\). This decrease is 3–4 K less than the cooling associated with MCS-P1 just before crossing Lake Michigan (Fig. 4.9a), and is consistent with the RUC sounding that showed only 800 J kg\(^{-1}\) of DCAPE in the atmosphere (Fig. 4.14). Additionally, between 17Z/04 and 19Z/04, MCS-D1 passed through cross-section line B–B’. The potential temperature cooling over those 2 h only extended up to about 925 hPa in the center of the cold pool (Fig. 4.18b). The low-level cooling resulted in a cold pool velocity (C) of around 8 m s\(^{-1}\) (using \(\bar{\theta} = 303 \text{ K}, \theta'_{\text{stc}} = 3 \text{ K}, \) and \(H = 0.7 \text{ km estimated from Figs. 4.15 and 4.18a,b}\)), which when coupled with the 12.5 m s\(^{-1}\) of 0–3-km line-perpendicular environmental shear, led to sub-optimal, shallow ascent at the leading edge of MCS-D1. This convective ascent was likely much weaker than the ascent associated with MCS-P1.

As MCS-D1 moved near Lake Michigan, recall that there was a shallow horizontal buoyancy gradient associated with the Lake Michigan cold dome. Using the two dimensional horizontal vorticity equation (e.g., Weisman 1993)
\[
\frac{\partial \eta}{\partial t} = -\frac{\partial B}{\partial x}
\]

where B is buoyancy and \( \eta \) is the horizontal vorticity represented as

\[
\frac{\partial u}{\partial z} - \frac{\partial w}{\partial z}
\]

the horizontal buoyancy gradient on the upstream side of Lake Michigan would result in a circulation in the same direction as the RKW shear circulation associated with MCS-D1. Since the convective cold pool was very shallow in conjunction with MCS-D1, the horizontal circulation imparted by the ambient Lake Michigan cold pool was likely similar in depth to the oppositely directed circulation associated with the convective cold pool. Thus, the Lake Michigan-induced horizontal buoyancy gradient and associated circulation had the potential to disrupt the already sub-optimal convective ascent associated with MCS-D1, and the disruption and weakening of MCS-D1 occurred almost immediately upon intersecting the upstream shore of Lake Michigan (Figs. 4.12d–f). An AMDAR sounding from 1955Z/04 from an airplane that landed at Chicago O’Hare airport, in the center of the MCS cold pool, corroborates the RUC cross section, showing that the MCS-induced inversion only extended upward from the surface to about 925 hPa.

A meteogram from Aurora, IL (ARR), in the path of MCS-D1, shows that at this station there was about 7°C of cooling at the surface as MCS-D1 passed, more than in the RUC cross-section (Figs. 4.18b and 4.19a). The dewpoint also dropped about 1°C as MCS-D1 passed (Fig. 4.19a). However, note that the SLP only rose about 1 hPa between 18Z/04 and 20Z/04 as MCS-D1 moved over ARR, confirming the much shallower cold pool than was associated with MCS-P1 (Figs. 4.10a and 4.19a). At buoy 45007, there was little disruption in the general 5-m temperature trend as MCS-D1 passed between
21Z/04 and 23Z/04 (Fig. 4.19b). However, MCS-D1 was dissipating by the time it crossed over the buoy (Figs. 4.11f,g). Furthermore, the pressure actually dropped as MCS-D1 passed (Fig. 4.19b), possibly in association with a remnant wake-low. The strength of the near-surface inversion over Lake Michigan at the time of MCS-D1 passage was only 2.1°C, about one-third to one-fourth as weak as was associated with MCS-P1 passage (Figs. 4.19a,b). So, while MCS-D1 also likely went up and over this near-surface inversion, its weaker strength may have allowed for some of the unfavorable near-lake air to be ingested into the MCS-D1 updraft. When any unfavorable air was coupled with minimal downstream MUCAPE, no supporting 850-hPa LLJ, and a shallow-weak convective cold pool, MCS-D1 collapsed quickly upon intersecting Lake Michigan.

4.5 Persisting MCS of 18–19 June 2010

The second set of case study comparisons begins with the 18–19 June 2010 persisting MCS (hereafter MCS-P2). This second set of cases features two well-developed bow echo MCSs at the time of Lake Michigan crossing. At 12Z/18, MCS-P2 organized into a north–south-oriented line along the Nebraska/Iowa border, and MCS-P2 produced numerous severe reports (especially severe wind gusts) upon progressing toward Lake Michigan (Fig. 4.20). By 21Z/18, MCS-P2 contained a bow structure and the northern half of MCS-P2 was about to cross Lake Michigan. Three-hours later at 00Z/19, the northern half of MCS-P2 had crossed Lake Michigan and continued to produce severe reports until the bow-echo MCS dissipated just after 03Z/19. Note that the northern half of MCS-P2 that crossed Lake Michigan produced severe wind reports
on the immediate eastern shore of Lake Michigan in equal to or greater number than the portion of MCS-P2 that did not cross Lake Michigan (Fig. 4.20).

At 14Z/18 June 2010, MCS-P2 was already organized into an intense MCS, with a coherent leading convective line (Fig. 4.21a). By 16Z/18, MCS-P2 had progressed into central Iowa (Fig. 4.21b), and by 18Z/18, the leading edge of MCS-P2 was beginning to bow out (Fig. 4.21c). Two-hours later at 20Z/18, the well-developed bow echo associated with MCS-P2 was located in northwestern Illinois (Fig. 4.21d), and the northern portion of MCS-P2 intersected Lake Michigan before 22Z/18 (Fig. 4.21e). At 00Z/19, MCS-P2 was located in Michigan and Indiana, still producing significant wind damage (Figs. 4.20 and 4.21f), before beginning to weaken around 02Z/19 (Fig. 4.21g). Behind MCS-P2, another robust bow echo MCS formed along a convectively induced outflow boundary from MCS-P2 (Fig. 4.21g). However, this second MCS is not the focus of this section since it did not cross Lake Michigan.

At 16Z/18, MCS-P2 was located in central Iowa about 6 h before crossing Lake Michigan and was positioned to the south-southeast of a strong 992-hPa surface low (Fig. 4.22a) and to the east of a weaker 1005-hPa surface low along the Nebraska/Kansas border (an inverted surface trough in the region of this second low). Additionally, MCS-P2 was situated on the poleward side of a 576-hPa 1000–500-hPa thickness ridge and located under 14–16 g kg$^{-1}$ 925-hPa mixing ratio values. Aloft, MCS-P2 was downstream of a 200-hPa trough (Fig. 4.22b), consistent with the dynamic bow echo synoptic-scale pattern (Johns 1993). A 15 m s$^{-1}$ 850-hPa LLJ was impinging upon MCS-P2 at 16Z/18 (Fig. 4.22b). Furthermore, MCS-P2 was located to the southeast of a broad 400-hPa trough (Fig. 4.22c). There was no distinct 400-hPa short-wave trough in the
height field coincident with MCS-P2, even though there was a region of $16\sim20 \times 10^{-5} \text{s}^{-1}$ of 400-hPa absolute vorticity immediately downstream of MCS-P2. Additionally, at 16Z/18, MCS-P2 was moving eastward along the poleward edge of a strong gradient of MUCAPE, with MUCAPE values above 3500 J kg$^{-1}$ located just to the south of MCS-P2 (Fig. 4.22d). The 0–6-km shear over MCS-P2 at the same time was around 20 m s$^{-1}$, contributing to a robust MUCAPE/0–6-km shear environment.

A sounding from GRB at 18Z/18 confirms the favorable convective environment that MCS–P2 was moving into on this day (Fig. 4.23). The 18Z/18 GRB sounding had a surface temperature of 28°C with a dry-adiabatic layer extending up to around 900 hPa. The surface dewpoint was about 21°C and an elevated quasi-mixed layer was present between 850 and 750 hPa, resulting in DCAPE of about 1150 J kg$^{-1}$. Recall that when elevated mixed layers are present in the environment, the potential for evaporatively cooled downdrafts is enhanced (e.g., Browning and Ludlam 1962). MUCAPE values in this GRB sounding were 2200 J kg$^{-1}$ with 0–6-km shear values of 12 m s$^{-1}$. The 18Z/18 ILX sounding, taken farther to the south and deeper into the richer moisture, had almost 4500 J kg$^{-1}$ of MUCAPE (not shown), consistent with a region of increased MUCAPE in the RUC (Fig. 4.22d). The vertical shear at GRB was unidirectional, although not very strong (Fig. 4.23), conducive for severe wind-producing MCSs such as MCS-P2.

MCS-P2 was located in northwestern Illinois at 20Z/18. At the surface, MCS-P2 had produced an intense cold pool, with temperatures below 20°C in the center of this cooling in extreme eastern Iowa (Fig. 4.24). Surface temperatures increased 14°C across the southern boundary of this cold pool from eastern Iowa into central Illinois. The elevated mixed layer and associated dry air that was present in the 18Z/18 GRB sounding
likely contributed to this intense cooling from evaporatively cooled downdrafts. A very similar elevated mixed layer was associated with MCS-P1, which also produced a similar intense cold pool (Fig. 4.5). To the south of this convectively generated cold pool, surface mixing ratios in excess of 18 g kg$^{-1}$ were present in central Illinois and Missouri (Fig. 4.24). The bow-echo MCS that formed behind MCS-P2 initiated near this high mixing ratio air, and progressed eastward along the cold-pool boundary created by MCS-P2 (Figs 4.21d–g). The buoy at the southern end of Lake Michigan measured a surface temperature of 20°C at 20Z/18, about 8–12°C cooler than the surrounding land stations. However, the SLP of 1012.4 hPa at this buoy is consistent with the synoptic-scale gradient in this region, indicating a very-shallow Lake Michigan cold dome, as was present with the previous MCSs discussed in this chapter.

At 18Z/18, MCS-P2 was located under 0.5–0.75 Pa m$^{-2}$ s$^{-1}$ of 850-hPa Q-vector convergence (Fig. 4.25). This is very similar to the magnitude of 850-hPa Q-vector convergence associated with both MCS-P1 and MCS-D1 as each approached Lake Michigan (Figs. 4.8 and 4.16). The Q-vector convergence and forcing for ascent concomitant with MCS-P2 resulted from both cyclonic vorticity advection and low-level warm advection. The Q-vectors to the west of MCS-P2 in southern Minnesota crossed the 850-hPa isotherms towards warm air, indicative of weak frontogenetical forcing (Fig. 4.25). The strong bow echo that formed behind MCS-P2 organized near this region of frontogenetical forcing (Figs. 4.21e–g) and 700 hPa Q-vector convergence and forcing for ascent (Fig. 4.25).

At 22Z/18, the northern half of MCS-P2 was located over Lake Michigan, while the remainder was just to the south over northern Indiana (Fig. 4.21e). At this time,
MCS-P2 was situated to the southeast of a 996-hPa surface low (Fig. 4.26a) and to the east of a weaker 1006-hPa surface low near the southern Nebraska/Iowa border (not shown). Additionally, MCS-P2 was positioned under a 576-dam 1000–500-hPa thickness ridge and 16–18 g kg$^{-1}$ 925-hPa mixing ratio values, with mixing ratio values greater than 18 g kg$^{-1}$ located immediately to the south (Fig. 4.26a). At 22Z/18, MCS-P2 remained downstream of a 200-hPa trough and was about 200-km away from a 30–40 m s$^{-1}$ 200-hPa jet stream (Fig. 4.26b). Furthermore, MCS-P2 was situated near the terminus of a 12.5–15 m s$^{-1}$ 850-hPa LLJ that was transporting high MUCAPE air and over the shallow Lake Michigan cold dome. At 400-hPa, a weak short-wave trough was present in the height field at 22Z/18 over northern Illinois (Fig. 4.26c). This apparent convectively generated trough formed in situ over MCS-P2 (not shown). The 400-hPa vorticity that was ahead of MCS-P2 at 16Z/18 (Fig. 4.22c) was now located in lower Michigan (Fig. 4.26c). At 22Z/18, MCS-P2 continued to progress eastward on a strong MUCAPE gradient, tapping into values in excess of 4500 J kg$^{-1}$ immediately to the south of Lake Michigan (Fig. 4.26d). To complement this robust MUCAPE environment, about 15 m s$^{-1}$ of 0–6-km shear was located over MCS-P2.

The convectively induced cold pool associated with MCS-P2 was notable. Two-hour 975-hPa potential temperature differences along the path of MCS-P2 prior to crossing Lake Michigan between 20Z/18 and 22Z/18 reveal temperatures that fell between 8 and 10 K upon MCS-P2 passage (Fig. 4.27a). This robust cold pool was consistent with the DCAPE in the GRB sounding (Fig. 4.23). Between 20Z/18 and 22Z/18, MCS-P2 passed through cross section C–C’, and the depth of the convective cooling extended well above 800-hPa (Fig. 4.27b) and occurred in about 12.5 m s$^{-1}$ of 0–
3-km shear (Fig. 4.27a). This amount of 0–3-km shear was less than that associated with MCS-P1 and MCS-D1 upon approaching Lake Michigan (Figs. 4.9a and 4.18a). As MCS-P2 approached Lake Michigan, the cold pool velocity (C) was estimated to be around 25 m s^{-1} (using $\bar{\theta}$=303 K, $\theta'_s$=10 K, and H=2 km estimated from Figs. 4.24 and 4.27a,b). This cold pool velocity coupled with only 10–12.5 m s^{-1} of 0–3-km line-perpendicular shear results in a ratio of C/ΔU that exceeds 2, likely causing MCS-P2 to tilt upshear, and form a robust bow echo (e.g., Rotunno and Weisman 2004). The convective ascent at the leading edge of MCS-P2 was much deeper and more robust than any circulation associated with the shallow Lake Michigan horizontal buoyancy gradient, and there was no observable disruption in the implied upshear tilt and resulting bow echo structure of MCS-P2 upon intersecting Lake Michigan. An AMDAR sounding from an aircraft landing at MSN just behind MCS-P2 shows the convective inversion extending to almost 900 hPa (Fig. 4.27c), in agreement with the RUC cross section at the same location (Fig. 4.27b).

As MCS-P2 passed over Rockford, IL (RFD), a 9°C temperature decrease was recorded between 20Z/18 and 21Z/18 (Fig. 4.28a). This temperature decrease is consistent with the magnitude of the 975-hPa temperature decrease noted in the RUC as MCS-P2 progressed across northern Illinois (Fig. 4.27a). Additionally, the surface dewpoint dropped about 3°C and the SLP rose between 3.5 and 4 hPa coincident with MCS-P2 (Fig. 4.28a). This large pressure rise from the cold pool of MCS-P2 is consistent with the extremely deep cold pool noted in the RUC analyses (Fig. 4.28b). As MCS-P2 crossed the shallow Lake Michigan cold dome, the surface temperature only dropped about 3°C between 21Z/18 and 22Z/18 at buoy 45007 (Fig. 4.28b). This
temperature drop was much less than the 9°C at RFD (Fig. 4.28a) because the air
temperature at the buoy was already suppressed by the thermal stability of Lake
Michigan. Additionally, as MCS-P2 passed over buoy 45007, the SLP increased almost
2 hPa. Again, as with previous MCS cases, there was a strong near-surface inversion
over Lake Michigan as MCS-P2 passed. The temperature increased 2.9°C over the
lowest 5 m of the atmosphere just before MCS-P2 passed (Fig. 4.28b). This inversion
strength was not as strong as with MCS-P1, but still was stable enough to allow MCS-P2
to persist upon crossing Lake Michigan. Furthermore, MCS-P2 was moving extremely
quickly while passing over southern Lake Michigan (e.g., Figs. 4.21d–f). The southern
portion of Lake Michigan is approximately 150-km wide, and MCS-P2 covered this
distance in 1.5–2 h, moving at a speed of nearly 100 km h⁻¹. Thus, MCS-P2 had much
less time to ingest any unfavorable air associated with Lake Michigan than a more slowly
moving MCS. This typical fast translation speed associated with most of the bow echoes
in the Lake Michigan-crossing MCS climatology likely contributes to the high percentage
of bow echo persistence (Fig. 2.14). In general, MCS-P2 persisted in a very favorable
synoptic-scale environment upon crossing Lake Michigan.

Lake Michigan can have other effects on MCSs as well. By 2130Z/18, MCS-P2
was crossing Lake Michigan (Fig. 4.29a). Out ahead of MCS-P2 at 2130Z/18,
convective cells formed along the Lake Michigan/Indiana border, and by 22Z/18, these
cells merged with the bowing line, creating a reflectivity bulge right along the
Indiana/Michigan border. As soon as this reflectivity bulge from MCS-P2 moved
onshore, a high concentration of severe wind reports was noted along the
Michigan/Indiana border (Fig. 4.20). In fact, some of the highest convective gusts
associated with MCS-P2 occurred as this reflectivity bulge moved onshore, including a 90-kt gust in La Port, Indiana\textsuperscript{33}. The convective cells that eventually merged with MCS-P2, formed coincident with a convergence boundary that was located along the southeastern shore of Lake Michigan (Fig. 4.29a). Offshore west-southwesterly flow along the extreme southeastern Lake Michigan shore of around 5 m s\textsuperscript{−1}, impinged on ambient south-southwesterly flow in northern Indiana and southwestern Michigan of 7.5 m s\textsuperscript{−1}, resulting in surface convergence values along the southern Lake Michigan shore that exceeded $-15\times10^{-5}$ s\textsuperscript{−1} (Fig. 4.29a). Earlier in the day at around 17Z/18, the surface wind field at South Haven, MI (LWA) on the eastern shore of Lake Michigan quickly veered to westerly (Fig. 4.29b). This wind shift at LWA was coincident with a 4°C (1°C) temperature (dewpoint) drop. Coincidently, the SLP rose about 0.5–0.75-hPa, all indicative of onshore flow associated with the onset of a lake breeze. This lake breeze continued throughout the day as the air temperature over southern Michigan remained warmer than the air temperature over Lake Michigan (Fig 4.24). By 2130Z/18, convective cells had formed along this convergent, lake-breeze boundary near the intersection of Lake Michigan, lower Michigan, and Indiana (Fig. 4.29a). These cells remained relatively stationary until 22Z/18 when MCS-P2 overtook them from the west, forming the aforementioned reflectivity bulge. While the lake breeze itself did not disrupt the convection associated with MCS-P2, the convection that formed along this lake-breeze boundary appeared to contribute to the wind damage caused by an already robust MCS-P2.

\textsuperscript{33} Source: http://www.spc.noaa.gov/climo/reports/100618_rpts.html
4.6 Dissipating MCS of 24 June 2003

Just before 04Z/24 June 2003, a west–east oriented MCS (hereafter MCS-D2) formed out of a small region of disorganized convection. By 04Z/24, MCS-D2 had already produced a number of severe reports across western Iowa and eastern Nebraska (Fig. 4.30). Three-hours later, MCS-D2 had reoriented into a north-south convective line and by 11Z/24 had a bowing shape. The severe wind reports associated with MCS-D2 ceased around 11Z/24. By 14Z/24, MCS-D2 was located over Lake Michigan and the leading 45-dBZ line was beginning to break up. Shortly after 14Z/24, MCS-D2 completely dissipated.

At 02Z/24, MCS-D2 was just beginning to organize along the Nebraska/South Dakota border (Fig. 4.31a). Two-hours later at 04Z/24, MCS-D2 had a west–east orientation and two distinct embedded bow echoes on NOWrad composite reflectivity (Fig. 4.31b). By 06Z/24, MCS-D2 had progressed into central Iowa (Fig. 4.31c), and by 08Z/24, the convective line took on a north–south orientation (Fig. 4.31d). Additionally at 08Z/24, MCS-D2 had an extensive region of trailing stratiform precipitation. North–south-oriented MCS-D2 traversed into extreme western Wisconsin by 10Z/24 (Fig. 4.31e), and at 12Z/24, had a well-defined bow echo shape prior to crossing Lake Michigan (Fig. 4.31f). Two hours later as MCS-D2 began to traverse Lake Michigan, significant convective weakening occurred (Fig. 4.31g), and by 16Z/24, little convection remained concomitant with MCS-D2 (Fig. 4.31h).

MCS-D2 was about 6-h from intersecting Lake Michigan at 08Z/24, and was positioned in a region of implied geostrophic southerly flow at the surface under a 576-hPa 1000–500-hPa thickness ridge (Fig. 4.32a). In addition, at 08Z/24, MCS-D2 was in a
region of 16–18 g kg$^{-1}$ 925-hPa mixing ratio values. Farther aloft, MCS-D2 was situated
downstream of a 200-hPa trough, again indicative of the dynamic bow echo synoptic-
scale pattern (Fig. 4.32b). A 40–50 m s$^{-1}$ 200-hPa jet stream was located to the
immediate northwest of MCS-D2, and MCS-D2 was positioned near the broad
equatorward-entrance region of this 200-hPa jet stream (Fig. 4.32b). Furthermore, MCS-
D2 was located in the core of a 22.5 m s$^{-1}$ 850-hPa LLJ, stronger than the LLJ associated
with MCS-P2 6 h prior to crossing Lake Michigan (Figs. 4.22a and 4.32a)\(^{34}\). At 400 hPa,
there is quite a bit of noise in the height field (Fig. 4.32c) and none of these features in
the height field were coherent short-wave troughs associated with MCS-D2. Recall that
the environment near MCS-P2 also did not feature a well-developed upper-level short-
wave trough (Fig. 4.22c). Finally, MCS-D2 was positioned in a region of 2500–3500 J
kg$^{-1}$ of MUCAPE and 12.5 m s$^{-1}$ of 0–6-km shear, a robust convective environment that
was similar to the one MCS-P2 was positioned in 6-h prior to crossing Lake Michigan
(Figs. 4.22d and 4.32d).

A sounding from GRB reveals the vertical profile of the environment that MCS-
D2 was progressing into at 12Z/24 (Fig. 4.33). Nearly calm southerly surface winds
veered to southwesterly at 12.5 m s$^{-1}$ just above the boundary layer. Since this sounding
was taken at 12Z/24, a surface nocturnal inversion was present that extended up to
around 925 hPa. MCS-D2 was likely feeding off unstable air that was elevated above
this near-surface nocturnal inversion. For many years, researchers thought that nocturnal
convection often exhibited bore-like behavior (e.g., Carbone et al. 1990) above these

\(^{34}\) Although note MCS-D2 was 6-h prior to Lake Michigan crossing at 08Z while MCS-
P2 was 6-h prior to crossing Lake Michigan at 16Z. Climatologically, the LLJ tends to
be more robust during the overnight hours (e.g., Blackadar 1957; Stensrud 1996).
nocturnal inversions, and that the RKW cold-pool viewpoint on convective maintenance did not hold. However, Parker (2008) found that simulated MCS cold pools could lift surface parcels into MCS updrafts in a dynamically consistent manner with RKW theory until nocturnal cooling exceeded 10°C. In this case, the surface temperature had only cooled from 29°C at 00Z/24 (not shown) to 22°C at 12Z/24, a difference of 7°C. Thus, it is possible that surface parcels were still being ingested into the MCS updraft. The dewpoint at GRB was 18°C (Fig. 4.33), resulting in almost 2700 J kg\(^{-1}\) of MUCAPE located above the nocturnal inversion. However, the vertical wind profile only resulted in about 8 m s\(^{-1}\) of 0–6-km shear. Additionally, about 900 J kg\(^{-1}\) of DCAPE was available for downdrafts. Thus, even though ample MUCAPE was still available for MCS-D2 upon approaching Lake Michigan, the 0–6-km shear necessary for organization was extremely weak near Lake Michigan.

At 12Z/24, MCS-D2 was about 1–2 h away from intersecting Lake Michigan. At the surface, there was only 2–4°C cooling as MCS-D2 passed over southern Wisconsin. However, recall that 12Z is early in the morning in Wisconsin. Thus, significant surface cooling would not be expected. Additionally, the surface mixing ratios near MCS-D2 were only in the 12–14 g kg\(^{-1}\) range, with the MCS likely ingesting unstable air above the nocturnal inversion (Fig. 4.33). In addition, at the surface, a 1008-hPa wake low formed behind MCS-D2 in southern Minnesota. Even though MCS-D2 was approaching Lake Michigan in the early morning hours, the buoy temperature at the southern end of Lake Michigan was only 15°C, about 7–8°C colder than the surrounding land stations (Fig. 4.34). The SLP at this buoy was 1015.9 hPa, similar to the other SLP readings nearby, again indicating the MCS-D2 was approaching a very shallow Lake Michigan cold dome.
MCS-D2 was situated under 0.25–0.5 Pa m$^{-2}$ s$^{-1}$ of 850-hPa Q-vector convergence at 12Z/24 (Fig. 4.35). This Q-vector convergence magnitude was slightly lower than that associated with MCS-P2 (Fig. 4.25). However, very weak low-level forcing for ascent was still present over MCS-D2 (Fig. 4.35). The Q-vector convergence over MCS-D2 was mostly driven by warm advection but there was a small contribution from cyclonic vorticity advection (not shown). Farther to the west over Minnesota and South Dakota, there was a stronger region of 850-hPa Q-vector forcing for ascent, and implied frontogenetical forcing near the surface wake low (Fig. 4.34) and a stratiform region of precipitation behind MCS-D2 (Figs. 4.31e–g). Again, MCS-P2 did not have significantly more Q-vector convergence when compared to MCS-D2, further corroborating climatological and compositing results that the magnitude of Q-vector convergence does not typically differentiate between MCSs that persist and dissipate upon crossing Lake Michigan.

By 14Z/24 June 2003, MCS-D2 was positioned over Lake Michigan and was beginning to dissipate (Fig. 4.31g). At this time MCS-D2 was located to the east of a weak surface low that had formed in central Minnesota (Fig. 4.34) and under implied south-southwesterly geostrophic surface flow (Fig. 4.36a). Additionally, MCS-D2 was near the apex of a 576-hPa 1000–500-hPa thickness ridge and beneath 12–14 g kg$^{-1}$ 925 mixing ratio values (Fig. 4.36a). These 925-hPa mixing ratio values are almost 4 g kg$^{-1}$ lower than those concomitant with MCS-P2 upon crossing Lake Michigan (Fig. 4.26a). Aloft at 200-hPa, MCS-D2 remained downstream of an upper-level trough, and a 50 m s$^{-1}$ jet stream (Fig. 4.36b). However, MCS-D2 had distanced itself from this 200-hPa jet stream compared to 6 h earlier (Fig. 4.32b). This increased distance from the upper-level
jet stream likely explains why the 12Z/24 GRB sounding had only 8 m s$^{-1}$ of 0–6-km shear (Fig. 4.33). There was still a weak 10 m s$^{-1}$ 850-hPa LLJ near MCS-D2 (Fig. 4.36b). However, this LLJ was not as extensive as that associated with MCS-P2 upon crossing Lake Michigan (Fig. 4.32b).

At 400-hPa, there was a weak short-wave-trough in the height field in eastern Wisconsin beneath MCS-D2, but as with MCS-P2 this feature was convectively generated in-situ over MCS-D2 (not shown). Finally, MCS-D2 was still located in a region of between 500 and 1500 J kg$^{-1}$ of MUCAPE and around 5–7.5 m s$^{-1}$ of 0–6-km shear (Fig. 4.36d). However, there was negligible MUCAPE in lower Michigan. Climatological and compositing results revealed that MCSs typically dissipate if there is limited MUCAPE in lower Michigan, and this case corroborates these results. The low (< 10 m/s) 0–6-km shear was also not enough to maintain convective organization. In total, MCS-D2 existed in a favorable synoptic-scale environment 6-h prior to Lake Michigan crossing, much like MCS-P2 (Figs. 4.22a–d and 4.32a–d). However, MCS-P2 remained in this favorable environment upon crossing Lake Michigan (Figs. 4.26a–d) while MCS-D2 dissipated upon entering a very unfavorable synoptic-scale environment (Figs. 4.36a–d).

As MCS-D2 was about to intersect Lake Michigan, two hour 975-hPa potential temperature differences between 11Z/24 and 13Z/24 revealed only between 2–4 K cooling as the MCS passed (Fig. 4.37a). This amount of 975-hPa cooling was significantly less than that associated with MCS-P2 just before intersecting Lake Michigan (Fig. 4.27a). However, MCS-D2 crossed southern Wisconsin during the early morning hours when there is typically not as much potential for surface cooling. Marshall
et al. (2011) found that surface cold pool properties can vary widely in regions affected by nocturnal MCSs. MCS-D2 was only in the presence of about 10–12.5 m s\(^{-1}\) of 0–3-km shear and the depth of the aforementioned cooling only extended up to about 900 hPa (Figs. 4.37a,b)\(^{35}\). Thus, in an RKW sense, there was much weaker, shallower ascent at the leading edge of MCS-D2 when compared to MCS-P2.

If the theoretical cold pool speed was calculated in the same manner as for previous MCSs in this chapter, \(C\) would equal approximately 11 m s\(^{-1}\) at 12Z/24 (using \(\tilde{\theta}=297\) K, \(\theta_{sc}=4\) K, and \(H=1\) km estimated from Figs. 4.34 and 4.37a,b) just before crossing Lake Michigan. A \(C\) value of 11 m s\(^{-1}\) coupled with only 10–12.5 m s\(^{-1}\) of 0–3-km line-perpendicular shear would lead to a ratio of \(C/\Delta U\) that was nearly optimal. Even though the ratio of \(C/\Delta U\) was nearly optimal, MCS-D2 still took on a bow-echo shape upon crossing Lake Michigan (Figs. 4.31f,g). Given that MCS-D2 crossed Lake Michigan near 13Z/24, the MCS was likely feeding off of high MUCAPE air above the nocturnal inversion (Fig. 4.33), and thus, the amount of near-surface cooling might be less important to convective evolution, especially when compared to the unfavorable synoptic-scale environment MCS-D2 progressed into at 14Z/24 (Figs. 4.36a–d). Even with the nocturnal inversion present, near-surface air parcels were still potentially being ingested into the MCS updraft via RKW convective cold pool dynamics since Parker (2008) found that a simulated MCS cold pool could lift surface parcels until nocturnal surface cooling exceeded 10°C. This temperature cutoff had not quite been reached in the MCS-D2 environment. However, at about 11Z/24 MCS-D2 began moving at a speed faster than the calculated cold-pool speed around the time that surface severe wind

\(^{35}\) There was no AMDAR sounding available to confirm the cold pool depth.
reports ceased (Figs. 4.30 and 4.31d–g), indicating the MCS-D2 may have become elevated and bore-like (e.g., Parker 2008) a couple of hours prior to crossing Lake Michigan.

A meteogram from Oshkosh, WI, (OSH) reveals that as MCS-D2 passed over the site, only a 1–2°C temperature drop occurred between 12Z/24 and 13Z/24 (Fig. 4.38a). The SLP had been slowly rising at OSH between 09Z/24 and 12Z/24, but there was a slight increase in the rate of SLP rise between 12Z/24 and 13Z/24. Immediately after MCS-D2 passage, there was a dramatic 5-hPa SLP drop in conjunction with a wake low. A meteogram from buoy 45007 does not show any influence from MCS-D2 passage because the southern extent passed just to the north of this location (Figs. 4.31g,h and 4.38b). However, the near surface inversion strength over Lake Michigan was 3.4°C, even stronger than that associated with MCS-P2 (Figs. 4.28b and 4.38b). Recall that the top of the nocturnal inversion in the GRB sounding was around 925 hPa (Fig. 4.33). Therefore, MCS-D2 may not have seen Lake Michigan below because the nocturnal inversion was deeper than the inferred Lake Michigan cold dome depth. This interesting result reveals another reason why the strength of the Lake Michigan near-surface inversion is not a perfect discriminator between MCSs that persisted and dissipated over Lake Michigan even though this metric did show a statistically significant differentiation between these two MCS categories (Fig. 2.11a–c). If an MCS is elevated over a nocturnal inversion and feeding off elevated MUCape, the MCS may not be significantly influenced by atmospheric conditions below this inversion, regardless of the strength of the Lake Michigan near-surface inversion. Furthermore, the conditions associated with Lake Michigan are irrelevant for MCS persistence if the synoptic-scale
environment becomes unfavorable for convective evolution. MCS-D2 dissipated upon moving into a very unfavorable synoptic-scale environment, and Lake Michigan likely played little to no role in this dissipation as MCS-D2 just happened to be positioned over Lake Michigan when synoptic-scale support for persistence waned.

4.7 Persisting MCS of 22 August 2007

The third and final pair of real-data case study comparisons begins with the 22 August 2007 MCS (hereafter MCS-P3) that persisted upon crossing Lake Michigan. These case studies both occurred during the late warm season (defined as JAS) and were associated with the progressive synoptic-scale pattern. By 02Z/22, MCS-P3 was located in northern Iowa and had already produced numerous high wind and large hail reports across Nebraska and Iowa (Fig. 4.39). Around 08Z/22, MCS-P3 neared Lake Michigan and continued to produce high wind reports until intersecting the western shore of Lake Michigan. After MCS-P3 persisted upon crossing Lake Michigan, a cluster of severe wind reports were recorded in southwestern lower Michigan around 12Z/22 before the leading 45 dBZ line of MCS-P3 began to break up around 14Z/22.

At 00Z/22, MCS-P3 was organizing into a northeast–southwest-oriented narrow line of severe convection in northwestern Iowa (Fig. 4.40a). Two-hours later at 02Z/22, MCS-P3 had progressed eastward, and contained a well-developed convective line (Fig. 4.40b). At 04Z/22, MCS-P3 was located in northeastern Iowa and the stratiform precipitation shield associated with MCS-P3 had grown in size (Fig. 4.40c). At 06/22, MCS-P3 briefly contained a bow-echo structure, and by 08Z/22, MCS-P3 was beginning to intersect Lake Michigan (Fig. 4.40e). At 10Z/22, MCS-P3 was over Lake Michigan.
(Fig. 4.40f), and by 12Z/22 had emerged into lower Michigan, continuing to cause severe wind reports (Figs. 4.39 and 4.40g). Finally, at 14Z/22, MCS-P3 was beginning to weaken near the U.S./Canadian border to the east of Detroit (Fig. 4.40h).

At 04Z/22 August 2007, MCS-P3 was about 6-h away from crossing Lake Michigan. At this time, MCS-P3 was positioned immediately to the east of an inverted trough that extended across Nebraska into western Iowa (Fig. 4.41a). This inverted trough was collocated with a weak quasi-stationary boundary at the surface (not shown). Additionally, MCS-P3 was positioned under a 573 dam 1000–500-hPa thickness ridge and was located near robust 925-hPa mixing ratios, with values in excess of 20 g kg$^{-1}$ immediately equatorward of MCS-P3 (Fig. 4.41a). Aloft, the 200-hPa height pattern was associated with a generally zonal-oriented 50 m s$^{-1}$ jet stream, indicative of the progressive synoptic-scale pattern identified in chapter 3 (Johns 1993). MCS-P3 was positioned on the anticyclonic shear side of this 200-hPa jet stream. A 15 m s$^{-1}$ 850-hPa LLJ was located across Iowa, and MCS-P3 was positioned near its terminus (Fig. 4.41b). Although there were no distinct 400 hPa short-wave troughs progressing eastward with MCS-P3 (Fig. 4.41c), MCS-P3 was persisting in a robust convective environment with over 5500 J kg$^{-1}$ of MUCAPE and 25–35 m s$^{-1}$ of 0–6-km shear located nearby (Fig. 4.41d).

There were no rawinsondes launched that sampled environmental air in the path of MCS-P3, so a representative RUC sounding from the central Wisconsin/Illinois border is presented to get a sense of the convective environment MCS-P3 was moving into at 06Z/22. This sounding features a relatively unidirectional wind profile above the surface, with 500-hPa winds of around 12.5 m s$^{-1}$ (Fig. 4.42). The surface temperature at the
sounding location was approximately 27°C and there was a shallow nocturnal inversion that extended up to around 950-hPa. MCS-P3 was likely feeding off MUCAPE above this nocturnal inversion. The surface temperature at the sounding location had only fallen about 4°C from 00Z/22, so surface parcels were likely still being ingested into the updrafts of MCS-P3 via cold pool dynamics (e.g., Parker 2008). The surface dewpoint at the sounding location was around 25°C\(^{36}\) (Fig. 4.42), and there was a dry layer centered around 800 hPa. In total, about 3400 J kg\(^{-1}\) of MUCAPE and 13 m s\(^{-1}\) 0–6-km were present in the sounding along with approximately 1100 J kg\(^{-1}\) of DCAPE. MCS-P3 was progressing into a robust convective environment, although the 0–6-km shear at the sounding location was not as high as the values over MCS-P3 at 04Z/22 (Figs. 4.41d and 4.42).

As MCS-P3 progressed eastward at 06Z/22, surface temperatures cooled about 4–5°C in its wake (Fig. 4.43). This surface cooling led to a convectively enhanced surface boundary across central Iowa where surface temperatures increased from 20°C in northeastern Iowa to 27°C in southeastern Iowa. Along and to the south of this boundary, robust surface mixing ratio values approaching 20 g kg\(^{-1}\) were positioned across central Iowa and northern Illinois (Fig. 4.43). At 06Z/22, MCS-P3 was about 2–3 h away from intersecting Lake Michigan, and at this time, buoy 45007 in the south-central portion of Lake Michigan recorded a surface temperature of 22°C. This temperature was consistent with the air temperatures at the other land stations around Lake Michigan, indicating that unlike the environments associated with the previously discussed earlier-season MCSs in this chapter, the environment over Lake Michigan was likely very similar to the

\(^{36}\) The surface dewpoint in this sounding was about 1–2°C higher than actual nearby surface observations.
surrounding land environment. Thus, MCS-P3 did not have to navigate a lake-cooled
dome of air in order to persist upon crossing Lake Michigan.

At 06Z/22, MCS-P3 was located in southeastern Wisconsin coincident with a
region of slightly less than $0.25\times10^{-12}$ Pa m$^{-2}$ s$^{-1}$ of 850-hPa Q-vector convergence (Fig.
4.44). Although the 850-hPa Q-vector forcing for ascent was weak, there was still
forcing for ascent located over MCS-P3. This Q-vector convergence is almost
exclusively the result of low-level warm advection as the implied 850-hPa geostrophic
wind crosses the 850-hPa isotherms from warm to cold. The magnitude of Q-vector
convergence in the vicinity of MCS-P3 was generally less than the magnitude associated
with the previous persisting MCSs presented in this chapter (e.g., Figs. 4.8 and 4.24).

Recall that MCSs that occurred in conjunction with the progressive synoptic-scale pattern
generally were coincident with a smaller magnitude of 850-hPa Q-vector convergence
and forcing for ascent than those that occurred in conjunction with the dynamic synoptic-
scale pattern (Figs. 3.4 and 3.9).

At 14Z/22, MCS-P3 was positioned directly over Lake Michigan under broad
south-southwest geostrophic surface flow (Fig. 4.45a). Additionally, MCS-P3 was
situated under a 576-hPa 1000–500-hPa thickness ridge and near 16–18 g kg$^{-1}$ 925-hPa
mixing ratio values. Aloft, the synoptic-scale pattern continued to be generally zonal
along and to the north of MCS-P3 (Fig. 4.45b). MCS-P3 remained situated along the
anticyclonic-shear side of a 50 m s$^{-1}$ 200-hPa jet stream, a typical location for
progressive MCSs in the Great Lakes region (e.g., Metz and Bosart 2010). A 15–17.5 m
s$^{-1}$ 850-hPa LLJ curved anticyclonically into the vicinity of MCS-P3 at 14Z/22. A weak
400-hPa short-wave trough was present in the height field to the west of MCS-P3 over
southwestern Wisconsin, but this trough lacked temporal coherence and was likely generated in situ (Fig. 4.13). As MCS-P3 crossed Lake Michigan, the convective environment still featured 500–1500 J kg\(^{-1}\) of MUCAPE and 12.5–15 m s\(^{-1}\) of 0–6-km shear (Fig. 4.45d). However, unlike the environments associated with MCS-P1 and MCS-P2 that persisted upon crossing Lake Michigan (Figs. 4.7c and 4.26d), there was not a robust amount of MUCAPE located in lower Michigan for MCS-P3 to feed off of and persist. However, there was a small region of MUCAPE values of between 500 and 1500 J kg\(^{-1}\) along the Michigan/Indiana border and MCS-P3 moved along the poleward periphery of this adequate convective environment after crossing Lake Michigan (Figs. 4.40f,g). MCS-P3 finally dissipated after progressing eastward from this small region of enhanced MUCAPE (Figs. 4.40h and 4.45d).

As MCS-P3 was about to cross Lake Michigan, 975-hPa potential temperatures fell in excess of 5 K between 06Z/22 and 09Z/22 associated with convectively induced downdrafts (Fig. 4.46a). Keep in mind that Marshan et al. (2011) found that surface cold pool properties associated with nocturnal MCSs can vary widely. This relatively strong cold pool existed in 12.5–15 m s\(^{-1}\) of 0–3-km line-perpendicular shear. Additionally, between 06Z/22 and 09Z/22, MCS-P3 passed through cross section E–E’ and deep cooling was noted with this cold pool extending up to around 850-hPa (Fig. 4.46b). Additionally, this cross section reveals mid-level cooling in the 700–800 hPa layer, consistent with the dry layer located at about the same level in the 06Z/22 RUC sounding (Figs. 4.42 and 4.46b).

This strong, deep convective cold pool had a velocity (C) that was calculated to be about 16 m s\(^{-1}\) (using \(\overline{\theta}=299\) K, \(\theta'_{sfc}=5\) K, and H=1.5 km estimated from Figs. 4.43
and 4.46a,b), slightly larger than the magnitude of the environmental 0–3-km shear. These values result in a ratio of $C/\Delta U$ slightly larger than optimal, indicating the potential for a slightly upshear-tilted MCS (e.g., Weisman and Rotunno 2004). MCS-P3 did feature a bow echo structure at 06Z/22. However, this bow echo structure transitioned to a linear structure by 09Z/22, consistent with a ratio of $C/\Delta U$ that was borderline between indicating vertically upright and upshear-tilted updrafts (Fig. 4.46a). This combination of a strong convective cold pool and moderate 0–3-km shear likely resulted in strong, deep ascent based on RKW theory as MCS-P3 approached Lake Michigan. Recall, that the 06Z/22 RUC sounding did show a shallow, nocturnal inversion (Fig. 4.42). However, the magnitude of surface cooling was far less than the 10°C that simulated MCSs required before transitioning into a purely elevated convective system (Parker 2008) and there was still surface-based CAPE in the environment (Fig. 4.42). Thus, this cold pool likely lifted surface parcels into the MCS-P3 updraft in a manner dynamically consistent with RKW theory. Additionally, MCS-P3 never significantly changed its forward translation speed in a manner similar to MCS-D2. Given that Lake Michigan was similar in temperature to the surrounding land stations (Fig. 4.43), there was no shallow horizontal buoyancy gradient on the upstream shore of Lake Michigan. MCS-P3 persisted upon crossing Lake Michigan through normal convective processes, as the environmental conditions over Lake Michigan were almost identical to the surrounding land conditions. An AMDAR sounding from an aircraft landing at MKE behind MCS-P3 shows the depth of the convective cold pool extending up to almost 900 hPa (Fig. 4.46c), consistent with the depth of the convective cold pool in the RUC cross section at the same approximate location.
A meteogram from RFD (Fig. 4.47a), in the path of MCS-P3, corroborates the magnitude of convective cooling indicated by the RUC (Figs. 4.46a,b). As MCS-P3 passed over RFD beginning just before 08Z/22, the surface temperature dropped about 4°C with a similar magnitude drop in the dewpoint (Fig. 4.47a). The SLP increased almost 3 hPa between 07Z/22 and 08Z/22, indicative of a deep convective cold pool, especially given the relatively modest amount of surface cooling. The 5-m air temperature trace at buoy 45007 did not indicate a decrease as MCS-P3 passed between 09Z/22 and 10Z/22 (Fig. 4.47b). The air temperature did not drop because the surface air was already saturated (not shown) and very near to the Lake Michigan water temperature. The 2.5–3 hPa SLP rise at the buoy was very similar to the SLP rise at RFD. Additionally, note that there was almost no near-surface inversion at buoy 45007. At 08Z/22 as MCS-P3 was beginning to intersect Lake Michigan, the 5-m air temperature was only 0.1°C warmer than the water temperature. This inversion was nowhere near as stable as the over-lake inversions associated with MCS-P1 and MCS-P2 (Figs. 4.10 and 4.28). However, MCS-P3 was able to persist anyways, because in late August, Lake Michigan is much warmer than in early June (e.g., Fig. 2.4). Thus, MCS-P3 persisted in a favorable synoptic-scale environment on a day when the conditions over Lake Michigan were very similar to those over the surrounding land locations. Lake Michigan did not create any environmental conditions that were unfavorable to MCS-P3 persistence.

4.8 Dissipating MCS of 6 July 2003
At 03Z/06 July 2003, a robust MCS (hereafter MCS-D3) was progressing across eastern Nebraska (Fig. 4.48). By 03Z/06, MCS-D3 had already produced a wide swath of severe reports. As MCS-D3 moved eastward, the severe reports continued, except for a lull between 08Z/06 and 12Z/06. After 12Z/06, a few severe wind and hail reports were recorded in southeastern Wisconsin, just before MCS-D3 began crossing Lake Michigan. MCS-D3 dissipated just after 15Z/06 while still over Lake Michigan and did not cause any severe reports in lower Michigan (Fig. 4.48).

At 04Z/06, MCS-D3 contained a robust convective line and a large region of stratiform precipitation in northeastern Nebraska (Fig. 4.49a), and by 06Z/06, MCS-D3 had progressed into western Iowa (Fig. 4.49b). By 08Z/06, MCS-D3 had lost much of its associated convection over central Iowa, and began to take on a mesoscale convective vortex (MCV) structure on radar (e.g. Zhang and Fritsch 1987; Galarneau et al. 2009; Fig. 4.49c). By 10Z/06, MCS-D3 had reorganized with a region of 45-dBZ radar echoes along the northeastern periphery of the MCV (Fig. 4.49d). This convective reorganization and MCV development coincided with the time when MCS-D3 did not produce any severe reports (Fig. 4.48). Two hours later at 12Z/06, MCS-D3 had moved out ahead of the parent MCV into Wisconsin, and by 14Z/06 (Fig. 4.49e), was situated immediately to the west of Lake Michigan (Fig. 4.49f). A region of stratiform precipitation connected MCS-D3 back to the parent MCV over northeastern Iowa, in a fashion visually reminiscent of a warm front associated with an occluded synoptic-scale

37 After the MCV formed, MCS-D3 is identified as the mesoscale region of convection that moved out ahead of the parent MCV. This mesoscale feature produced the severe reports that occurred between 12Z/06 and 15Z/06 on the upstream side of Lake Michigan (Fig. 4.48) and met the climatological criteria laid out in chapter 2. The MCV existed primarily as stratiform precipitation once MCS-D3 progressed to the east.
cyclone. MCS-D3 crossed Lake Michigan as a relatively small MCS, and weakened significantly by 16Z/06 (Fig. 4.49g). By 18Z/06, there was only a weak reflectivity signature of MCS-D3 remaining along the Lake Michigan/lower Michigan border (Fig. 4.49h), and even these remaining precipitation echoes would disappear soon after 18Z/06. Behind MCS-D3, the relatively innocuous stratiform precipitation region associated with the parent MCV continued progressing eastward through western Wisconsin. This MCV eventually fired off new convection and morphed into another MCS in lower Michigan. However, this subsequent MCS development occurred primarily after the MCV intersected Lake Michigan, and thus, the MCV itself will not be a major focus of this section.

At 10Z/06 July 2003, MCS-D3 was about 6 h from crossing over Lake Michigan. A persistent west–east-oriented quasi-stationary boundary had been set up across Nebraska, Iowa and Illinois over much of the week leading up to MCS-D3 (Metz and Bosart 2010; their Fig. 15), and this boundary was still present at 10Z/06 (not shown). In addition, at the surface at 10Z/06, MCS-D3 was located in a region of implied surface southwesterly geostrophic flow (Fig. 4.50a). Furthermore, MCS-D3 was positioned on the poleward periphery of a 573-dam 1000–500-hPa thickness ridge and was in a region of 12–14 g kg$^{-1}$ 925-hPa mixing ratio values. Aloft at 200-hPa, the flow was generally zonal (Fig. 4.50b), consistent with the progressive synoptic-scale pattern identified by Johns (1993). MCS-D3 was positioned near the equatorward-entrance region of a 40–50 m s$^{-1}$ 200-hPa jet streak and was located on the anticyclonic shear side of this jet stream, a common location for progressive MCSs in the Great Lakes region (e.g., Metz and Bosart 2010; Fig. 4.50b). At 10Z/06, MCS-D3 was situated near the northern terminus of
a 17.5 m s$^{-1}$ anticyclonically curving 850-hPa LLJ. Additionally, MCS-D3 was located immediately downstream of a robust 400-hPa short-wave trough that contained up to 28×10$^{-5}$ s$^{-1}$ of 400-hPa absolute vorticity (Fig. 4.50c). The MUCAPE/0–6-km shear environment was adequate for MCS-D3 persistence at 10Z/06 with 1500–2500 J kg$^{-1}$ of MUCAPE and 12.5 m s$^{-1}$ of 0–6-km shear located near MCS-D3 (Fig. 4.50d).

A 12Z/06 sounding from GRB reveals the vertical structure of the atmosphere immediately to the north of the pathway traversed by of MCS-D3 (Fig. 4.51). This sounding shows relatively weak winds below 500 hPa, with only 10 m s$^{-1}$ of westerly flow at 500 hPa. The 12Z/06 surface temperature (dewpoint) at GRB was 20°C (19°C) and a strong nocturnal inversion was present extending up to almost 925 hPa. MCS-D1 was likely feeding off of MUCAPE that was elevated above this nocturnal inversion$^{38}$. Parker (2008) showed that simulated MCSs could still ingest surface parcels through RKW cold pool dynamics until surface nocturnal cooling exceeded 10°C and a surface convective cold pool was no longer identifiable. The 00Z/06 temperature at GRB was about 10°C warmer than the 12Z/06 temperature (not shown). Thus, MCS-D3 was near the cutoff where these cold pool processes no longer govern MCS maintenance. The temperature, dewpoint, and wind profiles in the 12Z/06 GRB sounding yielded approximately 1000 J kg$^{-1}$ of MUCAPE and 11 m s$^{-1}$ of 0–6-km shear, along with only about 700 J kg$^{-1}$ of DCAPE. Thus, the convective environment associated with MCS-D3 just before crossing Lake Michigan (Fig. 4.51) was not as favorable as that associated with MCS-P3 (Fig. 4.42).

$^{38}$ However, this sounding was taken at 12Z and MCS-D3 intersected Lake Michigan at 14Z. The surface temperature at GRB warmed up about 2°C in this 2 hour period.
At 12Z/06, MCS-D3 was located in southwestern Wisconsin. There was relatively weak cooling at the surface associated with convectively driven downdrafts, as surface temperatures in northeastern Iowa dropped to 18°C (Fig. 4.52). Temperatures increased about 4–5°C across the southern edge of this convectively induced cold pool boundary in eastern Iowa. Surface mixing ratio values were only around 14 g kg\(^{-1}\) near and ahead of MCS-D3 (Fig. 4.52), about 4–6 g kg\(^{-1}\) lower than those located near MCS-P3 (Fig. 4.43). The buoy in the south-central portion of Lake Michigan recorded a 5-m air temperature of 19°C at 12Z/06. This temperature was 2–3°C lower than the surface temperatures upstream of Lake Michigan, but similar to those downstream of Lake Michigan (Fig. 4.52). In fact, there was a weak east–west-oriented synoptic-scale temperature gradient across Lake Michigan and the buoy air temperature was generally consistent with other surface temperatures nearby. Thus, Lake Michigan was not significantly colder than the surrounding land stations as MCS-D3 approached from the west.

At 12Z/06, MCS-D3 was situated in southwestern Wisconsin near the equatorward edge of a region of 0.25×10\(^{-12}\) Pa m\(^{-2}\) s\(^{-1}\) of 850-hPa Q-vector convergence and forcing for ascent (Fig. 4.53). This magnitude of 850-hPa Q-vector convergence over MCS-D3 was weak and similar to the magnitude found over MCS-P3 prior to crossing Lake Michigan (Fig. 4.44). This similarity again supports climatological and compositing results that revealed that the magnitude of 850-hPa Q-vector convergence directly over Lake Michigan-crossing MCSs did not discriminate between MCSs that persisted and dissipated upon crossing Lake Michigan (e.g., Fig. 2.14). The parent MCV associated with MCS-D3 was located over central Iowa at this time and Q-vector
convergence values over this MCV were around $0.5 \times 10^{-12} \text{ Pa m}^{-2} \text{ s}^{-1}$ (Figs. 4.49e and 4.53). Just as with MCS-P3, much of the 850-hPa Q-vector convergence directly over MCS-D3 resulted from warm advection (Figs. 4.44 and 4.53). There was also a 850-hPa Q-vector convergence maximum farther to the west along the North Dakota/Minnesota border where cyclonic vorticity advection made a large contribution to low-level forcing for ascent, but MCS-D3 was located well to the west of the region at 12Z/06 (Fig. 4.53).

By 16Z/06, MCS-D3 was located over Lake Michigan. At the surface broad southerly to south-southwesterly flow impinged on MCS-D3 (Fig. 4.54a). Additionally, MCS-D3 was situated under a 570-hPa 1000–500-hPa thickness ridge and in a region of 14–16 g kg$^{-1}$ 925-hPa mixing ratio values. These 925-hPa mixing ratio values were about 2 g kg$^{-1}$ less than those associated with MCS-P3 upon crossing Lake Michigan (Fig. 4.45a). Aloft, the 200-hPa flow remained zonal, consistent with the progressive synoptic-scale pattern (Johns 1993; Fig. 4.54b). MCS-D3 was located on the broad anticyclonic shear side of a 40 m s$^{-1}$ 200-hPa jet stream near the equatorward-entrance region of this jet. An anticyclonically curving 850-hPa LLJ was close to the MCS-D3 region, but MCS-D3 had pulled away from its core (Fig. 4.54b). The 850-hPa winds over MCS-D3 were 5–7.5 m s$^{-1}$ weaker than those over MCS-P3 at the time of lake crossing (Figs. 4.45b and 4.54b). A 400-hPa trough that contained $28 \times 10^{-5} \text{ s}^{-1}$ of absolute vorticity was positioned to the west of MCS-D3 (Fig. 4.54c). Recall that the MCS-P3 environment did not contain a similar 400-hPa short-wave trough (Fig. 4.45c). MCS-D3 remained in an environment of 500–1500 J kg$^{-1}$ of MUCAPE at 16Z/06 (Fig. 4.54d). However, there was only about 5 m s$^{-1}$ of 0–6-km shear, not enough for convective organization. The juxtaposition of MCS-P3 and MCS-D3 is interesting because MCS-D3 actually had more
MUCAPE available in lower Michigan, contrary to the general result of the overall climatology (Fig. 2.7b). However, the dynamical support for MCS-D3 was almost non-existent upon crossing Lake Michigan and resulted in the demise of MCS-D3. This result illustrates the caution that needs to be observed when only looking at climatological results and the reason why in-depth case studies are an appropriate complement to climatological and compositing results.

As MCS-D3 was progressing across southern Wisconsin and northern Illinois, 975-hPa potential temperatures dropped about 3 K between 12Z/06 and 14Z/06 (Fig. 4.55a). Again, recall that Marshan et al. (2011) showed that surface cold pools associated with nocturnal MCSs can have widely varying properties. Furthermore, this temperature decrease occurred in the presence of 10–12.5 m s$^{-1}$ of 0–3-km line-perpendicular shear. Between 12Z/06 and 14Z/06, MCS-D3 passed through cross-section line F–F’, and the convective cooling extended up to about 900-hPa (Fig. 4.55b$^{39}$). From an RKW standpoint, this relatively small and shallow amount of cooling results in a cold pool velocity ($C$) of about 10 m s$^{-1}$ (using $\bar{\theta}$=299 K, $\theta_{sfc}^*$=3 K, and $H$=1 km estimated from Figs. 4.52 and 4.55a,b). When this cold pool velocity is coupled with the environmental 0–3-km shear, the ratio of $C/\Delta U$ is approximately one. These values result in less vigorous and shallower ascent than was associated with MCS-P3 prior to crossing Lake Michigan (Figs. 4.46a,b and 4.55a,b). However, keep in mind that as MCS-D3 was moving along the Wisconsin/Illinois border, a strong nocturnal inversion was present at GRB at 12Z/06, and this convection was likely feeding off elevated MUCAPE above this inversion. Additionally, the surface temperature had cooled almost 10°C from the

\[39\] There was no AMDAR sounding available to confirm the cold pool depth.
daytime high and there was an extremely minimal amount of surface-based CAPE remaining in the atmosphere (Fig. 4.51), so it may have been difficult for the cold pool to lift surface parcels into the MCS updraft (Parker 2008). However, there was still identifiable cooling at the surface associated with MCS-D3 and Parker (2008) showed that simulated MCSs do not become truly elevated and exhibit bore-like behavior until this cooling abates. The convective cooling was likely larger with MCS-P3 than MCS-D3 due to the drier midlevel air and weaker nocturnal inversion associated with the MCS-P3 environment (Figs. 4.42 and 4.51). Regardless, of whether MCS-D3 was ingesting surface parcels upon intersecting Lake Michigan, recall that the air temperature over Lake Michigan was generally in the synoptic-scale temperature gradient, or perhaps only a couple of degrees cooler than upstream land stations (Fig. 4.52). Thus, MCS-D3 was not approaching any significant environmental horizontal buoyancy gradient.

As MCS-D3 passed over Juneau, WI (UNU) a 4°C surface temperature drop was recorded between 12Z/06 and 13Z/06 (Fig. 4.56a). This temperature drop is consistent with the values noted in the RUC analyses (Fig. 4.55a). During this same period, the dewpoint dropped about 1°C and the SLP rose 1.5–2 hPa. This SLP rise was smaller than the rise noted at RFD as MCS-P3 passed (Fig. 4.47a). At buoy 45007, the temperature did not change much as MCS-D3 passed overhead (Fig. 4.56b). Again, as with MCS-P3, the air over Lake Michigan was already nearly saturated (not shown) and very close to the Lake Michigan water temperature (Fig. 4.56b). Thus, there was not much potential for cooling at the buoy location. Also note that the near-surface inversion was very weak over Lake Michigan as the temperature only increased 0.8°C over the lowest 5 m immediately before MCS-D3 crossed (Fig. 4.56b). This inversion strength is
slightly stronger than that associated with MCS-P3 (Fig. 4.47b). However, the Lake Michigan water temperature was also very close to the surface air temperature on land, especially over lower Michigan (Figs. 4.52 and 4.56b). Thus, Lake Michigan likely played little role in the dissipation of MCS-D3, especially since the convection may have been elevated over the lake cold dome upon crossing Lake Michigan\textsuperscript{40}. MCS-D3 dissipated upon moving into a synoptic-scale environment that was less favorable than that associated with MCS-P3, especially in terms of dynamical support (Figs. 4.45a–d and 4.54a–d).

\textsuperscript{40} Although by 14Z/06, about 2°C of surface heating had occurred.
FIG. 4.1: Isochrones every 3 h of the leading 45-dBZ line of MCS-P1 (purple) overlaid upon severe storm reports of wind (blue), hail (green), and tornadoes (red) between 12Z/07 and 06Z/08 June 2008. Source: Severe Plot v.3.0 and UAlbany radar archive.
FIG. 4.2: NEXRAD level-III composite reflectivity data (in dBZ) at (a) 1503Z/07, (b) 1701Z/07, (c) 1900Z/07, (d) 2105Z/07, (e) 2304Z/07, (f) 0104Z/08, (g) 0302Z/08, and (h) 0503Z/08 June 2008. The arrows in panels (a), (d), and (g) illustrate the location of MCS-P1. Green and orange shading begin at 20 and 45dBZ, respectfully. Source: UAlbany radar archive.
FIG. 4.3: Water-vapor imagery from (a) 1515Z/07, (b) 1815Z/07, and (c) 2115Z/07 June 2008. The green arrows in panels (a) and (b) indicate the location of a first upper-level short-wave trough, while the white arrows in panels (b) and (c) indicate the location of a second short-wave trough. Source: RAP weather.
FIG. 4.4: (a) SLP (black lines every 4 hPa), 1000–500-hPa thickness (red dashed lines every 6 dam), and 925-hPa mixing ratio (shaded according to color bar in g kg$^{-1}$), (b) 200-hPa wind (shaded according to color bar in m s$^{-1}$), 200-hPa height (black lines every 12 dam), and 850-hPa wind (barbs greater than 12.5 m s$^{-1}$, where one long barb equals 5 m s$^{-1}$), and (c) CAPE (shaded according to color bar in J kg$^{-1}$) and 0–6-km shear (barbs in m s$^{-1}$, where one long barb equals 5 m s$^{-1}$) for 18Z/07 June 2008. The “P1” in each panel illustrates the location of MCS-P1. IR satellite imagery from 1810Z/07 June 2008 is inset into the lower-right corner of panel (a). Source: 20-km RUC analyses and UCAR MMM image archive.
FIG. 4.5: Atmospheric sounding of temperature (red line in °C), dewpoint (blue line in °C), and wind (black barbs in m s$^{-1}$, where one long barb equals 5 m s$^{-1}$) for 18Z/07 June 2008 from DVN. The inset in the upper-right corner shows the location of DVN. Source: UAlbany sounding archive.
FIG. 4.6: Manual surface analysis of SLP (black lines every 2 hPa), temperature (red dashed lines every 3°C), and mixing ratio equal to 18 g kg⁻¹ (green line) for 20Z/07 June 2008. The “P1” illustrates the location of MCS-P1. IR satellite imagery from 2010Z/07 June 2008 is inset into the lower-left corner of the image. Source: UAlbany surface archive and UCAR MMM image archive.
FIG. 4.7: As in Fig. 4.4 except for 00Z/08 June 2008. The “P1” in each panel illustrates the location of MCS-P1. IR satellite imagery from 0010Z/08 June 2008 is inset into the lower-right corner of panel (a). Source: 20-km RUC and UCAR MMM image archive.
FIG. 4.8: 850-hPa Q-vectors (arrows greater than $5 \times 10^{-7}$ Pa m$^{-1}$ s$^{-1}$), Q-vector convergence (shaded according to color bar in $\times 10^{-12}$ Pa m$^{-2}$ s$^{-1}$), temperature (green dashed lines every 3°C), and height (black lines every 3 dam) for 00Z/08 June 2008. A reference vector is located in the lower-right corner of the image. IR satellite imagery from 0010Z/08 June 2008 is inset into the upper-left corner of the image. Source: 1.0° GFS analysis and UCAR MMM image archive.
FIG. 4.9: (a) The 975-hPa previous 4-h potential temperature decrease (shaded according to color bar in K) and 0–3-km shear (barbs in m s\(^{-1}\), where one long barb equals 5 m s\(^{-1}\)) for 23Z/07 June 2008, (b) cross section A–A’ of the previous 4-h potential temperature decrease (shaded according to color bar in K), potential temperature (black lines every 1 K), and wind (barbs in m s\(^{-1}\), where one long barb equals 5 m s\(^{-1}\)) for 23Z/07 June 2008, and (c) AMDAR sounding of temperature (red line in °C), dewpoint (blue line in °C), and wind (red barbs in m s\(^{-1}\), where one long barb equals 5 m s\(^{-1}\)) for 2317Z/07 June 2008. NEXRAD level-III composite radar imagery for 19Z/07 and 23Z/07 June 2008 are inset into the lower-left and right corners of (a), respectively, with arrows indicating MCS-P1 location. The red hexagon in (a) indicates the AMDAR sounding location. Source: 20-km RUC analyses, UAlbany radar archive, and AMDAR.
FIG. 4.10: (a) Meteogram from 15Z/07 to 02Z/08 June 2008 of surface temperature (red line in °C), surface dewpoint (green line in °C), SLP (blue line in hPa), and current weather symbols at MSN. (b) Meteogram from 20Z/07 to 03Z/08 June 2008 of 5-m air temperature (red line in °C), surface water temperature (magenta line in °C), and SLP (blue line in hPa) at buoy 45007, and (c) the locations of the meteograms in (a) and (b). Source: UAlbany surface archive and NDBC.
FIG. 4.11: As in Fig. 4.1 except for between 06Z/04 and 21Z/04 June 2005 with MCS-D1. Source: Severe Plot v.3.0 and UAlbany radar archive.
FIG 4.12: NOWrad composite reflectivity data (in dBZ) at (a) 14Z/04, (b) 16Z/04, (c) 18Z/04, (d) 19Z/04, (e) 20Z/04, (f) 21Z/04, and 22Z/04 June 2005. The arrows in panels (a), (c), and (e) illustrate the location of MCS-D1. Green and orange shading begin at 20 and 45 dBZ, respectively. Source: UCAR MMM image archive.
FIG. 4.13: (a) SLP (black lines every 4 hPa), 1000–500-hPa thickness (red dashed lines every 6 dam), and 925-hPa mixing ratio (shaded according to color bar in g kg$^{-1}$), (b) 200-hPa wind (shaded according to color bar in m s$^{-1}$), 200-hPa height (black lines every 12 dam), and 850-hPa wind (barbs greater than 12.5 m s$^{-1}$, where one long barb equals 5 m s$^{-1}$), (c) 400-hPa absolute vorticity (shaded according to color bar in $\times 10^{-5}$ s$^{-1}$), 400-hPa height (black lines every 6 dam), and 400-hPa wind (barbs in m s$^{-1}$, where one long barb equals 5 m s$^{-1}$), and (d) CAPE (shaded according to color bar in J kg$^{-1}$) and 0–6-km shear (barbs in m s$^{-1}$, where one long barb equals 5 m s$^{-1}$) for 15Z/04 June 2005. The “D1” in each panel illustrates the location of MCS-D1. IR satellite imagery from 1510Z/07 June 2005 is inset into the lower-right corner of panel (a). Source: 20-km RUC analyses and UCAR MMM image archive.
FIG. 4.14: As in Fig. 4.5 except for a RUC sounding at 40.96°N latitude and 87.62°W longitude for 18Z/04 June 2005. The inset in the upper-right corner shows the location of the sounding. Source: 20-km RUC analysis.
FIG. 4.15: Manual surface analysis of SLP (black lines every 2 hPa), temperature (red dashed lines every 3°C), and mixing ratio equal to 16 g kg$^{-1}$ (green line) for 18Z/07 June 2005. The “D1” illustrates the location of MCS-D1. IR satellite imagery from 1810Z/07 June 2005 is inset into the upper-left corner of the image. Source: UAlbany surface archive and UCAR MMM image archive.
FIG. 4.16: As in Fig. 4.8 except for 18Z/04 June 2005. The “D1” illustrates the location of MCS-D1. IR satellite imagery from 1810Z/04 June 2005 is inset into the upper-left corner of the image. Source: 1.0° GFS analysis and UCAR MMM image archive.
FIG. 4.17: As in Fig. 4.13 except for 21Z/04 June 2005. The “D1” in each panel illustrates the location of MCS-D1. IR satellite imagery from 2110Z/04 June 2005 is inset in the lower-right corner of panel (a). Source: 20-km RUC and UCAR MMM image archive.
Fig. 4.18: (a) The 975-hPa previous 2-h potential temperature decrease (shaded according to color bar in K) and 0–3-km shear (barbs in m s\(^{-1}\), where one long barb equals 5 m s\(^{-1}\)) for 19Z/04 June 2005, (b) cross section B–B’ of the previous 2-h potential temperature decrease (shaded according to color bar in K), potential temperature (black lines every 1 K), and wind (barbs in m s\(^{-1}\), where one long barb equals 5 m s\(^{-1}\)) for 19Z/04 June 2005, and (c) AMDAR sounding of temperature (red line in °C), dewpoint (blue line in °C), and wind (red barbs in m s\(^{-1}\), where one long barb equals 5 m s\(^{-1}\)) for 1955Z/04 June 2005. NOWrad composite radar imagery for 17Z/04 and 19Z/04 June 2005 are inset into the lower-left and right corners of (a), respectively, with arrows indicating MCS-D1 location. The red hexagon in (a) indicates the AMDAR sounding location. Source: 20-km RUC analyses, UAlbany radar archive, and AMDAR.
FIG. 4.19: (a) As in Fig. 4.10a except at ARR from 12Z/04 to 23Z/04 June 2005, (b) as in Fig. 4.10b except for from 12Z/04 to 23Z/04 June 2005, and (c) the locations of the meteograms in (a) and (b). Source: UAlbany surface archive and NDBC.
FIG. 4.20: As in Fig. 4.1 except for between 10Z/18 and 03Z/19 June 2010 with MCS-P2. Source: Severe Plot v.3.0 and UAlbany radar archive.
FIG. 4.21: As in Fig. 4.2 except for (a) 14Z/18, (b) 16Z/18, (c) 18Z/18, (d) 20Z/18, (e) 22Z/18, (f) 00Z/19, and (g) 02Z/19 June 2010. The arrows in panels (a), (c), and (e) illustrate the location of MCS-P2. Source: UAlbany radar archive.
FIG. 4.22: As in Fig. 4.13 except for 16Z/18 June 2010. The “P2” in each panel illustrates the location of MCS-P2. IR satellite from 1610Z/18 June 2010 is inset in the lower-right corner of panel (a). Source: 20-km RUC and UCAR MMM image archive.
FIG. 4.23: As in Fig. 4.5 except for a sounding from GRB at 18Z/18 June 2010. The inset in the upper-right corner shows the location of the sounding. Source: UAlbany sounding archive.
FIG. 4.24: As in Fig. 4.6 except for 20Z/18 June 2010. The “P2” illustrates the location of MCS-P2. IR satellite imagery from 2010Z/18 June 2010 is inset into the upper-left corner of the image. Source: UAlbany surface archive and UCAR MMM image archive.
FIG. 4.25: As in Fig. 4.8 except for 18Z/18 June 2010. The “P2” illustrates the location of MCS-P2. IR satellite imagery from 1810Z/18 June 2010 is inset into the upper-left corner of the image. Source: 1.0° GFS analysis and UCAR MMM image archive.
FIG. 4.26: As in Fig. 4.13 except for 22Z/18 June 2010. The “P2” in each panel illustrates the location of MCS-P2. IR satellite from 2210Z/18 June 2010 is inset in the lower-right corner of panel (a). Source: 20-km RUC and UCAR MMM image archive.
FIG. 4.27: (a) The 975-hPa previous 2-h potential temperature decrease (shaded according to color bar in K) and 0–3-km shear (barbs in m s$^{-1}$, where one long barb equals 5 m s$^{-1}$) for 22Z/18 June 2010, (b) cross section C–C’ of the previous 2-h potential temperature decrease (shaded according to color bar in K), potential temperature (black lines every 1 K), and wind (barbs in m s$^{-1}$, where one long barb equals 5 m s$^{-1}$) for 22Z/18 June 2010, and (c) AMDAR sounding of temperature (red line in °C), dewpoint (blue line in °C), and wind (red barbs in m s$^{-1}$, where one long barb equals 5 m s$^{-1}$) for 2208Z/18 June 2010. NEXRAD level-III composite radar imagery for 20Z/18 and 22Z/18 June 2010 are inset into the lower-left and right corners of (a), respectively, with arrows indicating MCS-P2 location. The red hexagon in (a) indicates the AMDAR sounding location. Source: 20-km RUC analyses, UAlbany radar archive, and AMDAR.
FIG. 4.28: (a) As in Fig. 4.10a except at RFD from 16Z/18 to 23Z/18 June 2010, (b) as in Fig. 4.10b except for from 16Z/18 to 23Z/18 June 2010, and (c) the locations of the meteograms in (a) and (b). Source: UAlbany surface archive and NDBC.
FIG. 4.29: (a) 10-m convergence (shaded according to color bar in $\times 10^{-5}$ s$^{-1}$) and 10-m wind (barbs in m s$^{-1}$, where one long barb equals 5 m s$^{-1}$) at 20Z/18 June 2010 and (b) as in Fig. 4.10 except at LWA from 16Z/18 to 23Z/18 June 2010. NEXRAD level-III composite radar images for 2130Z/18 and 22Z/18 June 2010 are inset into the upper-left and right corners of (a), respectively, with arrows indicating MCS-P2 location. The red hexagon in (a) indicates the location of LWA. Source: 20-km RUC analyses, UAlbany radar archive, and UAlbany surface archive.
FIG. 4.30: As in Fig. 4.1 except for between 00Z/24 and 16Z/24 June 2003 with MCS-D2. Source: Severe Plot v.3.0 and UAlbany radar archive.
FIG. 4.31: As in Fig. 4.12 except for (a) 02Z/24, (b) 04Z/24, (c) 06Z/24, (d) 08Z/24, (e) 10Z/24, (f) 12Z/24, (g) 14Z/24, and (h) 16Z/24 June 2003. The arrows in panels (a), (d), and (g) illustrate the location of MCS-D2. Source: UCAR MMM image archive.
FIG. 4.32: As in Fig. 4.13 except for 08Z/24 June 2003. The “D2” in each panel illustrates the location of MCS-D2. IR satellite imagery from 0808Z/24 June 2003 is inset in the lower-right corner of panel (a). Source: 20-km RUC and BAMEX field catalog.
FIG. 4.33: As in Fig. 4.5 except for a sounding from GRB at 12Z/24 June 2003. The inset in the upper-right corner shows the location of the sounding. Source: UAlbany sounding archive.
FIG. 4.34: As in Fig. 4.6 except for 12Z/24 June 2003. The “D2” illustrates the location of MCS-D2. IR satellite imagery from 1209Z/24 June 2003 is inset into the upper-left corner of the image. Source: UAlbany surface archive and BAMEX field catalog.
FIG. 4.35: As in Fig. 4.8 except for 12Z/24 June 2003. The “D2” illustrates the location of MCS-D2. IR satellite imagery from 1209Z/24 June 2003 is inset into the upper-left corner of the image. Source: 1.0° GFS analysis and BAMEX field catalog.
FIG. 4.36: As in Fig. 4.13 except for 14Z/24 June 2003. The “D2” in each panel illustrates the location of MCS-D2. IR satellite imagery from 1409Z/24 June 2003 is inset in the lower-right corner of panel (a). Source: 20-km RUC and BAMEX field catalog.
FIG. 4.37: (a) The 975-hPa previous 2-h potential temperature decrease (shaded according to color bar in K) and 0–3-km shear (barbs in m s$^{-1}$, where one long barb equals 5 m s$^{-1}$) for 13Z/24 June 2003 and (b) cross section D–D’ of the previous 2-h potential temperature decrease (shaded according to color bar in K), potential temperature (black lines every 1 K), and wind (barbs in m s$^{-1}$, where one long barb equals 5 m s$^{-1}$) for 13Z/24 June 2003. NOWrad composite radar images for 11Z/24 and 13Z/24 June 2003 are inset into the lower-left and right corners of (a), respectively, with arrows indicating MCS-D1 location. Source: 20-km RUC analyses and UAlbany radar archive.
FIG. 4.38: (a) As in Fig. 4.10a except at OSH from 09Z/24 to 16Z/24 June 2003, (b) as in Fig. 4.10b except for from 09Z/24 to 16Z/24 June 2003 and (c) the locations of the meteograms in (a) and (b). Source: UAlbany surface archive and NDBC.
FIG. 4.39: As in Fig. 4.1 except for between 00Z/22 and 16Z/22 August 2007 with MCS-P3. Source: Severe Plot v.3.0 and UAlbany radar archive.
FIG. 4.40: As in Fig. 4.12 except for (a) 00Z/22, (b) 02Z/22, (c) 04Z/22, (d) 06Z/22, (e) 08Z/22, (f) 10Z/22, (g) 12Z/22, and (h) 14Z/22 August 2007. The arrows in panels (a), (d), and (g) illustrate the location of MCS-P3. Source: UCAR MMM image archive.
FIG. 4.41: As in Fig. 4.13 except for 04Z/22 August 2007. The “P3” in each panel illustrates the location of MCS-P3. IR satellite imagery from 0410Z/22 August 2007 is inset in the lower-right corner of panel (a). Source: 20-km RUC and UCAR MMM image archive.
FIG. 4.42: As in Fig. 4.5 except for a RUC sounding at 42.23°N latitude and 89.03°W longitude for 06Z/22 August 2007. The inset in the upper-right corner shows the location of the sounding. Source: 20-km RUC analysis.
FIG. 4.43: As in Fig. 4.6 except for 06Z/22 August 2007. The “P3” illustrates the location of MCS-P3. IR satellite imagery from 0610Z/22 August 2007 is inset into the lower-right corner of the image. Source: UAlbany surface archive and UCAR MMM image archive.
FIG. 4.44: As in Fig. 4.8 except for 06Z/22 August 2007. The “P3” illustrates the location of MCS-P3. IR satellite imagery from 0610Z/22 August 2007 is inset into the lower-left corner of the image. Source: 1.0° GFS analysis and UCAR MMM image archive.
FIG. 4.45: As in Fig. 4.13 except for 10Z/22 August 2007. The “P3” in each panel illustrates the location of MCS-P3. IR satellite imagery from 1010Z/22 August 2007 is inset in the lower-right corner of panel (a). Source: 20-km RUC and UCAR MMM image archive.
FIG. 4.46: (a) The 975-hPa previous 3-h potential temperature decrease (shaded according to color bar in K) and 0–3-km shear (barbs in m s\(^{-1}\), where one long barb equals 5 m s\(^{-1}\)) for 09Z/22 August 2007, (b) cross section E–E’ of the previous 3-h potential temperature decrease (shaded according to color bar in K), potential temperature (black lines every 1 K), and wind (barbs in m s\(^{-1}\), where one long barb equals 5 m s\(^{-1}\)) for 09Z/22 August 2007, and (c) AMDAR sounding of temperature (red line in °C), dewpoint (blue line in °C), and wind (red barbs in m s\(^{-1}\), where one long barb equals 5 m s\(^{-1}\)) for 1033Z/22 August 2007. NOWrad composite radar imagery for 06Z/22 and 09Z/22 August 2007 are inset into the lower-left and right corners of (a), respectively, with arrows indicating MCS-P3 location. The red hexagon in (a) indicates the AMDAR sounding location. Source: 20-km RUC analyses, UAlbany radar archive, and AMDAR.
FIG. 4.47: (a) As in Fig. 4.10a except at RFD from 06Z/22 to 13Z/22 August 2007, (b) as in Fig. 4.10b except for from 06Z/22 to 13Z/22 August 2007, and (c) the locations of the meteograms in (a) and (b). Source: UAlbany surface archive and NDBC.
FIG. 4.48: As in Fig. 4.1 except for between 00Z/06 and 16Z/06 July 2003 with MCS-D3. Source: Severe Plot v.3.0 and UAlbany radar archive.
FIG. 4.49: As in Fig. 4.12 except for (a) 04Z/06, (b) 06Z/06, (c) 08Z/06, (d) 10Z/06, (e) 12Z/06, (f) 14Z/06, (g) 16Z/06, and (h) 18Z/06 July 2003. The arrows in panels (a), (d), and (g) illustrate the location of MCS-D3. Source: UCAR MMM image archive.
FIG. 4.50: As in Fig. 4.13 except for 10Z/06 July 2003. The “D3” in each panel illustrates the location of MCS-D3. IR satellite imagery from 1009Z/06 July 2003 is inset in the lower-right corner of panel (a). Source: 20-km RUC and BAMEX field catalog.
FIG. 4.51: As in Fig. 4.5 except for a sounding from GRB at 12Z/06 July 2003. The inset in the upper-right corner shows the location of the sounding. Source: UAlbany sounding archive.
FIG. 4.52: As in Fig. 4.15 except for 12Z/06 July 2003. The “D3” illustrates the location of MCS-D3. IR satellite imagery from 1209Z/06 July 2003 is inset into the lower-left corner of the image. Source: UAlbany surface archive and BAMEX field catalog.
FIG. 4.53: As in Fig. 4.8 except for 12Z/06 July 2003. The “D3” illustrates the location of MCS-D3. IR satellite imagery from 1209Z/06 July 2003 is inset into the lower-left corner of the image. Source: 1.0° GFS analysis and BAMEX field catalog.
FIG. 4.54: As in Fig. 4.13 except for 16Z/06 July 2003. The “D3” in each panel illustrates the location of MCS-D3. IR satellite imagery from 1603Z/06 July 2003 is inset in the lower-right corner of panel (a). Source: 20-km RUC and BAMEX field catalog.
FIG. 4.55: a) The 975-hPa previous 2-h potential temperature decrease (shaded according to color bar in K) and 0–3-km shear (barbs in m s$^{-1}$, where one long barb equals 5 m s$^{-1}$) for 14Z/06 July 2003 and (b) cross section F–F’ of the previous 2-h potential temperature decrease (shaded according to color bar in K), potential temperature (black lines every 1 K), and wind (barbs in m s$^{-1}$, where one long barb equals 5 m s$^{-1}$) for 14Z/06 July 2003. NOWrad composite radar imagery for 12Z/06 and 14Z/06 July 2003 are inset into the lower-left and right corners of (a), respectively, with arrows indicating MCS-D3 location. Source: 20-km RUC analyses and UAlbany radar archive.
FIG. 4.56: (a) As in Fig. 4.10a except at UNU from 10Z/06 to 17Z/06 July 2003, (b) as in Fig. 4.10b except for from 10Z/06 to 17Z/06 July 2003, and (c) the locations of the meteograms in (a) and (b). Source: UAlbany surface archive and NDBC.
5. WRF Modeling Case Studies of Lake Michigan-Crossing MCSs

5.1 Purpose

The purpose of this chapter is to examine simulated data from two of the case studies presented in chapter 4. While the observational case study results shed light on many important environmental and lake features associated with persisting and dissipating Lake Michigan-crossing MCSs, the resolution of 20-km RUC and observational data constraints make direct attribution of persistence and dissipation to the synoptic-scale environment and/or Lake Michigan difficult. A numerical model allows for a more concrete examination of the relative roles of the synoptic-scale environment and the Lake Michigan marine layer to MCS persistence because Lake Michigan can be removed to determine how the evolution of a Lake Michigan-crossing MCS is altered when the shallow, cool marine layer is no longer in the MCS path. Thus, the main goal of this chapter will be to understand the relative roles of the synoptic-scale environment and the cool Lake Michigan marine layer to the persistence of MCSs.

5.2 Data and Methods

In order to simulate Lake Michigan-crossing MCSs, the WRF v.3.0 (e.g., Skamarock et al. 2005; Skamarock et al. 2008) was utilized. The WRF model has been employed by multiple authors in the past to successfully simulate convective activity (e.g., Martin and Johnson 2008; Schumacher et al. 2011). The two cases studies simulated were the 7–8 June 2008 MCS that persisted (MCS-P1) and the 4 June 2005 MCS that dissipated (MCS-D1; see chapter 4 for details on these case studies) upon crossing Lake Michigan. These two case studies were chosen for simulation because
realistic model runs were achieved for each MCS. Additionally, these two cases can be compared and contrasted with each other because both occurred at about the same time of day, same time of year, and under similar synoptic-scale patterns. The WRF was formulated with the Advanced Research (ARW) core (Skamarock et al. 2005), the Noah land surface model (Skamarock et al. 2005), and utilized the 32-km (NARR) for initial and boundary conditions. The NARR was employed for both cases over the 20-km RUC for initial and boundary conditions because multiple RUC hybrid-isentropic analysis files were either missing or corrupted from the ARM and NOMADS archives for the 4 June 2005 case.\(^{41}\)

Both of these cases were simulated using a 500×350 horizontal grid point domain centered on the upper Midwest and western Great Lakes (Fig. 5.1a). This particular domain was chosen since both of the MCSs considered in this section approached Lake Michigan from the west. Thus, a sizable region upstream of Lake Michigan was needed to give each MCS time to spin up in a realistic manner prior to intersecting Lake Michigan from the west. Additionally, the horizontal grid spacing in the WRF simulations was 4 km, as this grid spacing has been found to be sufficient to represent MCSs without the need for convective parameterization (e.g., Weisman et al. 1997). All WRF simulations were performed with 46 vertical levels, 13 of which were packed more tightly into the boundary layer in an attempt to realistically replicate the shallow Lake Michigan cold dome. However, it is important to remember that the WRF model is dependent upon the initial conditions (e.g., Weisman et al. 2008), and the NARR only has meteorological data every 25 hPa in the vertical. Other model options utilized included

\(^{41}\) The utilization of the NARR for boundary and initial conditions did produce realistic simulations of MCS-P1 and MCS-D1.
the MYJ boundary-layer scheme (Janjić 2001), WSM6 6-class graupel scheme (Hong and Lim 2006), Moin–Obukhov (Janjić Eta) surface-layer scheme (Janjić 1996, 2002), and explicit convection. Multiple boundary-layer and physics scheme formulations were tested in the process of producing the simulations in this chapter but the above combination was found to represent MCS-P1 and MCS-D1 in the most realistic fashion, and have been utilized by many authors in the past (e.g., Weisman 2008). These two MCSs crossed Lake Michigan between 20Z and 23Z. Thus, all simulations were initialized at 12Z, and run forward for 18 h. This lead time allowed each simulated MCS spin-up time before crossing Lake Michigan.

For each MCS, two different sets of simulations will be presented. The first is a control run that utilized all of the aforementioned WRF-model specifications (Fig. 5.1a). The second is a simulation that is similar in every way to the control run, except Lake Michigan was replaced by land (hereafter the noLM simulation; Fig. 5.1b). Very few authors have attempted to alter surface characteristics and note the effect on MCSs in real-data simulations (e.g., Srock 2011). In the noLM simulation, all of the model grid points that comprised Lake Michigan were turned into land at the initial model time with characteristics similar to surrounding land grid points. This transformation was accomplished by choosing a representative land point immediately upstream of Lake Michigan at the model initial time and imparting the surface properties from this point to all Lake Michigan water points (Fig. 5.1b). After the initial time, model characteristics were allowed to evolve through normal atmospheric processes. The noLM simulations were re-run numerous times, each time choosing a different grid point that was mapped to Lake Michigan. The exact point chosen to map onto Lake Michigan had a negligible
effect on the overall results presented as long as the point chosen was representative of
the near-surface land environment in proximity to Lake Michigan. Surface characteristics
that were altered to morph these water points into land included surface temperature, soil
characteristics, land-use index, and vegetation category.

Note that atmospheric conditions were not explicitly changed in the model. By
only changing land conditions, the WRF model did not require rebalancing. Given that
the Lake Michigan cold dome was so shallow, the cool inversion in the noLM
simulations was quickly scoured out by mixing effects within a couple of hours. After
these first two hours, the land over what was Lake Michigan in the noLM simulations
featured near-surface temperature, moisture, and wind characteristics that were consistent
with the nearby synoptic-scale environment. By running a control and noLM simulation
for both MCS-P1 and MCS-D1, a more definitive answer can be given as to the precise
role of Lake Michigan in the persistence and dissipation of these two MCSs. In all
simulations, the forecast synoptic-scale environment looks very similar to the analyzed
synoptic-scale environment identified in chapter 4 (except for one notable exception that
will be discussed later in this chapter). Thus, the purpose of this chapter will be to
compare and contrast reflectivity trends between the control and noLM simulations.
From these reflectivity trends, the relative importance of the synoptic-scale and Lake
Michigan to convective persistence can be inferred.

5.3 Simulated Persisting MCS of 7–8 June 2008 (MCS-P1)

Around 20Z/07 June 2008, MCS-P1 was in the process of becoming better
organized in south-central Wisconsin and was approximately 2–3 h from crossing Lake
Michigan (Figs. 4.2c,d). The 8-h forecast from the 12Z/07 WRF control run valid at 20Z/07 shows that the surface air temperature over most of Lake Michigan was between 11°C and 14°C (Fig. 5.2a). These simulated air temperatures agree nicely with buoy 45007 in the south-central portion of Lake Michigan which recorded an air temperature of 12°C at the same time (Fig. 4.6). On the western shore of Lake Michigan, surface simulated temperatures in the control run were approximately 29°C (Fig. 5.2a) again very similar to actual observations at the same time (Fig. 4.6). The noLM simulation forecast also valid at 20Z/07 reveals a very different near-surface environment over Lake Michigan (Fig. 5.2b). The surface air temperature in the noLM simulation over the region where Lake Michigan became land was around 29°C, very similar to the surface temperatures that appeared in eastern Wisconsin and western Michigan. Note that the surface air temperatures over the other Great Lakes were unchanged between the control and noLM simulations (Figs. 5.2a,b).

A WRF 8-h forecast sounding comparison valid at 20Z/07 for both the control and noLM runs shows the vertical environmental differences at the same location in south-central Lake Michigan (Figs. 5.2a,b and 5.3). The control and noLM profiles appear largely similar above 850 hPa (Fig. 5.3), consistent with both simulations initially being identical except for the Lake Michigan surface modification. Near the surface, the control simulation reveals a strong inversion associated with the cool Lake Michigan boundary layer. The surface temperature and dewpoint were around 13°C with a shallow, intense inversion where temperature increased upward about 12°C over the lowest 15–20 hPa of the sounding. There was 0 J kg$^{-1}$ of surface-based CAPE in the control sounding, and about 2500 J kg$^{-1}$ of MUCAPE above the surface inversion. The
noLM sounding contained a well mixed boundary layer, with a dry adiabatic layer extending up to about 875 hPa (Fig. 5.3), consistent with soundings from surrounding land locations (not shown). The surface temperature (dewpoint) in the noLM sounding was about 30°C (23°C), resulting in over 4500 J kg\(^{-1}\) of surface-based and MUCAPE\(^{42}\). The land modification in the noLM simulation did not alter a significant portion of the vertical profile, but did succeed in drastically increasing the atmospheric instability. The surface winds in the control simulation were backed compared to the noLM simulation (Fig. 5.3), consistent with a stable, decoupled boundary layer.

At 20Z/07 June 2008, the 8-h simulated reflectivity forecasts from both the control and noLM runs appear nearly identical (Figs. 5.4a,b). In both runs, MCS-P1 was beginning to organize in south-central Wisconsin in a very similar location to real MCS-P1 at the same time (Figs. 5.4a,b). Two hours later at 22Z/07, the 10-h simulated reflectivity forecasts for both runs again appear qualitatively similar (Figs. 5.4c,d). The only notable difference is to the northeast of MCS-P1 along the Lake Michigan/lower Michigan border, where convective cells in the noLM simulation were slightly more intense when compared to the control run (Figs. 5.4c,d). Again, both simulations replicated reality except for a region of convection in eastern Illinois, which in the actual environment was only a single supercell (not shown). By 00Z/08, simulated MCS-P1 in both the control and noLM runs was crossing Lake Michigan, and these two simulated MCSs look largely similar (Figs. 5.4e,f). Recall that as MCS-P1 crossed Lake Michigan on 7 June 2008, there was a strong near-surface inversion over the lake (Figs. 4.10b and 5.3). The noLM simulated reflectivity associated with MCS-P1 reveals a slightly larger

\(^{42}\) This robust amount of MUCAPE is similar to the amount of MUCAPE over northern Illinois in both the WRF control run and the RUC analysis (Figs. 4.4c and 4.7c).
region of simulated echoes above 50 dBZ (Figs. 5.4e,f), but the largest difference in the simulated reflectivity between the two runs remains to the east and northeast of simulated MCS-P1. Again, the two runs continued to do an excellent job at representing the actual reflectivity pattern at 00Z/08 (Fig. 5.4e), and captured the convection that formed to the west of MCS-P1 (Figs. 4.2e,f). Finally, the 14-h simulated reflectivity forecasts for the two runs reveal that MCS-P1 persisted upon crossing Lake Michigan in both simulations (Figs. 5.4g,h). However, simulated MCS-P1 in the noLM run was more intense. This intensity increase in the noLM simulation primary resulted from MCS-P1 merging with the region of increased convection that formed to the northeast of MCS-P1 beginning at 22Z/07 (Fig. 5.4d). Although there was adequate MUCAPE above the Lake Michigan stable cold dome in the control run (and in reality) to allow MCS-P1 to persist, there was even more MUCAPE in the noLM run (Fig. 5.3). This increased amount of MUCAPE in the noLM run allowed the relatively weak region of convection that formed to the northeast of MCS-P1 in reality and the control run, to intensify in the noLM run (Figs. 5.4c,d,f) resulting in a more robust MCS-P1 in the noLM run after crossing Lake Michigan.

Both the control and noLM versions of MCS-P1 had crossed Lake Michigan fifteen hours into each simulation. An examination of the difference in total accumulated precipitation over those 15 h reveals that in certain locations, MCS-P1 produced up to 50 mm more precipitation in the noLM simulation when compared to the control simulation (Fig. 5.5). A swath of this increased precipitation occurred from west to east across Wisconsin primarily result from slight phase differences in the location of simulated precipitation echoes between the two simulations.
central Lake Michigan as MCS-P1 was slightly more intense in the noLM simulation than in the control simulation (Figs. 5.4e,f). However, most of the precipitation difference occurred on the downstream side of Lake Michigan, where in the noLM simulation, simulated MCS-P1 merged with a more robust region of convection to the northeast (Figs. 5.4f,h and 5.5). These simulations reveal that given the synoptic-scale and Lake Michigan conditions on 7–8 June 2008, MCS-P1 was able to persist upon crossing Lake Michigan. However, just because MCS-P1 was able to persist over the stable inversion associated with Lake Michigan, does not mean that Lake Michigan had no effect on MCS-P1. When Lake Michigan was removed in the noLM run, MCS-P1 was more intense, primarily due to the increase in MUCAPE in an already favorable synoptic-scale environment. Thus, MCS-P1 had the potential to be even more robust if Lake Michigan had not been situated in its way. Although real MCS-P1 was able to feed off of elevated MUCAPE about the stable Lake Michigan cold dome, this shallow stable layer “cut off” a portion of the atmosphere from contributing to the robust MUCAPE environment on 7–8 June 2008. When the inversion was removed in the noLM simulation, a deep, relatively moist mixed layer resulted, consistent with surrounding land locations that resulted in almost double the amount of environmental MUCAPE.

5.4 Simulated Dissipating MCS of 4–5 June 2005 (MCS-P2)

Around 20Z/04 June 2005, MCS-D1 was beginning to dissipate upon intersecting the southwestern coast of Lake Michigan (Fig 4.12e). The 8-h control simulation forecast of MCS-D1 valid at this time shows surface temperatures between 11°C and 14°C over Lake Michigan. In reality, buoy 45007 in the south-central portion of Lake
Michigan recorded an air temperature of 14°C, in agreement with the WRF control simulation (Figs. 4.15 and 5.6). The noLM simulation also valid at 20Z/04 does not reveal any atmospheric imprint of Lake Michigan (Fig. 5.6b). In fact, the surface temperatures over the region where Lake Michigan was removed appear to be consistent with the synoptic-scale gradient. The noLM and actual temperatures both range from 26 to 29°C in eastern Wisconsin and western Michigan (Figs. 4.15 and 5.6b).

WRF environmental soundings from 20Z/04 near MCS-D1 in the south-central portion of Lake Michigan reveal a similar story as with the MCS-P1 simulations. The control and noLM soundings look nearly identical above 850 hPa (Fig. 5.7). Near the surface, the control sounding has a strong near-surface inversion where the temperature increases from 12°C to 21°C over the lowest 15–20 hPa. There was no surface-based CAPE in this control sounding and above this shallow Lake Michigan-induced inversion, approximately 600 J kg⁻¹ of MUCAPE was present. The noLM sounding contains a dry-adiabatic layer extending up to about 860 hPa with a surface temperature (dewpoint) of 28°C (20°C). The temperature and moisture profiles associated with the noLM sounding yield about 1150 J kg⁻¹ of surface-based and MUCAPE at the sounding location (Fig. 5.7). This is much smaller than the amount of MUCAPE associated with the noLM simulation of MCS-P1 (Fig. 5.3) but still almost double that associated with the control simulation of MCS-D1 (Fig. 5.7). The control run sounding contains a more backed surface wind consistent with a decoupled, stable surface layer.

At 20Z/04 June 2005, the 8-h reflectivity forecasts for both the control and noLM simulations look identical (Figs. 5.8a,b). In both model versions, simulated MCS-D2 is comparable to reality, and is situated in southeastern Wisconsin. Additionally at 20Z/04,
both model runs are a bit slow as real MCS-D1 had already begun to cross Lake Michigan and weaken (Fig. 5.8a). Two hours later at 22Z/04, MCS-D1 started crossing Lake Michigan in both simulations. The northern portion of MCS-D1 in the control run weakened almost immediately upon intersecting Lake Michigan, while the noLM MCS-D1 had a stronger northern extent (Figs. 5.8c,d). By 00Z/05, MCS-D1 had dissipated in reality, but simulated MCS-D1 was still progressing across Lake Michigan in both simulations (Fig. 5.8e,f). The noLM simulated MCS-D1 remained stronger than the control version of MCS-D1, especially on the northern periphery. Finally, at 03Z/05, both the control and noLM simulated versions of MCS-D1 began to weaken with the noLM MCS-D1 remaining slightly stronger than the control version (Figs. 5.8g,h).

In order to understand why the simulated versions of MCS-D1 persisted upon crossing lake Michigan, 925-hPa equivalent potential temperature ($\theta_e$) images are presented for the 11-h forecast of the control run, the 11-h forecast of the noLM run, and the RUC analysis all valid at 23Z/07. Both the 11-h control and noLM runs reveal 341K 925-hPa $\theta_e$ values in southwestern lower Michigan (Figs. 5.9a,b). However, in reality, the RUC analysis shows 329–335 K 925-hPa $\theta_e$ values in the same region (Fig. 5.9c). Before MCS-D1 progressed across Lake Michigan, a region of mid-level cloudiness was present in western-lower Michigan for a few hours, and this cloudiness likely reduced the instability in the real atmosphere (Fig. 5.9c). The WRF simulations likely did not identify this region of cloudiness, and thus, both of the simulated versions of MCS-D1 persisted upon crossing Lake Michigan. Other than the additional downstream instability, the synoptic-scale environments were almost identical between the WRF forecasts and the RUC analyses (not shown).
Fifteen-hour accumulated simulated precipitation differences between 12Z/04 and 03Z/05 reveal that the noLM run produced 40–50 mm more precipitation coincident with the northern portion of simulated MCS-D1 (Fig. 5.10). The fact that both WRF simulations allowed MCS-D1 to persist upon crossing Lake Michigan was actually a fortuitous mistake, allowing the relative contributions of synoptic-scale forcing and the Lake Michigan marine layer to the dissipation of MCS-D1 to be parsed. The added downstream instability in both simulations of MCS-D1 allowed for persistence into lower Michigan, until this region of increased instability was eventually outrun (Figs. 5.9g,h). This late model dissipation was likely the result of the synoptic-scale environment since the convective environment became unfavorable over lower Michigan (Figs. 4.17b,d), and complete dissipation over land could not be associated with Lake Michigan.

However, Lake Michigan also likely played a smaller but still significant role in the dissipation of MCS-D1. The control run of MCS-D1 produced less convection than the noLM run, given the smaller amount of environmental MUCAPE (Figs. 5.9e,f). These convective differences were especially notable farther north over central Lake Michigan and resulted in a significant simulated precipitation difference (Fig. 5.10). Given the southerly flow at the surface out ahead of MCS-D1 (e.g., Figs. 5.8a,b), air over Lake Michigan had a longer residence time at farther north locations. Thus, the cold air may have been deeper farther to the north. In total, the unfavorable synoptic-scale eventually limited MCS-D1 convection in both simulations. Before dissipation could occur, the noLM simulation contained convection that was more robust over Lake Michigan in the presence of added MUCAPE that resulted from a deep, well mixed boundary layer consistent with an over-land location. Thus, both sets of simulations
reveal the importance of the synoptic-scale to MCS persistence. However, within a favorable synoptic-scale environment, MCSs might have the potential to be even more robust if Lake Michigan was not in their path, given the possibility of added instability.
5.5 Chapter 5 Figures

FIG. 5.1: WRF horizontal domain and land mask for the (a) control and (b) noLM simulations. The magenta arrow in panel (b) illustrates a region where Lake Michigan was turned to land in the noLM run, while the yellow hexagon in panel (b) shows the location of the point that was mapped to Lake Michigan. Source: 4-km WRF–ARW.
FIG. 5.2: WRF 8-h forecast of 2-m temperature (shaded in °C according to color bar) and 10-m wind (barbs in m s$^{-1}$ where one long barb equals 5 m s$^{-1}$) valid at 20Z/07 June 2008 for the (a) control and (b) noLM simulations. The yellow hexagons in each panel reveal the location of the soundings in Fig. 5.3. The arrow in panel (b) shows the location of Lake Michigan. Source: 4-km WRF–ARW.
FIG. 5.3: WRF 8-h forecast sounding from the yellow hexagons in Figs. 5.2a,b of temperature (solid line in °C), dewpoint (dashed line in °C) and wind (barbs in m s$^{-1}$ where one long barb equals 5 m s$^{-1}$) valid 20Z/07 June 2008 for the (a) control and (b) noLM simulations. Source: 4-km WRF–ARW.
FIG. 5.4: WRF forecast simulated reflectivity (shaded according to color bar in dBZ), SLP (black lines every 4 hPa), and 10-m winds (black barbs in m s$^{-1}$ where one long barb equals 5 m s$^{-1}$) for the (a) control simulation 8-h forecast valid at 20Z/07 (b) noLM simulation 8-h forecast valid at 20Z/07, (c) control simulation 10-h forecast valid at 22Z/07, (d) noLM 10-h simulation forecast valid at 22Z/07, (e) control simulation 12-h forecast valid at 00Z/08, (f) noLM 12-h simulation forecast valid at 00Z/08, (g) control simulation 14-h forecast valid at 02Z/08, and (h) noLM simulation 14-h forecast valid at 02Z/08 June 2008. The arrows in each panel reveal the simulated location of MCS-P1. Radar imagery in the lower-left corner of panels (a), (c), (e), and (g) are from 2004Z/07, 22Z/07, 0001Z/08, and 0203Z/08 June 2008, respectively. Source: 4-km WRF–ARW and UAlbany radar archives.
FIG. 5.4: continued
FIG. 5.5: WRF 15-h forecast accumulated precipitation difference between the noLM and control runs (shaded according to color bar in mm). The precipitation difference was calculated by subtracting the amount of precipitation that fell between 12Z/07 and 03Z/08 June 2008 in the control simulation from the amount of precipitation that fell between 12Z/07 and 03Z/08 June 2008 in the noLM simulation. The arrow indicates a region where more accumulated precipitation fell in the noLM simulation associated with MCS-P1. Source: 4-km WRF–ARW.
FIG. 5.6: As in Fig. 5.2 except for WRF 8-h forecast valid at 20Z/04 June 2005. The yellow hexagons in each panel reveal the location of the soundings in Fig. 5.7. Source: 4-km WRF–ARW.
FIG. 5.7: As in Fig. 5.3 except for the WRF 8-h forecast soundings at the yellow hexagons in Figs. 5.6a,b valid at 20Z/07 June 2008. Source: 4-km WRF–ARW.
FIG. 5.8: As in Fig. 5.4 except for the (a) control simulation 8-h forecast valid at 20Z/04 (b) noLM simulation 8-h forecast valid at 20Z/04, (c) control simulation 10-h forecast valid at 22Z/04, (d) noLM simulation 10-h forecast valid at 22Z/04, (e) control simulation 12-h forecast valid at 00Z/05, (f) noLM simulation 12-h forecast valid at 00Z/05, (g) control simulation 15-h forecast valid at 03Z/05, and (h) noLM simulation 15-h forecast valid at 03Z/05 June 2005. The arrows in each panel reveal the simulated location of MCS-D1. Radar imagery in the lower-left corners of panels (a) and (c) are from 200Z/04 and 22Z/04 June 2005, respectively. Source: 4-km WRF–ARW and UCAR MMM image archive.
FIG. 5.8: continued
FIG. 5.9: 925-hPa $\theta_e$ (shaded according to color bar in K), height (black lines every 30 m), and wind (barbs in m s$^{-1}$, where one long barb equals 5 m s$^{-1}$) for the (a) WRF control simulation 11-h forecast, (b) WRF noLM simulation 11-h forecast, and (c) RUC analysis valid at 23Z/04 June 2005. An IR satellite image from 1810Z/04 June 2005 is inset into the lower-left corner of panel (c). Source: 4-km WRF–ARW, 20-km RUC analysis, and UCAR MMM image archive.
FIG. 5.10: As in Fig. 5.5 except for the WRF 15-h forecast precipitation difference between 12Z/04 and 03Z/05 June 2005. The arrow indicates a region where more accumulated precipitation fell in the noLM simulation associated with MCS-D1. Source: 4-km WRF–ARW.
6. Concluding Synthesis and Discussion

6.1. Overview

The purpose of this thesis was to investigate Lake Michigan-crossing MCSs to determine the environmental and lake conditions that differentiated between MCS persistence and dissipation. Many important results have been identified through climatological, compositing, and case study (both real and simulated) analysis. In order to best synthesize these results and relate them back to previous research, this chapter will be divided into multiple sections, each addressing one of the five main research questions posed in chapter 1. Following these five sections, potential directions for future research will be discussed.

6.2. Lake Michigan-Crossing MCS Persistence/Dissipation Percentage

The climatology of Lake Michigan crossing MCSs identified 110 coherent MCSs that intersected Lake Michigan during the warm seasons (April–September) of 2002–2007. Of these 110 MCSs, 47 (43%) persisted while 63 (57%) dissipated (Table 2.1). Graham et al. (2004) found a 68% persistence rate in their climatology of Lake Michigan-crossing MCSs from different years. However, Graham et al. (2004) focused on intense MCSs, which likely increased the persistence percentage. Strong bow-echo MCSs in this climatology persisted at a 69% rate, similar to the Graham et al. (2004) percentage (Fig. 2.14). Four of the six years in the Lake Michigan-crossing MCS climatology contained more dissipating MCSs than persisting MCSs, similar to the overall climatology (Fig. 2.3). However, two years, 2003 and 2004 contained more persisting MCSs than dissipating MCSs. One possible reason for these two anomalous
years might be the numerous long-track, well-organized MCSs that occurred (e.g., Ahijevych et al. 2004; Metz and Bosart 2010).

The largest number of MCSs in the Lake Michigan-crossing MCS climatology occurred during the summer months of July and August, a climatologically favored period for MCS occurrence in the Great Lakes region (e.g., Johns and Hirt 1987; Fig. 2.4). Additionally, MCS persistence was also most frequent during July and August when the Lake Michigan water temperature was warmer (e.g., Laird et al. 2001). However, a number of MCSs persisted during the early-warm season when the Lake Michigan water temperature was below 10°C (Fig. 2.4). Furthermore, Lake Michigan-crossing MCSs occurred and persisted during all hours of the day, favoring the evening and overnight hours (Fig. 2.5). The evening and overnight hours coincide with the time of day when the LLJ tends to be strongest (e.g., Bonner 1967; Stensrud 1996). The LLJ can enhance synoptic-scale vertical motion in the presence of these Lake Michigan-crossing MCSs and increase sensible heat and moisture transport into the MCS region (e.g., Uccellini and Johnson 1979; Schumacher and Johnson 2009; Figs. 3.5a,b). Additionally, nocturnal MCSs can feed off of elevated MUCAPE above a surface inversion (e.g., Colman 1990a,b), and may never see the shallow Lake Michigan cold layer below (e.g., Wylie and Young 1979).

6.3. Environmental and Lake Conditions Associated with Lake Michigan-Crossing MCSs

Common synoptic-scale and Lake Michigan environmental features were identified that differentiated between MCSs that persisted (Fig. 6.1) and dissipated upon crossing Lake Michigan. CAPE and 0–6-km shear have long been recognized as
necessary ingredients for MCS persistence (e.g., Weisman 1993; Coniglio et al. 2004). Persisting Lake Michigan-crossing MCSs were generally associated with at least 500–750 J kg\(^{-1}\) more MUCAPE along and downstream of Lake Michigan than dissipating MCSs (e.g., Fig. 2.7b). These results mirror those of Letkewicz and Parker (2010) who found that Appalachian Mountain-crossing MCSs persisted with almost 1500 J kg\(^{-1}\) more MUCAPE to the east of the mountains, illustrating the importance of CAPE to MCS persistence in regional climatologies. The amount of MUCAPE upstream of Lake Michigan averaged around 2500 J kg\(^{-1}\) for both persisting and dissipating MCSs (e.g., Figs. 2.7a, 4.7c, and 4.17d) as dissipating MCSs also needed enough instability to organize into relatively long-lived MCSs. Additionally, the amount of upstream and downstream 0–6-km shear averaged about 15 m s\(^{-1}\) for both persisting and dissipating MCSs also showing no discrimination between both categories of Lake Michigan-crossing MCSs (Figs. 2.7a,b).

In order to create realistic synoptic-scale MCS composites, climatological MCSs were subset based on the upper-level pattern (e.g., Johns 1993). Both the dynamic and progressive MCS composites also revealed that MCSs persisted with about 500–750 J kg\(^{-1}\) more downstream MUCAPE and that both persisting and dissipating MCSs averaged about 15 m s\(^{-1}\) of 0–6-km shear along their pathway (e.g., Figs. 3.6a,b). However, individual case study results showed that not all Lake Michigan-crossing MCSs dissipated with reduced downstream MUCAPE. For example, the 6 July 2003 MCS dissipated upon crossing Lake Michigan after moving into an environment that featured more than 1500 J kg\(^{-1}\) of MUCAPE (Fig. 4.54d). However, only about 5 m s\(^{-1}\) of 0–6-km shear remained over Lake Michigan, indicating that some Lake Michigan-
crossing MCSs are exceptions to climatological expectations as this lack of shear led to convective disorganization and collapse.

Furthermore, climatological, compositing, and case study results confirm the importance of a robust 850-hPa LLJ to MCS persistence (e.g., Figs. 2.8, 4.7b, and 4.17b). In a climatological sense, MCSs persisted with an 850-hPa LLJ that was about 5 m s\(^{-1}\) stronger than those that dissipated (Fig. 2.8). A strong LLJ is an ubiquitous feature associated with MCSs across the world (e.g., Stensrud 1996; Laing and Fritsch 1997), and previous authors have shown that LLJs occur with intense Lake Michigan crossing MCSs (e.g., Metz and Bosart 2010). The results in this thesis support this previous research, as the LLJ strength was one of the best discriminators between MCS persistence and dissipation as differences in the strength of the 850-hPa LLJ were statistically significant to the 99.9th percentile (e.g., Fig. 2.8). The stronger LLJs associated with persisting Lake Michigan-crossing MCSs contributed to low-level moisture convergence that approached \(-4.0 \times 10^{-7}\) \(s^{-1}\) (e.g., Figs. 3.5a,b), consistent with or greater than values near heavy-rain producing MCSs (e.g., Moore et al. 2003). As dissipating MCSs began to cross Lake Michigan, these MCSs typically moved eastward from the main core of 850-hPa LLJ dynamical support, and encountered decreased amounts of moisture transport necessary for persistence (e.g., Uccellini and Johnson 1979; Schumacher et al. 2009). Dynamic and progressive MCS composites confirmed the importance of a robust LLJ to MCS persistence (e.g., Figs. 3.3a,b).

Furthermore, composite analyses, along with individual case studies revealed that persisting Lake Michigan-crossing MCSs generally occurred near a 10 m s\(^{-1}\) stronger 200-hPa jet stream than dissipating MCSs. Many of the dynamic MCSs in this thesis,
crossed Lake Michigan while situated in the broad equatorward-entrance region of the 200-hPa jet stream, a region for favorable synoptic-scale ascent (e.g., Figs. 3.3a,b). Additionally, many of the progressive MCSs translated along the anticyclonic shear side of the 200-hPa jet stream after forming in the equatorward entrance region (e.g., Figs. 3.8a,b). Many previous authors have illustrated the importance of upper-level jet stream support to MCS evolution across the U.S. (e.g., Johns and Hirt 1987; Coniglio et al. 2004) and the Great Lakes region (e.g., Metz and Bosart 2010). However, this increased dynamical support for persisting MCSs was from jet-level and not reflected in the magnitude of 850-hPa Q-vector convergence. In general, the 850-hPa Q-vector convergence (and 700-hPa) magnitude did not differentiate between MCSs that persisted and dissipated upon crossing Lake Michigan (e.g., Fig. 2.13). Again, as with the amount of upstream MUCAPE, both persisting and dissipating MCSs were generally located in a dynamically favorable environment at formation time, consistent with areas of Q-vector convergence and forcing for ascent. The persisting MCS environment often featured a region of Q-vector convergence that was at least $0.5 \times 10^{-12} \text{ Pa m}^{-2} \text{ s}^{-1}$ greater than in the dissipating environment, but this region was usually displaced from the actual MCS location, often to the north closer to the jet stream (e.g., Figs. 3.4a,b).

Johns and Hirt (1987) found that derecho (intense damage-producing MCS) development typically occurs in conjunction with an upper-level short-wave trough. However, the MCS case studies shown in this thesis (and those not shown) were not always concomitant with a mesoscale short-wave trough. For example, MCS-P1 was associated with a distinct short-wave trough visible on water-vapor imagery (Figs. 4.2a–c), while MCS–P3 was not (Fig. 4.45c). Metz and Bosart (2010) found during an active
three day period in July of 2003 that multiple successive intense MCSs were associated
with a wide variety of short-wave trough strengths, consistent with the results presented
herein.

Climatological and case study analyses revealed that Lake Michigan-crossing
MCSs often persisted with at least 100–200 J kg\(^{-1}\) more environmental DCAPE than
those that dissipated (e.g., Fig. 2.9). The convective cold pool is important to MCS
maintenance and evolution (e.g., RKW; Weisman and Rotunno 2004), and DCAPE can
estimate the potential for strong-evaporatively cooled downdrafts (e.g., Emanuel 1994).
Cohen et al. (2007) found that DCAPE was a good predictor of MCS organization and
severity. The persisting Lake Michigan-crossing MCSs in this thesis were often
associated with elevated mixed layers between 800 and 600 hPa (e.g., Fig. 4.5). These
mixed layers can increase DCAPE and the potential for evaporatively cooled downdrafts
(e.g., Browning and Ludlam 1962). Not only did persisting Lake Michigan-crossing
MCSs tend to have at least 100–200 J kg\(^{-1}\) more DCAPE, but this DCAPE was realized,
as persisting MCSs had stronger and deeper convective cold pools than those that
dissipated (e.g., Fig. 4.9a,b and 4.18a,b). Strong, deep convective cold pools in the
presence of robust low-level shear led to convective ascent that was calculated to be
strong and deep (e.g., RKW; Weisman and Rotunno 2004). Thus, any horizontal
buoyancy gradients associated with a shallow, cool Lake Michigan were easily overcome
by the persisting MCSs (Fig. 6.2; RKW theory will be discussed more in section 6.4).

Finally, the strength of the near-surface inversion over Lake Michigan did
statistically discriminate between MCSs that persisted and dissipated upon crossing Lake
Michigan, especially during the early warm season when this inversion strength was
about 1.25°C stronger for persisting MCSs (e.g., Figs. 2.11a–c). The near-surface inversion (temperature at 5 m minus water temperature) that developed over Lake Michigan allowed MCSs to persist upon crossing a cool Lake Michigan (Fig. 2.4), and made the actual Lake Michigan temperature unimportant to MCS persistence (Fig. 2.10). These results mirror those by Graham et al. (2004) who found that MCSs persisted when the near-surface inversion strength was at least 2.5°C. Modeling results showed that removing Lake Michigan for both a persisting and a dissipating MCS resulted in increased amounts of 50 dBZ or greater convection in both cases (e.g., Figs. 5.4e,f). Thus, the actual strength of the near-surface inversion might not be as important to MCS persistence as climatological results suggest. Rather, the removal of Lake Michigan resulted in almost double the amount of MUCAPE (e.g., Fig. 5.3) and stronger resulting convection regardless of whether an MCS persisted or dissipated upon crossing Lake Michigan.

In total, synoptic-scale environmental conditions that are favorable for Lake Michigan-crossing MCS persistence (Fig. 6.1) are similar to those previously identified for the global population of MCSs. Persisting Lake Michigan-crossing MCSs were associated with increased MUCAPE over lower Michigan, a robust LLJ, upper-level jet stream support, significant DCAPE (elevated mixed layers and strong convective cold pools), and shallow, intense inversions over Lake Michigan (especially during the early warm season).

6.4. Application of RKW Theory to Lake Michigan-Crossing MCSs
RKW theory discusses the maintenance and evolution of MCSs in terms of the balance between the circulation concomitant with the horizontal buoyancy gradient from the cold pool and the circulation associated with the vertical shear (e.g., Weisman and Rotunno 2004). In an optimal state for convective ascent, the ratio of the convective cold pool velocity \( C \) equals the vertical shear magnitude over the depth of the convective cold pool. The ratio of \( C/\Delta U \) was calculated for each MCS case study in chapter 4, and the resulting value was related to the perceived tilt of the MCS. For example, both MCS-P2 had a ratio of \( C/\Delta U \) that exceeded 1, consistent with the upshear tilt of a bow echo MCS.

Early warm season MCSs that approached Lake Michigan had to also encounter a horizontal buoyancy gradient associated with Lake Michigan (e.g., Fig. 4.6). MCS-P1 persisted upon crossing Lake Michigan with a strong, deep convective cold pool that yielded a cold pool velocity estimate of \( 18 \text{ m s}^{-1} \) in the presence of \( 17.5 \text{ m s}^{-1} \) of 0–3-km line-perpendicular shear (Figs. 4.9a,b), resulting in a ratio of \( C/\Delta U \) that was nearly optimal. The shallow Lake Michigan buoyancy gradient that MCS-P1 encountered also had to cause an atmospheric circulation in agreement with the dimensional horizontal vorticity equation (e.g., Weisman 1993). Even though Lake Michigan is not a convective cold pool, \( C \) was calculated in order to compare the strength of the Lake Michigan-induced circulation to the convective cold-pool circulation. Wylie and Young (1979) found the shallow cold dome over Lake Michigan to be 150–200 m deep on a particular day. So as MCS-P1 approached Lake Michigan, if the marine-layer depth was assumed to be 250 m (generous estimate), with \( \theta = 303 \text{ K} \) and \( \theta'_{sfc} = 17 \text{ K} \) (Figs. 4.6 and 4.9b), the resulting cold pool velocity associated with the Lake Michigan cold dome would be
about 12 m \(s^{-1}\). Thus, the strength of the circulation associated with the Lake Michigan horizontal buoyancy gradient was only slightly less vigorous than the circulation associated with the cold pool from MCS-P1. However, since the convective cold pool was so much deeper than the Lake Michigan cold pool (about 1500 m; Fig. 4.9b), this robust shallow circulation was likely overwhelmed by the convective ascent that was at least six times as deep, and MCS-P1 crossed Lake Michigan uninterrupted (Figs. 4.1 and 6.2). Similarly, the convective cold pool associated with MCS-P2 was also very deep (Figs. 4.28,b), and again likely overwhelmed any circulation associated with the Lake Michigan horizontal buoyancy gradient.

On the other hand, MCS-D1 was only concomitant with a cold pool velocity (C) of 8 m \(s^{-1}\) in the presence of nearly 12.5 m \(s^{-1}\) of 0–3-km line-perpendicular shear (Figs. 4.18a,b). Additionally, the convective cold pool was only 700 m deep as MCS-D1 approached Lake Michigan (Fig. 4.18b). Again, if the Lake Michigan cold dome depth was assumed to be 250 m, with \(\overline{\theta} = 303\) K and \(\theta'_{sfc} = 12\) K (Figs. 4.15 and 4.18b), the Lake Michigan cold pool velocity would be 10 m \(s^{-1}\), stronger than the convective cold pool velocity. Additionally, the depth of the circulation associated with the convective cold pool was relatively close to that of the inferred circulation associated with Lake Michigan. Since these horizontal circulations are additive, the Lake Michigan circulation would combine with the shear circulation resulting in woefully suboptimal convective ascent, especially over the lowest portion of the convective updraft. As MCS-D1 moved into the vicinity of Lake Michigan, an immediate convective impact occurred. MCS-D1 was progressing eastward in an increasingly unfavorable synoptic-scale environment.
(e.g., Figs. 4.17a–d). However, MCS-D1 immediately weakened at the Lake Michigan shore upon intersecting this inferred circulation.

One caveat to the effect of the Lake Michigan horizontal buoyancy gradient to RKW theory is illustrated by MCS-D2 that progressed eastward during nocturnal hours, feeding off of MUCAPE mainly above a nocturnal inversion (e.g., Fig. 4.33). For many years, researchers thought that these nocturnal MCSs behaved in a bore-like manner (e.g., Carbone et al. 1990) and did not lift surface parcels via cold pool dynamics. However, Parker (2008) recently found in simulated MCSs, that the cold pool could lift surface parcels in a dynamically consistent manner with RKW theory until the surface cooling exceeded 10°C and there was no longer any evidence of a surface-based cold pool. MCS-D2 was coincident with nocturnal surface cooling of only about 7°C and still had a surface-based representation of a convective cold pool. However, just prior to intersecting Lake Michigan, the forward speed of MCS-D2 increased (Figs. 4.30e–h), consistent with a transformation into an elevated MCS, as the amount of surface-based environmental CAPE tended towards zero (Fig. 4.33). Thus, it is unclear whether MCS-D2 was ingesting surface parcels lifted by the convective cold pool. MCS-D2 did have a shallow, convective cold pool (Fig. 4.37b) upon approach to Lake Michigan, much like MCS-D1 and weakened soon after intersecting Lake Michigan. However, as with MCS-D1, MCS-D2 was moving into an increasingly unfavorable synoptic-scale environment and cold pool strength is related to the synoptic-scale environment. In addition, MCS-D2 was not immediately disrupted at the leading edge of Lake Michigan like MCS-D1, rather dissipating more gradually over the lake, perhaps because MCS-D2 was feeding off of elevated MUCAPE, dissipating mainly due to unfavorable synoptic-scale conditions.
MCS-P3 and MCS-D3 occurred in the late warm season during the overnight and early morning hours. MCS-P3 possessed a stronger, 1-km deeper convective cold pool than MCS-D3 (Figs. 4.46b and 4.55b), again in a more favorable synoptic-scale environment (Figs. 4.45a–d and 4.54a–d). However, there was no notable Lake Michigan horizontal buoyancy gradient associated with either of these MCSs as Lake Michigan was similar in temperature to the surrounding land environment (e.g., Fig. 4.43). Thus, these two MCSs persisted and dissipated through normal convective dynamics, whether they were elevated and bore-like or they ingested surface parcels consistent with RKW cold pool dynamics (e.g., Carbone et al. 1990; Weisman and Rotunno 2004).

6.5. Marine Layer Navigation and MCS Characteristic Alterations

Given the inhospitable Lake Michigan water temperatures during the early warm season (e.g., Laird et al. 2001), the only way that MCSs could persist during these months (AMJ) was to ingest MUCAPE above the intense shallow near-surface inversion (e.g., Wylie and Young 1979; Fig. 4.10b). Climatological results revealed that persisting MCSs were associated with a near-surface inversion than was about 1°C more intense than for dissipating MCSs (Figs. 2.11a–c), in agreement with results from Graham et al. (2004) in a separate climatology of Lake Michigan-crossing MCSs. Given a weaker near-surface inversion, an MCS might have a greater chance to ingest unfavorable near-surface air, given a long enough residence time over the water. However, WRF-modeling results revealed that MCS-P1 and MCS-D1 were both more intense when Lake Michigan was removed, producing up to 50 mm more precipitation (e.g., Figs. 5.4g,h,
This result implies that both persisting and dissipating MCSs could benefit from the removal of Lake Michigan since increased instability would result.

An examination of radar imagery from the climatological MCSs that occurred during the early warm season did not reveal any systematic alteration in organization or speed of MCSs upon crossing Lake Michigan. For example, if the MCSs transitioned into a bore-like phenomena on top of the lake inversion, a faster forward speed consistent with a gravity wave would be expected (e.g., Parker 2008), and no such increase in forward speed was observed. Given the extreme shallow layer of the Lake Michigan inversion, RKW cold pool dynamics likely continued to apply as the MCSs crossed Lake Michigan, and the MCSs treated the stable layer much like a “speed bump” (Fig. 6.2). In many instances, damage reports were noted as soon as a Lake Michigan-crossing MCS emerged into lower Michigan (e.g., Fig. 4.20).

During the later warm season, Lake Michigan was very similar in temperature to the surrounding land (e.g., Fig. 4.43). Thus, there was generally no stable layer to navigate upon crossing Lake Michigan. In fact, although RUC analyses did not capture this effect, locally the amount of MUCAPE over Lake Michigan might have been higher during the later warm season, given the increased surface fluxes off of the lake. The resulting moist boundary-layer air might locally improve the convective environment during the short period that an MCS was located over Lake Michigan. In general, late warm season MCSs persisted upon crossing Lake Michigan through normal convective processes (e.g., RKW; Weisman and Rotunno 2004).

Finally, convection that forms along a lake breeze (e.g., Laird et al. 2001) can alter the structure of an MCS. As MCS-P2 crossed Lake Michigan, 55–60 dBZ
Convection initiated along a convergent lake breeze boundary that had pushed onshore into southeastern Michigan (Figs. 4.29a,b). MCS-P2 eventually merged with this lake-breeze convection creating a reflectivity bulge at the center of bow-echo MCS-P2. (Fig. 4.29a). As this reflectivity bulge moved onshore, the most robust wind reports associated with MCS-P2 were noted, including a 90-kt gust (Fig. 4.20).

6.6. Relative Roles of Synoptic-Scale versus Lake Michigan

Perhaps the most important part of this thesis was determining the relative importance of the synoptic scale and Lake Michigan to MCS persistence, given the difficulty in forecasting Lake Michigan-crossing MCSs (R. Graham 2009, personal communication). MCS case studies revealed the importance of a favorable synoptic-scale configuration for MCS persistence over Lake Michigan. Each of the three dissipating MCS case studies presented weakened over Lake Michigan when large-scale features became unfavorable for MCS persistence. For example, MCS-D1 moved into an environment that lacked downstream MUCAPE in lower Michigan and 850-hPa LLJ support (Figs. 4.17b,d). MCS-D2 also progressed into an environment that lacked downstream MUCAPE to draw from and 850-hPa LLJ support (Figs. 36b,d). Finally, MCS-D3 moved into an environment that lacked 850-hPa LLJ support and had a nearly nonexistent amount of 0–6-km shear (Figs. 4.54b,d). In addition, climatological and case-study results revealed a tendency for MCSs to persist with a stronger, more stable near-surface inversion, especially during the early warm season (e.g., Fig. 2.11b). Thus, the relative importance of the synoptic scale and Lake Michigan needed to be parsed.
Control and noLM WRF-model simulations of MCS-P1 and MCS-D1 were produced. The synoptic-scale patterns in all WRF simulations were generally similar to the synoptic-scale patterns in reality. The main result from these simulations was that the synoptic-scale appeared to be the primary control between persistence and dissipation. However, both MCS-P1 and MCS-D1 were more intense in the noLM run than the control run. The noLM simulations both had almost twice as much MUCAPE over the Lake Michigan region than the control simulations (e.g., Fig. 5.3). This increased MUCAPE resulted from changing a shallow near-surface inversion into a well-mixed, relatively moist boundary layer. Given this doubling of MUCAPE, the noLM simulation produced more robust convection that resulted in regions that received 50 mm of enhanced precipitation (Figs. 5.5 and 5.10) within the envelope of the favorable synoptic-scale environment. Once the synoptic-scale became unfavorable though, both the control and noLM simulations dissipated quickly in tandem (e.g., Figs. 4.8g,h). The reduced MUCAPE over Lake Michigan in the control simulations compared to the rest of the environment might explain why many previous studies found a decrease in over-lake precipitation when compared to nearby land locations (e.g., Changnon 1968; Bolsenga 1977; Wilson 1977).

These results suggest that the synoptic-scale is the dominant factor for the persistence and dissipation of MCSs upon crossing Lake Michigan. Even though the strength of the Lake Michigan inversion was found to be an important factor in MCS persistence, the synoptic-scale environment likely plays a much more well defined, robust role. The importance of the synoptic-scale environment to MCS persistence is further illustrated by the nearly 70% of bow echo MCSs in the climatology that persisted
upon crossing Lake Michigan (Fig. 2.14). Bow echoes typically have a much more favorable convective environment to feed off of than weaker MCSs (e.g., Johns 1993; Weisman 1993). If Lake Michigan were not present, both persisting and dissipating MCSs would likely be somewhat stronger.

As MCSs move to the approximate longitude of Lake Michigan, they begin to outrun the climatologically favored region for the LLJ (e.g., Bonner 1968) that has been shown to be so vital for MCS persistence. Additionally, some of the MCSs that dissipated over Lake Michigan were part of long-lived convective episodes (e.g., Ahijevych et al. 2004; Carbone and Tuttle 2008; Fig. 1.4) that formed to the lee of the Rockies, often between 1800 and 2100 UTC, and then progressed eastward. Given the travel distance from convective episode formation at around 100°W, many of these long-lived MCSs crossed Lake Michigan between 1300 and 1800 UTC, during the time that the LLJ often weakens (e.g., Bonner 1968; Stensrud 1995) likely contributing to dissipation. While creating the MCS climatology for this thesis, a significant number of long-track MCSs were noted that formed to the lee of the Rockies and moved eastward too slowly to intersect Lake Michigan, completely dissipating immediately upstream of Lake Michigan as each moved into a less favorable synoptic-scale environment. Carbone and Tuttle (2008; their Fig. 10b) reveal that climatologically, the convective episode pathway terminates around 90°W, very near to Lake Michigan, consistent with this repeated dissipation upstream that sometimes extends over Lake Michigan.

Finally, the persisting MCSs relied on MUCAPE downstream of Lake Michigan for persistence in a manner reminiscence of Appalachian Mountain-crossing MCSs (e.g., Letkewicz and Parker 2010) illustrating that well-known convective ingredients are just
as important for the persistence of a local MCS subset as for the entire global population. Rieman-Campe et al. (2009) showed that the average summertime distribution of CAPE has a climatologically strong east–west gradient situated near Lake Michigan, with average values decreasing to the east. The location of this gradient again identifies the longitudes near Lake Michigan as a synoptically-favored location for MCS dissipation. Therefore, while Lake Michigan likely exacerbates the west-to-east decrease in convection across the longitudes of Lake Michigan (e.g., Fig. 2.1a), by reducing available MUCAPE, causing RKW ascent disruption, and/or perhaps allowing MCSs to ingest unfavorable lake-modified air, the synoptic-scale seems to play the primary dissipation role. Given the near-surface inversion over Lake Michigan during the early-warm season, numerous nocturnal MCSs, and the warmer temperature of the lake during the late-warm season, many MCSs likely do not even know Lake Michigan is located below. Instead, these MCSs primarily depend upon the configuration of the synoptic-scale for persistence and dissipation.

5.7. Pathways for Future Research

This thesis covered significant ground on elucidating the persistence and dissipation of Lake Michigan crossing MCSs. However, as with any project, other directions could be explored to further this research. The first possible addition to this research would be to extend the climatology to include more years and all the Great Lakes. The years 2002–2007 were utilized for this climatology because the NOWrad composite dataset was available in a consistent format for the entire period, and allowed for the development of a reliable MCS climatology. This NOWrad dataset is not readily
available for previous or subsequent years. However, as long as composite reflectivity is utilized, the climatology could be extended in a consistent manner. An extension of the climatology would increase the robustness of the results. Additionally, an examination of lake crossing MCSs over other lakes would allow environmental comparisons to occur between all lakes. In order to produce a complete climatology of the other Great Lakes, Canadian radar data would need to be acquired.

A second future direction for research centers upon Lake Michigan-crossing composites. In this thesis, MCSs were subset based upon the upper-level synoptic-scale environment into the dynamic and progressive synoptic-scale patterns (Johns 1993). These two categories were purposely left broad so that all composites would have a large number of members, while still representing synoptically identifiable patterns. In the future, the further separation of these composites into more specific large-scale patterns may be undertaken. For example, there might be preferentially two different dynamic composite patterns, one with deep southerly flow, and another with “shallower" southwesterly flow aloft. Additionally, insight into common synoptic-scale signatures of Lake Michigan-crossing MCSs could be gained by partitioning the composites into early warm season (AMJ) and late warm season (JAS) groups. This separation was not employed in this thesis because with only 110 MCS cases, further separation from the two synoptic-scale patterns would have resulted in some composite groups only containing a small number of cases. Expanding the years in the climatology, as noted above, would address this case number issue.

Furthermore, another pathway for future research would be to perform additional real data WRF simulations. The two cases presented in this thesis, offer significant
insight into the importance of the synoptic scale and Lake Michigan to MCS persistence. However, the simulations presented only represent two cases. Other preliminary simulation results appear to agree with the two MCS cases presented herein. However, simulating a significant number of other Lake Michigan-crossing MCS cases would bolster these results. These added simulations should be relatively straightforward, albeit time consuming, given that a successful simulation methodology already exists.

Finally, an eventual research goal is to perform idealized modeling of Lake Michigan-crossing MCSs. Lericos et al. (2007) studied idealized MCSs that traveled from the Gulf of Mexico to Florida, and using surface and radiation physics, prescribed differing land and water temperatures. By utilizing idealized modeling, the possibility arises to observe the MCS response to differing strengths of the low-level shear and the cold pool. Additionally, the intensity of the over-lake cold dome could be altered in an attempt to understand differing lake influences. Trajectories could be initiated from within the lake cold pool to show conclusively whether lake-crossing MCSs ingest only elevated air (e.g., Parker 2008), or marine boundary-layer. From a larger standpoint, an extension of the Parker (2008) simulations to determine how long nocturnal Lake Michigan-crossing MCSs continue to lift surface parcels in a manner consistent with RKW theory would augment the observational results presented in this thesis.
FIG. 6.1: Idealized schematic of common features associated with persisting Lake Michigan-crossing MCSs. A region of MUCAPE (purple shading), DCAPE (magenta shading), moisture convergence (green shading), as well as an upper-level jet stream (orange shading), a LLJ (blue barbs), and a cold-pool-induced boundary (red dashed line) are indicated in this image.
FIG. 6.2: Idealized west–east cross section schematic of an early warm-season persisting MCS. Circulations are those associated with the convective cold pool, vertical shear, and Lake Michigan marine layer at (a) a time prior to intersecting Lake Michigan and (b) a time after intersecting Lake Michigan. The green arrow indicates convective ascent.


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