A 22-Year Evaluation of Convection Reaching the Stratosphere Over the United States

Cameron R. Homeyer\(^1\) and Kenneth P. Bowman\(^2\)

\(^1\)School of Meteorology, University of Oklahoma, Norman, OK, USA, \(^2\)Department of Atmospheric Sciences, Texas A&M University, College Station, TX, USA

Abstract Stratosphere-reaching moist convection can significantly alter the dynamics, chemistry, and climate of the Earth system. This study seeks to add to the emerging understanding of the frequency, depth, and stratospheric impact of such events using 22 years (1996–2017) of ground-based radar observations in the contiguous United States. While most prior studies identify such storms using the temperature lapse-rate tropopause (LRT) as a tropopause-stratosphere boundary, this study is the first to identify convection that reaches into stratospheric air below the LRT (tropopause depressions, excluding folds) as well. It is found that tropopause depression (TD) overshooting and LRT overshooting occur at similar frequency over the United States, with TD overshooting being more episodic in nature than LRT overshooting. TD overshooting is also found more often throughout the cooler months of the year, while LRT overshooting dominates all overshooting in the summer months. Stratospheric residence of overshoot material, as estimated using trajectory calculations driven by large-scale winds, suggests that the vast majority of TD overshoot material does not remain in the stratosphere within 5 days downstream and rarely impacts altitudes more than 1 km above the LRT. Conversely, the majority of LRT overshoot material remains in the stratosphere downstream and routinely impacts altitudes >1 and >2 km above the tropopause.

1. Introduction

Moist convection that reaches the stratosphere provides a rapid transport mechanism that can alter the composition of the upper troposphere and lower stratosphere (UTLS). Cloud material, aerosols, and trace gases in convective plumes are irreversibly delivered to the UTLS through mixing with environmental air (e.g., Anderson et al., 2012; Chagnon & Gray, 2007; Dessler & Sherwood, 2004; Fischer et al., 2003; Frey et al., 2015; Fromm & Servanckx, 2003; Gray, 2003; Hanisco et al., 2007; Heglin et al., 2004; Herman et al., 2017; Homeyer et al., 2011; Homeyer, 2015; Homeyer, Pan, Dorsi, et al., 2014; Mullendore et al., 2005; Pan et al., 2014; Phoenix et al., 2020; Pouillard et al., 1996; Randel et al., 2012; Schwartz et al., 2013; Setvák et al., 2008; Smith et al., 2017; Tang et al., 2011; Tinney & Homeyer, 2021; Wang, 2003). Modification of the vertical distributions of water vapor and ozone is a significant concern due to their potential effects on radiative forcing, which could drive climate variability and change (e.g., Forster & Shine, 1999; Lacis et al., 1990; S. Solomon et al., 2010). Under environmental conditions with sufficiently low temperature and appreciable aerosol loading, activation of catalytic ozone destruction chemistry is also possible in the stratosphere from the convective delivery of copious amounts of water vapor (Anderson et al., 2012, 2017).

Stratosphere-reaching convection is more often referred to as tropopause-overshooting (or simply, overshooting) convection in the literature. In addition to UTLS composition impacts, overshooting convection has been associated with severe and/or hazardous weather at the Earth’s surface (e.g., K. M. Bedka, 2011; K. M. Bedka et al., 2015; K. Bedka et al., 2018; Dworak et al., 2012; Fujita, 1974; Homeyer et al., 2017; Line et al., 2016; Marion et al., 2019; Sandmæl et al., 2019). The hazards most commonly associated with overshooting convection are hail and tornadoes. When an overshooting storm irreversibly injects cloud material into the lower stratosphere (visually identified in satellite imagery and referred to as “above-anvil cirrus plumes”), it is often identified as one of the most important overshooting storms in relation to severe weather occurrence (K. Bedka et al., 2018; K. M. Bedka et al., 2015; Homeyer et al., 2017). Thus, storms that have the greatest impact on the stratosphere and the Earth’s surface are often one and the same. Given these potentially significant and wide-ranging impacts, understanding the frequency and magnitude of convection reaching the stratosphere is of high importance.
Here we use the unqualified term "tropopause" to mean the boundary between air with tropospheric and stratospheric characteristics. The environmental parameters commonly used to identify the tropopause are changes in static stability (temperature lapse rate $\Gamma$ or Brunt-Väisälä frequency $N$), changes in composition (e.g., water vapor, ozone, or other trace gases with sources predominantly in one of the two layers), and potential vorticity (PV). High-resolution composition measurements show that the tropopause is a transition layer of finite depth ranging from as thin as a few tens of meters to as much as several kilometers, depending on the synoptic and local scale environmental state and history (Hegglin et al., 2009; Tilmes et al., 2010). A number of studies have investigated the strengths and limitations of different diagnostic parameters for identifying the boundary between tropospheric and stratospheric air (e.g., Berthet et al., 2007; Bethan et al., 1996; Fischer et al., 2000; Gettelman et al., 2011; Homeyer et al., 2010; Hoor et al., 2002; Kunz et al., 2011; Pan et al., 2004; Zahn & Brenninkmeijer, 2003).

In this study we focus on the extratropical tropopause, which in many situations can be reliably identified by changes in the static stability (temperature lapse rate). Lapse-rate tropopause diagnostics can fail, however, in and around stratospheric intrusions, which are also referred to as tropopause folds. This is illustrated in the bottom panels of Figure 1. In a stratospheric intrusion stratospheric air can reach altitudes as low as the top of the atmospheric boundary layer, and the tropopause itself becomes folded, with tropospheric air layered between the stratospheric air in the fold and the overlying main stratosphere (e.g., Bourqui & Trépanier, 2010; Browell et al., 1987; Kuang et al., 2012; Skerlak et al., 2015; Sprenger et al., 2003; Sprenger & Wernli, 2003; Wernli & Bourqui, 2002). Adjacent to the folded part of the intrusion, the actual tropopause is frequently depressed relative to the LRT. Previous studies have generally neglected convection that overshoots into this part of the stratosphere through reliance on the lapse-rate diagnostic for the tropopause. Here we investigate the occurrence of overshooting convection in regions where the tropopause is depressed relative to the LRT.

Stratospheric intrusions and tropopause depressions can be identified in meteorological analyses and simulations by using PV isosurfaces and PV gradients or through observations of stratospheric tracers such as ozone (see Section 2.3 and Figure 4). These features are driven by ageostrophic circulations near upper tropospheric jet streams (Keyser & Shapiro, 1986), characterized by downward transport on the cyclonic side of the jet where it is accelerating from west to east. Since the circulations driving these depressions of stratospheric air below the LRT reverse when the jet decelerates from west to east, much of the air within the tropopause depression region has the potential to return adiabatically to above-LRT altitudes downstream within 1–2 days (in contrast to much of the air in the folded elements being irreversibly mixed into the troposphere). Thus, convection that reaches into tropopause depressions (hereafter “TD overshooting”, which excludes folds) represents an often overlooked population of storms that could have long-lasting impacts on UTLS composition. While it is possible for some storms to overshoot both tropopause depression PV boundaries and the LRT, such events will be shown to be rare. This study compares the relative importance of these two pathways (LRT vs. TD overshooting) for troposphere-to-stratosphere transport by convection.

Climatological analysis of LRT overshooting has been the focus of many prior studies using satellite imagery, satellite- and ground-based radar observations, and satellite-based lidar observations (K. Bedka et al., 2010; Cooney et al., 2018; C. Liu & Zipser, 2005; N. Liu & Liu, 2016; N. Liu et al., 2020; Pan & Munchak, 2011; D. L. Solomon et al., 2016). Some general results of these studies are: (a) LRT overshooting is most frequent over land, (b) overshooting occurs primarily during the warm season, and (c) the deepest and most frequent regions of LRT overshooting are found in the midlatitudes. The contiguous United States (CONUS) is one of

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**Figure 1.** Schematic illustrating (top) lapse-rate tropopause (LRT) overshooting in a situation where the LRT accurately identifies the tropopause, and (bottom) tropopause depression (TD) overshooting, where the LRT is at a higher altitude than the actual tropopause. In the vertical sections (left), white indicates tropospheric air; light gray indicates stratospheric air; medium gray indicates TD overshooting; and dark gray indicates LRT overshooting. Vertical profiles of temperature in the convective environments (right; at the locations marked by the $X$s at the base of the sections) are shown in blue, with LRT and TD altitudes superimposed.
the most active regions for LRT overshooting found anywhere around the globe (e.g., N. Liu & Liu, 2016; N. Liu et al., 2020). The CONUS has been the focus of several recent long-term studies of LRT overshooting using gridded observations from the operational ground-based radar network (the GridRad data set, which is used here). D. L. Solomon et al. (2016) analyzed LRT overshooting using GridRad at 3-h intervals for a single year (2004), revealing the geographic distribution and diurnal and seasonal variability of LRT overshooting over the CONUS and demonstrating that observed overshooting depth increases with decreasing lower stratosphere stability, as hypothesized in Homeyer, Pan, Dorsi, et al. (2014). Cooney et al. (2018) extended the GridRad LRT overshooting analysis to 10 warm seasons (March–August 2004–2013) while increasing the analysis frequency to hourly. They examined the frequency and depth of overshooting in great detail, revealing that approximately half of LRT overshooting events over the CONUS reach the so-called ‘stratospheric overworld’ (that is, above the 380-K potential temperature surface, which is entirely located in the stratosphere).

Studies of TD overshooting have been more limited. While many case studies exist (Cooper et al., 2005; Gray, 2003; Homeyer et al., 2011; Langford & Reid, 1998; Reid & Vaughan, 2004; Schroeder et al., 2014), there have been (to the authors’ knowledge) no prior climatological studies. Some climatological analyses have used the two potential vorticity unit (PVU; 1 PVU = 10^{-6} K m^2 kg^{-1} s^{-1}) isosurface as a troposphere-stratosphere boundary for overshooting diagnosis (e.g., N. Liu & Liu, 2016), which would include both TD and LRT overshooting events. However, it is impossible to assess the relative frequencies of each overshooting type in that case. Our limited knowledge of the frequency of TD overshadowing and its potential impact on UTLS composition motivates this study.

As previously alluded to, the stratosphere is commonly classified into two distinct layers for UTLS studies (as defined in Hoskins [1991] and Holton et al., [1995]): (a) the lowermost stratosphere (LMS), which extends vertically from the altitude of the extratropical tropopause to the 380-K potential temperature level, and (b) the overworld, which extends from the 380-K level to the stratopause. The use of 380 K as a marker of the transition from LMS to overworld stratosphere reflects its common coincidence with the tropopause altitude in the tropics, where it reaches its highest altitude globally (e.g., Highwood & Hoskins, 1998). Since the large-scale overturning circulation (the Brewer-Dobson circulation) in the stratosphere is downward in the extratropics, and quasi-isentropic transport between the tropical upper troposphere and LMS is possible, the residence time of air in the stratosphere decreases with decreasing altitude (e.g., Stohl et al., 2003). As such, overshooting that reaches higher altitudes in the stratosphere will typically have longer-lasting impacts on UTLS composition and climate, though recent work has shown that climate sensitivity to UTLS composition change (especially that for water vapor) is greatest in the LMS (Banerjee et al., 2019; Riese et al., 2012).

In this study, we present the longest observational analysis of stratosphere-reaching convection over the United States to date by using 22 years (1996–2017) of GridRad data. LRT and TD overshadowing is analyzed for all months of the year, in contrast to the prior GridRad multi-year analysis, which focused on the warm season (March–August). The climatology of TD overshadowing is the first of its kind, revealing occurrences of stratosphere-reaching convection that have been overlooked in prior work. For both overshooting types, the short-term stratospheric residence time of overshoot material is estimated using trajectories driven by three-dimensional wind fields from a modern reanalysis model. Together, the overshooting climatologies and transport calculations improve our understanding of the frequency of stratosphere-reaching convection and its expected impact on UTLS composition.

2. Data and Methods

2.1. Radar Observations

This study uses hourly, instantaneous, Version 3.1, gridded NEXRAD WSR-88D Radar (GridRad) analyses to identify overshooting convection (Bowman & Homeyer, 2017). GridRad is a large-area, three-dimensional merger of single-radar observations from operational systems throughout the majority of the CONUS (115°-69°W, 25°-49°N). The operational systems contributing to GridRad volumes are S-band (10-cm wavelength), Doppler, polarimetric (2013-Present) radars from the National Weather Service’s Next Generation Weather Radar (NEXRAD) Weather Surveillance Radar - 1988 Doppler (WSR-88D) network (Crum & Alberty, 1993). When measuring precipitation, individual NEXRAD radars obtain volumes on a
spherical grid (range, azimuth, and elevation relative to the radar location) every 4–7 min that include, at a minimum, radar reflectivity at horizontal polarization ($Z_H$), radial velocity ($V_R$), and velocity spectrum width ($\sigma_V$). The V3.1 GridRad data set is built using a binning algorithm that weights contributing NEXRAD WSR-88D observations based on their distance from radar location (range) and time relative to the target analysis time, resulting in volumes on a 0.02° longitude-latitude grid with vertical grid spacing of 1 km from 1–24 km above sea level (Homeyer & Bowman, 2017). Observations out to 300 km in range from each radar are used for GridRad creation. While GridRad analyses can be made at higher frequencies, the long-term archive used here exists at hourly intervals for the years 1995–2017 and includes $Z_H$ only. All data used here (1996–2017) are quality-controlled prior to analysis following the recommendations outlined in Homeyer and Bowman (2017). Analyses from 1995 are excluded here due to relatively poor multi-radar sampling.

The NEXRAD WSR-88D network was built mostly during the years 1995 and 1996, with some slight expansion and changes to network thereafter. At present, the NEXRAD network observes nearly all of the contiguous U.S. land mass within 300 km range from each radar (Figure 2, left). The areal extent and density of the network, which are generally unmatched by alternative networks outside of the United States, allow most regions to be observed by multiple radars, especially outside areas of complex terrain. Such overlapping radar coverage often reaches the threshold of three or more radars (Figure 2, right) that commonly enables vertical sampling in the UTLS three times finer than that possible from a single radar (Homeyer, 2014). The method for archiving NEXRAD WSR-88D data originally involved writing data to physical storage devices that were shipped to a central location. Storage of these data in a unified long-term archive was often delayed considerably from the time of observation and sometimes failed, with the process eventually replaced by real-time data transfer and archival via the internet in the early 2000s. Analysis of the area observed by $\geq 1$ and $\geq 3$ radars over time, as assessed via the v3.1 GridRad data, reveals this evolution well and is a useful measure of the overall data coverage and quality (Figure 3). Substantial day-to-day variability in data coverage is observed prior to 2004, with generally reduced data availability and coverage in years 1996–2005 compared to that in more recent years. Data dropouts (where most data were lost for a day) can also be found in these time series, but fortunately, they are rare.

2.2. Meteorological Analyses

To enable overshoot identification, tropopause diagnostics for the analysis region and time period are necessary. In addition, three-dimensional wind fields are needed to carry out transport calculations. For this purpose we use 6-hourly analyses from the ERA-Interim reanalysis produced by the European Center for Medium-Range Weather Forecasts (ERA-Interim; Dee et al., 2011). The ERA-Interim reanalysis typically has the smallest error in tropopause altitude among modern reanalyses (e.g., Xian & Homeyer, 2019), and calculations with alternative reanalyses for short periods were found to be nearly identical (not shown). ERA-Interim is available on a Gaussian grid with ~0.75° longitude-latitude spacing and 60 vertical levels,
providing temperature, humidity, and winds at a vertical resolution ranging from 750 to 1250 m in the UTLS. For our analysis we interpolate all ERA-Interim output to a regular 1° longitude-latitude grid.

Note that while ECMWF has recently released a new reanalysis (ERA5) with finer spatial and temporal resolution, such increased resolution is not necessary for adequate statistical assessment of overshooting since the uncertainty in tropopause altitude from ERA-Interim is lower than the uncertainty in the storm top altitudes from GridRad and tropopause altitudes typically do not change much over time intervals less than 12–24 h or distances less than 1000 km. One exception to this is near the subtropical jet, where there is a sharp discontinuity in tropopause altitude known as the tropopause break (e.g., Birner, 2010; Castanheira & Gimeno, 2011; Randel et al., 2007). Increased temporal resolution can improve the representation of some tropopause depressions and folds, as well as trajectory calculations used for transport analysis (e.g., Bowman et al., 2013), but trajectory statistics in the UTLS using 6-hourly ERA-Interim and hourly ERA5 reanalyses show little difference (Hoffmann et al., 2019).

2.3. Tropopause Identification

As outlined in the Introduction, while LRT overshooting and TD overshooting both represent convection that reaches into stratosphere air, tropopause depressions (and stratospheric intrusions) are large-scale excursions of air to altitudes well below the LRT. It is common practice to identify the boundaries of tropopause depressions using a specific PV isosurface, which is often referred to as the “dynamical” tropopause. Here, for clarity, we refer to PV-based boundaries between tropospheric and stratospheric air as tropopause depression altitudes.

The primary tropopause altitudes in this study are thus LRT altitudes, as identified by applying the World Meteorological Organization (WMO) definition (World Meteorological Organization, 1957). In regions where the LRT is relatively flat, it typically coincides very closely with sharp vertical gradients of static stability and composition (e.g., Gettelman et al., 2011; Pan et al., 2018). Near jet streams, however, the LRT can exhibit abrupt changes in altitude over short horizontal distances. The limited vertical resolution of reanalysis data in the UTLS and the nonlinear nature of the WMO tropopause definition can lead to localized spurious jumps in the LRT altitude. To reduce these artifacts, the LRT altitude on the ERA-Interim grid is smoothed using an 8° × 8° longitude-latitude median filter (e.g., Homeyer et al., 2010). While this may be considered to be a relatively - or unnecessarily - large spatial filter, the choice of a neighborhood median approach is less aggressive than common alternatives and limits the introduction of altitude biases (which may be as large as +1 km in some instances). Smoothing with a median filter also improves performance of the PV-based tropopause depression diagnosis. Following LRT identification and quality control, depressions are identified using a PV gradient approach to avoid biases from seasonality in

![Figure 3](image_url). History of Next Generation Weather Radar (NEXRAD) WSR-88D areal data coverage in the hourly GridRad archive. The thin black and blue lines show the total daily area covered by at least one and three radars, respectively, with the thicker horizontal black and blue lines indicating the maximum possible areal coverage (i.e., the shaded areas in Figure 2).
the PV isosurface that best coincides with the troposphere-stratosphere transition (e.g., see Homeyer & Bowman, 2013; Kunz et al., 2011, 2015). In particular, the scaled magnitude of the three-dimensional PV gradient is calculated as follows:

$$
PV_{grad} = \sqrt{150 \left( \frac{\partial (PV)}{\partial x} \right)^2 + 150 \left( \frac{\partial (PV)}{\partial y} \right)^2 + \left( \frac{\partial (PV)}{\partial z} \right)^2}
$$

(1)

where the factor of 150 for horizontal gradient terms is chosen to account for the difference in horizontal and vertical scales in the atmosphere (approximated as the ratio of the Brunt-Väisälä frequency $N$ to the Coriolis parameter $f$, which typically ranges from 100 to 200 in the UTLS). Factors ranging from...
100 to 200 were tested; little sensitivity in the diagnosed tropopause depression boundary was found (not shown). The tropopause depression boundary is identified as the minimum altitude at which $PV_{\text{grad}}$ exceeds 1.5 PVU km$^{-1}$ below the LRT, which enables identification of the primary, near-tropopause gradient rather than detached PV features in the troposphere that are common when using PV isolines. Tropopause depression boundaries are retained for analysis if they are identified at least 2.5 km below the LRT where the LRT is 14 km or lower, which we have found to reliably allow depression identification near upper tropospheric jets. As a result, TD overshooting can only be identified where a significant vertical separation occurs between the PV gradient boundary and the LRT. In other locations only LRT overshooting identification is possible, though the PV gradient is often nearly coincident with the LRT there.

Figure 4 shows example tropopause depression diagnosis for a 3-days sequence in April 2011. Here, a longwave trough, jet streak, and attendant tropopause depression translate from west to east through the analysis domain. To demonstrate how the depression diagnosis corresponds to the below-LRT PV feature and related diagnostics in the UTLS, vertical sections from north-to-south in the center of the domain are shown. The identified tropopause depression boundaries clearly indicate regions where high PV extends well below the LRT altitude and lie approximately near the 2-PVU isosurface at these times, which is a common PV threshold used to represent the dynamic tropopause and stratospheric intrusion edges in prior work. ERA-Interim LRT altitudes with and without the aforementioned 8° × 8° longitude-latitude median filter applied are also shown in the vertical sections and are generally found within ±1 km of each other in regions without large latitudinal changes in altitude.

### 2.4. Overshoot Identification

Overshooting is identified in this study by using a combination of GridRad $Z_H$ echo-top altitudes and the ERA-Interim LRT or tropopause depression boundaries identified as outlined in Section 2.3. A threshold of $Z_H = 10$ dBZ is used to define the echo-top altitude, with the additional requirement that $Z_H \geq 20$ dBZ at the altitude of the LRT or depression boundary. Overshoots are identified at each GridRad pixel where the echo-top altitude exceeds the collocated LRT or tropopause depression altitude by at least 1 km, so long as $Z_H \geq 20$ dBZ at the altitude of the LRT or depression boundary. Requiring a minimum overshoot distance of 1 km is motivated by the uncertainties of the GridRad echo-top altitudes, which are ±1 km and ±0.5 km, respectively (Cooney et al., 2018; D. L. Solomon et al., 2016). Following overshooting identification, all points are examined and further quality-controlled to remove likely radar artifacts following the procedure described in Cooney et al. (2018). Briefly, bogus overshooting echoes have large overshoot depths and weak column-maximum $Z_H$. These constitute a small fraction of initial overshoot identifications (∼4% and ∼2% for LRT and TD events, respectively) and are easily removed prior to final analysis.

### 2.5. Trajectory Calculations

Trajectories of air from overshoot locations are computed kinematically with the TRAJ3D trajectory model (Bowman, 1993; Bowman & Carrie, 2002; Bowman et al., 2013) driven by three-dimensional ERA-Interim winds ($u$, $v$, $\omega$). Particles are initialized for LRT and TD overshoots at the grid spacing of the GridRad data (i.e., every 0.02° longitude-latitude) and at 1-km altitude increments from either the LRT or the tropopause depression boundary up to the top of the radar echo. Trajectories are integrated forward in time for 5 days, with ERA-Interim LRT altitudes and PV linearly interpolated to the particle locations. Results from two-dimensional isentropic trajectories (not shown) differ little from the three-dimensional trajectories used here, indicating that diabatic effects are small.

### 3. Results

Here, we summarize results of LRT and TD overshoot identification and transport diagnosis for the entire 22-year record (1996–2017). Since the GridRad archive used for this analysis is comprised of instantaneous, hourly observations of storm activity (i.e., snapshots), the results are an underestimate of true overshooting activity. Namely, the lifetime of an individual overshoot (from ascent to collapse) is on
the order of minutes, with common lifetimes of \( \sim 5 \) min and few lasting more than 10 min (e.g., Dauhut et al., 2018; Fujita, 1974). Thus, true overshooting activity will be 4–12 times higher than that diagnosed here, but remaining climatological features such as the spatial patterns, relative frequency, and temporal evolution of overshooting are expected to be well characterized (especially in years 2006–Present, when radar coverage is consistently near the maximum possible based on the current NEXRAD WSR-88D radar network). Red and blue color scales are used for all analyses to represent LRT and TD overshooting, respectively.

### 3.1. Geographic Distributions and Seasonality

To first provide an understanding of both where and how often overshooting events occur, we present geographic distributions of annual hourly LRT and TD overshooting in Figure 5. These maps reveal key similarities and differences between the two types of overshooting analyzed. Both overshooting types have similar magnitudes of peak overshooting frequency, but the area of the LRT overshooting maximum is larger than that for TD overshooting. Both types of overshooting are most frequent in the Great Plains, consistent with prior analyses of overshooting and the most common location of severe weather in the CONUS (e.g., Brooks et al., 2003; Doswell et al., 2005). While LRT overshooting is mostly confined to the Great Plains and upper midwest U.S., there is a weak region of enhanced occurrence in the southeast U.S., except along the main axis of the Appalachian Mountains. In comparison, TD overshooting maximizes in the central Great Plains (mostly over Kansas), and moderate frequencies of overshooting extend eastward to the Atlantic coast north of 30°N. This confinement of TD overshooting to more northern areas of the CONUS reflects the constraint of tropopause depressions being north of the subtropical jet, while folded elements of intrusion events are commonly found slightly equatorward (e.g., Manney et al., 2014; Škerlak et al., 2015; Sprenger et al., 2003).

Figure 6 shows the geographic distributions of LRT and TD overshooting as a function of season. LRT overshooting is rare during fall and winter, as first described in D. L. Solomon et al. (2016). The vast majority of overshooting occurs during spring and summer, shifting northward as the warm season progresses, consistent with the seasonal migration of the subtropical jet and lower tropopause environments (cf. Figure 10 in Cooney et al., 2018). LRT overshooting in the southeast U.S. maximizes in the summer months, where short-lived deep convection shallowly overshoots the high tropical tropopause altitudes commonly found at that time. By comparison, TD overshooting is modest during fall and winter and occurs most often in the central Great Plains. TD overshooting activity peaks in spring, then weakens and shifts northward during the summer months. For context, seasonal maps of TD frequency over the study domain are provided in the appendix (Figure A1).
The seasonality of overshooting is revealed further by the climatological annual cycle of daily overshooting volume for the 22-year analysis period (Figure 7). Annual cycles are shown in Figure 7c as the fraction of the total annual volume of LRT overshooting to normalize the results and allow for comparison of the relative frequency of each overshooting type. Consistent with the seasonal maps, the annual maximum overshooting occurs during spring and summer. The increased detail here shows that the peaks of LRT and TD overshooting have similar magnitudes, and both occur in late spring to early summer. The largest differences in the annual cycles of LRT and TD overshooting occur after this peak, with LRT overshooting higher than TD overshooting during the summer, then falling nearly to zero during the fall and winter. TD overshooting over the CONUS decreases rapidly in the early summer, then slowly from mid-July to the end of January. There is an indication of a minor secondary peak in TD overshooting in October, signaling the southward march of the subtropical jet during the
fall and the return of tropopause depressions (and related tropopause folds) to the CONUS. In addition to these broad differences, TD overshooting exhibits considerably larger day-to-day variability, reflecting the more episodic nature of these events (as also found in annual cumulative distributions - not shown).

The final broad climatological analysis presented here is the diurnal cycle of overshooting (Figure 8). Results are presented in UTC time since the time of peak diurnal heating and surface temperature east of the Rocky mountains that drives much of the deep convection analyzed here spans an approximately 3-h period. Examining the diurnal cycle in UTC time is also consistent with prior overshooting analyses over the CONUS, which allows for direct comparison with such studies. The diurnal cycle of LRT overshooting is nearly identical to that in prior work, indicating a large peak near 00 UTC (often the time of maximum surface temperature and convective potential) and a broad minimum from local midnight to the midday hours (06–18 UTC). The diurnal cycle of TD overshooting indicates a slightly earlier increase in activity (offset ∼2 h prior to that for LRT overshooting), with a dominant peak near 00 UTC and a minor secondary peak near 12 UTC. The shift to earlier times likely reflects the easier path to overshooting facilitated by the lower altitudes of tropopause depression boundaries (compared to LRT altitudes). Moreover, because tropopause depressions (and stratospheric intrusions) are a synoptic-scale phenomenon, they are not expected to have substantial diurnal cycles, so it is not surprising that both quantities have similar diurnal cycles consistent with the strong diurnal cycle in convection that peaks in late afternoon and early evening.

3.2. Relationships Between LRT and TD Overshooting

To explore the relationships between LRT and TD overshooting, Figure 9 contains scatterplots of annual totals of several quantities. Annual overshooting statistics exhibit a noticeable shift between the earlier part of the record (2005 and earlier) the later period (2006 and later). Though connections to large-scale climate variability are a possible explanation for this shift, it is most likely due to changes in the NEXRAD observing system. The most significant observing system changes are the increases in the number of active radars and the reliability of the network coverage, which can be seen in Figure 3. With the exception of a few added radars around 2013, the network was largely complete by 2006. Therefore, the scatterplots in Figure 9 use data only for the 12-year period from 2006 to 2017. The overshooting diagnostics are normalized by the annual-mean radar coverage for each year to adjust for the small changes that occur during this period.

Figure 9a shows TD overshooting area as a function of tropopause depression area. There is a strong positive correlation between the two quantities (that is significantly different from zero for a 95% confidence limit), which is consistent with the idea that greater tropopause depression area provides more opportunities for TD overshooting. There is also a relatively strong relationship between TD overshooting area and LRT overshooting area (though not statistically significant), as seen in Figure 9b. Again, this is not surprising, as greater intense convection, as measured by LRT overshooting, should also lead to
greater TD overshooting. Figure 9c shows that LRT overshooting is also positively correlated with tropopause depression area (though also not statistically significant). Because tropopause depressions and stratospheric intrusions are associated with strong jets, this suggests that an overall increase in baroclinity (or frequency of baroclinic disturbances) contributes to the occurrence of LRT overshooting, which likely also contributes to the occurrence of TD overshooting seen in Figure 9a.

### 3.3. Large-Scale Transport and Stratospheric Residence

One element of overshooting convection that has been examined minimally in prior work is the residence time of overshoot material in the stratosphere following an event. Vertical transport in the midlatitude lower stratosphere is characterized by slow downwelling, resulting in timescales of weeks to several months for air in the LMS to enter the troposphere (based on typical reanalysis descent rates). Stratosphere-to-troposphere transit times generally increase with increasing altitude in the lower stratosphere. Thus, the impact of convection on UTLS composition is expected to increase with increasing overshooting depth. The importance of TD overshooting relative to LRT overshooting with respect to stratosphere composition depends on how much of the tropopause depression air remains in the stratosphere. Here we measure that by evaluating how much air from both TD and LRT overshooting is above the LRT 5 days after the overshooting event. That interval is several times longer than the time required for air to flow through a tropopause depression region. Note that the trajectory calculations analyzed here do not account for the modifications to the environment resulting from convection, which may enhance or reduce the diagnosed amounts of cross-tropopause transport and subsequent long-range transport pathways. However, they provide a statistical assessment of the probability (or amount) of overshoot material retained in the lower stratosphere based on large-scale motion alone, which will serve as a measure of the relative impact of LRT and TD overshooting on stratosphere composition.

Figure 10 presents joint frequency distributions of TD overshoot PV at trajectory initialization and LRT-relative altitude both at initialization and 5 days downstream. These frequency distributions reveal how much tropopause depression material returns to the LRT-diagnosed stratosphere downstream and how it relates to the “depth” of overshooting (PV magnitude). The distributions also summarize how much of the original TD overshoot volume would also be classified as LRT overshooting, which is small (3.7%), indicating that TD overshooting events are indeed largely distinct from LRT overshooting events. At 5 days downstream, the amount of TD overshoot material that is above the LRT is 17.5%, an increase of ~14% following the event and appearing to be generally insensitive to how deep the overshooting reached into the tropopause depression. Despite the return of a substantial amount of TD overshoot air to above-LRT altitudes, the frequency distributions demonstrate that little TD overshoot air reaches more than 1 km above the LRT. Conversely, these distributions demonstrate that the vast majority of intrusion overshoot air (i.e., 82.5%) is located below the LRT several days downstream, so most air injected into tropopause depres-

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**Figure 9.** (a) Normalized tropopause depression (TD) overshooting area as a function of tropopause depression area, (b) normalized TD overshooting area as a function of normalized lapse-rate tropopause (LRT) overshooting area, and (c) normalized LRT overshooting area as a function of tropopause depression area. Overshooting areas are normalized by the annual-mean daily area of GridRad coverage by three or more radars (see Figure 2). Correlation coefficients \( r \) are given for each graph.
sions by convection returns to the troposphere. As a complement to this analysis, the PV change experienced by TD overshoot particles 5 days downstream is given in the appendix and similarly suggests limited deep stratospheric impact (i.e., slightly skewed to modest PV increase for above-LRT particles; Figure A2).

Figure 11 shows the amount of overshoot material above the LRT for both LRT and TD overshooting in the 5 days following trajectory initialization. The quick return of ~14% of the TD overshooting air above the LRT is evident here 1 day after the event. The minimal amount of TD-overshoot material that makes it more than 1 km above the LRT is also evident, with only ~2% above the LRT after 5 days. Long-range transport of LRT overshoot material is characterized by nearly the opposite behavior of that for TD overshooting. Namely, the amount of LRT-overshoot material that remains in the lower stratosphere decreases rapidly in the first 1–2 days following initialization to ~60% of that diagnosed at the time of the event, generally plateauing beyond day 2 (the fraction at day 5 is ~61.5%). The amount of all LRT overshoot air that remains at altitudes more than 1 km above the LRT is ~30% 5 days downstream, down ~24% from that at initialization. At more than 2 km above the LRT, the amount of all overshoot material decreases from ~11% initially to ~5% at days 1 and 2 downstream, increasing gradually to ~8.5% by 5 days downstream. It is not clear what drives this large-scale down-up dynamic behavior, but it is clear that the impact of LRT overshooting extends deeper into the stratosphere and remains longer than TD overshooting.

4. Summary and Discussion

This study examines 22 years (1996–2017) of ground-based radar observations of moist convection reaching the stratosphere over the United States. In addition to diagnosing LRT overshooting (as is common in prior work), a climatology of convective overshooting into tropopause depressions (stratospheric air
below the LRT) is described for the first time. It is demonstrated that TD overshooting and LRT overshooting are largely distinct events and occur with comparable frequency over the CONUS. Both overshooting types are most common in the Great Plains, but TD overshooting occurs over a broader range of longitudes, primarily poleward of 30°N (Figures 5 and 6). The seasonality of both overshooting types is similar, with a peak in late spring to early summer. While LRT overshooting is exceedingly rare in the cool months (October–February), modest amounts of TD overshooting are found during those months (Figures 6 and 7). Annual cycles of overshooting reveal well the more episodic nature of TD overshooting (Figure 7), while annual scatterplots indicated that TD overshooting is sensitive to interannual variability in both convection occurrence and baroclinity within the analysis domain (Figure 9). The diurnal cycle of LRT overshooting is 1–2 h later than TD overshooting. Finally, despite the similarity of the total annual amount of LRT and TD overshooting, analysis of the large-scale motion of overshoot material finds that ∼61.5% of LRT overshoot air is above the LRT after 5 days, compared to only ∼17.5% of TD overshoot air. TD overshoot material rarely reaches altitudes ≥1 km above the LRT downstream, while nearly one-third of LRT overshoot material reaches such altitudes. Thus, combining newfound understanding of the frequency and stratospheric staying power of overshoot material, these results suggest that (for the CONUS) LRT overshooting is roughly 3.5 times more impactful to lower stratosphere composition than TD overshooting, and that ∼22% of all stratosphere-impacting convection has been overlooked in most prior work.

A number of factors that have not been considered here could affect the results. First, while ground-based radar observations provide the best long-term observational record to assess overshooting (based on coverage, resolution, and three-dimensionality), the radars do not detect cloud particles and provide somewhat limited coverage in the northern Great Plains. As a result, the analysis conducted here probably under-represents the frequency and depth of overshooting, although to what extent is unknown. Alternative approaches with satellite imagery, lidar, or observations from other platforms that detect cloud particles may improve the results, but such techniques have their own limitations. Combining precipitation radar observations and cloud observations may be the most successful approach. Second, using environmental fields with higher spatial and temporal resolution may also be helpful for improving overshooting estimates, but any such improvement is expected to be minor for the radar observations analyzed here. Third, while analysis of year-to-year variability in overshooting was restricted to the time period of complete coverage by the observational network (2006 and later), exploring the role of large-scale climate variability in explaining the shift in character observed near 2005 is warranted and requires analysis of additional auxiliary data.

Another limitation of this work is the use of trajectories based on large-scale analyzed winds. Overshooting convection can impact the dynamics of the UTLS in myriad ways, from introducing anticyclonic flow anomalies on medium to large scales (e.g., Maddox et al., 1981; Ninomiya, 1971; Stensrud, 1996; Stensrud & Anderson, 2001; Zhang & Harvey, 1995) to generating gravity waves that can break and irreversibly mix overshoot material into the stratosphere (e.g., Homeyer et al., 2017; Luderer et al., 2007; Wang, 2003; Wang et al., 2016). The large-scale wind fields used here do not include most of these dynamical perturbations from convection (especially at small scales) and are thus incapable of capturing the entire envelope of transport outcomes and, ultimately, stratosphere-troposphere exchange. Numerical models that resolve convection are a promising pathway to developing better estimates of transport and improved understanding of transport mechanisms. Comparing the dynamics of LRT and TD overshooting, especially as it pertains to the expected range of UTLS composition impacts, is one additional motivating factor, since overshooting alone does not necessarily result in irreversible mixing and significant composition changes. Such comparisons have been rare in prior work (e.g., Homeyer, Pan, & Barth, 2014), but numerous numerical modeling case studies of either LRT or TD overshooting have been carried out. Nevertheless, there is considerable room for additional study including adding

**Figure 11.** Based on trajectory calculation, time series (in days since trajectory initialization) of the fraction of total overshoot volume located above the lapse-rate tropopause (LRT) for LRT overshooting (red lines) and tropopause depression overshooting (blue lines). Thick solid lines show the fraction of total overshoot volume at and above the LRT, short-dashed lines the fraction of total overshoot volume at least 1 km above the LRT, and long-dashed lines the fraction of total overshoot volume at least 2 km above the LRT.
complexity to the large-scale trajectory analyses used here, such as diagnosing the Lagrangian cold point of overshoot material to infer potential water loading of the stratosphere.

Finally, while this work leverages an extensive database to evaluate overshooting over the United States, further examination of LRT and TD overshooting globally is needed. This is especially true for TD overshooting, since this study is the first such dedicated climatology and the frequency of events and their long-range transport behavior outside of the United States is unknown. Use of the growing satellite-based precipitation measuring mission datasets is perhaps the most promising avenue for global climatological work (e.g., see recent overshooting diagnoses in N. Liu and Liu (2016) and N. Liu et al. (2020)).

**Appendix A: Supplemental Figures**

Two figures are included here for context in support of the analysis presented throughout the paper. First, seasonal maps of TD frequency over the study domain based on ERA-Interim reanalysis from 1996 to 2017 are given (Figure A1). Second, the change in PV value between initialization and 5 days downstream for TD overshoot particle trajectories is given to provide a dynamical estimate of stratospheric depth achieved for air returning to above-LRT altitudes (Figure A2). Each figure follows the same design as comparable analyses presented in the paper, which are referenced in the figure captions.

![Figure A1](image)  
*Figure A1.* As in Figure 6, but for ERA-Interim tropopause depression frequency during the 1996–2017 study period.
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References

Data Availability Statement
All data used in this study were obtained from and are publicly available at the National Center for Atmospheric Research (NCAR) Research Data Archive (RDA): GridRad (Bowman & Homeyer, 2017) and ERA-Interim (ECMWF, 2009).

Figure A2. As in Figure 10, but for the PV change between particle initialization and 5 days downstream, as a function of LRT-relative altitude. The bin size is 0.25 pvu by 0.5 km and frequency is computed relative to the total occurrence at each LRT-relative altitude.


