Climatology of the Elevated Mixed Layer Over the Contiguous United States and Northern Mexico: 1979–2021

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Elevated mixed layers (EMLs) are an important influence on the severe convective storm climatology in the contiguous United States (CONUS), playing a role in storm generation, sustenance, and suppression. A function of the topography in the western CONUS and northern Mexico, EMLs are elevated layers of nearly dry adiabatic lapse rates and high potential temperature, typically with a capping inversion at their base. Although it is well-established that EMLs are primarily a warm-season phenomenon most frequent in the Great Plains, no research to date has examined their variability in-depth, or whether they have changed through time. This study creates an updated, high-resolution climatology of the EML to analyze EML variability and changes in EML occurrence and characteristics over the last four decades. An objective algorithm is applied to ECMWF Reanalysis Version 5 (ERA5) to detect EMLs, defined in part as layers of steep lapse rates \((\geq 8.0 \, ^\circ C \cdot km^{-1})\) at least 200 mb thick, in the CONUS and northern Mexico from 1979 to 2021. The interannual, intra-annual, and seasonal variability of EMLs are investigated, as are the typical values and ranges of EML attributes including lapse rates, potential temperature, and convective inhibition (CIN). Long-term trends in these attributes and EML days are calculated and assessed for significance. Additionally, practically perfect
hindcasts of hail and tornadoes are used to assess whether severe thunderstorms favor certain regions relative to the EML center.

Results reveal that EMLs are most frequent over the Great Plains in spring and summer, with a standard deviation of 4–10 EML days per year highlighting sizable interannual variability. Mean CIN associated with the EML’s capping inversion suggests many EMLs prohibit convection, although, like nearly all EML characteristics, there is considerable spread and notable seasonal variability. In the western Great Plains, statistically significant increases in EML days (four to five more days per decade) coincide with warmer EML bases and steeper EML lapse rates, driven by warming and drying in the low levels of the western CONUS during the study period. Additionally, increases in EML base temperatures result in significantly more EML-related CIN over the Great Plains, which may continue to have implications for storm frequency, intensity, severe perils, and precipitation if this trend persists into the future.
CLIMATOLOGY OF THE ELEVATED MIXED LAYER OVER
THE CONTIGUOUS UNITED STATES AND
NORTHERN MEXICO: 1979–2021

BY

MARGO SICILIANO ANDREWS
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Thesis Director:
Vittorio A. Gensini
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CHAPTER 1

INTRODUCTION

The important role the elevated mixed layer (EML) plays in the generation, sustenance, and, in some cases, suppression of severe convective storms (SCS) in the contiguous United States (CONUS) was identified over seven decades ago by Fawbush and Miller (1954). Since that early work, conceptual models have illustrated the EML’s features, formation and migration, and influence on the SCS climatology in the CONUS. The EML is an elevated layer of steep lapse rates, and high, nearly constant, potential temperature atop a temperature inversion at its base referred to as the lid (Carlson and Ludlam 1968; Carlson et al. 1983; Lanicci and Warner 1991a). A function of the unique combination of topography and source regions present in the CONUS, the EML typically originates as planetary boundary layer (PBL) air in northern Mexico, the southwestern CONUS, the High Plains, and/or the high terrain of the Rocky Mountains (Carlson and Ludlam 1968; Carlson et al. 1983). Under certain synoptic patterns, this air mass is advected eastward and loses connection with the ground, flowing over and “capping” warm, moist lower-level southerly flow that typically originates in the Gulf of Mexico and the western Caribbean (Carlson and Ludlam 1968; Carlson et al. 1983; Lanicci and Warner 1991a). The steep mid-level lapse rates of the EML often enhance convective available potential energy (CAPE), a necessary ingredient for convective storms, while the lid at the base of the EML can prevent convection over large areas, permitting CAPE to accumulate with time (Carlson and Ludlam 1968; Carlson et al. 1983). As a result, any storms that do form have the potential to be
stronger than if no EML or capping inversion had been present (Carlson and Ludlam 1968; Carlson et al. 1983; Graziano and Carlson 1987). Thus, understanding the geographic distribution and frequency of this feature has long been of interest to operational forecasters and climatologists.

Early attempts at EML climatologies were limited in their spatiotemporal extent due to computing resource constraints and were subject to the relatively coarse spatial and temporal resolution of upper-air observations. Though limited to four years, Lanicci and Warner (1991a,b,c) generated the first comprehensive climatology of the EML over the Great Plains, exploring its frequency distribution, associated synoptic patterns, and relationship to severe storms. A recent climatology by Ribeiro and Bosart (2018) has updated and expanded the work of Lanicci and Warner while highlighting the benefits of using high-resolution reanalysis data. The development of a new reanalysis dataset presents the opportunity to examine the EML with greater vertical, spatial, and temporal resolution. While previous research has established that EMLs are primarily a warm-season phenomenon in the Great Plains (Carlson et al. 1983; Lanicci and Warner 1991a; Ribeiro and Bosart 2018; Li et al. 2020), the creation of a new climatology also permits exploring how much the EML deviates from its mean state on an annual and seasonal basis, and whether EMLs have exhibited notable changes over the past four decades. To date, no research has examined the interannual or intra-annual variability of the EML, nor how its frequency or characteristics have changed in the long-term, both of which could have major implications for SCS frequency, intensity, and location in the CONUS.

This project used the European Center for Medium-Range Weather Forecasts (ECMWF) Reanalysis Version 5 (ERA5; Hersbach et al. 2020), the most recent global atmospheric
reanalysis, to create an updated climatology of EMLs from 1979 to 2021 in the CONUS and northern Mexico. The following questions will be answered:

1) What is the frequency, spatial distribution, and spatial extent of the EML? How do these characteristics change throughout the year and the study period? How do these results compare to previous climatologies?

2) How do EML attributes such as lapse rates, potential temperature, base height, and thickness change with space and time? How do these results compare to previous climatologies?

3) How does the strength of the capping inversion associated with the EML change with space and time?

To address these questions, an objective algorithm was created to automatically identify the EML. The results of this algorithm were compared to previous climatologies, used to better understand the magnitude and distribution of the capping inversion at the base of the EML, and used to determine whether the EML or its capping inversion exhibited substantial changes within the study period. Additionally, because the EML’s presence was linked to a greater number of severe storms and other high-impact weather events (Cordeira et al. 2017), in addition to a disproportionate number of injuries and deaths from severe storms in parts of the CONUS (Banacos and Ekster 2010), the hope is that the knowledge derived from this new high-resolution climatology will provide a useful reference for forecasters. Furthermore, these results should be particularly relevant to climatologists seeking to examine the variability and long-term trends in the EML and SCSs.
CHAPTER 2

BACKGROUND

Conceptual Models of the EML

Conceptual models describing the formation of SCS environments in the Great Plains have long highlighted the importance of the Gulf of Mexico and the high terrain of the western CONUS. These early models reveal how the formation and structure of the EML are closely tied to these features, as well as how the EML influences the intensity and location of storms. The EML originates as a well-mixed, deep PBL with high, nearly constant, potential temperature due to intense surface heating over the elevated terrain of northern Mexico, the southwestern CONUS, the High Plains, and/or the Rocky Mountains (Carlson and Ludlam 1968; Carlson et al. 1983; Lanicci and Warner 1991a). In the conceptual model first introduced by Carlson and Ludlam (1968), a trough approaches the Great Plains from the west. Associated mid- and upper-level southwesterly flow advects the boundary layer air northeastward off of the higher terrain, causing it to become elevated. This air flows over the moist southerly low-level flow from the Gulf of Mexico, effectively capping the environment below. The EML is, therefore, an elevated air mass featuring steep lapse rates and high potential temperatures with a capping inversion at its base, often referred to as the lid (Carlson and Ludlam 1968; Carlson et al. 1983; Lanicci and Warner 1991a).

The lid generally prevents, or delays, deep convection over a region, allowing CAPE to
increase more than if the layer were allowed to convect or expand (Carlson and Ludlam 1968; Carlson et al. 1983). Therefore, any convection that does overcome the cap has a higher likelihood of being severe (Carlson and Ludlam 1968; Carlson et al. 1983; Graziano and Carlson 1987). Storms typically form in regions where the convective inhibition (CIN) associated with the lid is minimal or nonexistent, which tends to be along the lid edge (Carlson and Ludlam 1968; Carlson et al. 1983; Keyser and Carlson 1984; Lanicci and Warner 1991b; Cordeira et al. 2017). This is often the western or northern edges of the capping inversion, where warm moist air flows out from under the lid in a process called underrunning (Carlson et al. 1983; Benjamin and Carlson 1986; Lanicci and Warner 1991b; Farrell and Carlson 1989; Banacos and Ekster 2010). The lid can be weakened or breached via a number of other processes, including surface heating, orography, differential advection, and mesoscale or synoptic-scale lifting (Carlson et al. 1983; Farrell and Carlson 1989; Lanicci and Warner 1991b,c; Weckwerth and Parsons 2006; Cordeira et al. 2017).

Aspects of this model, including the important roles of the Gulf of Mexico and elevated topography in supporting SCS environments and EML formation have been tested and verified by subsequent studies (Benjamin and Carlson 1986; Li et al. 2021). In the absence of either feature, EML frequency is altered, as are the distributions and magnitudes of instability and CIN over the eastern two-thirds of the CONUS (Li et al. 2021). The synoptic patterns associated with EML formation also appear to be largely consistent with the early conceptual models, particularly in the spring (Lanicci and Warner 1991a; Ribeiro and Bosart 2018). However, EMLs can form under various synoptic-scale patterns, as noted by Lanicci and Warner (1991a, 1997) in their case study and EML climatology.
Existing Synoptic-Scale Climatologies of the EML

Identifying springtime EMLs in four years (1983–1986) of observed soundings from the southern and central Great Plains, Lanicci and Warner (1991a) found several synoptic setups favorable for its formation. Although the pattern described in the conceptual model was the most effective at producing EMLs throughout the spring, additional synoptic-scale patterns—such as anticyclonic and northwesterly mid-level flow—become favorable for EML formation later in the season. An EML was present over about 20–25% of the southern and central Great Plains at any given time. The highest geographic frequency shifted from south Texas in April and May to the north and west through June, with the northward shift a result of the low-level moisture and EML source region expanding poleward with time (Lanicci and Warner 1991a). The geographic location of this frequency maximum, and its northward progression, is consistent with a climatology by Farrell and Carlson (1989) in the same region. Individual EMLs were determined to have an average length of just under a week, with a pattern of recurrence seemingly dependent on EML size (Lanicci and Warner 1991b). The size of a severe weather event and the size of an EML prior to that convection were most highly correlated in the early spring, but, throughout the season, there were numerous instances of large EMLs producing only small severe weather events (Lanicci and Warner 1991c), likely leaving a substantial portion of the EML intact, allowing it to influence future environments, potentially downstream.

In the most complete climatology since Lanicci and Warner's efforts, Ribeiro and Bosart (2018) examined EMLs in South America and compared them to their North American counterparts using the NCEP Climate Forecast System Reanalysis (CFSR; Saha et al. 2010). Their study benefited from the higher spatial and temporal resolution of the CFSR compared to
that of available observed soundings used in earlier climatologies. As expected, they found the EML was most common in the spring, followed by the summer, and least common in the winter months. Consistent with Lanicci and Warner (1991a), the dominant EML-producing synoptic pattern in the springtime featured southwesterly flow aloft, the result of longwave troughing in the west and ridging downstream. Maximum EML frequency was found over southern Texas and northeastern Mexico in the spring, with a reduced frequency maximum centered over the central Great Plains in the summer. Average EML thickness and lapse rates also exhibited a northward shift in the summer from their springtime maximum just downstream of the Rocky Mountains in the central CONUS. In the warm seasons, EML base heights and EML potential temperatures averaged 700–750 mb and 316–320 K, respectively. Compared to South America, North American EML environments had higher CAPE, lower lifting condensation levels (LCLs), and greater 0–1 km storm-relative helicity (SRH), implying more favorable conditions for supercells and tornadoes if the greater CIN in North America can be overcome (Ribeiro and Bosart 2018).

Forecasting Convection with an EML and its Capping Inversion

Identifying environments favorable for severe storms is typically done using an ingredients-based approach (Johns and Doswell 1992; Doswell et al. 1996). The well-established ingredients required for severe storms, including supercells, are instability, moisture, vertical wind shear, and a lifting mechanism (e.g., McNulty 1985; Johns and Doswell 1992; Rasmussen and Blanchard 1998). The existence of the warm, moist boundary layer air from the Gulf of Mexico below the steep lapse rates of the EML often ensures the presence of sufficient low-level moisture and can contribute to significant instability. Combined with the vertical wind shear
often present with the synoptic patterns favoring EML formation, it follows that EMLs are often supportive of severe storms (Carlson and Ludlam 1968; Carlson et al. 1983; Farrell and Carlson 1989; Lanicci and Warner 1991a; Banacos and Ekster 2010; Cordeira et al. 2017; Ribeiro and Bosart 2018), assuming there is adequate forcing in the form of surface heating, differential advection, orography, or boundaries to overcome the cap (e.g., Carlson et al. 1983; McNulty 1995; Weckwerth and Parsons 2006). However, in some cases, the capping inversion at the base of the EML may be too strong to overcome, inhibiting convection (Carlson and Ludlam 1968; Carlson et al. 1983; Graziano and Carlson 1987; Cordeira et al. 2017). Therefore, the presence of an EML alone is not sufficient to produce SCSs (Cordeira et al. 2017).

Only approximately 6–8% of environments in the eastern half of the CONUS that are favorable for severe thunderstorms produce deep convection (Taszarek et al. 2020). To anticipate convective initiation (CI) in the presence of an EML and its capping inversion, various metrics have been suggested. Examining the 700–500 millibar (mb) lapse rates provides some indication of the location of instability, but lapse rates alone do not provide a sense of the strength of the lid (Doswell 1985). Temperatures at 700 mb have long been used as a proxy for whether the cap will break, with values greater than 10–12 °C historically thought to suppress convection (Means 1952; Miller 1967; Johns and Doswell 1992; Junker et al. 1999; Bunkers et al. 2010). However, 700 mb temperatures are related to elevation and vary seasonally, with over 25% of severe storms in the High Plains forming with temperatures above that threshold (Bunkers et al. 2010). Therefore, Bunkers et al. (2010) caution against relying on 700 mb temperatures alone to forecast CI.

The lid strength index (LSI) was developed by Carlson et al. (1980) to indicate the
potential for storms when a capping inversion is present. Together, its two components—buoyancy and lid strength—seek to narrow down the areas with thunderstorm potential better than instability alone (Carlson et al. 1980, 1983; Graziano and Carlson 1987). For a given value of buoyancy, a statistical analysis of the lid strength (LS) term found that, while storms are less likely to occur under stronger lids, particularly when LS values > 2, those that do manage to form are more likely to be severe (Graziano and Carlson 1987). The LSI, which considers only a limited number of vertical levels, is not widely used today. Instead, instability and lid strength are examined using the integrated quantities CAPE and CIN, respectively.

CAPE measures the amount of positive buoyant energy available to a rising air parcel once it reaches its level of free convection (LFC), while CIN provides an estimate of the strength of the capping inversion, measuring the negative buoyancy a parcel must overcome to reach the LFC and produce deep convection. Since most storms form in unstable environments with less than 75–100 \( \text{J} \cdot \text{kg}^{-1} \) of absolute CIN, CIN also provides some indicator of the likelihood of CI (Bunkers et al. 2010; Gensini and Ashley 2011; Hoogewind et al. 2017; Taszarek et al. 2021a). Too much CIN can prevent convection, but some CIN is generally favorable for SCSs as it can delay the time of initiation until instability has increased and limit the number of storms that form (Bunkers et al. 2010). However, factors such as soil moisture and differential heating mean that the EML and the CIN associated with the lid are not homogeneous (Benjamin and Carlson 1986; Lanicci et al. 1987; Lanicci and Warner 1997), often making it difficult to anticipate the location and timing of storms.

Climatologically, CIN is maximized over the central CONUS (Bunkers et al. 2010; Taszarek et al. 2020). This can likely be explained by the frequent advection of the EML over this region (Bunkers et al. 2010), particularly in the spring and summer months when CIN is
maximized annually (Taszarek et al. 2020). Ribeiro and Bosart (2018) looked specifically at
EML-related CIN in the CONUS, creating springtime composites in three fixed regions. The
median CIN associated with EMLs in all regions had a magnitude of approximately 200 J · kg⁻¹
(Ribeiro and Bosart 2018).

Examining the distribution of CIN associated with EMLs in all seasons reveals the
typical magnitude and seasonality of the lid and gives insight into how often EML environments
are favorable for deep convection. Furthermore, the length of our study permits exploring
whether the CIN associated with the EML changes long-term, and whether that is consistent with
the notable increase in CIN in ERA5 and observed soundings over most of the CONUS in the
last four decades (Taszarek et al. 2021a; Pilguj et al. 2022). The decadal increases in CIN, which
are most significant in the Great Plains in the spring and summer months (Taszarek et al. 2021a;
Pilguj et al. 2022), are likely related to the statistically significant increase in the magnitude of
the mid-level lapse rates in portions of the Great Plains over the same period (Taszarek et al.
2021a). This increase in mid-level lapse rates is likely driven by a significant trend towards a
hotter and drier boundary layer over the western CONUS, which is increasingly favorable for the
development of an EML (Taszarek et al. 2021a). Therefore, it is also of interest whether the
magnitude of the EML lapse rates changes within the study period, as stronger EMLs and
resulting larger capping inversions could have implications for thunderstorm frequency and
intensity in the future (Rasmussen et al. 2020; Taszarek et al. 2021a; Haberlie et al. 2022; Ashley
et al. 2023).
EMLs Outside the Great Plains of the CONUS

When instability is not released close to the initial location of the EML (such as the Great Plains), the EML can move to other regions of the CONUS, likely primarily via horizontal lapse rate advection (Banacos and Ekster 2010). Severe weather associated with the downstream advection of the EML has been noted in the upper Ohio Valley (Farrell and Carlson 1989), the Northeast (Johns and Dorr 1996; Banacos and Ekster 2010), and the Midwest (Cordeira et al. 2017). Farrell and Carlson's (1989) climatology found that EML events in the northeastern portion of the country are very rare, on the order of one day per month in the warm season. Banacos and Ekster (2010) focused solely on significant severe weather reports in the Northeast related to the EML, as they were responsible for a notable number of storm-related deaths and injuries, despite the limited number of occurrences. Troughing over the West Coast with ridging over the central CONUS appeared favorable for the advection of the EML that far north and east (Banacos and Ekster 2010; Cordeira et al. 2017). Cordeira et al. (2017) looked specifically at Minneapolis after a series of severe storms and high-temperature records were linked to an EML. Their brief climatology revealed that days with an EML had substantially more severe weather reports than the calculated climatology. Whether this is true of other regions in the CONUS is unknown. The number of EML days in regions such as the Midwest, the Northern Plains, and the Southeast, and how that varies year to year are also unknown, thus motivating this thesis.

EMLs also exist outside of the CONUS, downstream of other substantial mountain ranges, including in Southeast Asia (Das et al. 2014), South America (Rasmussen and Houze 2016; Ribeiro and Bosart 2018), Africa (Parker et al. 2005), and western Europe (Carlson and Ludlam 1968). Worldwide, the highest frequency of environments favorable for significant
severe storms are downstream of some of these mountain ranges (Brooks et al. 2003).

Differences compared to CONUS EMLs result from variations in the character and arrangement of the high terrain and moisture sources across continents. For example, EMLs in South America have much lower seasonal variation, a lower maximum frequency of occurrence, and form via different processes (Ribeiro and Bosart 2018).
CHAPTER 3

DATA AND METHODS

Data

A climatology of EMLs east of the continental divide in the CONUS and northern Mexico is created using ERA5 (Hersbach et al. 2020), the latest high-resolution global reanalysis data set available to researchers. Reanalysis combines numerous sources of meteorological data including surface-based observations, rawinsonde measurements, and information from satellites and aircraft into a numerical weather prediction model to provide the best possible approximation of past atmospheric conditions at a much higher spatiotemporal resolution than observations alone (e.g., Gensini et al. 2014; Taszarek et al. 2021b). ERA5 has hourly output with a horizontal grid spacing of 31 kilometers and 137 hybrid-sigma levels in the vertical, 28 of which are in the lowest two kilometers (Hersbach et al. 2020; Copernicus Climate Change Service 2017). The increased vertical resolution compared to other reanalyses results in an improved depiction of the vertical temperature profile of the EML, including the capping inversion, which earlier reanalyses struggled to resolve (Brooks et al. 2003; Gensini et al. 2014; Ribeiro and Bosart 2018; Taszarek et al. 2018). Reanalysis, however, is not a perfect representation of reality. For example, ERA5 still underestimates CAPE, and, to a lesser degree, CIN, and struggles in areas with topographic changes (Taszarek et al. 2021b). However, reanalyses, including ERA5, have proven to be an effective means to understand the climatology
and long-term trends of severe convective storms and their environments (Brooks et al. 2003; Riemann-Campe et al. 2009; Gensini and Ashley 2011; Gensini and Brooks 2018; Tang et al. 2019; Taszarek et al. 2020; Li et al. 2020; Taszarek et al. 2021a; Pilguj et al. 2022). Therefore, ERA5 is well suited to the task of updating the EML climatology, particularly in the 1979 to 2021 study period, where it has demonstrated the ability to represent both weather systems and convective environments, including their synoptic features, reasonably well (Hersbach et al. 2020; Li et al. 2020; Taszarek et al. 2021b; Pilguj et al. 2022).

Methods

EML Identification Algorithm

Various definitions of the EML have been proposed for the purpose of identification (Table 1). In this study, an algorithm identified EML soundings using a modified version of the criteria presented by previous authors including Lanicci and Warner (1991a), Ribeiro and Bosart (2018), and Li et al. (2020). This algorithm was applied to vertical profiles of temperature, pressure, specific humidity, and geopotential from ERA5 to locate grid points within the study period and domain that met the following criteria (Fig. 1):

1) An environmental lapse rate of at least 8 °C · km⁻¹ over a depth of at least 200 mb,
2) Greater than 0 J · kg⁻¹ of MUCAPE,
3) An EML base, defined as the first level of the steep lapse rate layer, located above 1000 m AGL but below 500 mb,
### Table 1
Criteria Used by Previous Studies to Identify the EML

<table>
<thead>
<tr>
<th>Authors (Year)</th>
<th>Dataset Used</th>
<th>Lapse Rates</th>
<th>Instability Threshold</th>
<th>Inversion/ Lid Strength</th>
<th>Potential Temperature</th>
<th>Relative Humidity (RH)</th>
<th>EML Base</th>
<th>EML Depth</th>
</tr>
</thead>
<tbody>
<tr>
<td>Farrell and Carlson (1989)</td>
<td>Radiosondes</td>
<td>Static stability of &lt; 4.5°C/100 mb above the inversion</td>
<td>Buoyancy term: ((\theta_{sw5} - \theta_{wb}) &lt; 0.5)°C</td>
<td>LS ≥ 2°C and a layer with a saturation wet bulb potential temp. lapse rate of ≥ 0 within 100 mb of RH break</td>
<td>N/A</td>
<td>Sharp decrease with height (0.87% per mb) below 500 mb; RH must increase with height above the warmest part of the inversion</td>
<td>N/A</td>
<td>N/A</td>
</tr>
<tr>
<td>Lanicci and Warner (1991)</td>
<td>Radiosondes</td>
<td>Static stability of &lt; 4.5 °C/100 mb above the inversion</td>
<td>Buoyantly unstable at 500 mb: ((\theta_{sw5} - \theta_{wb}) &lt; 0.0)°C</td>
<td>LS ≥ 2°C and a layer with a saturation wet bulb potential temp. lapse rate of ≥ 0 within 100 mb of RH break</td>
<td>N/A</td>
<td>Sharp decrease with height (0.87% per mb) below 500 mb; RH must increase with height above the warmest part of the inversion</td>
<td>N/A</td>
<td>N/A</td>
</tr>
<tr>
<td>Banacos and Ekster (2010)</td>
<td>Radiosondes</td>
<td>Environmental lapse rate of ≥ 8°C over the minimum EML depth</td>
<td></td>
<td>N/A</td>
<td>N/A</td>
<td>RH at the top of the EML must be greater than at the EML base</td>
<td>Must be above the surface</td>
<td>≥ 200 mb</td>
</tr>
<tr>
<td>Cordeira et al. (2017)</td>
<td>Radiosondes</td>
<td>Environmental lapse rate of ≥ 8 °C⋅km⁻¹ over the minimum EML depth</td>
<td></td>
<td>N/A</td>
<td>N/A</td>
<td>≥ 30°C over the minimum EML depth</td>
<td>N/A</td>
<td>Between 900—400 mb</td>
</tr>
<tr>
<td>Ribeiro and Bosart (2018)</td>
<td>CFSR</td>
<td>Environmental lapse rate of ≥ 7.5 °C⋅km⁻¹ over the minimum EML depth and an average lapse rate below EML base of ≤ 7.5°C⋅km⁻¹</td>
<td>≥ 100 J/kg of MUCAPE</td>
<td>N/A</td>
<td>N/A</td>
<td>RH at the top of the EML must be greater than at the EML base</td>
<td>First level of the steep lapse rate layer; must be at least 1000 meters above the surface but below 500 mb</td>
<td>≥ 150 mb</td>
</tr>
<tr>
<td>Li et al. (2020)</td>
<td>ERA5</td>
<td>Environmental lapse rate of ≥ 8 °C⋅km⁻¹ over the minimum EML depth and an average lapse rate below EML base of ≤ 8 °C⋅km⁻¹</td>
<td></td>
<td>N/A</td>
<td>N/A</td>
<td>RH at the top of the EML must be greater than at the EML base</td>
<td>First level of the steep lapse rate layer; must be at least 1000 meters above the surface but below 500 mb</td>
<td>≥ 200 mb</td>
</tr>
</tbody>
</table>
4) A higher relative humidity (RH) at the top of the EML compared to the EML base, and

5) An average lapse rate below the EML base of less than \(8 \, ^\circ\text{C} \cdot \text{km}^{-1}\).

The depth and lapse rate thresholds (criteria 2 and 5) were consistent with Li et al.'s (2020) selection following testing of various lapse rate and thickness combinations after they tested various combinations.
determined the frequency of the EML in ERA5 was highly sensitive to the depth and stability criteria used. These thresholds were more restrictive than Ribeiro and Bosart (2018) because, unlike the CFSR, ERA5 accurately depicted both the magnitude and spatial distribution of mid-level lapse rates (Taszarek et al. 2021b). Criterion 3 was designed to eliminate the inclusion of steep lapse rates in the boundary layer and those that existed in the upper troposphere as the environmental lapse rate approached the dry adiabatic lapse rate (Ribeiro and Bosart 2018). The final criterion further ensured that mixed layers originating at the surface were not identified as EMLs.

Though the algorithm was not capable of discerning the physical processes responsible for soundings classified as EMLs, the MUCAPE requirement was added to eliminate some soundings that did not form via the downstream advection of a well-mixed boundary layer off of high terrain. Both Lanicci and Warner (1991a) and Ribeiro and Bosart (2018) also used a minimum convective instability threshold. Similar to Ribeiro and Bosart (2018), the inclusion of the instability criterion eliminated a number of the soundings with subsidence inversions. Many of these soundings occurred in the higher terrain during the cool season and had zero MUCAPE, owing to cold and/or dry low levels. Neither the $> 0 \text{ J} \cdot \text{kg}^{-1}$ nor the $\geq 100 \text{ J} \cdot \text{kg}^{-1}$ MUCAPE threshold tested eliminated all such soundings. The $> 0 \text{ J} \cdot \text{kg}^{-1}$ threshold was ultimately selected because it was less restrictive, while still eliminating many of the cases described above. Though adding the instability requirement did substantially reduce the frequency of detected EML days (Fig. 2), it had the added advantage of highlighting soundings with convective potential, which was relevant to the portion of our study interested in the relationship between the EML and severe storm reports.
To illustrate the application of the five presented criteria, six ERA5 soundings were plotted on skew T-log p diagrams (Fig. 3). Continuous layers of lapse rates at least 8 °C·km⁻¹ were labeled in blue to highlight candidate EMLs. Two of the six examples met all criteria and were therefore classified as EMLs by the algorithm (Fig. 3a,b). The sounding in Figure 3a featured nearly 1000 J·kg⁻¹ of MUCAPE, a lapse rate greater than 9 °C·km⁻¹, and an EML base around 750 mb. RH also rapidly decreased with height within the nearly 300 mb steep lapse rate layer. Due to the large capping inversion present, the lapse rate below the EML base was 1.8 °C·km⁻¹, well below the maximum allowed value of 8 °C·km⁻¹. The second valid EML extended through a deeper layer with a lapse rate of slightly lesser magnitude (Fig. 3b).
Figure 3. ERA5 soundings illustrating EML cases (a, b) and cases that fail to meet all five EML criteria (c, d, e, f). Layers at least 150 mb thick with lapse rates greater than 8 °C km⁻¹ are marked in blue to highlight candidate EMLs.
Although nearly dry adiabatic (9.5 °C · km⁻¹) through a 300 mb layer that featured decreasing RH with height, the profile out of southwestern Nebraska (Fig. 3c) did not have an EML, despite sufficient MUCAPE. Instead, the surface-based steep lapse rates were consistent with that of a sounding from the EML source region, such as west of a Great Plains dryline (Schaefer 1986). Criterion five ensured this layer was not labeled as an EML. The profile in Figure 3d met all EML criteria except for criterion 3, which required that the base of the steep lapse rate layer be located below 500 mb. The steep lapse rates, which started in the upper troposphere at approximately 435 mb, clearly did not result from the advection of a surface-based mixed layer off of the high terrain. The sounding out of eastern Texas had a 200 mb layer of sufficiently steep lapse rates above the inversion and over 1400 J · kg⁻¹ of MUCAPE (Fig. 3e). However, the relative humidity at the top of the candidate EML (6%) was slightly less than the relative humidity at the bottom (~14%) of the layer, so it was not classified as an EML by the algorithm. The final sounding failed to meet the depth criterion but satisfied all other requirements (Fig. 3f). Lapse rates over 9 °C · km⁻¹ extended through a layer approximately 160 mb thick, just short of the algorithm’s 200 mb depth requirement.

The algorithm employing these five criteria was applied every three hours (0000, 0300, 0600, 0900, 1200, 1500, 1800, 2100 Z) to each grid point in the domain over the full 43-year period. Grid points that met all criteria were marked as having an EML at that time. Attributes of all grid points, including CAPE, CIN, and where applicable, EML thickness, EML base height, EML base pressure, EML lapse rate, average EML potential temperature, and maximum EML temperature were also collected for further analysis. A list of all calculated variables and parameters is available in Table 2.
Table 2
Variables Calculated in this Study

<table>
<thead>
<tr>
<th>Parameter</th>
<th>Unit</th>
<th>Parameter</th>
<th>Unit</th>
</tr>
</thead>
<tbody>
<tr>
<td>EML depth</td>
<td>Pa, m</td>
<td>CAPE (SB, ML, MU)</td>
<td>J \cdot kg^{-1}</td>
</tr>
<tr>
<td>EML lapse rate</td>
<td>°C \cdot km^{-1}</td>
<td>CIN (SB, ML, MU)</td>
<td>J \cdot kg^{-1}</td>
</tr>
<tr>
<td>EML base height</td>
<td>m AGL</td>
<td>Normalized MUCAPE &amp; MUCIN</td>
<td>J \cdot kg^{-1} \cdot m^{-1}</td>
</tr>
<tr>
<td>EML base pressure</td>
<td>Pa</td>
<td>700–500 mb lapse rates</td>
<td>°C \cdot km^{-1}</td>
</tr>
<tr>
<td>Maximum, minimum, &amp; mean EML potential</td>
<td>K</td>
<td>2–5 km lapse rates</td>
<td>°C \cdot km^{-1}</td>
</tr>
<tr>
<td>temperature</td>
<td></td>
<td>700 mb temperature</td>
<td>K</td>
</tr>
<tr>
<td>Maximum, minimum, &amp; mean EML temperature</td>
<td>%</td>
<td>Lid strength</td>
<td>°C</td>
</tr>
<tr>
<td>Steepest EML lapse rate</td>
<td>°C \cdot km^{-1}</td>
<td>0–6 km wind shear</td>
<td>m \cdot s^{-1}</td>
</tr>
<tr>
<td>Mean boundary layer mixing ratio (below EML)</td>
<td>kg \cdot kg^{-1}</td>
<td>0–1 &amp; 0–3 km SRH</td>
<td>m^2 \cdot s^{-2}</td>
</tr>
<tr>
<td>Mean boundary layer equivalent potential</td>
<td>K</td>
<td>500 mb geopotential height</td>
<td>gpm</td>
</tr>
<tr>
<td>temperature (below EML)</td>
<td></td>
<td>500 mb geopotential height</td>
<td>K</td>
</tr>
<tr>
<td>Surface temperature</td>
<td>K</td>
<td>500 mb temperature</td>
<td>K</td>
</tr>
<tr>
<td>Surface pressure</td>
<td>Pa</td>
<td>500 mb wind speed</td>
<td>m \cdot s^{-1}</td>
</tr>
<tr>
<td>Surface specific humidity</td>
<td>kg \cdot kg^{-1}</td>
<td>500 mb wind direction</td>
<td>°</td>
</tr>
<tr>
<td>Surface geopotential</td>
<td>m^2 \cdot s^{-2}</td>
<td>850 mb geopotential height</td>
<td>gpm</td>
</tr>
<tr>
<td>Surface u/v winds</td>
<td>m \cdot s^{-1}</td>
<td>850 mb temperature</td>
<td>K</td>
</tr>
<tr>
<td>Surface dew point</td>
<td>°C</td>
<td>850 mb wind speed</td>
<td>m \cdot s^{-1}</td>
</tr>
<tr>
<td>Maximum temperature in the profile</td>
<td>K</td>
<td>850 mb wind direction</td>
<td>°</td>
</tr>
</tbody>
</table>
EML Occurrence and Variability

The frequency of the EML and the number of days with an EML at each grid point were calculated for the full study period. Mapping the frequency of the EML annually and by season revealed the EML’s spatial distribution and seasonality while also permitting a comparison to previous climatologies. Consistent with previous studies, the frequency of the EML at each grid point was given in terms of the percentage of time steps with an EML present. To provide a difference metric of expected occurrence, the average number of EML days were calculated monthly, seasonally, and annually. In this study, an EML day was defined as a day (1200–1200 Z) where an EML was present at a grid point for at least one of the three-hourly time steps in the 24-hour period. The diurnal distribution of EML soundings aggregated over the full study domain was also examined (Fig. 4).

EML days were then used to explore the variability of the EML. The maximum, minimum, and standard deviation were calculated annually and for each season at every grid point. Spatial plots of the maximum and minimum number of EML days within the period revealed the upper and lower bounds at each location, while the standard deviation indicated the magnitude and locations of greatest interannual variability. Both interannual and intra-annual variability in EML days were highlighted by cumulative frequency plots at eight cities: Fort Stockton, TX; Lubbock, TX; Oklahoma City, OK; Garden City, KS; Colorado Springs, CO; Valentine, NE; Bismarck, ND; and Kansas City, MO (Fig. 4). At the grid point containing each city center, cumulative frequencies were calculated for every year in the 43-year period, along with the period mean. The cities selected provided a roughly even geographic distribution within the >10 EML days per year contour.
Figure 4. The study domain (in white) and eight US cities (red dots) selected for further analysis. All cities are located within the >10 annual EML day contour.

EML Attributes

To explore how relevant characteristics of the EML varied with space and time, spatial plots of mean EML attributes were created, as were boxplots at the eight aforementioned cities. Mean annual and seasonal EML thickness, base height, base pressure, average potential temperature, and lapse rates were calculated at each grid point by compositing all time steps in the period with an EML at that location. Seasonal box plots of the same variables at each city contained all time steps with an EML at the city’s latitude and longitude, revealing the seasonality and distribution of these EML attributes within the study period. Spatial plots and
boxplots of CAPE and CIN were calculated the same way to understand the magnitude and variability of the instability and capping inversion associated with the EML. CAPE and CIN were calculated with the virtual temperature correction (Doswell and Rasmussen 1994) using the most unstable parcel in the lowest 3 km AGL (MU) and the lowest 100 mb mixed layer parcel (ML).

The areal extent of the EML was another attribute of interest. The total daily EML coverage over the study domain was calculated by multiplying the number of EML grid points each day by the area of one grid box (961 km$^2$). Using all days in the 1979–2021 period, including those with zero EML coverage, box plots were created to look at the seasonal distribution of daily EML area including measures of central tendency. The interannual variability of EML coverage is also shown via yearly time series of median daily EML area and total EML area.

**Long-Term Trends in the EML**

To determine if there were notable trends in EML occurrence during the 43-year period, EML days were first summed at each grid point for every individual year in the dataset. Trends were then calculated at each grid point on the yearly values using the Theil-Sen estimator (Wilcox 2010). This method is frequently used to calculate trends in the atmospheric sciences (e.g., Gensini and Brooks 2018; Tang et al. 2019; Taszarek et al. 2021a; Pilguj et al. 2022) because it is nonparametric and relatively insensitive to outliers (Wilcox 2010). Kendall’s $\tau$ statistic and a $p$ value at the 0.05 significance level were used to assess the significance of the Theil-Sen slope.
Trends in EML characteristics including lapse rates, potential temperature, and CIN were calculated similarly to trends in EML days. The Theil-Sen estimator was applied to the annual means at each grid point, which were calculated using the 3-hourly values from time steps with an EML. As with the EML day trends, the slopes denote change per year and were tested for significance using Kendall’s $\tau$ statistic and a $p$ value at the 95% confidence level.

**Severe Storms Relative to the EML**

To determine if there was a preferred location for severe storms relative to the EML, 2D histograms were created to show the relationship between individual EML centroids and the centroids of hail and tornado practically perfect hindcasts (PPH). PPH probabilities of hail and tornadoes were calculated using the method described in Gensini et al. (2020). A Gaussian filter was applied to daily (1200–1200 Z) severe and significant severe storm reports to create spatially smoothed probabilities of hail and tornado occurrence, consistent with that of a perfect Storm Prediction Center convective peril outlook. Polygons with $\geq 15\%$ and $\geq 30\%$ severe and significant severe hail PPH probabilities were used, as were $\geq 5\%$ and $\geq 10\%$ PPH tornado and significant tornado probabilities. The same Gaussian smoother was applied to daily EML grid points, resulting in smoothed probabilities of EML occurrence ranging from 0 to 100%. For this portion of the study, polygons with probabilities of an EML $\geq 80\%$ were considered EMLs. Slightly higher (lower) thresholds resulted in fewer (greater) EMLs but did not notably impact the results. The centroids of all EML, PPH hail, and PPH tornado polygons were calculated. A spatial join was then used to relate the EML polygons to the hail and tornado PPH geometries. PPH hail and tornado polygons were joined to an EML polygon if they intersected or were
within a 250 km search radius. For pairs of joined polygons, the distance (km) and compass bearing between the two were calculated. The distances and directions for all pairs were aggregated into bins and plotted on a 2D polar histogram, the middle of which represents the EML centroid. While this methodology cannot determine whether the storms occurred outside of the EML bounds or within it, it does provide insight into the typical proximity of severe storms relative to the EML.
CHAPTER 4

RESULTS

EML Occurrence and Variability

Mean annual EML days were most frequent along and just east of the EML source regions in the western CONUS and northern Mexico, with 15–30 EML days per year over most of the Great Plains (Fig 5a). Seasonally, peak EML occurrence was found in the spring (MAM), with the greatest number of EML days concentrated in the southern Great Plains and northeastern Mexico (12–21 days \(\cdot\) yr\(^{-1}\); Fig 5c). This springtime maximum and its location are consistent with previous climatologies including Lanicci and Warner (1991a), Ribeiro and Bosart (2018), and Li et al. (2020). While the EML source region was primarily confined to northern Mexico in the early spring, it expanded northward following maximum solar radiation later in the warm season to include the high terrain of the western CONUS (Lanicci and Warner 1991a). The corridor of peak EML days followed, with a northward shifted maximum of slightly lesser magnitude along the western extent of the central and northern Great Plains in the summer (JJA; Fig. 5d). The northward shift was also evident on a monthly basis from April to August (Fig. 6d-h). In all months, EMLs east of the Mississippi River were rare, with one or fewer EML days per month (Fig. 6a-l). EMLs were also very rare in the winter (DJF) and fall (SON), owing to less intense surface heating and limited instability (Fig 5b,e).
Figure 5. Mean EML days a) annually, and by season for b) winter, c) spring, d) summer, and e) fall. An EML day was defined as a day (starting at 1200 Z) where an EML was present at a grid point for at least one hour.
Figure 6. Mean number of EML days per month for a) January, b) February, c) March, d) April, e) May, f) June, g) July, h) August, i) September, j) October, k) November, and l) December.
EML frequency largely mirrored the spatial distribution and seasonal variability of EML days. The locations of the spring and summer maxima (Fig 7c,d) aligned with Ribeiro and Bosart (2018) and were quantitatively very similar to Li et al. (2020; cf. their Fig. 12) in these two seasons. Magnitudes of annual, winter, and fall EML frequency (Fig. 7a,b,e) were less than that of Li et al. (2020) due to the addition of our minimum MUCAPE threshold. The difference was most notable in the winter when this criterion resulted in the removal of the greatest number of soundings (Fig. 2). Compared to Ribeiro and Bosart (2018), EML frequencies were less in all seasons, likely due to the sensitivity of EMLs to the chosen criteria, including the lapse rate and thickness thresholds (Li et al. 2020). Electing to use objective criteria of any threshold is a limitation because no one set of criteria can detect all EMLs. It is possible that, for example, EMLs over certain regions were shallower or had lower lapse rates on average than those over the Great Plains, meaning the thresholds employed here, and by similar studies, have largely failed to capture them. Regardless of the threshold selected, some borderline cases (e.g., Fig. 3e and 3f) are always going to be excluded. Additionally, due to averaging when fitting observations to the reanalysis grid, some errors may also be introduced in detecting the EML. However, consistency among climatologies using a range of reanalyses and observations supports the methodology and choice of criteria herein, although continuing to evaluate and adjust the criteria used to identify EMLs remains important.

Although there is a well-established seasonal cycle for the EML, there was also substantial variability from year-to-year and season-to-season. The greatest year-to-year variability was found in the Great Plains, where the standard deviation of EML days was largest (4–10 days; Fig. 8c). Parts of southwest Texas and northwestern Oklahoma experienced the greatest range in annual EML days within the 1979–2021 period, with as many as 50–56 and as
Figure 7. Percentage of three-hourly time steps with an EML present a) annually, and by season for b) winter, c) spring, d) summer, and e) fall.
Figure 8. Maximum (a, d, g, j, m), minimum (b, e, h, k, n), and standard deviation (c, f, i, l, o) of EML days annually (a, b, c) and in winter (d, e, f), spring (g, h, i), summer (j, k, l), and fall (m, n, o).
few as 9–15 EML days in a year (Fig. 8a,b). On a seasonal basis, the range between the period maximum and period minimum was as high as 38 EML days in spring in southern Texas and northeastern Mexico (Fig. 8g,h) and 24–32 EML days in the western portion of the central and northern Great Plains in summer (Fig. 8j,k). Interannual variability of this magnitude may have had implications for the hazards supported by EMLs, such as severe convection. While the EML is not sufficient for convection, periods with above or below-average EML days could have contributed to greater or fewer favorable environments compared to normal. The relatively small variability in the fall and winter (Fig. 8d,e,f,m,n,o) reflected the limited number of EML days in both seasons. Interannual variability is also shown with annual time series of EML days at the eight selected cities (Fig. A1).

Intra-annual variability differed by location. Most EML days in the southern Great Plains occurred before July, after which variability noticeably increased (Fig. 9a-d). This was true of Fort Stockton, Lubbock, Oklahoma City, and Garden City, which saw a relatively small spread in year-to-year cumulative EML days from January through June, followed by increased variability. Increased variability after June may have been a function of a few factors, all of which make it more difficult to consistently get EMLs. In the summer, this was likely a northward displaced jet and a less focused EML source region compared to spring, while in the fall and winter, less intense surface heating and instability were likely the major contributors. Further north in Valentine and Bismarck, EML days increased relatively linearly from April through October, with slightly more consistent variability throughout the year compared to cities further south (Fig. 9f,g). The steady increase in EMLs and fairly constant variability in the northern Great Plains may reflect the fact that although EMLs are most common in the summer, the region lacks a source region as focused as that of the southern Great Plains in the spring.
Figure 9. Annual cumulative sums of EML days at eight US cities. Cumulative sums for each year are shown in grey, while the thick black lines denote the 1979–2021 means.
Cumulative frequency plots for additional cities are found in Figure A2.

The diurnal distribution of EMLs across the study domain differed throughout the year. In spring, EML soundings were most frequent from 0600–1500 Z and least common from 2100–0000 Z (Fig. 10c). Fewer EML soundings in the afternoon and early evening appeared to be the result of a couple of factors. In some cases, storms initiated and eroded the EML locally in the afternoon and/or evening. In other instances, typical afternoon surface-based heating resulted in steep low-level lapse rates, such that the algorithm’s fifth criterion, requiring lapse rates below the EML base of \(<8 \, ^\circ C \cdot km^{-1}\), was not satisfied if the steep lapse rates continued above the PBL without a break (i.e., there was no inversion present). The summertime distribution of EMLs was similar to the spring distribution; however, the difference between the number of soundings from 0600–1500 Z and those from 2100–0000 Z was slightly more pronounced (Fig. 10d). In contrast, winter and fall saw very minor differences in the number of EMLs throughout the day (Fig 10b,e).

### EML Attributes

Most EML attributes varied seasonally, including EML area. When all days in the study period were considered, including those with no EMLs within the domain, spring had the largest daily mean and median EML coverage, followed by summer (Fig 11). The total daily EML area in the spring averaged just under 462,000 km², with a median area of approximately 260,000 km², slightly smaller than the state of Colorado (Fig. 11). In the summer, the mean and median daily EML area were about 286,000 km² and 147,000 km², respectively (Fig. 11). These measures of central tendency indicate that while some EMLs were expansive, most only covered
Figure 10. Cumulative number of 1979–2021 EML soundings by hour of the day a) annually, and by season for b) winter, c) spring, d) summer, and e) fall.
a small portion of the study domain. Although winter and fall had relatively limited EML coverage due to the less frequent occurrence of the EML, there was a substantial range in daily EML area in all seasons, with 2,368 days in the period with zero EML grid points within the domain, and 104 days with EML coverage that exceeded two million km² (Fig. 11).

Figure 11. Daily EML area in millions of km² from 1979–2021 for winter, spring, summer, and fall (includes days with no EML coverage). Boxes indicate the interquartile range, while dots and lines within the boxes denote the mean and medians, respectively. Whiskers indicate the 1st and 99th percentiles and circles represent the outliers.

In addition to seasonal differences, there was also sizable interannual variability in EML area. Both the median daily EML area (Fig. 12a) and the cumulative annual EML area (Fig. 12b) varied considerably from one year to the next in all seasons. Median daily EML coverage in the spring ranged from 103,000 km² in 2015 to 588,000 km² in 2012, while summer had an interannual range of nearly 363,000 km² between 2011 and 1981 (Fig. 12a). For cumulative
annual EML area, spring had the greatest year-to-year range (54 million km$^2$), followed by summer (39 million km$^2$; Fig. 12b). Once again, this variability emphasizes that although there may be an established climatology for the EML, individual years can deviate substantially from the mean and median. Years with a large daily median EML area tended to correspond to years with high cumulative EML coverage. However, there were periods, such as the spring of 1996, where the two did not match up due to the sensitivity of the cumulative annual area to outliers.

Figure 12. a) Median daily EML area (million km$^2$) and b) cumulative EML area (ten million km$^2$) each year from 1979–2021 by season. Both calculations include all days in the study period including those with zero EML grid points.

EML base pressure and base height also varied throughout the year. Annually, mean EML bases were located between 700–800 mb and 1500–2500 m AGL across the Great Plains (Fig. 13a, Fig. 14a). EML base pressure averaged 700–750 mb in the spring in the same region (Fig. 13c), consistent with Ribeiro and Bosart's (2018) climatology. Compared to spring, EML bases were found at higher pressures in summer and fall over most of the Great Plains and parts of the Midwest (750–800 mb; Fig. 13c-e). The same was true of mean base heights in the central and northern Great Plains, with bases located closer to the surface in summer and fall.
Figure 13. Mean EML base pressure a) annually, and by season for b) winter, c) spring, d) summer, and e) fall.
Figure 14. Mean EML base height a) annually, and by season for b) winter, c) spring, d) summer, and e) fall.
(1500–2000 m AGL; Fig. 14d,e) compared to spring (2000–2500 m AGL; Fig. 14c). This result is counterintuitive. Warmer temperatures in the summer suggest deeper PBLs, implying EML bases should instead rise from spring to summer. However, Lanicci and Warner (1991a) noted the same decrease throughout the warm season and determined it was due to the more frequent occurrence of synoptic patterns favoring subsidence in the late spring and summer. In all seasons, EML base heights increased with eastward extent across the domain, while mean base pressures decreased, consistent with the conceptual model of EML formation (Carlson and Ludlam 1968; Carlson et al. 1983). This eastward increase in EML base height in the Great Plains is physically reasonable because the surface elevation decreases eastward from the Rocky Mountains. However, since there were relatively few EMLs east of the Mississippi River, these values, particularly in the winter and fall, should be interpreted with caution.

EML depth was largely consistent across space and time, with mean depths of 225–250 mb at nearly all selected cities in all seasons (Fig. 15). The exceptions were Valentine and Bismarck in the winter and Fort Stockton and Oklahoma City in the fall, the latter three of which had mean EML depths below 225 mb. Since these four cities had very few EMLs in winter and fall, their mean values may not be particularly meaningful. For example, the wintertime average at Valentine appeared to be the result of a few deep EMLs skewing the mean. In other cases, because it was so rare to get EMLs in the cool season, the EMLs that did occur fell on the very low end of the permitted EML thickness threshold. Although the means were generally consistent, there was still a sizable spread in EML depth, with thicknesses that exceeded 300 and even 350 mb in all seasons (Fig. 15). It is worth noting that some of the consistency in EML depth was likely due in part to the relatively strict EML depth criterion ($\geq$200 mb) within the EML detection algorithm.
Figure 15. Boxplots of seasonal EML depth at eight US cities. Boxes indicate the interquartile range, while dots and lines within the boxes denote the means and medians, respectively. Whiskers indicate 1.5 times the interquartile range and circles represent the outliers.

An east-to-west oriented potential temperature gradient was present between the Rockies and the Great Plains annually, with the highest mean potential temperatures tucked along and just east of the mountains, close to the EML source regions (Fig. 16a). In the winter, spring, and fall, mean EML potential temperature generally increased from north to south (Fig. 16b,c,e), with springtime values that ranged from 322 K over parts of Mexico to 306 K near the US-Canadian border. A function of the range of EML source region temperatures, this north-south gradient was also found in spring by Lanicci and Warner (1991a). Due to the decreased meridional temperature gradient in summer, mean potential temperature was nearly constant.
Figure 16. Mean EML potential temperature a) annually, and by season for b) winter, c) spring, d) summer, and e) fall.
across the domain. In addition to being the least variable, summer also had the highest mean potential temperatures, with values of 318–320 K at all locations except for the far northern CONUS (Fig. 16d). Mean warm season values in the Great Plains were similar to Ribeiro and Bosart's (2018) range of 316 to 320 K.

Annually, the steepest mean EML lapse rates (8.9–9.0 °C·km⁻¹; Fig. 17a) were located close to the high terrain with decreasing values to the east, similar to various climatologies of 700–500 mb lapse rates in the CONUS (e.g., Brooks et al. 2003; Taszarek et al. 2021b). The orientation of the steepest lapse rates in spring were consistent with a primary source region of northern Mexico. From spring to fall, the steepest mean lapse rates (8.8–9.1 °C·km⁻¹; Fig. 17c-e) expanded northward with time, following maximum EML frequency. This northward expansion from spring to summer was consistent with a much broader source region extending into the high terrain of the western CONUS later in the warm season (Lanicci and Warner 1991a) and aligned well with Ribeiro and Bosart (2018; cf. their Fig. 5). Quantitatively, the magnitude of the steepest lapse rates in spring and summer were approximately 0.8 °C·km⁻¹ larger than Ribeiro and Bosart's (2018) values. This difference may be related to the much higher vertical resolution of ERA5 compared to CFSR. Ribeiro and Bosart (2018) noted that CFSR EML lapse rates were an average of 0.7 °C·km⁻¹ less than lapse rates calculated from observed soundings, in part due to limited vertical levels. While this study did not perform a comprehensive comparison of ERA5 and observed EML soundings, Taszarek et al. (2021b) found a correlation of 0.94 between the midlevel lapse rates in ERA5 and observed soundings, providing confidence in ERA5’s ability to represent EML lapse rates reasonably well.

Seasonally, the highest mean absolute MLCIN (375–450 J·kg⁻¹) associated with EMLs was confined to far southern Texas and northern Mexico in the winter and spring (Fig. 18b,c),
Figure 17. Mean EML lapse rate a) annually, and by season for b) winter, c) spring, d) summer, and e) fall.
Figure 18. Mean EML MLCIN a) annually, and by season for b) winter, c) spring, d) summer, and e) fall.
the same areas where EML days were maximized (Fig. 5b,c). The magnitude of the capping decreased with northward extent in both seasons, with a mean of 225–375 J · kg\(^{-1}\) of absolute EML MLCIN in spring over the remaining portions of the southern and central Great Plains (Fig. 18c). This northward decrease in the strength of the cap was anticipated because the magnitude of the cap is partially a function of the surface temperatures in the EML source regions. As a result, the warmer source regions further south in the spring typically had warmer EML bases and larger inhibition. From spring to summer, the region of greatest inhibition shifted northward and expanded, with mean absolute values of 300–375 J · kg\(^{-1}\) of MLCIN over the majority of the Great Plains (Fig. 18d). These warm season means in the Great Plains were well above the threshold (absolute CIN of 75–100 J · kg\(^{-1}\)) under which most convective storms in favorable environments form (Bunkers et al. 2010; Gensini and Ashley 2011; Hoogewind et al. 2017; Taszarek et al. 2021a). Additionally, absolute CIN above 200 J · kg\(^{-1}\) is generally considered prohibitive (Rasmussen et al. 2020), meaning that a substantial number of EML environments were not supportive of convection, even if all other ingredients were favorable. Although Ribeiro and Bosart (2018) used a different methodology to look at MLCIN associated with EMLs, they also found mean absolute values in spring of greater than 200 J · kg\(^{-1}\) over the Great Plains and Mississippi Valley. Despite large mean inhibition, there were plenty of instances where EMLs may have been supportive of convection, including severe storms. For example, while all cities had prohibitive mean and median absolute EML MLCIN in at least one season, they also saw a sizable number of EMLs with moderate and even weak inhibition (Fig. 19). Some of these smaller values may have occurred near the edge of the EML, where storms are more likely because CIN is generally less compared to the EML center (Carlson and Ludlam 1968; Carlson et al. 1983; Keyser and Carlson 1984; Lanicci and Warner 1991b). Mean spatial plots and
seasonal boxplots for additional EML variables are available in Appendix B.

Figure 19. Boxplots of seasonal EML MLCIN at eight US cities. Boxes indicate the interquartile range, while dots and lines within the boxes denote the means and medians, respectively. Whiskers indicate 1.5 times the interquartile range and circles represent the outliers.

Long-Term Trends in the EML

Statistically significant increases in EML days were found annually and in all four seasons (Fig. 20). On an annual basis, increases were concentrated in the high terrain of the Great Plains, extending from Montana and western North Dakota to northern Mexico (Fig. 20a).
Figure 20. Theil-Sen slope of the yearly grid-point sum of EML days a) annually, and by season in b) winter, c) spring, d) summer, and e) fall. Hatching indicates statistical significance ($p$ value $\leq 0.05$) using Kendall’s $\tau$ statistic.
The most notable trend in annual EML days occurred from northern Nebraska to southwest Texas, with increases on the order of 0.4–0.5 EML days per year, equivalent to approximately one more EML day every two years, or four to five more days per decade. Spring and summer (Fig. 20c,d) were largely responsible for the annual trend, with significant springtime increases of 0.15–0.30 EML days each year in northern Mexico and western Texas, Oklahoma, and Kansas (Fig. 20c). In the summer, the largest year-to-year changes were located further north, with 0.2–0.45 more EML days each year over the high terrain of the central and northern Great Plains (Fig. 20d). These warm season increases in EML days were likely responsible for the differential warming with height noted in observations and ERA5 from 1980–2018 (Pilguj et al. 2022). Maximized in the spring and summer between 2–3 km AGL over the Great Plains (Pilguj et al. 2022), the significant warming trend was consistent with the typical height of the EML base in these seasons (Fig. 14c,d), and the more frequent appearance of the EML over this region with time (Fig. 20c,d).

The upward trend in EML days was likely driven by warming and drying in the western CONUS during the study period. Increasing near-surface temperatures in the western and southwestern CONUS during this time were well documented by both observations and reanalysis, with observed increases between 0.1–0.5 °C per decade (IPCC 2022) and statistically significant increases in the 95th percentile of ERA5 2-meter temperatures (Taszarek et al. 2021a). Likewise, drying trends in the western CONUS were observed via decreases in 0–500 m mixing ratio and 0–4 km relative humidity in observations and ERA5 (Taszarek et al. 2021a; Pilguj et al. 2022). Warming surface temperatures and steepening low-level (0–3 km) lapse rates in the EML source regions of both datasets (Taszarek et al. 2021a; Pilguj et al. 2022) strongly suggest that
the warming and drying in the west contributed to the increase in EML days by increasing the maximum EML temperature and steepening EML lapse rates.

Since the EML was defined as a continuous layer of steep lapse rates, the maximum EML temperature was located at the EML base. The temperature at the EML base was a direct reflection of the surface temperatures in the western CONUS, since the PBL was advected downstream during EML formation. Due to the low-level warming and drying out west, statistically significant trends in maximum EML temperature were found annually and in spring and summer (Fig. 21a,c,d). Spring saw significant increases in maximum EML temperature in the western portions of the central and southern Great Plains (0.1-0.2 K · yr\(^{-1}\)), with a secondary area of warming centered over the middle Mississippi Valley (0.15-0.25 K · yr\(^{-1}\); Fig. 21c). Summertime increases in EML base temperature extended further north, from eastern Colorado through Montana (~0.05 K · yr\(^{-1}\); Fig. 21d). Since potential temperature is nearly constant in a well-mixed layer such as the PBL of the EML source regions, warming surface temperatures in the western CONUS should also result in a higher potential temperature through the depth of the layer. On an annual basis, the increases in mean EML potential temperature were greatest just east of the Rockies (0.1-0.15 K · yr\(^{-1}\); Fig. 22a), largely collocated with the increases in maximum EML temperature (Fig. 21a). Upward trends in the warm season (Fig. 22c,d) overlapped with those of maximum EML temperature, although the statistically significant increases in mean EML potential temperature in the summer were more widespread.

Annually, the western corridor of increasing EML base temperatures (Fig. 21a) matched well with the increase in EML days (Fig. 20a). This was anticipated because warmer EML bases should support steeper EML lapse rates, thereby increasing the number of EMLs, since EMLs were defined as lapse rates of a certain magnitude. Spatially, the locations of statistically
Figure 21. Theil-Sen slope of mean yearly EML maximum temperature a) annually, and by season for b) winter, c) spring, d) summer, and e) fall. Hatching indicates statistical significance ($p$ value $\leq 0.05$) using Kendall’s $\tau$ statistic.
Figure 22. Theil-Sen slope of yearly mean EML potential temperature a) annually, and by season for b) winter, c) spring, d) summer, and e) fall. Hatching indicates statistical significance ($p$ value $\leq 0.05$) using Kendall’s $\tau$ statistic.
significant trends in maximum EML temperature and mean EML lapse rates were very similar, as were the increases in mean EML lapse rates and EML days, further supporting the notion that the increase in EML days was driven by the steepening of lapse rates by warmer near-surface temperatures in the EML source regions. Like maximum EML temperatures, mean EML lapse rates increased in the western Great Plains annually (0.002-0.01 °C·km⁻¹·yr⁻¹; Fig. 23a), with a secondary area of increases over eastern Kansas and the western Middle Mississippi Valley. Although small in magnitude, the largest increases in spring and summer were found in the central Great Plains (Fig. 23c,d). East of the Rockies, the annual increases were similar in magnitude and location to the increases in the 95th percentile of mid-level lapse rates in ERA5 over approximately the same period (Taszarek et al. 2021a).

In addition to supporting steeper lapse rates, warmer EML base temperatures also supported a more strongly capped environment below the EML base, as evidenced by significant increases in absolute EML MLCIN (Fig. 24) collocated with increases in maximum EML temperature (Fig. 21). Annually, the absolute MLCIN associated with EML soundings increased 1.5–4 J·kg⁻¹, or about 15–40 J·kg⁻¹ each decade over most of the western Great Plains (Fig. 24a). Seasonally, the most substantial and widespread increases in EML-related capping occurred in the spring over the southern half of the Great Plains (2–4 J·kg⁻¹ yr⁻¹; Fig. 24c). In the summer, the largest increases in inhibition were concentrated over eastern Montana (Fig. 24d). Increased stability could have major implications for the spatial distribution and frequency of the SCS climatology in the CONUS. Increasing inhibition could prevent convection entirely, or slow the timing of CI, in some cases enhancing the likelihood of stronger storms through the buildup of instability with time (e.g., Bunkers et al. 2010; Rasmussen et al. 2020). The trend towards more absolute CIN is consistent with previous work which found significant increases in
Figure 23. Theil-Sen slope of yearly mean EML lapse rates a) annually, and by season for b) winter, c) spring, d) summer, and e) fall. Hatching indicates statistical significance ($p$ value $\leq 0.05$) using Kendall’s $\tau$ statistic.
Figure 24. Theil-Sen slope of yearly mean EML MLCIN a) annually, and by season for b) winter, c) spring, d) summer, and e) fall. Hatching indicates statistical significance (p value ≤0.05) using Kendall’s τ statistic.
inhibition over the Great Plains in observations and ERA5 over the last four decades in spring and summer (Taszarek et al. 2021a; Pilguj et al. 2022). While these studies did not isolate EML-related CIN, the increases in stability were more than likely influenced by warmer EML bases and more frequent EMLs.

Spatially, the regions that saw increased inhibition associated with EMLs (Fig. 24) were similar to those with increased EML days (Fig. 20), implying that the stronger capping was directly linked to the warmer EML base temperatures and steeper lapse rates that supported the increase in EMLs. The increased number of EML days and warmer EML base temperatures may help explain the rise in environments that inhibited CI (absolute CIN > 75 J kg⁻¹) and the decrease in the number of favorable environments that produced precipitation over the Great Plains from 1979 to 2019 (Taszarek et al. 2021a). The increased magnitude of EML MLCIN may have been particularly impactful in the far southern Great Plains and northern Mexico in the spring, which also saw a decreasing trend in MLCAPE associated with EML soundings (Fig. 25b). This region experienced a corresponding decrease in the number of environments initiating in spring (Taszarek et al. 2021a), representing a decreased frequency of storms in a region that has generally been considered the climatological maximum of springtime severe convection in North America. Theil-Sen slopes of monthly EML days and additional variables are available in Appendix C.

If the warming and drying trends in the western CONUS continue to drive an increase in PBL lapse rates that are subsequently advected downstream, steeper mid-level lapse rates will likely continue to cause an increase in EML frequency, while the warmer temperatures at the base of this advected layer and more frequent EMLs will continue to increase the magnitude of CIN, particularly across the Great Plains. Climate change simulations suggest that this increase
Figure 25. Theil-Sen slope of yearly mean EML MLCAPE a) annually, and by season for b) winter, c) spring, d) summer, and e) fall. Hatching indicates statistical significance (p value ≤0.05) using Kendall’s τ statistic.
in inhibition will persist in the CONUS (Trapp and Hoogewind 2016; Hoogewind et al. 2017; Rasmussen et al. 2020; Haberlie et al. 2022; Ashley et al. 2023), which could be in part due to a continuing trend towards more EMLs in a future climate. However, the latter half of the 1979–2021 period experienced a particularly noteworthy drought in the western and southwestern CONUS, as indicated by the Palmer Drought Severity Index (Fig. 26). While it is likely that the warming and drying trends will continue to persist due to climate change, it is also possible that any future years with wetter conditions in the western CONUS may somewhat mitigate the increasing trend in EMLs and EML-related CIN, at least in the short term.

Figure 26. Time series of the Palmer Drought Severity Index from 1895 to 2022 (from NOAA 2023) for the southwestern CONUS. Negative values indicate below-average soil moisture and positive values indicate above-average soil moisture conditions. The red line is the LOESS curve, and the black box indicates the study period.

Severe Storms Relative to the EML

Severe storms associated with EMLs appeared to favor certain locations. Relative to EML centroids—located in the middle of the 2D polar histograms in Figure 27—the
Figure 27. The locations of PPH probabilities of a) ≥ 15% severe hail, b) ≥ 15% significant hail, c) ≥ 5% tornado, d) ≥ 5% significant tornado centroids relative to EML centroids (located in the center of the polar plots). Darker colors indicate more instances within each distance (km) and direction bin.
centroids of most PPH hail and tornado polygons were clustered between 250 and 750 kilometers (~155–466 miles), with compass bearings ranging from 345 to 90 degrees. These angles correspond to north-northwest and east, indicating that most PPH centroids were found to the north and east of their EML centers. This was true of severe (Fig. 27a,c) and significant severe (Fig. 27b,d) cases for both hail (Fig. 27a,b) and tornadoes (Fig. 27c,d), suggesting that the magnitude and type of peril had little impact on the location of the severe weather relative to the EML. Increasing the PPH thresholds to \( \geq 30\% \) for hail and \( \geq 10\% \) for tornadoes also had little impact on the distribution (not shown).

There were very few instances of PPH probabilities located to the south of their associated EML centers, likely because capping was stronger to the south compared to locations further north. Similarly, there were limited cases with PPH probabilities near the center of EMLs (within 250 km of the EML centroid) where capping was likely strongest. Instead of forming near the EML center, previous work has shown that storms tend to form near the EML edge where CIN tends to be lower, with the northern or western edges of the EML most common (Carlson and Ludlam 1968; Carlson et al. 1983; Keyser and Carlson 1984; Benjamin and Carlson 1986; Farrell and Carlson 1989; Lanicci and Warner 1991a,b; Banacos and Ekster 2010; Cordeira et al. 2017). Consistent with our results, more recent work has provided additional evidence that storms favor the northern EML boundary, finding the strongest ascent on the northern edge in spring (Ribeiro and Bosart 2018).

The relatively infrequent occurrence of storms along the EML’s western flank compared to previous work may be related to several factors including limitations in our methodology. In some cases, any convection east of a dryline (on the western edge of the EML) may have been too limited in coverage to reach the selected PPH thresholds. Alternatively, if storms also formed
from mechanisms other than the dryline and/or grew upscale to the east, severe storm reports may have been averaged into one large area of PPH probability with its centroid far east of the initial convection. This is of particular concern because both the PPH and EML probabilities were aggregated on a daily (1200–1200 Z) scale. The coarse temporal resolution likely also had implications for EML centroids, which generally advected to the east during the day. Due to this eastward shift with time, daily EML centroids may have been located further west than where the EML was located when convection occurred, making the region of severe storms appear further east relative to the EML than was actually the case. The use of a daily time scale also meant it was not possible to determine whether the storms and the EML were present at the same time. For example, severe storms could have occurred before an EML was present over a region, yet if both existed within the same 24-hour period in close proximity, they were assumed to be associated. Additionally, it is also possible that known biases in severe storm reports, such as population bias (e.g., Verbout et al. 2006; Potvin et al. 2019), may have skewed the results.

While there are a substantial number of limitations, this section introduced important considerations for future work looking at the relationship between the EML and severe storms. Using a sub-daily time scale would have provided a more accurate determination of the favored locations for hail and tornado reports relative to EMLs. Using the outlines of the PPH and EML probability polygons in addition to their centroids would have helped confirm whether most storms occurred within or outside of the EML boundary. Furthermore, using individual storm reports, despite their inherent biases, would have permitted more detailed analysis of CI relative to the EML edge.
CHAPTER 5

CONCLUSIONS

This study sought to create an updated high-resolution climatology of the EML to analyze EML variability and changes in EML occurrence and characteristics over the last four decades. First, spatial climatologies of EML days and EML frequency were created by applying an objective algorithm to ERA5 soundings. The period (1979–2021) maximum, minimum, and standard deviation of annual and seasonal EML days were then used, along with yearly cumulative sums, to investigate the interannual and intra-annual variability of the EML for the first time. Next, expected values and typical ranges of EML attributes including base height, depth, potential temperature, lapse rates, and CIN were calculated to understand their spatial distribution and seasonal variability. Trends in these attributes and EML occurrence were determined and tested for significance to evaluate how EMLs have changed through time. Finally, EML centroids were related to PPH hail and tornado centroids to determine if there were preferred locations for severe weather relative to the EML.

EML days were maximized in the spring, with the highest frequency concentrated over the southern Great Plains and northeastern Mexico (mean of 12–21 days). Following the poleward expansion of the EML source region from spring to summer, the peak shifted northward to the high terrain of the central and northern Great Plains (mean of 10–18 days). EML days were relatively rare outside of the Great Plains, owing to the increased distance from source regions, while limited surface heating and instability resulted in very few EML days in
the winter and fall. EML frequency generally aligned well with previous studies, with the most notable differences a result of the EML’s sensitivity to the criteria used to define it. While no set of criteria can capture all EMLs, the similarities amongst results, despite different datasets, combined with ERA5’s improved resolution of the cap compared to previous reanalyses (Taszarek et al. 2021b), provided confidence in the ability of the selected criteria to produce reasonably accurate and meaningful results.

Despite well-established normals, substantial interannual variability was found, with a difference of 15–40 EML days between the period maximum and minimum in spring throughout the southern and central Great Plains. Variability of this magnitude may have influenced severe storm frequency since EMLs can both suppress and enhance convection (e.g., Carlson and Ludlam 1968; Carlson et al. 1983; Graziano and Carlson 1987). Future work could determine whether years with a particularly high or low number of EML days were correlated with more or less severe storm reports compared to climatology. PPH probabilities of hail and tornadoes suggested severe storms were most common to the north and east of the EML centroid, and very infrequent close to or south of the EML center. While preferences for the northern flank and regions outside the EML center were consistent with previous studies (e.g., Lanicci and Warner 1991a; Ribeiro and Bosart 2018), additional work could relate EMLs and severe storm reports on a sub-daily basis to avoid biases caused by the movement of the convection and EMLs throughout the day.

EML soundings were most frequent late at night through mid-morning in the spring and summer, with far fewer EML occurrences at 2100 Z and 0000 Z, often due to the erosion of portions of the EML by convection. The daily EML coverage over the domain was largest in the spring, although there was substantial spread and interannual variability in all seasons. Mean and
median daily values emphasized that while EMLs can cover a sizable portion of the CONUS, they are often much more localized phenomena. All additional EML variables except EML depth exhibited notable seasonal variability. The steepest mean EML lapse rates were located in the Great Plains and expanded northward from spring to fall. Mean EML potential temperature decreased with northward extent in all seasons except for summer, which had the least variable and highest mean values across the domain. Finally, the average MLCIN associated with EML soundings in all seasons indicated that a sizable proportion of EMLs were prohibitive (absolute MLCIN >200 J/kg) for convection, although there were many other EMLs with capping inversions that may have been more easily overcome through forcing.

The most notable changes in EML days were found in the regions where EMLs were climatologically maximized, with statistically significant increases in the southern Great Plains in spring and the western portions of the central and northern Great Plains in summer. These seasonal trends drove significant increases on the order of four to five more EML days per decade throughout the western Great Plains. More frequent EML occurrence is aligned with previous findings noting increased warming between 2–3 km AGL and significant increases in absolute CIN over the same period and region in the warm season (Taszarek et al. 2021a; Pilguj et al. 2022). Increases in EML days appeared to be driven by warming and drying trends in the western and southwestern CONUS over the study period, resulting in warmer surface temperatures and steeper PBL lapse rates which were subsequently advected downstream. Significant increases in EML base temperatures and EML lapse rates resulted, with the increased lapse rates supporting higher EML frequency. Warmer EML bases also significantly increased the MLCIN associated with EML soundings, which appeared to have implications for CI (Taszarek et al. 2021a). Since ERA5 is available back to 1940, future work could extend the
study period to determine if these trends are persistent or relatively recent developments.

Additional work could also examine the impact of drought on EML frequency to determine if warm seasons with, or proceeded by, exceptionally dry conditions in the western CONUS saw increased EML occurrence. Finally, because continued trends towards more EMLs and stronger inhibition would have substantial implications for future thunderstorm frequency and intensity (e.g., Rasmussen et al. 2020; Taszarek et al. 2021a; Haberlie et al. 2022; Ashley et al. 2023), similar methodology could be applied to climate simulations, such as those described in Gensini et al. (2022), to examine future projections of the EML and its capping inversion.
REFERENCES


APPENDIX A

EML VARIABILITY
Figure A1. Time series of cumulative annual EML days at a) Fort Stockton, TX, b) Lubbock, TX, c) Oklahoma City, OK, d) Garden City, KS, e) Colorado Springs, CO, f) Valentine, NE, g) Bismarck, ND, and h) Kansas City, MO.
Figure A2. Annual cumulative sum of EML days for 25 US cities. The thicker lines denote the 1979-2021 means.
APPENDIX B

ADDITIONAL EML ATTRIBUTE BOXPLOTS AND SPATIAL PLOTS
Figure B1. Boxplots of seasonal EML base height in m AGL at eight US cities. Boxes indicate the interquartile range, while dots and lines within the boxes denote the means and medians, respectively. Whiskers indicate 1.5 times the interquartile range and circles represent the outliers.
Figure B2. Same as Figure B1, except for EML base pressure in mb.
Figure B3. Same as Figure B1, except for EML potential temperature in Kelvin.
Figure B4. Same as Figure B1, except for EML lapse rate in °C · km⁻¹.
Figure B5. Same as Figure B1, except for EML MLCAPE in J \cdot kg^{-1}.
Figure B6. Same as Figure B1, except for EML MUCAPE in J·kg$^{-1}$. 
Figure B7. Same as Figure B1, except for EML MUCIN in J·kg$^{-1}$.
Figure B8. Same as Figure B1, except for EML lid strength in °C. Lid strength is calculated using the equation presented in Graziano and Carlson (1987), $\theta_{sw1} - \bar{\theta}_w$, where $\theta_{sw1}$ is the saturation wet-bulb potential temperature at the warmest part in the inversion, and $\bar{\theta}_w$ is the average wet-bulb potential temperature between 30 and 80 mb.
Figure B9. Same as Figure B1, except for EML 700 mb temperature in Kelvin.
Figure B10. Mean EML depth in mb a) annually, and by season for b) winter, c) spring, d) summer, and e) fall.
Figure B11. Mean EML MLCAPE in J kg⁻¹ a) annually, and by season for b) winter, c) spring, d) summer, and e) fall.
Figure B12. Mean EML MUCAPE in J kg$^{-1}$ a) annually, and by season for b) winter, c) spring, d) summer, and e) fall.
Figure B13. Mean EML MUCIN in J·kg⁻¹ a) annually, and by season for b) winter, c) spring, d) summer, and e) fall.
Figure B14. Mean EML 700 mb temperature in Kelvin a) annually, and by season for b) winter, c) spring, d) summer, and e) fall.
Figure B15. Mean EML lid strength in °C a) annually, and by season for b) winter, c) spring, d) summer, and e) fall.
Figure C1. Monthly Theil-Sen slope of the yearly grid-point sum of EML days for a) January, b) February, c) March, d) April, e) May, f) June, g) July, h) August, i) September, j) October, k) November, l) December. Hatching indicates statistical significance ($p$ value $\leq 0.05$) using Kendall’s $\tau$ statistic.
Figure C2. Theil-Sen slope of yearly mean EML depth a) annually, and by season for b) winter, c) spring, d) summer, and e) fall. Hatching indicates statistical significance \((p \text{ value } \leq 0.05)\) using Kendall’s \(\tau\) statistic.
Figure C3. Theil-Sen slope of yearly mean EML MUCAPE a) annually, and by season for b) winter, c) spring, d) summer, and e) fall. Hatching indicates statistical significance (p value ≤0.05) using Kendall’s τ statistic.
Figure C4. Theil-Sen slope of yearly mean EML MUCIN a) annually, and by season for b) winter, c) spring, d) summer, and e) fall. Hatching indicates statistical significance ($p$ value $\leq 0.05$) using Kendall’s $\tau$ statistic.
Figure C5. Theil-Sen slope of yearly mean EML 700 mb temperatures a) annually, and by season for b) winter, c) spring, d) summer, and e) fall. Hatching indicates statistical significance ($p$ value $\leq 0.05$) using Kendall’s $\tau$ statistic.
Figure C6. Theil-Sen slope of yearly mean EML lid strength a) annually, and by season for b) winter, c) spring, d) summer, and e) fall. Hatching indicates statistical significance (p value ≤0.05) using Kendall’s τ statistic.