APPALACHIAN LEE TROUGHS AND THEIR ASSOCIATION
WITH SEVERE CONVECTIVE STORMS

by

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ABSTRACT

Forecasting convective storms in the mid-Atlantic region of the U.S. (hereafter mid-Atlantic) is important because of the proximity of the convective initiation region to major East Coast cities. These forecasts can be challenging because of the weak synoptic-scale forcing for ascent and abundant convective instability typical of the mid-Atlantic warm season (May–September). As a result, mesoscale boundaries such as the Appalachian lee trough (ALT) play an important role in initiating convective storms. This thesis examines the association between ALTs and warm-season severe convective storms in the mid-Atlantic in order to understand how ALTs modulate the frequency and distribution of severe convective storms and to provide enhanced situational awareness for forecasters.

A climatology of warm season ALTs was developed for the region to the lee of the Appalachians. The climatology used an objective algorithm that searched for prominent lower-tropospheric features common to ALTs, which include an MSLP trough and a thermal ridge in the lee of the Appalachians, as well as downslope winds over the Appalachians. The climatology results indicate that ALT formation is favored during times of peak diurnal and seasonal heating. ALTs identified by the climatology were categorized according to their relationship to synoptic-scale cold fronts. A composite of the category in which ALTs occur in advance of cold frontal passages showed a plume of convective available potential energy that extended from the Carolinas to southeastern Pennsylvania. This plume was collocated with > 30 kt surface to 500-hPa vertical wind shear (VWS) over the Washington, DC, to Philadelphia corridor. This corridor also coincided with a maximum in severe local storm reports for events in this category.
A case study of a high-impact severe convective storm event was performed. This event featured prefrontal storms forming along a wind-shift boundary and intensifying upon approaching an axis of high equivalent potential temperature (345–365 K) collocated with an ALT. It is hypothesized that storms approaching an ALT in an environment of > 30 kt VWS can be expected to intensify, since ALTs contain lower-tropospheric thermal ridges, which correspond to convective instability maxima near the ALT.
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1. Introduction

1.1 Motivation

Accurately forecasting the location, mode, and severity of convective storms in the mid-Atlantic region of the U.S. (hereafter the mid-Atlantic; defined here as the region to the east of the Appalachian Mountains from northern South Carolina to southeastern Pennsylvania) is important because of the proximity of the convective initiation region to major East Coast cities (Fig. 1.1). Forecasting convective storms in the mid-Atlantic can be challenging during the warm season because of weak synoptic-scale forcing in the presence of abundant convective instability, which increases the importance of mesoscale boundaries (relative to synoptic-scale forcing) in initiating convective storms. According to Koch and Ray (1997), examples of mesoscale boundaries that can initiate convective storms in the mid-Atlantic include convective storm outflow boundaries, sea-breeze fronts, horizontal rolls in the boundary layer, shallow fronts (i.e., fronts that are not associated with upper-tropospheric jets), differential heating boundaries, and surface troughs (e.g., lee troughs and prefrontal troughs).

Prior research on the association between Appalachian lee troughs (ALTs) and convective storms has either been part of a larger climatology of convective storms in the northeastern U.S. [e.g., Lombardo and Colle 2010, 2011; Murray and Colle 2011 (hereafter MC11)] or was conducted prior to the advent of gridded regional and global reanalysis datasets (e.g., Weisman 1988, 1990a,b). Because of the scarcity of these studies, isolating the role of the ALT among the wide variety of factors that modulate the convective environment of the mid-Atlantic is essential so that forecasters can more
accurately predict convective storms associated with ALTs. A better understanding of the structure of ALTs and under which environmental conditions ALT-associated convective storms are favored can lead to increased situational awareness for forecasters, and ultimately to more accurate convective storm forecasts and more timely severe weather warnings.

1.2 Literature Review

1.2.1 Formation and Structure of Lee Troughs

The *AMS Glossary of Meteorology* defines a lee trough as, “A pressure trough formed on the lee side of a mountain range in situations where the wind is blowing with a substantial component across the mountain ridge …” (Glickman 2000, 447–448). Lee trough formation can be explained from both dynamical and thermodynamic perspectives. From a dynamical perspective, lee trough formation can be explained by two equivalent processes: 1) the generation of a cyclonic circulation by horizontal convergence associated with vertical stretching of air columns; and 2) the conservation of potential vorticity, where decreasing static stability through air column stretching is compensated by generation of relative vorticity (Fig. 1.2). From a thermodynamic perspective, lee troughs are formed when air parcels descending the lee side of a mountain range warm because of adiabatic compression. To satisfy hydrostatic balance, the pressure to the lee of the mountains must decrease and troughing results at the surface.

The formation of lee troughs is similar to the formation of lee cyclones in some respects and is different in others. Both lee troughs and lee cyclones form in the lee of
major mountain ranges in association with downsloping winds. Lee troughs can be precursors for cyclogenesis. Lee cyclogenesis has been associated with an upper-tropospheric jet approaching a preexisting lee trough (e.g., Hovanec and Horn 1975; Achtor and Horn 1986; Marshment and Horn 1986). Deformation of a preexisting lower-tropospheric baroclinic zone by circulations associated with an upper-tropospheric disturbance has also been found to contribute to lee cyclogenesis (e.g., Buzzi and Tibaldi 1978; McGinley 1982; Smith 1984, 1986). However, lee trough formation does not require an upper-tropospheric disturbance or a preexisting lower-tropospheric baroclinic zone; lee trough formation only requires a wind component normal to and downslope of a mountain range (Weisman 1990a). O’Handley and Bosart (1996) formulated a conceptual model for lee cyclogenesis south of a cyclonic weather system (the “primary”) crossing the Appalachians. Initially, the upper- and lower-tropospheric reflections of the primary are in phase. As the primary approaches the mountains, enhanced lower-tropospheric downsloping easterly flow causes pressure falls in an area north of the primary and west of the Appalachians through vortex stretching, whereas a similar process occurs with westerly flow in an area south of the primary and east of the Appalachians (Fig. 1.3). The result is the northward deflection of the primary surface cyclone and lee cyclogenesis to the south of the primary. A process evolving according to this conceptual model could produce a lee trough instead of a lee cyclone in the absence of preexisting baroclinicity.

Both lee troughs and surface cold fronts possess wind shifts and pressure troughs. However, although lee troughs are sometimes misrepresented as surface cold fronts on synoptic analyses (Hobbs et al. 1990), the two differ in their thermal structure. A surface
cold-frontal pressure trough is coincident with the leading edge of a zone of enhanced baroclinicity, whereas a lee trough is coincident with a lower-tropospheric thermal ridge (Carlson 1961; Fig. 1.4). Lee troughs can also show characteristics similar to the dryline (Schaefer 1974; Benjamin and Carlson 1986; Martin et al. 1990, 1995; Steenburgh and Mass 1994). Specifically, the lee trough can act as a boundary between warm, dry air with a history of subsidence to the west of the trough and moist air that is advected poleward by southerly winds to the east of the trough (Benjamin and Carlson 1986; Steenburgh and Mass 1994).

Benjamin (1986) performed two-dimensional model simulations of flow normal to an idealized 2000 km-wide plateau. A trough formed both in the absence of surface heating and when the plateau was removed but differential surface heating was allowed. The former and latter experiments illustrate the dynamical and thermodynamic factors leading to lee trough formation, respectively. When the two experiments were combined (i.e., a heated plateau subject to a normally-oriented wind field), the lee trough that resulted had a greater amplitude than the lee trough that resulted in either of the two prior experiments. When a diurnal surface heating profile over the plateau was allowed, the lee trough attained its greatest amplitude in the afternoon.

Palmén and Newton (1969, section 11.8) suggested that the lee trough is a stationary feature that remains anchored to the mountains until it is overtaken by a cold front. However, Carlson’s (1961) theory states that a lee trough will move away from the mountains once its associated lower-tropospheric thermal ridge strengthens to a critical amplitude. The presence of an approaching upper-tropospheric ridge (trough) increases (decreases) the critical amplitude. By Carlson’s theory, a lee trough would strengthen
and remain stationary in the presence of an approaching upper-tropospheric ridge because of terrain-tied downsloping effects, whereas the lee trough would dynamically move away from the mountains in association with quasigeostrophic ascent induced by an approaching upper-tropospheric trough. A possible example of the latter mechanism occurred in a case study by Gaza and Bosart (1985) in which a lee trough was “dragged off” the lee of the Rocky Mountains by an upper-tropospheric short-wave trough. In this case, the surface pressure falls associated with the approaching short wave overwhelms the terrain-tied downsloping effects and causes the northern portion of the lee trough to dynamically propagate away from the mountains.

There are also many examples of mobile lee troughs embedded within synoptic-scale midlatitude cyclones. When lee troughs occur in the presence of approaching cold fronts (e.g., in the warm sector of a midlatitude cyclone), they can be manifested as prefrontal troughs (PFTs; Hutchinson and Bluestein 1998; Schultz 2004, 2005). Steenburgh and Mass (1994) simulated a lee trough/PFT that evolved according to the conceptual model of Hutchinson and Bluestein (1998; Fig. 1.5). In this model, lower-tropospheric flow normal to a mountain range initially produces a lee trough (Fig. 1.5a). At a later time, low-level winds to the east of the lee trough begin to veer; meanwhile, a cold front approaches from the northwest. Warm advection associated with the veering winds to the east of the lee trough causes the trough to propagate eastward away from the mountains (Fig. 1.5b). Eventually, the cold front begins to overtake the lee trough/PFT so that the two features merge at the surface (Fig. 1.5c). A few case studies have noted the similarities of merged cold fronts/lee troughs and occluded fronts (e.g., Locatelli et al. 1989; Steenburgh and Mass 1994). Alternatively, lower-tropospheric warm advection
ahead of an approaching cold front can induce surface pressure falls within the lee trough/PFT, whereas the front itself can be blocked or slowed by upwind mountains (Dickinson and Knight 1999; Schultz 2004, 2005). Schumacher et al. (1996) showed that fronts that were not slowed by the Appalachians were associated with strong upper-tropospheric dynamics. Furthermore, Schumacher et al. (1996) showed that the Froude number, which has often been used to determine whether flow will be blocked by a barrier, was of lesser importance than the dynamical forcing in determining whether fronts crossing the Appalachians would be slowed.

1.2.2 Appalachian Lee Troughs

The mid-Atlantic is bounded on the west by the Appalachians and on the east by the Atlantic Ocean (Fig. 1.1). The Appalachians crest at Mount Mitchell, NC, 2037 m MSL. Southeast of the Appalachians, the terrain slopes down to the Piedmont and Coastal Plain regions, respectively, before reaching the Atlantic coast ~400 km southeast of Mount Mitchell. Assuming hydrostatic balance, the region to the lee of the Appalachians is favorable for surface troughs in the warm season because of its location between the climatologically cooler Appalachians and Atlantic Ocean.

Weisman (1988) constructed a climatology of ALTs in the mid-Atlantic and found that an ALT occurred on 40% of warm-season (May–September) days in 1984 and 1985. In an observational study of ALTs during the 1956 warm-season, Weisman (1990a) used pilot balloon data averaged over 45 ALT cases to construct a composite circulation. His composite showed southerly lower-tropospheric flow east of the ALT and blocked lower-tropospheric flow west of the Appalachians. He also found that the
ALT was associated with lower-tropospheric convergence that increased during the daytime, as well as ascent at all times of the day.

A few other studies have documented ALTs. Seaman and Michelson (2000) performed a case study of an ALT that was associated with high ozone levels in the northeastern U.S. They found that the ALT was associated with lower-tropospheric convergence, ascent, and south-southwesterly surface winds to its east, in agreement with Weisman (1990a). The ALT was coincident with a wind-shift line at 925-hPa, with westerly (southwesterly) flow to the west (east) of the ALT (Fig. 1.6). A deep surface mixed layer was found to the west of the ALT, whereas a shallower mixed layer existed to the east of the ALT. The difference in mixed-layer depth was attributed to the partial modification of the air to the east of the ALT by the cooler Atlantic Ocean. Koch and Ray (1997) found five instances of PFTs during the 1994 warm season in North Carolina, but did not classify these as ALTs because of weak cross-mountain flow.

It is instructive to consider a conceptual basis for the role of the ALT in initiating convective storms. Lower-tropospheric convergence and ascent analyzed in the vicinity of ALTs by Weisman (1990a) and Seaman and Michelson (2000) could provide the lifting mechanism needed to realize convective instability. Additionally, lower-tropospheric warm advection and frontogenesis on the leading edge of a thermal ridge collocated with an ALT were found to provide lifting mechanisms for convective storm initiation by Lombardo and Colle (2011). Benjamin and Carlson (1986) simulated a Rocky Mountain lee trough that induced southeasterly lower-tropospheric flow to its east, advecting warm, moist air from the Gulf of Mexico poleward. If we assume that the lee trough forms parallel to the major axis of the mountains, the process described by
Benjamin and Carlson (1986) would result in southerly lower-tropospheric flow to the east of an ALT (since the Appalachians are oriented ~45° right of the Colorado Rocky Mountain Front Range). This flow was observed by Weisman (1990a) and Seaman and Michelson (2000). In the mid-Atlantic, this flow could result in the poleward advection of warm, moist air originating from the Gulf of Mexico and/or the southeastern U.S. Such warm, moist air would increase convective instability, and the aforementioned lifting mechanisms could provide ascent, possibly resulting in the development of convective storms in the vicinity of the ALT. However, if trajectories originated from the Atlantic Ocean, the process described by Benjamin and Carlson (1986) could have a stabilizing effect (Seaman and Michelson 2000), especially early in the warm season when SSTs are still relatively cool.

1.2.3 Distribution of Warm-Season Convective Storms in the Mid-Atlantic

Although severe convective storms (i.e., severe hail, severe convective wind, and tornadoes) are not as common in the eastern U.S. as in the Great Plains (Kelly et al. 1985; Brooks et al. 2003; Doswell et al. 2005), warm-season severe convective storm occurrence is still frequent enough to warrant further investigation, especially because of the dense population of the eastern U.S. (Lombardo and Colle 2011). The mid-Atlantic has been included within the domains of several studies of the distribution of convective storms. These studies have been performed using proxies such as radar reflectivity, lightning data, and severe local storm reports. Many studies have discussed problems associated with the National Climatic Data Center (NCDC) severe thunderstorm and tornado data (e.g., Weiss and Vescio 1998; Weiss et al. 2002; Brooks et al. 2003; Doswell et al. 2005; Trapp et al. 2006). Such problems include population bias,
nonuniform reporting methods, changing reporting methods over time, over- or under-
reporting the impact of an event, and imprecise measurements and estimates of wind
speeds. However, spatially smoothing and temporally filtering the data can compensate
for some of the problems (Brooks et al. 2003; Doswell et al. 2005).

Severe convective wind has a much higher annual frequency than either severe
hail or tornadoes over the mid-Atlantic (Brooks et al. 2003; Doswell et al. 2005). Annual
occurrences of severe convective wind in the mid-Atlantic are maximized over southern
Pennsylvania, with a secondary maximum over northwestern South Carolina and a
minimum over the high elevations of southern West Virginia and far western Virginia
(Kelly et al. 1985; Doswell et al. 2005; Asuma 2010; Fig. 1.7). Meanwhile, annual
occurrences of severe hail are maximized in northern South Carolina and southern North
Carolina and are minimized near the Atlantic coast (Doswell et al. 2005). Annual
tornado occurrences are maximized over southeastern Pennsylvania and minimized over
the Appalachians (Brooks et al. 2003). The frequency of annual lightning strikes over the
mid-Atlantic decreases from south to north, and relatively fewer lightning strokes are
observed over the Appalachians compared with adjacent areas to the east and west
(Orville and Huffines 2001; Zajac and Rutledge 2001).

The frequency of severe hail over the mid-Atlantic is maximized in June, whereas
the frequency of severe convective wind is maximized during June, July and August.
The earlier peak of the severe hail season relative to the severe convective wind season is
consistent with the relative occurrence of those seasons across the rest of the U.S. (Kelly
et al. 1985; Doswell et al. 2005). Doswell et al. (2005) showed that the frequency of
severe convective wind occurrence is as high in the mid-Atlantic as anywhere else in the
U.S. during July and August. In the northern mid-Atlantic, lightning in May and June is enhanced (diminished) over portions of the Appalachians (Coastal Plain and coastal waters) compared with July and August (MC11). MC11 attributed this temporal distribution to greater convective instability over the mountains early in the warm season compared with eastward locations, and to a reversal of the instability gradient and more sea breeze-related convective storms near the coast in July and August.

Numerous studies have shown that the diurnal cycle is the dominant temporal mode of convective storm variability in the mid-Atlantic, with a maximum in the afternoon and a minimum overnight [e.g., Landin and Bosart 1985; Reap and Orville 1990; Weisman 1990b; Koch and Ray 1997; Orville and Huffines 2001; Zajac and Rutledge 2001; Murphy and Konrad 2005 (hereafter MK05); Parker and Ahijevych 2007 (hereafter PA07); Lombardo and Colle 2010; MC11]. An eastward-propagating convective signal is also evident, with convective storms often forming over the Appalachians and pushing eastward into the Piedmont, the Coastal Plain, and occasionally reaching the Atlantic coast (Orville and Huffines 2001; MK05; PA07; MC11). Convective storms also have been shown to preferentially form earlier near a sea-breeze boundary compared with areas farther inland (Koch and Ray 1997; Orville and Huffines 2001; MC11). However, Landin and Bosart (1985) found that the sea breeze suppressed afternoon convection over coastal portions of Long Island and New England, owing to the stabilizing influence of marine air. Several studies have also found an overnight peak in convective storms over the coastal waters of the Atlantic Ocean (e.g., Weisman 1990b; Orville and Huffines 2001; Zajac and Rutledge 2001; PA07; Lombardo and Colle 2010; MC11).
Several studies have identified common synoptic-scale features and convective parameters that are associated with convective storms in the mid-Atlantic. The identification of these features and parameters in an operational setting can aid in the forecasting process. PA07 found organized convective episodes in the mid-Atlantic during the warm season to be associated with anomalously high vertical wind shear. MC11 found that active severe convective days in the northeastern U.S. (including northern portions of the mid-Atlantic) are associated with the following factors: 1) an upstream upper-tropospheric trough and associated southwesterly upper-tropospheric flow over the northeastern U.S. (Fig. 1.8a); 2) positive lower-tropospheric equivalent potential temperature advection (Fig. 1.8b); 3) a meridionally elongated axis of high CAPE (Fig. 1.8c); and 4) a right entrance region of a 200-hPa jet. Factors common to strong tornadoes in the northern mid-Atlantic were found by Giordano and Fritsch (1991) to include high surface dewpoints, strong directional vertical wind shear, proximity to an upper-tropospheric jet streak, and upper-tropospheric diffluence. Other studies have further stratified convective episodes in the mid-Atlantic by their radar reflectivity structure (e.g., Lombardo and Colle 2010, 2011) and by strength of synoptic-scale forcing (e.g., Weisman 1990b).

The only papers in the literature that specifically studied the role of ALTs in modulating the distribution of convective storms in the mid-Atlantic were by Weisman (1988, 1990b). He stratified the spatial distribution of lightning on ALT days by synoptic-scale forcing (i.e., strong, moderate, weak and suppressed) and found that the ALT was the main focus of convective storms only in the moderate forcing category. He attributed the convective storm maximum in the moderate category to 850-hPa warm
advection in the vicinity of the ALT. An ALT also appeared in a composite of days when the areal distribution of convective storms exceeded one standard deviation above the 12-yr areal mean over portions of the northeastern U.S. and northern mid-Atlantic presented by MC11 (Fig. 1.8c). Lombardo and Colle (2010, 2011) stratified convective storms according to whether they exhibited cellular, linear, or nonlinear structures in radar reflectivity and found an ALT in composites of each category. They found that 53% of severe linear convective storms in coastal Rhode Island, Connecticut, New York, and New Jersey form within the ALT, whereas only 13% form along a trailing cold front.

1.3 Research Goals and Thesis Structure

A few of the papers mentioned in the previous sections will provide the framework upon which this thesis will build. This thesis will use contemporary reanalysis datasets and observations to update the findings of Weisman (1988, 1990a,b) with respect to climatological, structural, and convective characteristics of ALTs. The results of this thesis will also be compared with the convective storm distributions found by PA07, Lombardo and Colle (2011), and MC11. Some of the smoothing and filtering techniques used by Brooks et al. (2003), Doswell et al. (2005), and Asuma (2010) will be employed in order to compensate for the inconsistencies in severe local storm report data.

The main goals of this thesis are to: 1) establish a climatology of warm-season ALTs in the mid-Atlantic; 2) document the spatial and temporal distributions of severe convective storms in the mid-Atlantic with particular emphasis on the role of the ALT in modulating those distributions; 3) investigate the convective environments characteristic of the ALT Zone; 4) construct composites of ALT events associated with severe
convective storms; and 5) study a specific ALT event associated with severe convective storms.

This thesis will be divided into the following chapters: Chapter 1 has provided an overview of research related to lee troughs and to convective storm distributions in the mid-Atlantic; Chapter 2 will present the data sources and methodology; Chapter 3 will present the results of the climatology, composites, convective storm distributions, and case study; and Chapter 4 will discuss the results and offer conclusions.
Fig. 1.1. Topography and selected major cities of the mid-Atlantic.
Fig 1.2. Schematic diagram showing a lee trough developing via dynamical mechanisms. Isentropes (degrees K) are contoured in black. “L” represents the location of the lee trough [adapted from Martin (2006)].
Fig. 1.3. Schematic of a mechanism for altering the path of a cyclonic weather system (the “primary”) crossing the Appalachians. The Appalachians are represented by an ellipse. The solid lines with arrows represent the surface circulation of the primary, which moves due east along the dashed path. At location 1, the upper- and lower-tropospheric reflections of the primary are in phase. When the primary is between locations 1 and 2, increasing downsloping easterly flow causes surface pressure falls and vorticity production at point A, and the primary surface cyclone is deflected northward (upper-left dot-dashed line). Meanwhile, increasing downsloping westerly flow forces lee cyclogenesis at point B. Thereafter, the lee cyclone gradually comes into phase with the primary upper-tropospheric cyclone (lower-right dot-dashed line) [from O’Handley and Bosart (1996)].
Fig. 1.4. Schematic diagrams of cold and pseudo fronts. The pseudo front corresponds to a lee trough [from Carlson (1961)].
Fig. 1.5. Idealized evolution of a lee/prefrontal trough (dashed line). Arrows indicate lower-tropospheric flow, triangles indicate mountains, and “WA” indicates warm advection [from Hutchinson and Bluestein (1996)].
Fig. 1.6. Streamline analysis of observed 925-hPa winds (kt) at 1200 UTC 14 Jul 1995. Dashed line indicates ALT [from Seaman and Michelson (2000)].
Fig. 1.7. Smoothed annual probability of severe convective wind reports (percent) [from Doswell et al. (2005)].
Fig. 1.8. Composite maps of days when convective frequency exceeded one standard deviation above the areal 12-yr mean over the region outlined by the box in (a). Composites show (a) 500-hPa geopotential height (contours, m) and absolute vorticity (shaded, $10^{-5}$ s$^{-1}$); (b) 925-hPa equivalent potential temperature (shaded, K), winds (full barb = 10 m s$^{-1}$), and geopotential height (contours, m); and (c) sea level pressure (contours, hPa), surface winds (full barb = 10 m s$^{-1}$), and most unstable CAPE (shaded, J kg$^{-1}$) calculated from the 30-hPa layer average with the highest equivalent potential temperature within 180 hPa of the surface [from Murray and Colle (2011)].
2. Data and Methodology

2.1 Data Sources

A climatology of and categorization scheme for ALTs during the warm season (defined here as May–September) were developed using gridded data from the four times daily (0000, 0600, 1200, and 1800 UTC) 0.5° × 0.5° National Centers for Environmental Prediction (NCEP) Climate Forecast System Reanalysis (CFSR; Saha and Coauthors 2010). Mean sea level pressure (MSLP) analyses, created by the Hydrometeorological Prediction Center (HPC), were manually checked for the presence of an ALT when gridded data analysis was inconclusive. These analyses were obtained from the Colorado State NWS DIFAX Weather Map Archive (http://archive.atmos.colostate.edu/).

Severe convective storm reports (hereafter storm reports; i.e., severe convective wind reports, severe hail reports, and tornado reports) were extracted from the NCDC Storm Data database for the warm seasons of 2000–2009 in order to determine the spatial and temporal distributions of severe convective storms in the mid-Atlantic. Severe convective winds are defined as any winds that occur within 30 minutes of lightning being observed or detected that produce a fatality, injury, or damage, and/or have a measured or estimated surface wind speed of ≥ 25.7 m s⁻¹. Severe hail is defined as hail with diameter ≥ 1 in (2.5 cm).

Most unstable CAPE (MUCAPE) and surface to 500-hPa vertical wind shear (hereafter VWS) were calculated at the location of the first daily storm report throughout the climatology using the eight times daily (0000, 0300, 0600, 0900, 1200, 1500, 1800, and 2100 UTC) 32 km × 32 km NCEP North American Regional Reanalysis (NARR;
Mesinger and Coauthors 2006). NARR MUCAPE is a built-in gridded variable that is calculated using the 30-hPa layer in the lowest 180 hPa with the highest mean equivalent potential temperature. VWS was calculated by determining the magnitude of the vector difference between the 500-hPa and 10-m winds. Observed MUCAPE values derived from radiosonde data were obtained from the Storm Prediction Center (SPC) in order to investigate how convective instability at sounding sites changes in the presence of ALTs. Observed MUCAPE was calculated in the same manner as NARR MUCAPE, except using the parcel with the highest equivalent potential temperature in the lowest 300 hPa, and with equivalent potential temperature calculated using the virtual temperature. In order to assess the accuracy of reanalysis-derived MUCAPE values, observed MUCAPE values were also compared with NARR and CFSR MUCAPE values. CFSR MUCAPE is a built-in gridded variable that is calculated in the same manner as NARR MUCAPE, except using the parcel with the highest equivalent potential temperature in the lowest 70 hPa. The different methods to calculate MUCAPE in each reanalysis were not altered to conform to one another in order to compare the built-in gridded MUCAPE values for each reanalysis with observed MUCAPE values.

Composite averages of geopotential height, horizontal wind, vertical motion, temperature, and specific humidity were calculated on isobaric and constant height surfaces using the NARR. Derived variables, such as VWS, equivalent potential temperature, $Q$ vectors, and 1000–850-hPa thickness, were calculated using the aforementioned composited variables. $Q$ vectors were calculated using the built-in $Q$-vector parameter in the General Meteorological Package (GEMPAK; DesJardins et al. 1991), using:
Regions of forcing for vertical motion were determined from the right-hand side of the $Q$-vector form of the QG omega equation:

$$\ddot{\overline{Q}} = - \left( \frac{\partial \overline{\vec{v}}_g}{\partial x} \times \overline{\vec{v}}_p \theta \right) = \left( \begin{array}{c} Q_1 \\ Q_2 \end{array} \right)$$

(1)

as defined in Keyser et al. (1992), where $h = (R/p_o)(p_o/p)^c_v/c_p$ and $p_o$ is a reference pressure (1000 hPa). The quantity $\sigma = -h\partial \Theta/\partial p$ is the static stability coefficient, where $\Theta$ is a pressure-dependent reference profile of potential temperature. Here, $Q$-vector convergence (divergence) indicates regions of forcing for ascent (descent).

A case study was analyzed using NARR data, composite National Operational Weather radar reflectivity (NOWrad) imagery (obtained from the University Corporation for Atmospheric Research and the WSI Corporation), radiosonde data, and hourly Automated Surface Observing System (ASOS) observations (obtained from the Iowa State University Data Archive at [http://mtarchive.geol.iastate.edu](http://mtarchive.geol.iastate.edu)). Backward trajectories were produced using the National Atmospheric and Oceanic Administration (NOAA) Air Resources Laboratory Hybrid Single-Particle Lagrangian Integrated Trajectory (HYSPLIT) model (Draxler and Rolph 2012; Rolph 2012; available online at [http://ready.arl.noaa.gov/HYSPLIT.php](http://ready.arl.noaa.gov/HYSPLIT.php)). State variables from the NARR were smoothed
using a Gaussian filter. Derived variables were calculated from the smoothed state variables.

2.2 Methodology

2.2.1 ALT Climatology

In order to construct a climatology of ALTs, it is necessary to determine atmospheric features common to ALTs that can be objectively identified within a gridded reanalysis dataset. To determine these common features, 13 warm-season ALTs were analyzed using plots of CFSR data. Once common features representative of the 13 ALTs were determined, an algorithm was designed to check for the presence of these features at each CFSR analysis time during the warm seasons of 2000–2009, which is the duration of the ALT climatology. The three common features for which the algorithm checked were: 1) a 925-hPa wind component direction orthogonal to and downslope of the Appalachians, 2) a negative MSLP anomaly with respect to a zonal mean, and 3) a positive 1000–850-hPa thickness anomaly with respect to a zonal mean. In order to perform the algorithm, three quantitative criteria were developed for the three aforementioned common features, respectively. The three criteria are:

1) Meteorological wind directions computed from zonally averaged wind components along each 0.5° latitude within an area encompassing the
Appalachians (i.e., “Wind Zone”; Fig. 2.1) must be between 223° and 43°\(^1\).

2) The difference along each 0.5° latitude of the minimum MSLP found in the region to the lee of the Appalachians (i.e., “ALT Zone”; Fig. 2.1) and the zonally averaged MSLP within an area surrounding the ALT Zone (i.e., “Domain”; Fig. 2.1) must be less than a threshold value.

3) The difference along each 0.5° latitude of the maximum 1000–850-hPa thickness within the ALT Zone and the zonally averaged 1000–850-hPa thickness within the Domain must be greater than a threshold value.

The determination of the threshold values will be discussed in section 3.1. If the three criteria are met for six consecutive 0.5° latitudes, an ALT is considered to be present.

The algorithm was run for each CFSR analysis time over the duration of the climatology. Plots of MSLP and 1000–850-hPa thickness produced from CFSR data for each time an ALT was identified by the algorithm were examined to check for false alarms. Common false alarms included frontal troughs (i.e., surface pressure troughs coincident with frontal boundaries), tropical and extratropical cyclones, and times when the MSLP gradient was large. HPC MSLP analyses were consulted when the plots of CFSR data were inconclusive in determining the existence of a false alarm.

\(^1\) Since the Appalachians in the mid-Atlantic are oriented roughly 43° right of true north, meteorological wind directions that are orthogonal to and downslope of the Appalachians exist between 223° and 43°.
2.2.2 ALT Categorization

ALTs that were both identified by the algorithm and not considered false alarms were categorized according to their relationship to synoptic-scale cold fronts. To categorize ALTs, plots of MSLP and 1000–850-hPa thickness produced from CFSR data were consulted to identify the position of both the ALT and the nearest upstream synoptic-scale cold front. It was subjectively determined that an ALT can be placed into one of the following four categories: 1) Inverted, 2) No PFT, 3) PFT with Partial FROPA (frontal passage), and 4) PFT with Total FROPA. Category 1 ALTs (e.g., Fig. 2.2a) are typified by a trough extending northward into the ALT Zone from an area of surface low pressure located to the south of the ALT Zone. At times, a stationary front exists north of the northern extent of the ALT. Category 2 ALTs (e.g., Fig. 2.2b) occur in the absence of a synoptic-scale cold front; i.e., they are not directly tied to synoptic-scale frontal cyclones. Category 3 and category 4 ALTs (e.g., Figs. 2.2c,d) are also PFTs. For an ALT to fall into category 3 or 4, the front must be south of the New York/Pennsylvania border or east of the western third of Pennsylvania. Fronts associated with category 3 ALTs either do not enter the ALT Zone or do not pass through the entire ALT Zone within 24 h of first entering the ALT Zone. Fronts associated with category 4 ALTs pass through the entire ALT Zone within 24 h of first entering the ALT Zone. Table I summarizes the four ALT categories.
2.2.3 Distributions of Severe Convective Storms in the Mid-Atlantic

In an attempt to correct for the population bias with respect to storm reports that is inherent in the Storm Data dataset, a “clustering” technique was employed. This technique is similar to that used by Brooks et al. (2003) in their climatology of tornadoes, Doswell et al. (2005) in their climatology of severe convective wind and severe hail reports, and Asuma (2010) in his climatology of severe convective wind and gradient high-wind reports. First, a $0.5 \times 0.5$° grid was overlaid on the region of analysis. For each 6-h period in the climatology beginning at each CFSR analysis time, if a storm report occurred in a grid box, that grid box was considered “active.” Any subsequent storm reports that occurred within the active grid box during the 6-h period were discarded. The number of active grid boxes in the ALT Zone was recorded for each 6-h period to measure how widespread the storm reports were during that 6-h period. At each grid point located at the center of each grid box, the number of 6-h periods with an active grid box was divided by the number of 6-h periods in the climatology. The resulting value at each grid point represents the percentage of 6-h periods in the climatology in which an active grid box occurred (i.e., at least one storm report occurred within the grid box surrounding the grid point). Maps of the spatial distributions of the temporal frequency of storm reports at each grid point were created by contouring the resulting gridded values using GEMPAK. The clustering technique was repeated for all 24-h periods in the climatology to measure the spatial distribution of storm reports on each calendar day for use in creating severe and nonsevere composites of each ALT category (discussed in section 2.2.5). The 24-h periods (hereafter days) were chosen to begin at
0400 UTC to correspond to the beginning of the calendar day in the Eastern time zone [i.e., 0000 Eastern Daylight Time (EDT)].

Maps of the spatial distributions of the temporal frequency of storm reports were also created for each ALT category. Storm reports were binned according to the ALT category present at the next earliest CFSR analysis time. For example, if ALT category 2 was present at 1800 UTC on a certain day, all storm reports from 1800 to 2359 UTC on that day would correspond to ALT category 2. The procedure to create maps of the spatial distributions of the temporal frequency of storm reports was then repeated for each ALT category in the manner described in the previous paragraph, except that, for each ALT category, at each grid point, the number of 6-h periods with an active grid box was divided by the number of times in the climatology that each ALT category occurred.

2.2.4 Convective Environments Characteristic of the ALT Zone

In order to quantify convective environments characteristic of the ALT Zone during the warm season, MUCAPE and VWS were calculated from the NARR at the location of the first daily storm report in three distinct latitudinally divided sectors within the ALT Zone (i.e., North, Center, and South; Fig. 2.3). This division was done to minimize the likelihood that the convective environment at the first daily storm report was unrepresentative of the entire ALT Zone. Since the diurnal cycle is the dominant mode of temporal variability in the mid-Atlantic (discussed further in section 3.3.1), only days in which the first daily storm report occurred between 1530 and 0029 UTC were used. These days accounted for 79.7%, 82.4%, and 82.1% of all of the days in the climatology with at least one storm report in the North, Center, and South Sectors,
respectively. The calculations of MUCAPE and VWS were made at the location of the first daily storm report in each sector using the nearest earlier NARR analysis time, as long as the analysis time was at least 30 minutes prior to the first daily storm report (Table II). The use of an analysis time at least 30 minutes prior to the first daily storm report minimizes the likelihood that the MUCAPE value at the first daily storm report is contaminated by convection and thus is unrepresentative of the larger-scale environment in which severe convection initiated.

To investigate how convective instability changes in the presence of ALTs, observed MUCAPE values from sounding times when an ALT was present (hereafter ALT times) were compared with MUCAPE values from sounding times when an ALT was absent (hereafter non-ALT times). All observed MUCAPE values were also compared with MUCAPE values from the NARR and CFSR. Observed MUCAPE values and lifted parcel levels (i.e., level of the parcel with the highest equivalent potential temperature in the lowest 300 hPa) were obtained for four sounding sites located within or in close proximity to the ALT Zone at 1200 UTC and 0000 UTC on all days in the climatology. The four sites are: 1) Greensboro, NC (GSO); 2) Blacksburg, VA (RNK); 3) Wallops Island, VA (WAL); and 4) Washington-Dulles International Airport, VA (IAD). The locations of the four sites are shown in Fig. 2.3. Sounding times were eliminated if the observed MUCAPE value equaled zero. In order to facilitate comparison between the observed values and the two gridded datasets, sounding times were also eliminated if the lifted parcel level was greater than 180 hPa above the surface, since this level is the uppermost lifted parcel level used to calculate MUCAPE in the NARR. The remaining sounding times were partitioned according to whether there was
an ALT present at the time of the sounding. Box-and-whisker plots were generated to compare observed MUCAPE distributions for each site at ALT times and non-ALT times, and to compare distributions between the observations, the NARR, and the CFSR.

2.2.5 Composite Analyses

Severe and nonsevere composites of ALT categories 2, 3, and 4 were created. Category 1 was excluded from the composite analyses due to the low frequency of occurrence of category 1 ALTs. For each day that exhibited an ALT of a particular category, the number of active grid boxes on that day was used to determine if that day would be included in either the severe or nonsevere composites. If a day exhibited more than one ALT category, the ALT category present at 1800 UTC was chosen to represent that day. The threshold criteria of the number of active grid boxes per day for the severe and nonsevere composites were designed to both maximize the difference in severe convective activity between the severe and non-severe composites and to yield similar numbers of members for both severe and nonsevere composites. The analysis time of each member of the category 2 and category 3 composites was chosen to be 1800 UTC in order to facilitate comparisons between the severe and nonsevere composites. The active grid box threshold criteria were relaxed for category 4 relative to categories 2 and 3 due to the relatively low frequency of occurrence of category 4 ALTs. The number of members, active grid box threshold criterion, and analysis time used for each composite category are summarized in Table III. Each composite category will be referred to in the text by its ALT category number followed by either “N” or “S”, denoting nonsevere and severe, respectively. For instance, “3S” refers to ALT category 3, severe composite.
2.2.6 Case Study

The 6 June 2002 severe convective storm event was chosen as a case study because it resulted in extensive damage and five injuries, it contained features similar to the composite analyses, and both the ALT and a trailing cold front played roles in initiating and sustaining severe convective storms. Analyses constructed from NARR data for this case were compared to the composite analyses. Manual analyses of ASOS observations of altimeter setting and equivalent potential temperature ($\theta_e$) were overlaid on plots of NOWrad data annotated with storm report locations in order to determine the exact positions of the ALT, the trailing cold front, and axes of high $\theta_e$, relative to convective initiation and storm report locations.
<table>
<thead>
<tr>
<th>Category Number</th>
<th>Category Name</th>
<th>Category Description</th>
</tr>
</thead>
<tbody>
<tr>
<td>1</td>
<td>Inverted</td>
<td>ALT extends northward into the ALT Zone.</td>
</tr>
<tr>
<td>2</td>
<td>No PFT</td>
<td>ALT occurs in the absence of a synoptic-scale cold front.</td>
</tr>
<tr>
<td>3</td>
<td>PFT with Partial FROPA</td>
<td>ALT occurs in advance of a synoptic-scale cold front. Front must be south of the New York/Pennsylvania border or east of the western third of Pennsylvania. Front either does not enter the ALT Zone or does not pass through the entire ALT Zone within 24 h of first entering the ALT Zone.</td>
</tr>
<tr>
<td>4</td>
<td>PFT with Total FROPA</td>
<td>Same as category 3, except that the front passes through the entire ALT Zone within 24 h of first entering the ALT Zone.</td>
</tr>
</tbody>
</table>

Table I. Shows the four ALT categories referred to in the text, as well as a description of each category.

<table>
<thead>
<tr>
<th>Time of First Daily Storm Report (UTC)</th>
<th>Corresponding NARR Analysis Time (UTC)</th>
</tr>
</thead>
<tbody>
<tr>
<td>1530–1829</td>
<td>1500</td>
</tr>
<tr>
<td>1830–2129</td>
<td>1800</td>
</tr>
<tr>
<td>2130–0029</td>
<td>2100</td>
</tr>
</tbody>
</table>

Table II. Shows the NARR analysis time used to calculate MUCAPE and VWS for the first daily storm report whose time falls within the indicated range.
Table III. Summarizes each ALT composite category, the number of members used in each composite category, the number of active grid boxes per day needed for a particular day to qualify for each composite category, and the NARR analysis time used for each member of each composite category. ALT category numbers are explained in Table I. Letters “N” and “S” denote nonsevere and severe composites, respectively.

<table>
<thead>
<tr>
<th>ALT Composite Category</th>
<th>Number of Members</th>
<th>Number of Active Grid Boxes per Day</th>
<th>NARR Analysis Time (UTC)</th>
</tr>
</thead>
<tbody>
<tr>
<td>2N</td>
<td>22</td>
<td>0</td>
<td>1800</td>
</tr>
<tr>
<td>2S</td>
<td>22</td>
<td>&gt; 20</td>
<td>1800</td>
</tr>
<tr>
<td>3N</td>
<td>17</td>
<td>0</td>
<td>1800</td>
</tr>
<tr>
<td>3S</td>
<td>17</td>
<td>&gt; 20</td>
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<td>4N</td>
<td>17</td>
<td>&lt; 4</td>
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<tr>
<td>4S</td>
<td>17</td>
<td>&gt; 9</td>
<td>Variable</td>
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</tbody>
</table>
Fig. 2.1. Geographical areas over which the ALT identification algorithm was run. “Domain” refers to the area enclosed by the blue trapezoid, “Wind Zone” refers to the area enclosed by the red trapezoid, and “ALT Zone” refers to the area enclosed by the black trapezoid.
Fig. 2.2. Plots of MSLP (black contours, hPa), and 1000–850-hPa thickness (fills, dam) showing examples of (a) ALT category 1; (b) ALT category 2; (c) ALT category 3; and (d) ALT category 4. ALT is denoted by dashed line.
Fig. 2.3. The ALT Zone divided into sectors by latitude. Locations of major cities are indicated with black dots. Sounding locations are indicated with a red star.
3. Results

3.1 ALT Climatology

A composite of 13 warm-season ALTs occurring at 1800 UTC (Fig. 3.1a) shows the three common lower-tropospheric features for which the climatology algorithm searched: 1) wind components over the Appalachians orthogonal to and downslope of the Appalachians, 2) the ALT itself in the lee of the Appalachians with a surface pressure ridge on the windward side of the Appalachian crest, and 3) a thermal ridge collocated with the ALT. In addition, an area of negative 1000–850-hPa thermal vorticity is located in the vicinity of the ALT. This negative thermal vorticity has two distinct minima in the lee of the Appalachians from central Virginia to South Carolina: 1) in the immediate lee of the Appalachian crest, and 2) just inland of the Atlantic Coast. The negative thermal vorticity implies a warm core in the 1000–850-hPa layer. Six hours later, at 0000 UTC, the composite shows that the thermal vorticity minimum near the coast has moved farther inland, likely because of the cooling of coastal areas as a result of the sea breeze, whereas the thermal vorticity minimum in the immediate lee of the Appalachians has remained quasi-stationary (Fig. 3.1b).

The climatology of ALTs is sensitive to the threshold values used for the MSLP anomaly and 1000–850-hPa layer-mean temperature anomaly (hereafter thermal anomaly). The algorithm to identify ALTs was run keeping the wind criteria (defined in section 2.2.1) constant but varying the MSLP and thermal anomaly thresholds. The loosest thresholds used (i.e., an MSLP anomaly of < −0.25 hPa combined with a thermal anomaly of > 0.5°C) resulted in an ALT being recorded 37.5% of the time during the
climatology after false alarms (e.g., frontal troughs, cyclones, and large MSLP gradients) were removed (Fig. 3.2). In contrast, the strictest thresholds used (i.e., an MSLP anomaly of $<-1.75$ hPa combined with a thermal anomaly of $>3^\circ$C) resulted in an ALT being recorded just 0.6% of the time during the climatology after false alarms were removed. It was subjectively determined to use an MSLP anomaly threshold of $<-0.75$ hPa and a thermal anomaly threshold of $>1^\circ$C. These thresholds were chosen because they were strict enough to omit weaker ALTs and relaxed enough to include enough ALTs with which to construct a robust climatology. With these thresholds, an ALT was recorded 26.6% of all the CFSR analysis times during the climatology (i.e., 1629 times) after false alarms were removed. The temporal distribution of ALT occurrence shows that ALTs occur most often at 1800 and 0000 UTC (33.3% and 31.9% of ALTs occur at these times, respectively; Fig. 3.3a). The monthly distribution of ALT occurrence shows that ALTs occur most often during June, July, and August (23.0%, 27.8%, and 25.0% of ALTs occur during these months, respectively; Fig. 3.3b). The temporal and monthly distributions of ALTs suggest that ALT formation is partially tied to the diurnal and seasonal heating cycles, a result which agrees with the findings of the idealized modeling study performed by Benjamin (1986).

3.2 ALT Categorization

Figure 3.4 shows the percentage of occurrence of each ALT category with respect to all 1629 ALTs. Category 2 (no PFT) occurs the most frequently (50.8%), whereas categories 3 and 4 (i.e., the two PFT categories) account for 44.8% of ALTs (36.8% and 8.0%, respectively). Category 1 occurs the least frequently (4.4%).

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Figures 3.5a–d show the monthly percentage of occurrence of each ALT category. Category 1 ALTs are spread out fairly evenly between June, July, and August (JJA), and May and September (MS) (54.8% and 45.2%, respectively; Fig. 3.5a). Category 2 and category 3 ALTs occur more often during JJA than during MS. The months of JJA account for 77.4% (82.6%) of category 2 (3) ALTs (Figs. 3.5b,c). The peak month of occurrence of category 2 (3) ALTs is July (August), which accounts for 29.9% (30.2%) of category 2 (3) ALTs. In contrast to categories 2 and 3, category 4 ALTs occur more often during MS than during JJA. The months of MS account for 53.1% of category 4 ALTs (Fig. 3.5d). The peak month of occurrence of category 4 ALTs is May, which accounts for 30.8% of category 4 ALTs. The preference for category 4 ALT occurrence during MS is likely tied to stronger baroclinicity associated with a stronger mean upper-tropospheric jet, which promotes more frequent frontal passages through the entire ALT Zone during MS compared with JJA.

3.3 Distributions of Severe Convective Storms in the Mid-Atlantic

3.3.1 Temporal Distribution

At least one storm report occurred within the ALT Zone on 754 out of the 1530 days in the climatology (49.3%). Of the 12 330 storm reports examined, severe convective winds, severe hail, and tornadoes accounted for 79.1%, 16.5%, and 4.4% of the reports, respectively. If only the 8116 storm reports occurring < 6 h after an ALT are considered, the proportion of severe convective wind reports increases whereas the proportion of tornado reports decreases (i.e., severe convective winds, severe hail, and tornadoes account for 82.0%, 16.3%, and 1.7% of the reports, respectively). There were
99 days (6.5% of all the days in the climatology) with at least 20 active grid boxes, which is a subjective metric of a fairly widespread severe convective storm event. This result means that roughly 10 fairly widespread severe convective storm events can be expected in the ALT Zone per warm season. The day with the most active grid boxes (90) was 13 May 2002 (a day in which an ALT was absent), although this day was a statistical anomaly since the second-most active grid boxes on a single day was 49. A pronounced peak in storm report frequency occurs during the midafternoon and early evening hours, with maxima at 2100 and 2200 UTC, whereas a minimum in storm report frequency occurs during the overnight and early morning hours (Fig. 3.6). During 6-h periods beginning at 1800 UTC with at least one active grid box, nearly twice as many storm reports occurred, on average, when an ALT was present at 1800 UTC (17.0) than when an ALT was absent at 1800 UTC (9.3). A possible reason for this important result will be discussed in section 3.4.

The temporal distribution of the first daily storm report for each ALT Zone sector (comprising 1272 storm reports) is shown in Fig. 3.7. A pronounced peak in first daily storm report frequency occurs during the early afternoon and early evening hours, with maxima at 1800, 1900, and 2000 UTC. This peak leads the peak in the time of all storm reports (Fig. 3.6) by 2–3 h. A secondary maximum in first daily storm report frequency occurs during the 0400 UTC hour, which likely exists simply because the day begins at that hour by definition. Differences in the peak time of the first daily storm report are evident for each sector. The South Sector exhibits a peak during the 1800 UTC hour, whereas the North Sector exhibits a peak during the 2000 UTC hour. The earlier peak in
the South Sector is likely due to stronger surface heating, which leads to the South Sector reaching the convective temperature earlier (on average).

3.3.2 Spatial Distribution

The spatial distribution of the percentage of 6-h periods with an active grid box for all ALT categories combined is shown in Fig. 3.8. Two local maxima in active grid boxes exist within the ALT Zone: 1) from northwestern South Carolina through southwestern Virginia, and 2) from northeastern Virginia through southeastern Pennsylvania. Overall, the percentage of active grid boxes is higher within the ALT Zone than outside of the ALT Zone, suggesting that the ALT Zone is a favored location for severe convective storms when an ALT is present.

Figure 3.9 shows the spatial distribution of the percentage of 6-h periods with an active grid box for each ALT category. Active grid boxes are favored over the southern half of the ALT Zone during category 1 periods (Fig. 3.9a). The distributions of active grid boxes within the ALT Zone in categories 2 and 3 (Figs. 3.9b,c) are qualitatively similar, with local maxima in the same locations as in Fig. 3.8. Two differences between the distributions of categories 2 and 3 are evident: 1) more active grid boxes occur over central North Carolina and central South Carolina in category 3, and 2) more active grid boxes occur northwest of the ALT Zone from Indiana and northeastern Kentucky through central Pennsylvania in category 2. The latter difference is likely a result of the varying position of the cold front between the two categories (i.e., in category 3, on average, the front has already passed through the area northwest of the ALT Zone that exhibits variation between the two categories). In category 4, a distinct maximum exists from
northeastern Virginia through southeastern Pennsylvania, including the Washington, DC,
and Philadelphia metropolitan areas (Fig. 3.9d). Also, relatively more storm reports
extend toward the Atlantic Coast in category 4 than in categories 2 and 3, which is likely
tied to frontal forcing for ascent spreading across the entire ALT Zone in category 4.

3.4 Convective Environments Characteristic of the ALT Zone

Figures 3.10a,b show the statistical distributions of MUCAPE and VWS,
respectively, calculated at the location of the first daily storm report for each ALT Zone
sector. Since the diurnal cycle is the dominant mode of temporal variability of storm
reports, MUCAPE and VWS were calculated only for first daily storm reports that
occurred between 1530 and 0029 UTC [comprising 1037 of the 1272 first daily storm
reports (81.5%)]. The distribution of MUCAPE (VWS) is highest in the South (North)
Sector and lowest in the North (South) Sector. These distributions are likely a reflection
of the ambient conditions representative of each sector during the warm season (i.e., the
North Sector experiences the highest mean VWS since it is located closest to the mean
upper-tropospheric jet, whereas the South Sector experiences the highest mean MUCAPE
since it is associated with the highest mean lower-tropospheric temperature and
humidity). MUCAPE and VWS at the location of the first daily storm report are plotted
in MUCAPE/VWS phase space for each sector in Figs. 3.11a–c. For each sector,
MUCAPE and VWS at the location of the first daily storm report does not appear to be a
good predictor of the number of active grid boxes on a particular day. However, for each
sector, a line can be drawn within MUCAPE/VWS phase space below which few first
daily storm reports occur. The monthly distributions of MUCAPE and VWS at the first
daily storm report are inversely related. For each of the sectors, the distributions of
MUCAPE are higher during JJA than during MS (Figs. 3.12a,c,e), whereas the distributions of VWS are higher during MS than during JJA (Figs. 3.12b,d,f). This inverse relationship is intuitive since the strongest surface heating occurs in JJA, whereas the mean upper-tropospheric jet is shifted equatorward (i.e., closer to the ALT Zone) in MS compared with JJA.

Figures 3.13a,b show observed MUCAPE statistical distributions for GSO, RNK, WAL, and IAD compared at ALT times and non-ALT times for 0000 and 1200 UTC. The observed MUCAPE distributions are greater at ALT times compared with non-ALT times for all four sites at both 0000 and 1200 UTC. The difference in MUCAPE between ALT times and non-ALT times is linked to the increased number of storm reports during the 6 h following the observation of an ALT compared with the 6 h following the absence of an ALT. Increased MUCAPE at ALT times relative to non-ALT times is intuitive since ALTs, by definition, contain lower-tropospheric thermal anomalies.

A comparison of the statistical distributions of MUCAPE at 0000 UTC between the observations, the NARR, and the CFSR for all four sites is shown in Figs. 3.14a–d. For each of the four sites, the MUCAPE distributions in the NARR and the CFSR are significantly less than the observations, and the MUCAPE distribution in the CFSR is less than the NARR at the median and at the 25th, 75th, and 90th percentiles. This discrepancy between the reanalyses and the observations demonstrates that caution should be used when attempting to interpret derived reanalysis variables such as MUCAPE quantitatively.
3.5 Composite Analysis

An absence of strong QG forcing for ascent in the immediate vicinity of the ALT Zone, as indicated by the absence of $Q$-vector convergence, is a characteristic of 2N, 2S, 3N, and 3S (Figs. 3.15a, 3.16a, 3.17a, 3.18a). Despite the similar absence of strong QG forcing for ascent, differences between the severe and nonsevere composites of categories 2 and 3 are evident in the position of broad upper troughs and ridges. A broad 500-hPa ridge is in place over the ALT Zone in 2N (Fig. 3.15a), whereas a broad 500-hPa trough is located over and to the northwest of the ALT Zone in 2S (Fig. 3.16a). The broad 500-hPa ridge is apparent in 21 out of the 22 2N composite members, whereas it is apparent in only 9 out of the 22 2S composite members. This difference implies that the presence of a broad 500-hPa ridge is a relatively good discriminator between category 2 active severe convective days and null days. A broad 500-hPa trough is present in 3N downstream of the ALT Zone (Fig. 3.17a), with weak QG forcing for descent over the northern portion of the ALT Zone. In contrast, a well-defined 500-hPa trough exists upstream of the ALT Zone in 3S, with weak QG forcing for ascent over portions of West Virginia and western Virginia (Fig. 3.18a). A broad, low-amplitude 500-hPa trough exists upstream of the ALT Zone in 4N, as well as two areas of QG forcing for ascent that are maximized over eastern Ohio and northeastern New York, respectively (Fig. 3.19a). In contrast, 4S features a sharper 500-hPa trough upstream of the ALT Zone, as well as an area of QG forcing for ascent just upstream of the ALT Zone from central Pennsylvania through southwestern Virginia (Fig. 3.20a). This area of QG forcing for ascent is located farther south in 4S than in 4N. The 500-hPa trough in 4S (Fig. 3.20a) is deeper and better
defined than the 500-hPa trough in 3S (Fig. 3.18a), leading to stronger QG forcing for ascent near the western boundary of the ALT Zone in 4S compared with 3S. Category 4S is unique in that it is the only category to feature strong QG forcing for ascent in the immediate vicinity of the ALT Zone.

The appearance of broad troughs and ridges in the composites is at least partially an artifact of the compositing procedure. A significant subset of the members of each category except 2N do not display the broad trough or ridge evident in the composite. To minimize the artificial broadening of the troughs and ridges in the composites, an alternative approach could involve compositing based on the position of the ridge or trough instead of based on the presence or absence of a frontal passage. However, this alternative approach may artificially broaden other signals evident in the composites. Despite the shortcomings of the compositing procedure with respect to the upper-tropospheric geopotential height field, the locations of troughs, ridges, and QG forcing for ascent relative to the ALT Zone are consistent with the presence or absence of severe convective storms in each composite.

Differences between the severe and nonsevere composites also arise with respect to the center of the composite surface low pressure. The surface low pressure is located northeast of Maine in 3N (Fig. 3.17b), whereas it is weaker and located near the triple point of New York, Ontario, and Quebec in 3S (Fig. 3.18b). However, the position of the surface low pressure in 3S appears to be a result of two preferred surface low pressure locations. In 3S, 7 (5) of the 17 members have surface low pressure centers in a preferred location near Lake Ontario (northeast of Maine). The number of 3S members with surface low pressure centers east and west of the composite surface low pressure center is
9 and 8, respectively, which explains the location of the composite surface low pressure.

In contrast, 11 of the 17 3N members had surface low pressure centers located in a similar location as the composite (northeast of Maine), and all but one of the members had surface low pressure centers east of 75°W. Knowledge of the preferred location of the surface low pressure in 3N can be useful in identifying possible null events on days that also exhibit other characteristics of the 3N composite. The surface low pressure is centered over southern Quebec in both 4N and 4S; however, it is roughly 5 hPa deeper in 4S than in 4N (Figs. 3.19b, 3.20b). In addition, the MSLP gradient and 10-m winds are greater in 4S than in 4N over nearly the entire area pictured in Figs. 3.19b and 3.20b. These facts suggest that, on average, a dynamically stronger system is present in 4S compared with 4N. Both 2N and 2S exhibit similar MSLP patterns with a windward ridge over the Appalachians and an ALT to the east of the ridge (Figs. 3.15b, 3.16b).

Winds at 10 m have a greater southerly component in 2S than in 2N over the Atlantic Ocean south of 40°N and east of 75°W, which is related to a stronger Bermuda High in 2S (Figs. 3.15b,c, 3.16b,c).

A feature common to 2S, 3S, and 4S is the presence of a plume of high MUCAPE extending poleward through the ALT Zone (Figs. 3.16d, 3.18d, 3.20d). A similar plume exists in 4N (Fig. 3.19d), although it is 250–750 J kg⁻¹ less than in 4S in the ALT Zone. Such a plume does not exist in 2N or 3N (Figs. 3.15d, 3.17d). In 2S, 3S, and 4S, higher MUCAPE in the ALT Zone exists to the east of the ALT (Figs. 3.16d, 3.18d, 3.20d). Higher MUCAPE in the severe composites compared with the nonsevere composites is related to higher 2-m dewpoints. Over the ALT Zone, 2-m dewpoints are 1–2°C greater in 2S compared with 2N (Figs. 3.15c, 3.16c) and 2–3°C greater in 3S compared with 3N.
(Figs. 3.17c, 3.18c). In 3S, 10-m winds over the ALT Zone have a greater southerly component than in 3N, especially over the northern portion of the ALT Zone (Figs. 3.17b,c, 3.18b,c). This greater southerly component allows moister air to be advected poleward in 3S, accounting for the increased dewpoints and MUCAPE plume. The westerly winds in 3N allow air that has subsided and dried from crossing the Appalachians to spread eastward across a larger portion of the ALT Zone compared with 3S.

VWS in the ALT Zone is weak (< 25 kt) in categories 2N, 2S, 3N, and 3S (Figs. 3.15d, 3.16d, 3.17d, 3.18d). Weak VWS in these categories is consistent with the lack of strong QG forcing for ascent described above. VWS is stronger in 4N and 4S than in the other categories. The 20-kt VWS contour is located across central Virginia in 4N (Fig. 3.19d), whereas it is located farther south over northern South Carolina and southern North Carolina in 4S (Fig. 3.20d). VWS in the ALT Zone north of northern Virginia increases to 20–30 kt in 4N, and 30–35 kt in 4S. A poleward-extended plume of high MUCAPE juxtaposed with > 30 kt VWS is a signature of 4S and is present in 11 out of the 17 members of the composite. This juxtaposition is also collocated with the category 4 storm report maximum (Fig. 3.9d). The stronger QG forcing and higher VWS in 4S compared with all other categories suggest that 4S severe convective storm events are more spatially organized and favor the Washington, DC, to Philadelphia corridor.

The equivalent potential temperature ($\theta_e$) deficit (TeD; Pryor and Ellrod 2004), defined here as the difference between $\theta_e$ at the surface and the minimum $\theta_e$ value between 750 and 450 hPa, is shown in panel (e) of Figs. 3.15–3.20. As part of the Microburst and Severe Thunderstorm (MIST) project in northern Alabama, Atkins and
Wakimoto (1991) found that wet microbursts were associated with TeD values of > 20 K in a high CAPE, low VWS environment. Wheeler and Roeder (1996) performed a similar study for Cape Canaveral, FL, and found that wet microbursts were associated with TeD values of > 30 K. TeD values in 2N (2S) range from 20–25 K (25–30 K) over the vast majority of the ALT Zone (Figs. 3.15e, 3.16e). Both 3N and 3S are characterized by TeD values > 20 K over the ALT Zone, with > 25 K in 3S over much of eastern Virginia and eastern and central North Carolina (Figs. 3.17e and 3.18e). TeD values over the ALT Zone in 2N, 2S, 3N, and 3S appear to be conducive to wet microbursts using the criterion of Atkins and Wakimoto (1991), with slightly higher TeD values in the severe composites compared with the nonsevere composites.

Differences between the severe and nonsevere composites are also evident with respect to midtropospheric lapse rates. The maximum midtropospheric lapse rate (between 850 and 500 hPa) with a depth of 200 hPa (hereafter MaxLR) is shown in panel (e) of Figs. 3.15–3.20. MaxLR values are greater in the severe composites compared with the nonsevere composites, which projects onto the greater MUCAPE values evident in the severe composites.

3.6 Case Study

3.6.1 Event Overview

The 6 June 2002 severe convective storm event was chosen for analysis because it was a high-impact event, it contained features similar to the ALT composites, and both the ALT and a trailing cold front played roles in initiating and sustaining severe convective storms. From 0400 UTC 6 June to 0400 UTC 7 June, 93 storm reports and 48
active grid boxes were recorded within the ALT Zone, making it the category 4 day with the most active grid boxes, and the day with the third-most active grid boxes on all days during the climatology. Overall, 112 storm reports were associated with this event (Fig. 3.21). Five injuries were reported in the ALT Zone due to felled trees, and Maryland’s 460-yr-old state tree, the largest white oak in the world, was toppled by severe convective winds (Baltimore Sun, 7 June 2002). Notable storm reports from this event included a 61-kt (31.4-m s\(^{-1}\)) wind gust recorded at Easton, MD; hail with a diameter of 2.75 in (7 cm; baseball-sized) measured at Cedar Run, NJ; and an EF-1 tornado reported in Orange, VA. The initial (final) storm report of the day within the ALT Zone occurred at 1830 UTC 6 June (0215 UTC 7 June).

3.6.2 Synoptic Overview

At 1800 UTC 6 June, a 300-hPa trough axis ran from eastern Upper Michigan southward toward the triple point of Illinois, Indiana, and Kentucky (Fig. 3.22a). An area of upper-tropospheric divergence downstream of the trough and within the equatorward jet entrance region was located just upstream of the ALT Zone (Fig. 3.22a). At the surface, a cold front extended from Nova Scotia to northern New Jersey to central West Virginia to central Tennessee (Fig. 3.22b). Southeast of the front, an ALT was evident within the ALT Zone running from the intersection with the cold front in New Jersey southwestward to Georgia (Fig. 3.22b). By 0000 UTC 7 June, the 300-hPa trough and equatorward jet entrance region had translated eastward, placing much of the ALT Zone into an area of upper-tropospheric divergence (Fig. 3.23a). This divergence was maximized over far southeastern Pennsylvania (Fig. 3.23a), where a weak surface frontal wave had developed (Fig. 3.23b). The surface cold front extended from New Jersey
southwestward through central North Carolina and had overtaken the northern portion of the ALT (Fig. 3.23b). Meanwhile, the southern portion of the ALT extended from southeastern Virginia southwestward into Georgia. The deformation of the baroclinic zone in the vicinity of the frontal wave occurred as the cold front accelerated from the Appalachians to the Atlantic Coast. This acceleration was likely aided by convectively-driven cool outflow, discussed further in section 3.6.3.2.

3.6.3 Summary of Radar and Storm Reports

Two distinct regimes of severe convective storms were evident via radar imagery on 6 June 2002: (1) a prefrontal regime and (2) an alongfront regime.

3.6.3.1 Prefrontal Convective Storms

Discrete prefrontal convective storms began to form near a wind-shift boundary over western Virginia by 1700 UTC (Fig. 3.24a). These storms (inferred from reflectivity values > 50 dBZ) strengthened by 1800 UTC west of the ALT, which was collocated with an axis of high $\theta_e$ (355–365 K; Figs. 3.24b,c). The ALT was located west of a NARR-analyzed MUCAPE maximum (Fig. 3.25a), TeD values of 25–35 K, and MaxLR values of 6.5–7 K km$^{-1}$ (Fig. 3.25b). In reality, the MUCAPE and TeD maxima were likely located closer to the ALT within the axis of high observed $\theta_e$ (Figs. 3.24b,c). A 24-h backward trajectory analysis (Fig. 3.26) shows that parcel trajectories originating in the lee of the Appalachians at 100 m AGL at 1700 UTC did not pass over high elevations of the Appalachians, and therefore did not experience any downslope drying. The lack of downslope drying was a major contributor to the moist near-surface air mass (e.g., surface dewpoints of 20–24 °C) that characterized the ALT Zone at the time of
convective initiation. Steep lower-tropospheric lapse rates were evident from 1200 UTC soundings at WAL and IAD, which exhibited 950–700-hPa lapse rates of 7.9 and 6.7 K km$^{-1}$, respectively (Figs. 3.27a,b). These steep lapse rates contributed to the high MUCAPE values east of the ALT. The discrete prefrontal convective storms grew upscale into clusters and line segments and moved eastward (Figs. 3.24d–i), continuing to be located west of both the ALT and an axis of $\theta_e$ values ranging from 345–360 K at 2100 UTC (Figs. 3.24f,g). These storms reached the Atlantic Coast by 2300 UTC and accounted for 51 storm reports from 1830 to 2300 UTC, including all 16 hail reports from the event.

A region of MUCAPE > 1000 J kg$^{-1}$ at 1800 UTC extended from South Carolina northward into central Delaware (Fig. 3.25a). The northern portion of this region of MUCAPE > 1000 J kg$^{-1}$ was collocated with a region of 20–40-kt VWS (Fig. 3.25a). This juxtaposition of a poleward-extended plume of MUCAPE with > 30 kt of VWS is a signature of the 4S composite (Fig. 3.20d). The discrete prefrontal convective storms that initiated over central and northern Virginia were in a region of 25–40-kt of VWS, which likely helped sustain and intensify the storms from 1800 to 2000 UTC (Figs. 3.24b–e, 3.25a). These storms produced 11 storm reports from 1800 to 2000 UTC. Meanwhile, the discrete prefrontal convective storms that initiated farther south over southern Virginia and northern North Carolina were in a region of < 20 kt of VWS, and these storms did not produce any storm reports from 1800 to 2000 UTC (Figs. 3.24b–e, 3.25a).
3.6.3.2 Alongfront Convective Storms

A broken line of convective storms along the eastward-advancing cold front was ongoing at 1800 UTC from southern West Virginia south-southwestward through eastern Tennessee (Figs. 3.24b,c). Ample lower-tropospheric moisture continued to exist over central Virginia at 2100 UTC in the wake of the prefrontal convective storms, as manifested by dewpoints of 21–23°C and a plume of $\theta_e$ values of 350–355 K between the prefrontal and alongfront convective storms (Fig. 3.24g). The alongfront convective storms intensified upon interacting with this warm, moist air from 2000 to 2200 UTC (Figs. 3.24e–h). These storms accounted for 20 storm reports from 2100 to 2200 UTC, including both of the tornado reports from the event (Fig. 3.24h).

The cold front had accelerated southeastward and had overtaken the northern portion of the ALT by 0000 UTC 7 June (Fig. 3.24j). Meanwhile, the ALT was still evident ahead of the cold front from southeastern Virginia to Georgia (Fig. 3.24j). The northern portion of the alongfront convective storms had weakened by 2300 UTC (Fig. 3.24i). Meanwhile, a cluster of nonsevere convective storms, which was a remnant of the initial nonsevere discrete prefrontal convective storms, was located near the Virginia/North Carolina border at 2300 UTC (Fig. 3.24i). These storms began to assume a linear mode over southeastern Virginia and northeastern North Carolina by 0000 UTC 7 June (Figs. 3.24j,k). This linear mode occurred as the cold front caught up with the cluster of storms. An increase in VWS to 20–30 kt, associated with the approaching front, was evident between 1800 UTC 6 June (Fig. 3.25a) and 0000 UTC 7 June (Fig. 3.28) in the vicinity of the storms assuming a linear mode. The increase in VWS, along
with convergence along the cold front, likely helped strengthen the storms and maintain
their linear organization as they produced nine storm reports from 2300 UTC 6 June to
0100 UTC 7 June (Figs. 3.24j–l). The storms also interacted with a moist lower
troposphere (e.g., θ_e values of 350–355 K at 0000 UTC; Fig. 3.24k) and advanced to the
Atlantic Coast by 0600 UTC (not shown). The southeastward acceleration of the cold
front from the Appalachians to the Atlantic Coast throughout this event was likely aided
by convectively driven cool outflow from the alongfront storms, as evidenced by severe
convective winds and one-hour temperature drops of 5–10°C as the alongfront storms
passed (e.g., Figs. 3.24e–l).

3.6.4 Diagnosis of Severe Convective Storms

Severe convective winds are primarily a result of downdrafts, which are often
initiated by condensate loading and diabatic cooling from precipitation phase changes
(Wakimoto 2001). Midtropospheric dry air can aid in initiating a downdraft by
enhancing evaporational cooling from precipitation falling into the dry air (Wakimoto
2001). Midtropospheric dry air at 550–600 hPa was in place over the ALT Zone on 6
June as sampled by the 1200 UTC WAL and IAD soundings (Figs. 3.27a,b). In contrast,
dry air can be detrimental to sustaining a downdraft by decreasing the virtual temperature
difference between the descending parcel and the environmental air (Srivastava 1985;
Wakimoto 2001). A relatively moist layer in the 650–800-hPa layer is evident below the
midtropospheric dry air in the WAL and IAD soundings (Figs. 3.27a,b). A NARR model
sounding was taken at 2100 UTC at Fort Eustis, VA (FAF; location plotted on Fig.
3.24g), just east of a cluster of storms that would produce nine severe convective wind
reports from 2100 to 2200 UTC in the vicinity of FAF (Fig. 3.24h). The sounding shows
a layer of dry air from 500 to 700 hPa, with a moist layer from 750 to 900 hPa, and a well-mixed layer from 900 hPa to the surface (Fig. 3.29a). This sounding is similar to a conceptual model of environments conducive for wet microbursts (Fig. 3.29b) created by Atkins and Wakimoto (1991) from the MIST field project. The NARR-analyzed vertical $\theta_e$ profile at FAF (Fig. 3.29c) is also similar to the conceptual model (Fig. 3.29d). The NARR-analyzed TeD value at FAF was 30 K, equal to the wet microburst criterion of Wheeler and Roeder (1996) and 10 K greater than the wet microburst criterion of Atkins and Wakimoto (1991).

The two tornadoes that occurred during this event were located within the broken line of alongfront convective storms in the area between the discrete prefrontal convective storms and the approaching cold front (Fig. 3.24h). An EF-0 tornado with a path length of 2 miles (3.2 km) occurred near Frederick, MD (FDK; location plotted on Fig. 3.24g), from 2105 to 2113 UTC, and an EF-1 tornado with a path length of 1 mile (1.6 km) occurred near Orange, VA (OMH; location plotted on Fig. 3.24g), from 2145 to 2150 UTC. Figures 3.30a,b show a 1700 UTC dewpoint depression of 8°C at FDK and 12°C at OMH. As the discrete prefrontal convective storms passed the two sites from 1800 to 1900 UTC, the temperature decreased and the dewpoint stayed constant or increased slightly. As a result, the dewpoint depression decreased to 2°C at both sites by 2100 UTC. In addition, the wind direction at 2100 UTC backed to south-southeasterly at FDK and southerly at OMH. Backed winds were observed at several stations between the discrete prefrontal convective storms and the approaching alongfront storms from 2000 to 2100 UTC (Fig. 3.31). Replacing the NARR-derived boundary-layer winds at FDK and OMH with the observed winds would result in a vertical wind profile that veers...
with height (Figs. 3.32a,b). The resulting wind profile, along with low LCL heights inferred from the small dewpoint depressions, likely provided an environment favorable for brief tornadoes as the alongfront convective storms approached.

3.6.5 Summary

The 6 June 2002 severe convective storm event featured two distinct regimes of severe convection. The first regime initiated as discrete prefrontal storms along a wind-shift boundary just west of an ALT. These storms grew upscale into clusters and line segments and produced widespread storm reports from central Virginia to the Atlantic Coast as the storms were sustained by a convectively unstable environment east of the initiation region. The second regime occurred as a broken line of storms along an eastward-advancing cold front intensified upon encountering an air mass within the ALT Zone that had not convectively stabilized in the wake of the prefrontal convective storms. These alongfront storms produced several storm reports from southern Pennsylvania through north-central North Carolina from 2100 to 2200 UTC, and sporadic storm reports from 2200 UTC 6 June to 0000 UTC 7 June as the broken line of storms advanced southeastward across southern Virginia and central North Carolina. Remnant storms from the southern portion of the first regime strengthened and assumed a linear mode from 2300 UTC 6 June to 0100 UTC 7 June over southeastern Virginia and northeastern North Carolina as VWS values increased (associated with the approaching cold front). This line of storms advanced to the Atlantic Coast by 0600 UTC and was responsible for several more storm reports.
An upstream 300-hPa trough and equatorward jet entrance region, in addition to convergence along the surface cold front and wind shift boundary west of the ALT, provided forcing for ascent during this event. Both regimes of convective storms were sustained by the aforementioned convective instability along with VWS values of 20–40 kt. Midtropospheric dry air located above moist air likely initiated and sustained convective downdrafts, and vertical $\theta_e$ profiles were reminiscent of conceptual models of storms that produce wet microbursts. A decrease in the dewpoint depression in association with the passage of the prefrontal convective storms, along with locally backed lower-tropospheric wind directions, were associated with the occurrence of two brief tornadoes within the broken line of alongfront convective storms.
Fig. 3.1. Composite of warm-season ALTs (N=13) at (a) 1800 UTC and (b) 0000 UTC. MSLP (black contours, hPa), 1000–850-hPa thickness (fills, dam), 1000–850-hPa thermal vorticity (white contours, only negative values contoured every $2\times10^{-5}$ s$^{-1}$), and 10-m wind (barbs, kt).
Fig. 3.2. Percentage of occurrence of ALTs during the climatology with respect to varying MSLP and thermal anomaly thresholds. Bubbles are sized proportionally to the percentage of ALT occurrence. Red boxes indicate the anomaly thresholds adopted to define ALTs and the percentage of occurrence of ALTs with respect to those thresholds.

Fig. 3.3. Percentage of occurrence of ALTs by (a) CFSR analysis time and (b) month.
Fig. 3.4. Percentage of occurrence of each ALT category.

Fig. 3.5. Monthly percentage of occurrence for (a)–(d) ALT categories 1–4.
Fig. 3.6. Histogram of all storm reports in the ALT Zone (N=12 330) by hour of occurrence.

Fig. 3.7. Histogram of normalized first daily storm reports in each ALT Zone sector by hour of occurrence.
Fig. 3.8. Spatial distribution of the percentage of 6-h periods with an active grid box for all ALT categories combined.
Fig. 3.9. Spatial distribution of the percentage of 6-h periods with an active grid box for (a)–(d) ALT categories 1–4.
Fig. 3.10. Box-and-whisker plots of the distributions of (a) MUCAPE and (b) VWS calculated at the location of the first daily storm report of each ALT Zone sector. Only storm reports occurring between from 1530 to 0029 UTC are shown. Horizontal line denotes median, box edges denote 25th and 75th percentiles, and whiskers denote 10th and 90th percentiles.
Fig. 3.11. Scatterplots of MUCAPE and VWS calculated at the location of the first daily storm report in the (a) South Sector, (b) Center Sector, and (c) South Sector. Red (blue) dots indicate days in which at least (less than) 5 active grid boxes occurred. Median MUCAPE and VWS for days in which at least (less than) 5 active grid boxes occurred are listed in red (blue) type. The black line was arbitrarily drawn to delineate MUCAPE/VWS phase-space in which first daily storm reports are rare. In the South, Center, and North Sectors, 1.8%, 0.9%, and 0% of days with storm reports had MUCAPE and VWS values located below the black line, respectively.
Fig. 3.12. Box-and-whisker plots of the distributions of (a) South Sector MUCAPE, (b) South Sector VWS, (c) Center Sector MUCAPE, (d) Center Sector VWS, (e) North Sector MUCAPE, and (f) North Sector VWS calculated at the first daily storm report and partitioned by month. Plots are according to the convention in Fig. 3.10.
Fig. 3.13. Box-and-whisker plots of the distributions of observed MUCAPE on days with observed MUCAPE > 0 at sounding sites within or near the ALT Zone compared at ALT times (orange) and non-ALT times (purple) for (a) 0000 UTC and (b) 1200 UTC. Plots are according to the convention in Fig. 3.10.
Fig. 3.14. Box-and-whisker plots of the distributions of MUCAPE on days with observed MUCAPE > 0 compared between observations (Obs), NARR, and CFSR, for (a) GSO, (b) RNK, (c) WAL, and (d) IAD. Plots are according to the convention in Fig. 3.10.
Fig. 3.15. ALT category 2N composites of (a) 500-hPa geopotential height (black contours, hPa), 500-hPa $Q$ vectors (arrows, $> 5 \times 10^{-11}$ K m$^{-1}$ s$^{-1}$; reference vector at the bottom of the panel), and 500-hPa $Q$-vector divergence (fills, $10^{-16}$ K m$^{-2}$ s$^{-1}$); (b) MSLP (black contours, hPa), 1000–850-hPa thickness (fills, dam), and 10-m winds (barbs, kt); (c) MSLP (black contours, hPa), 2-m dewpoint (fills, °C; 20°C isotherm contoured in white for reference), and streamlines of the 10-m wind; (d) VWS (black contours, $\geq$ 20 kt) and MUCAP (fills, J kg$^{-1}$); and (e) TeD [contours, K; values greater than or equal to (less than) 20 K contoured in blue (black)] and MaxLR (fills, K km$^{-1}$).
Fig. 3.16. As in Fig. 3.15 but for ALT category 2S.
Fig. 3.17. As in Fig. 3.15 but for ALT category 3N.
Fig. 3.18. As in Fig. 3.15 but for ALT category 3S.
Fig. 3.19. As in Fig. 3.15 but for ALT category 4N.
Fig. 3.20. As in Fig. 3.15, but for ALT category 4S.
Fig. 3.21. Location of storm reports occurring from 0400 UTC 6 June to 0400 UTC 7 June 2002.
Fig. 3.22. NARR analysis at 1800 UTC 6 June 2002 of (a) 300-hPa geopotential height (black contours, dam), divergence of the horizontal wind (green contours, only positive values contoured every $2 \times 10^{-5} \text{ s}^{-1}$), and wind speed (fills, m s$^{-1}$); (b) MSLP (black contours, hPa), 1000–850-hPa thickness (fills, dam), and 10-m winds (barbs, kt). In (b), ALT is denoted by black dashed line.
Fig. 3.23. As in Fig. 3.22 except at 0000 UTC 7 June 2002.
Storm initiation along wind-shift boundary
Fig. 3.24. Radar reflectivity (dBZ) and surface observations of (counterclockwise from top right) abbreviated altimeter setting (hPa), temperature (°C), dewpoint (°C), $\theta_e$ (K), and winds (barbs, kt) from 6 June 2002 at (a) 1700, (b), (c) 1800, (d) 1900, (e) 2000, (f), (g) 2100, (h) 2200, and (i) 2300 UTC; and from 7 June 2002 at (j), (k) 0000, and (l) 0100 UTC. Severe convective wind reports (blue dots), severe hail reports (green dots), and tornado reports (red dots) occurring 0–59 min prior to the time of each image are shown. A manual analysis of altimeter setting (black contours), frontal position, and ALT position (dashed line) is shown in panels (b), (f), and (j). A manual analysis of $\theta_e$ (black contours) is shown in panels (c), (g), and (k).
Fig. 3.25. NARR analysis at 1800 UTC 6 June 2002 of (a) VWS (black contours, kt) and MUCAPE (fills, J kg\(^{-1}\)), and (b) TeD [contours, K; values greater than or equal to (less than) 20 K contoured in blue (black)] and MaxLR (fills, K km\(^{-1}\)). ALT is denoted by black dashed line.
Fig. 3.26. NOAA HYSPLIT 24-h backward parcel trajectories originating at 100 m AGL at 1700 UTC 6 June 2002. Red, blue, and yellow correspond to trajectories originating at 36.0°N, 37.4°N, and 38.8°N, respectively.
Fig. 3.27. Skew $T$–log $p$ diagram of temperature (°C), dewpoint (°C), and wind (barbs, kt) at 1200 UTC 6 June 2002 for (a) WAL and (b) IAD.
Fig. 3.28. As in Fig. 3.25a except at 0000 UTC 7 June 2002.
Fig. 3.29. Skew $T$–log $p$ diagram of NARR-derived (a) temperature (black line, °C), dewpoint (red line, °C), and wind (barbs, kt) for FAF at 2100 UTC 6 June; (b) a conceptual model of an environment conducive to wet microbursts from Atkins and Wakimoto (1991) showing temperature (solid line, °C) and dewpoint (dashed line, °C). Panels (c) and (d) are the same as (a) and (b), respectively, except that they are log-$p$ diagrams of the vertical profile of $\theta_e$. 
Fig. 3.30. Meteogram from 1700 to 2200 UTC 6 June 2002 showing wind (barbs, kt), temperature (black line, °C), and dewpoint (red line, °C) at (a) FDK and (b) OMH. Red arrows correspond to the approximate tornado touchdown time for each station.
Fig. 3.31. Shows the stations between the prefrontal and alongfront convective storms that exhibited surface winds that backed from 2000 to 2100 UTC. Wind bars (kt) at 2000 (2100) UTC plotted in red (blue).
Fig. 3.32. As in Fig. 3.29a but for (a) FDK and (b) OMH. Observed 2100 UTC surface winds are shown to the right of the NARR-derived winds.
4. Discussion, Conclusions, and Suggestions for Future Work

4.1 Discussion

4.1.1 ALT Climatology

Results from a 10-yr climatology indicate that the frequency of ALT occurrence is 26.6% during the warm season (May–September). In a 2-yr climatology, Weisman (1988) found that ALTs occur on 40% of all days during the warm season. Most of the discrepancy between the ALT frequency value from Weisman (1988) and the value shown in the present study can be attributed to how often ALTs were searched for in the climatology. Weisman (1988) recorded the number of “ALT days” during the climatology (i.e., the number of calendar days during which an ALT was recorded at least once), whereas the present study checked for the presence of an ALT every 6 h during the climatology and recorded the total number of ALTs. Results from the present study indicate that 46.1% of all calendar days during the climatology (i.e., 706 out of 1530) would be classified as “ALT days”. The remaining discrepancy can probably be attributed to sampling differences (i.e., a difference in the length of the climatology between the two studies with no overlap in the period of time examined), as well as different data sources and methodology used to define an ALT between the two studies. Weisman (1988) used manually analyzed surface charts from the National Meteorological Center (NMC) to identify ALTs, recording an ALT whenever one of the following three criteria was met: 1) the NMC had analyzed a surface trough over the mid-Atlantic, 2) “baggy isobars” existed over the mid-Atlantic, or 3) an MSLP difference of ≥ 1 hPa existed between the central Carolinas and stations to the east and west. In contrast,
the present study performed an algorithm on CFSR gridded reanalysis data to objectively identify an ALT whenever certain criteria were met (see section 2.2.1). Although the ALT identification algorithm used in this thesis was objective, some subjectivity in the establishment of the criteria and the identification of false alarms could not be avoided. Nevertheless, the results from the ALT climatology seem to reflect an appropriate frequency of warm-season ALTs and are consistent with the ALT climatology of Weisman (1988) once the differences in data and methodology are taken into account.

In the CFSR composite of 13 ALTs whose features were used to derive the climatology criteria (Figs. 3.1a,b), two distinct 1000–850-hPa thermal vorticity minima were evident in the lee of the Appalachians from central Virginia to South Carolina: 1) in the immediate lee of the Appalachians, and 2) just inland of the Atlantic Coast. The negative thermal vorticity implies a warm core in the 1000–850-hPa layer. Assuming hydrostatic balance, the ALT would be expected to be collocated with the two thermal vorticity minima. However, the composite ALT is located between the two thermal vorticity minima. One possible reason for this apparent inconsistency is that the composite procedure, by definition, will smooth signals evident in each of the composite members. An alternative explanation is linked to the unique geography of the mid-Atlantic. The ALT Zone is located between the Appalachians and Atlantic Ocean, which both tend to be climatologically cooler than the ALT Zone during the warm season. As a result, a larger-scale trough would be expected to form within the ALT Zone to satisfy hydrostatic balance. It is likely that smaller-scale troughs appear in the immediate lee of the Appalachians and along the coast. On average, these smaller-scale troughs are likely
superimposed upon the larger-scale ALT, and are better resolved by the 32 km × 32 km NARR than the 0.5° × 0.5° CFSR.

4.1.2 ALT Categorization

ALTs identified in the climatology were categorized according to their relationship to synoptic-scale cold fronts. ALTs that occurred in advance of synoptic-scale cold fronts were also considered prefrontal troughs (PFTs). Manual inspection of instances of the two PFT categories (i.e., ALT categories 3 and 4) revealed that the PFTs were mobile and tied to the movement of the parent cyclone and upstream cold front. Many PFTs evolved according to the conceptual model of prefrontal wind shifts in the lee of the Rocky Mountains shown in Hutchinson and Bluestein (1998), where a lee trough that is initially anchored to the terrain (Fig. 1.5a) moves eastward in association with warm advection to the east of the lee trough (Fig. 1.5b), while an approaching cold front begins to overtake northern portions of the lee trough so that the two features merge at the surface (Fig. 1.5c). Similar lee trough movement has been observed in other studies of troughs forming in the lee of the Rocky Mountains and moving onto the Great Plains (e.g., Gaza and Bosart 1985; Locatelli et al. 1989; Steenburgh and Mass 1994; Martin et al. 1995).

Sanders (1999) presented a conceptual model that was similar to the conceptual model of Hutchinson and Bluestein (1998), except that quasigeostrophic theory was used to predict that the PFT would move faster than the cold front, thereby increasing the separation between the two features. Both conceptual models feature a PFT that moves eastward, but the difference between the two models lies in the initial formation of the
PFT and the subsequent movement of the cold front. In the model presented by Hutchinson and Bluestein (1998), the PFT formed orographically (i.e., downsloping winds caused lee troughing), and at a later time, a cold front overtook northern portions of the PFT. In contrast, orography did not play an explicit role in the conceptual model presented by Sanders (1999). Also, the conceptual model presented by Sanders (1999) showed an initial front without a PFT, while at a later time the PFT formed and moved eastward relative to the front as the front dissipated.

Both of these conceptual models could apply to the ALTs observed in the present study. During the climatology, many category 2 ALTs (i.e., non-PFTs) became category 3 ALTs (i.e., PFTs) as a synoptic-scale cold front approached. Also, as mentioned above, portions of some category 3 and 4 ALTs were overtaken by the approaching cold front. Both of these interactions between the ALT and the cold front fit the conceptual model of Hutchinson and Bluestein (1998). Alternatively, some of the cold fronts associated with category 3 ALTs stalled or dissipated while the ALT itself moved eastward, a progression that fits the conceptual model of Sanders (1999). However, category 4 ALTs cannot fit the conceptual model of Sanders (1999), since by definition, in category 4, the cold front passes through the entire ALT Zone and does not dissipate.

4.1.3 Distributions of Severe Convective Storms in the Mid-Atlantic

A pronounced maximum in storm report frequency was found during the midafternoon and evening hours, with maxima at 2100 and 2200 UTC, whereas a minimum was found during the overnight and early morning hours (Fig. 3.6). Many prior studies have found similar diurnal variability in severe convective storms in the
mid-Atlantic using radar (e.g., Koch and Ray 1997; PA07; Lombardo and Colle 2010; MC11) and/or lightning data (e.g., Reap and Orville 1990; Weisman 1990b; Orville and Huffines 2001; Zajac and Rutledge 2001; MK05; MC11) as a proxy for convective storms. The present study, however, is the first one to specifically examine the diurnal variability of severe convective storms over the mid-Atlantic. Zajac and Rutledge (2001) showed that the diurnal lightning activity cycle was very pronounced between the Appalachians and the Atlantic Coast during the warm season, whereas the diurnal cycle is not the dominant mode of temporal variability in areas farther west such as the Great Plains and Upper Midwest (see their Fig. 10b). In fact, the diurnal maximum in lightning over the Great Plains and Upper Midwest occurs during the late evening and overnight hours, which can be attributed to the numerous nocturnal MCSs that move across these regions during the warm season (McAnelly and Cotton 1989). The lack of storm reports during the overnight hours in the present study suggests that severe nocturnal MCSs are rare to the lee of the Appalachians. This assertion is backed by the results of MK05, who studied warm-season lightning distributions in an area centered on the Appalachians in North Carolina. They found that warm-season MCSs were rare in their region of study, and most of those that did occur stayed to the west of the Appalachians.

Despite the diurnal cycle being the dominant mode of variability in convective storms in the mid-Atlantic, several studies have found an eastward-propagating convective signal (e.g., Orville and Huffines 2001; MK05; PA07; MC11). In particular, PA07 and MC11 each found a favored convective initiation region along and in the immediate lee of the Appalachians in the early afternoon hours, with the convection progressing eastward through the afternoon and evening (see their Figs. 7 and 11,
respectively). MK05 attributed the preferred convective initiation region over the Appalachians to upslope flow and lower-tropospheric convergence. The present study found storm report maxima during the 6 h following an ALT in the immediate lee of the Appalachians (i.e., just inside the western boundary of the ALT Zone; Fig. 3.8). These storm report maxima are likely related to the preferred convective initiation regions and eastward-propagating convective signal found by PA07 and MC11. In an average sense, storms that initiate over the Appalachians and move eastward can be expected to strengthen upon reaching a more strongly heated boundary layer and a more convectively unstable air mass at lower elevations compared to over the Appalachians.

During 6-h periods beginning at 1800 UTC with at least one active grid box, nearly twice as many storm reports occurred, on average, when an ALT was present at 1800 UTC (17.0) than when an ALT was absent at 1800 UTC (9.3). This result is linked to increased MUCAPE at sounding locations near the ALT Zone at times when an ALT was present compared with times when an ALT was absent (Figs. 3.13a,b). Furthermore, the spatial distribution of storm reports during the 6 h following an ALT shows that the ALT Zone is a favored location for storm reports compared with locations outside of the ALT Zone (Fig. 3.8). MC11 showed a similar result in their composite of convectively active warm-season days over the mid-Atlantic and northeastern U.S., since an ALT was evident in their composite (Fig. 1.8c). Furthermore, MC11 found that 65% of days with lightning frequency over two standard deviations above the mean in northern Chesapeake Bay were associated with a PFT to the lee of the Appalachians. The present study hypothesizes that, at first-order, a surface trough in the lee of the Appalachians (i.e., an ALT or PFT) can be considered a “marker” for an axis of convective instability, due to
the collocation of the ALT or PFT with a lower-tropospheric thermal maximum. As a result, the presence of an ALT or PFT can serve as an alert to forecasters for the potential of active severe convection.

4.1.4 Convective Environments Characteristic of the ALT Zone

A comparison of MUCAPE values between observations derived from radiosondes and two reanalysis datasets (CFSR and NARR) revealed that both reanalyses consistently underestimated MUCAPE values relative to the observations (Figs. 3.14a–d). Part of the discrepancy between the observations and the reanalyses could be due to the differing methods of calculating MUCAPE between the three sources. It is also possible that the MUCAPE underestimation in the reanalyses is partially due to difficulties in accurately parameterizing the boundary layer. If that is the case, the reanalyses may also underestimate surface temperature, dewpoint, and derived variables such as TeD. As a result, caution should be exercised when attempting to interpret these reanalysis-derived boundary layer variables quantitatively.

MUCAPE and VWS were calculated at the location of the first daily storm report for each ALT Zone sector. These values did not show much skill in differentiating between days with more or less than five active grid boxes; however, for each sector, a line can be drawn within MUCAPE/VWS phase space below which few to no storm reports occur (Figs. 3.11a–c). The underestimation of MUCAPE in the NARR gives added confidence that this line separates environments that may be conducive for severe convective storms from environments that are not conducive for severe convective storms.
4.1.5 Composite Analysis

Prior works (e.g., Benjamin and Carlson 1986; Steenburgh and Mass 1994; Hobbs et al. 1996; see their sections 4c, 6d, and Fig. 1, respectively) have shown that Rocky Mountain lee troughs can act as boundaries between dry air to the west of the trough associated with westerly downslope winds and moist air to the east of the trough that is advected poleward by southerly winds. It is reasonable to assume that the downsloping westerlies would suppress convection to the west of the lee trough. However, the present study does not show suppressed convection in the immediate lee of the Appalachians; in fact, the immediate lee of the Appalachians is the location of the storm report maxima during the 6 h following the presence of an ALT (Fig. 3.8). A possible reason why severe convection is not suppressed in the immediate lee of the Appalachians during ALT days is evident upon examination of the 10-m wind fields of the initial ALT composite (Figs. 3.1a,b). In this composite, winds between the Appalachians and the ALT are southwesterly, not westerly. Manual inspection reveals that 8 of the 13 1800 UTC composite members (62%) exhibited 10-m meteorological wind directions of 180°–240° (i.e., between southerly and west-southwesterly) over the majority of the portion of the ALT Zone that lies between the Appalachians and the ALT. The lack of westerly winds between the Appalachians and the ALT likely means that drier air that has descended from the Appalachians does not make it far, if at all, into the ALT Zone (although a trajectory analysis would be needed to confirm this hypothesis).

Meteorological wind directions of 180°–240° at 10 m in the immediate lee of the Appalachians are evident to a degree in the 2S and 3S composites as well (Figs. 3.16b,c, 3.17b,c). Manual inspection of the members of 2S and 3S reveals that 77% and 71% of
the members, respectively, exhibit 10-m meteorological wind directions of 180°–240° over the majority of the portion of the ALT Zone that lies between the Appalachians and the ALT. The 6 June 2002 case study also showed meteorological wind directions of 180°–240° over the majority of the ALT Zone both to the east and the west of the ALT (discussed further in the next section).

The present study shows that the ALT does not always mark the location of a sharp wind-shift boundary. In fact, most of the composite members for all ALT categories did not exhibit a sharp wind shift anywhere within the ALT Zone; rather, many showed a gradual backing of the winds from westerly to southerly from the Appalachians to the Atlantic Coast. Manual analysis of the composite members for each of the six severe and nonsevere ALT categories reveals that only 15 of the 112 total members (13%) exhibited a wind shift of > 45° over 100 km anywhere between the Appalachians and the Atlantic Coast. As a result, in almost all of the composite members, the ALT did not act as a sharp dewpoint boundary, as lee troughs originating from downslope flow across the Rocky Mountains sometimes do. Rather, the dewpoint often gradually increased from west to east across the ALT Zone, as depicted in the composite analyses [panel (c) of Figs. 3.15–3.20]. Despite the lack of sharp wind-shift boundaries evident in the reanalyses it is likely that there are often localized wind-shift boundaries whose horizontal scales are too fine to be reproduced by reanalyses. Such a localized wind-shift boundary was evident in the 6 June 2002 case study.

A few speculative comments about the difference between Rocky Mountain lee troughs and ALTs with respect to the sharpness of the wind-shift boundary and the magnitude of the dewpoint gradient are presented below. These differences may be
related to the different geographic characteristics of the Rocky Mountains and the Appalachians. Two main geographical differences are evident: 1) the Rocky Mountains are higher and broader than the Appalachians, and 2) the orientation of the Rocky Mountains is roughly north-northwest to south-southeast from the U.S./Canada border to Colorado, and north to south from Colorado to the U.S./Mexico border, whereas the orientation of the Appalachians is roughly northeast to southwest. As a result, the Rocky Mountains may represent a quasi-infinite barrier, whereas the Appalachians represent a finite barrier around whose southern end air parcels in the lower troposphere may travel [as shown by Weisman (1988, 1990a)]. Because the Appalachians represent a finite barrier, it seems possible for lower-tropospheric trajectories terminating west of the ALT (i.e., between the Appalachians and the ALT) to not have crossed the Appalachians. In this case, the ALT and its associated gradual wind shift may mark the difference between ending locations of southwesterly lower-tropospheric trajectories originating from the Gulf Coast and terminating in the ALT Zone west of the ALT, and southerly lower-tropospheric trajectories originating from the Atlantic Ocean or eastern Carolinas and terminating in the ALT Zone east of the ALT. Because both of these source regions tend to be relatively moist during the warm season, there is not a marked dewpoint drop to the west of the ALT. In contrast, because of the height and orientation of the Rocky Mountains, it seems unlikely to have trajectories terminating to the west of the Rocky Mountain lee trough that do not originate over the higher terrain and dry upon descending to the Great Plains. In addition, since the elevation change from the Rocky Mountains to the Great Plains is much greater than the elevation change from the Appalachians to the
Coastal Plain, the degree of downslope drying is likely greater over the Rocky Mountains/Great Plains region than over the ALT Zone.

Considerable variation was evident among the members of both the severe and nonsevere composites with respect to the locations and amplitudes of upper-tropospheric troughs and ridges, as well as the locations and magnitudes of QG forcing and the nearest surface low pressure center. However, manual inspection of the members of each composite yields certain features that were present in the majority of the members. Some of these features can be used to discriminate between severe and nonsevere days for each category. The features that are evident in the majority of the severe composite members, as well as those that can be used to discriminate between severe and nonsevere days, are shown in conceptual models for ALT categories 2–4 (Figs. 4.1a–c, respectively). For categories 2 and 3, a plume of CAPE that extends poleward throughout the ALT Zone (i.e., approximately 900 km) was the best discriminator between the severe and nonsevere composites. In both 2S and 3S, 82% of the members showed the CAPE plume, whereas in 2N and 3N, only 36% and 12% of the members, respectively, showed the CAPE plume (Figs. 4.1a,b). Other features that appeared in the majority of the members of 2S and 3S were TeD > 20 K over at least half of the ALT Zone (shown in 100% and 88% of the members, respectively) and 10-m meteorological wind directions of 180°–240° over at least half of the portion of the ALT Zone between the Appalachians and the ALT (77% and 71%). The CAPE plume and high TeD values demonstrate some of the convective instability parameters that are necessary for the development of severe convective storms on 2S and 3S days. The 10-m wind direction parameter is linked to
the convective instability parameters for reasons discussed in the previous paragraph (i.e., drier air is likely prevented from penetrating into the ALT Zone).

In 4S, the juxtaposition of a poleward-extended CAPE plume with $> 30$ kt of VWS occurred in 65% of the composite members, whereas this juxtaposition occurred in just 18% of the 4N members (Fig. 4.1c). Greater VWS in category 4 compared with category 2 and 3 is a signal that convective storms occurring during category 4 days are more likely to be spatially organized and longer-lived than convective storms occurring during category 2 or 3 days. Furthermore, $\text{TeD} > 20$ K over at least half of the ALT Zone occurred in 82% of the members of 4S, whereas it occurred in just 47% of the members of 4N. In addition to demonstrating the convective instability present in 4S, this result shows that the thermodynamic environment may still be conducive for wet microbursts within areas of the ALT Zone that do not have high VWS on 4S days. Recognition of the aforementioned features that often occur during severe convective storm days, as well as those that discriminate between severe and nonsevere days, can allow for increased situational awareness in an operational forecasting setting.

4.1.6 Case Study

The 6 June 2002 severe convective storm event was chosen for analysis in order to identify mechanisms responsible for severe convective storms and to determine what role the ALT plays in initiating and sustaining these storms. Several aspects of this event were similar to the 4S composite (Figs. 3.20a–e). For instance, an upper-tropospheric trough was present upstream of the ALT Zone at 1800 UTC when the storms were initiating within the ALT Zone (Fig. 3.22a), although the trough axis and associated QG
forcing for ascent were farther west than in the composite and did not affect the ALT Zone until 2100 UTC (not shown). In addition, the event exhibited a plume of high CAPE that extended poleward throughout the ALT Zone and was juxtaposed with > 30 kt VWS (Fig. 3.25a), and the event exhibited > 20 K TeD over the majority of the ALT Zone (Fig. 3.25b), both of which are features that are evident in the 4S composite (Figs. 3.20d,e).

Prefrontal storms initiated along a wind-shift boundary in the immediate lee of the Appalachians (Fig. 3.24a). This wind-shift boundary occurred where westerly winds over the Appalachians converged with southwesterly winds between the Appalachians and the ALT. Downslope drying did not occur since backward trajectories originating in the lee of the Appalachians did not pass over the Appalachians (Fig. 3.26). As a result of the lack of downslope drying, dewpoints reached 20–24°C and \( \theta_e \) values reached 350–365 K in the vicinity of the ALT (Figs. 3.24b,c). As the prefrontal storms moved eastward, they moved into an increasingly convectively unstable air mass in the vicinity of the ALT. A conceptual model is presented for this event in Fig. 4.2. The main features of the conceptual model are: 1) storms initiating along a wind-shift boundary in the immediate lee of the Appalachians, 2) an ALT collocated with a surface \( \theta_e \) maximum located east of the storm initiation region, and 3) storms intensifying as they approach the ALT and its associated surface \( \theta_e \) maximum. Since ALTs are associated with lower-tropospheric thermal anomalies, given an appropriate amount of VWS, storms moving toward the ALT can be expected to intensify upon encountering an air mass with more convective instability in the vicinity of the ALT compared with the convective initiation region. The fact that the severe composites of ALT categories 2–4 show MUCAPE
increasing from west to east over the ALT Zone adds confidence to this conceptual model.

4.1.7 Applications of Research to Operational Forecasting

This research has attempted to classify ALTs and investigate their association with severe convective storms in order to provide forecasters with enhanced situational awareness when severe convective storms are possible in the presence of an ALT. Knowledge of the expected ALT category on a particular day (i.e., the location of the nearest upstream cold front and its expected progression) can provide forecasters with the expected spatial organization of severe convective storms, and it can point forecasters toward other parameters to examine in order to assess the likelihood of severe convective storms. For instance, on days featuring ALT categories 2 or 3, convection is less likely to be spatially organized relative to category 4, owing to the lack of VWS and QG/frontal forcing for ascent in categories 2 and 3 (with the exception of the area along the portion of the front that passes through the ALT Zone in category 3). Despite the expected lack of spatially organized convection in categories 2 and 3, certain steps can be taken to assess the likelihood and favored locations of severe convective storms. For instance, observed morning soundings and forecast afternoon soundings can be examined to assess the potential for wet microbursts. TeD values > 20 K along with midtropospheric dry air above a column of moist air both support the possibility of wet microbursts in a high-CAPE, low-VWS environment. In addition, deep convection in general will be more likely to occur if there is no capping inversion present. Climatologically favored areas of convective initiation should be considered [i.e., along and in the immediate lee of the Appalachians and along the sea-breeze boundary (when present)], along with favored
areas for storm reports for categories 2 and 3 (Figs. 3.9b,c). Throughout the afternoon as convective storm occurrence becomes more favorable because of increasing convective instability, detection of mesoscale surface boundaries and convergence zones is important, especially if the environment is uncapped and convectively unstable. In that case, all that is needed to initiate deep convection is a triggering mechanism.

If a category 4 ALT day is anticipated, the convection can be expected to be more spatially organized due to greater VWS and QG/frontal forcing for ascent than in categories 2 or 3. However, severe convection can still develop ahead of the strongest QG/frontal forcing as seen in the case study (i.e., because of convergence along a mesoscale surface boundary in an uncapped, convectively unstable environment). As a result, assessment of wet microburst potential and detection of mesoscale surface boundaries in the prefrontal air mass (i.e., in the vicinity of the ALT) remains important in order to identify areas located ahead of the strongest QG/frontal forcing that are still favorable for severe convective storms. In addition, identifying the collocation of the poleward-extended CAPE plume with > 30 kt VWS is important in order to pinpoint a possible threat region for the strongest, longest-lived convective storms (this collocation occasionally appears on category 3 days, as well). As shown in the case study conceptual model (Fig. 4.2), knowledge of the location of the convective initiation region with respect to the location of the ALT and any associated warm/moist axes can be used to predict whether convective storms will intensify given enough VWS.

In addition to assessing the potential for severe convection based upon ALT category, consideration of MUCAPE and VWS calculated at the first daily storm report can be useful in placing a certain day into climatological context. For instance, forecast
values of MUCAPE and VWS valid during the afternoon hours can be compared with box-and-whisker plots of MUCAPE and VWS calculated at the location of the first daily storm report for each ALT Zone sector and month (Figs. 3.12a–f). Specifically, since convective instability is usually present during the mid-Atlantic warm season, VWS values greater than the 75th percentile for each sector during JJA should be taken as a signal that a severe convective storm event is likely, given the presence of a triggering mechanism to initiate deep convection. PA07 support this assertion by finding that “even though instability is usually present in the southeastern United States during the summer, organized convective episodes only occur on days with relatively higher shear” (p. 3722; see also their Fig. 16). In fact, PA07 find that even during the month with the lowest shear (August), the mean VWS value in the mid-Atlantic during convectively active days is still greater than the median VWS value for the entire year. Despite the importance of VWS, however, severe convective storms are still possible in environments of low VWS (especially in environments conducive for wet microbursts), though the convection is less likely to be spatially organized or long-lived. During MS, when VWS is likely to be higher relative to JJA, the expected MUCAPE value may play a more important role relative to its role during JJA. For each sector during MS, MUCAPE values above the 75th percentile (recall that the MUCAPE values presented in Figs. 3.12a–f are calculated from the NARR and are thus likely underestimates) coupled with VWS values above the median should serve as a signal that a severe convective storm event is likely, given the presence of a triggering mechanism to initiate convection.

Additionally, comparison of forecast afternoon MUCAPE and VWS values with scatterplots of MUCAPE and VWS calculated at the location of the first daily storm
report for each sector (Figs. 3.11a–c) can yield insight into the possibility of severe convective storm occurrence. Especially on days in which the environment for severe convective storms seems marginal, comparison of forecast MUCAPE and VWS values with the line on the scatterplot below which severe convective storms are rare can increase confidence as to whether or not severe convective storms are possible.

4.2 Conclusions

The main goals of this thesis were to: 1) establish a climatology of warm-season ALTs in the mid-Atlantic; 2) document the spatial and temporal distributions of severe convective storms in the mid-Atlantic with particular emphasis on the role of the ALT in modulating those distributions; 3) investigate the convective environments characteristic of the ALT Zone; 4) construct composites of ALT events associated with severe convective storms; and 5) study a specific ALT event associated with severe convective storms. To accomplish these objectives, 13 cases of ALTs associated with severe convective storms were analyzed using CFSR data, and three prominent features representative of the 13 cases were identified. The three features were: 1) lower-tropospheric wind components over the Appalachians orthogonal to and downslope of the Appalachians, 2) the ALT itself in the lee of the Appalachians, and 3) a lower-tropospheric thermal ridge collocated with the ALT. An ALT identification algorithm that was based on these three features was performed using CFSR data during 10 warm seasons. ALTs were found to form most often during times of daily and seasonal peak heating. The ALTs identified by the algorithm were categorized based on their relationship to synoptic-scale cold fronts. The frequency of occurrence of each ALT category was dependent on the month. ALTs associated with a total frontal passage from
the Appalachians to the Atlantic Coast (i.e., category 4) were more likely to occur in May and September, whereas ALTs not associated with a total frontal passage (i.e., categories 2 and 3) were more likely to occur during June, July, and August.

A climatological context for severe convective storms was established by determining the spatial and temporal distributions of storm reports in the mid-Atlantic. The spatial distributions of storm reports varied by ALT category. The spatial distributions of storm reports for categories 2 and 3 were qualitatively similar within the ALT Zone, with local maxima in the immediate lee of the Appalachians. Storm reports associated with category 4 ALTs were found to favor the Washington, DC, to Philadelphia corridor. This corridor was collocated with the juxtaposition of a CAPE plume that extended poleward throughout the ALT Zone and > 30 kt VWS. In addition, a larger proportion of storm reports extended to the Atlantic Coast in category 4 than in categories 2 and 3, a result that is likely tied to frontal forcing spreading across the entire ALT Zone in category 4. A marked midafternoon and early evening maximum in storm reports was found in the ALT Zone, as well as a minimum in the overnight and early morning hours. The time of the first daily storm report varied by latitude, with the peak time of the first daily storm report occurring 2 h earlier in the southern portion of the ALT Zone compared with the northern portion.

The convective environments characteristic of the ALT Zone were examined by calculating MUCAPE and VWS at the location of the first daily storm report. Severe convective storms developed in an environment characterized by higher MUCAPE in the southern portion of the ALT Zone compared with the northern portion, whereas severe convective storms developed in an environment characterized by higher VWS in the
northern portion of the ALT Zone compared with the southern portion. In addition, a comparison of observed MUCAPE values revealed that times when an ALT was present were associated with higher MUCAPE values than times when an ALT was absent. This result was linked to the result that nearly twice as many storm reports occurred during the 6 h following the presence of an ALT compared with the 6 h following the absence of an ALT.

Severe and nonsevere composites were constructed for ALT categories 2–4 in order to pinpoint dynamical processes and thermodynamic environments conducive to severe convective storms, and to find features that discriminate between situations that favor and do not favor severe convective storms for each category. A plume of CAPE that extended poleward throughout the ALT Zone was evident in the severe composites of categories 2–4, and in category 4, that plume was juxtaposed with > 30 kt VWS. This juxtaposition was collocated with the maximum in storm reports for category 4 (i.e., the Washington, DC, to Philadelphia corridor). All three severe composites were also characterized by TeD values > 20 K, a threshold value found by Atkins and Wakimoto (1991) to be favorable for the occurrence of wet microbursts. A majority of members of the severe composites for categories 2 and 3 were characterized by 10-m meteorological wind directions of 180°–240° between the Appalachians and the ALT. It is hypothesized that the lack of westerly (downslope) winds in the immediate lee of the Appalachians means that drier air descending from the Appalachians does not progress far into the ALT Zone, so convection is not suppressed in the immediate lee of the Appalachians.

A severe convective storm event associated with an ALT was studied in order to clearly elucidate the dynamical processes and thermodynamic environments associated
with severe convective storms in the presence of an ALT. The event showed features similar to the category 4 severe composite, such as a CAPE plume that extended poleward throughout the ALT Zone juxtaposed with > 30 kt VWS. An extremely convectively unstable air mass was in place over the ALT Zone, characterized by TeD values of 20–35 K and \( \theta_e \) values of 345–365 K. Severe prefrontal storms initiated along a wind-shift boundary in the immediate lee of the Appalachians and strengthened upon approaching the ALT and an axis of high \( \theta_e \) that was collocated with the ALT. Observed 1200 UTC soundings and model-derived 2100 UTC soundings ahead of the prefrontal storms showed a layer of midtropospheric dry air in an otherwise moist environment, a setup that is comparable to conceptual models for wet microburst-producing storms. A second round of severe alongfront storms developed west of the prefrontal storms and strengthened within the ALT Zone in an environment that had not convectively stabilized in the wake of the prefrontal storms.

4.3 Suggestions for Future Work

An unanswered problem that arose from this research is quantifying the contribution of orographic effects relative to diabatic effects in ALT formation. On one hand, the presence of lower-tropospheric downslope wind components was a criterion used to define ALTs in the present study. On the other hand, ALTs are more likely to form during times of peak heating, suggesting a link to the diurnal and seasonal heating cycles. Additionally, the downslope flow over the Appalachians was often fairly weak. In cases of weak lower-tropospheric cross-mountain flow, the ALT is less likely to be a result of orographic forcing. A WRF modeling study where terrain and heating could be turned on and off [similar to Benjamin (1986), but specifically with terrain and boundary
conditions characteristic of the mid-Atlantic] could be a step toward answering this problem. It would also be useful to composite ALTs based upon wind direction between the Appalachians and the ALT. It seems likely that the dewpoint gradient in the vicinity of the ALT would be larger with northwest flow (i.e., downslope flow) to the west of the ALT than with southwest flow (i.e., flow not passing over the Appalachians), but this hypothesis should be tested. In such a test, the distribution of severe convective storms could be compared between the two composite categories.

Opportunities for future research also exist in examining PFTs in the northeastern U.S. (i.e., expanding the domain north of the ALT Zone). Anecdotal evidence and conventional wisdom from National Weather Service forecasters in the northeastern U.S. supports the hypothesis that severe convective storms in the northeastern U.S. often occur along or within a PFT. A climatology of PFTs in the northeastern U.S. would yield the frequency of PFT occurrence, and PFTs identified in the climatology could be compared to radar data and/or storm reports to determine the association of PFTs with severe convective storms. A categorization scheme similar to the one developed in the present study could be used to categorize PFTs in the northeastern U.S.

Additional analyses that quantify wet microburst potential could be performed in the ALT Zone. Prior work (e.g., Atkins and Wakimoto 1991; Wheeler and Roeder 1996) has found that the potential for wet microbursts becomes more likely as certain threshold TeD values are exceeded. Since these threshold values appear to be location-specific, a study of TeD values observed before the occurrence of wet microbursts within the three ALT Zone sectors could be performed in order to determine if there is a certain TeD threshold value above which wet microbursts become increasingly likely.
Fig. 4.1. Conceptual model of (a)–(c) key features of the severe composites of ALT categories 2–4, respectively. Blue shading indicates areas of MUCAPE > 1000 J kg\(^{-1}\), yellow shading indicates areas of VWS > 30 kt, and red shading indicates areas of \(Q\)-vector divergence < \(-3 \times 10^{-15}\) K m\(^{-2}\) s\(^{-1}\). Red (blue) numbers indicate the percentage of severe (nonsevere) composite members that exhibit the indicated feature. Blue numbers are only included for features that discriminate between the severe and nonsevere composites.
Fig. 4.2. Conceptual model based upon the 6 June 2002 severe convective storm event showing MSLP (black contours, hPa), VWS (dark blue contours, ≥ 25 kt), and 10-m winds (barbs, kt). Prefrontal storms (dots colored according to the key in the bottom left of the image) initiate along a wind-shift boundary in the immediate lee of the Appalachians west of the ALT (dashed black line) at $t = t_0$. The ALT marks the location of an axis of high $\theta_e$ (green shading). The storms intensify at $t = t_0 + \Delta t$ upon approaching the ALT and collocated high $\theta_e$ axis.
REFERENCES


