INVESTIGATING STABILITY EVOLUTION OF SNOW STORMS FEATURING LIGHTNING

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Public Abstract
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Wintertime forecasting can be a difficult venture for the operational forecaster. Problems arise in predicting how strong or significant a winter storm will be. Lightning is a good indicator of a strong storm capable of producing heavy banded snowfall. These convective bands can produce up to and greater than six inches of snowfall in a short period of time. Therefore, understanding the nature of these lightning-producing snowstorms will be beneficial to the operational forecasting community.

This study is an attempt to determine discriminators between thundersnow (TSSN) events and non-thundering snow events. Traditional forecasting tools, indices for forecasting the stability of the atmosphere, will be tested to determine if any thresholds exist in the atmospheric stability between TSSN and non-thundering events. Additionally, this study will examine the implications of the height at which the -10°C isotherm, the temperature of the first atmospheric charge reversal occurs, is located with respect to the most unstable lifted parcel (MU LPL). If the level at which the -10°C isotherms is located is within a statically stable layer of the atmosphere, lightning may not occur.

This study shows that certain stability indices are better discriminators for TSSN than others. It will also be shown that the stability tendency of TSSN events is quite different than that found in non-thundering snowstorms. In order to illustrate the results eight TSSN events are examined from several hours prior to lightning onset until several hours after the cessation of lightning in terms of stability and the results are compared to a set of seven non-thundering snow events in which a similar analysis is performed. Paradoxically, TSSN environments area prone to stabilization while non-thundering environments are typically destabilizing.

Ultimately the main goal of this study is to benefit the community through a better understanding of this wintertime phenomena. By further understanding what triggers thundersnow events, forecasters will be better prepared to warn communities of a potentially crippling snow event. The forecasters will be able to access and disseminate this information to emergency management and community leaders who will then be able to take proper actions to prepare for these types of snow events.
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# TABLE OF CONTENTS

Acknowledgements........................................................................ii
List of Figures................................................................................v
List of Tables..................................................................................xi

Chapter 1.......................................................................................1
Introduction......................................................................................1
  1.1 Purpose & Objectives ...................................................................2
    1.1.1 Purpose..............................................................................2
    1.1.2 Objectives..........................................................................2
  1.2 Statement of Thesis....................................................................3

Chapter 2.......................................................................................4
Literature Review.............................................................................4
  2.1 Climatology...............................................................................4
  2.2 Synoptic Scale..........................................................................5
  2.3 Banded Snow..........................................................................5
  2.4 CSI.........................................................................................6
  2.5 Lightning.................................................................................12

Chapter 3.....................................................................................16
Methodology...................................................................................16
  3.1 Event Selection.......................................................................16
  3.2 Data.......................................................................................17
  3.3 Indices Examined.....................................................................18
    3.3.1 Traditional Indices..............................................................18
    3.3.2 Static Stability .................................................................19
<table>
<thead>
<tr>
<th>Section</th>
<th>Page</th>
</tr>
</thead>
<tbody>
<tr>
<td>3.3.3 CSI</td>
<td>21</td>
</tr>
<tr>
<td>3.4 Methodology Summary</td>
<td>22</td>
</tr>
<tr>
<td>Chapter 4</td>
<td>23</td>
</tr>
<tr>
<td>Analysis</td>
<td>23</td>
</tr>
<tr>
<td>4.1 Stability Tendency</td>
<td>23</td>
</tr>
<tr>
<td>4.1.1 DADTP</td>
<td>23</td>
</tr>
<tr>
<td>4.1.2 VADVS</td>
<td>31</td>
</tr>
<tr>
<td>4.1.3 STRCH</td>
<td>37</td>
</tr>
<tr>
<td>4.1.4 Total Tendency</td>
<td>45</td>
</tr>
<tr>
<td>4.2 Total Totals</td>
<td>55</td>
</tr>
<tr>
<td>4.3 K-index</td>
<td>63</td>
</tr>
<tr>
<td>4.4 Mid-level Lapse Rate</td>
<td>72</td>
</tr>
<tr>
<td>4.5 CAPE</td>
<td>82</td>
</tr>
<tr>
<td>4.6 MU Lifted Index</td>
<td>83</td>
</tr>
<tr>
<td>4.7 The -10°C Isotherm</td>
<td>89</td>
</tr>
<tr>
<td>4.8 Symetric instability</td>
<td>96</td>
</tr>
<tr>
<td>Chapter 5</td>
<td>104</td>
</tr>
<tr>
<td>Summary</td>
<td>104</td>
</tr>
<tr>
<td>5.1 Discussion</td>
<td>104</td>
</tr>
<tr>
<td>5.2 Conclusions</td>
<td>116</td>
</tr>
<tr>
<td>Apendix A</td>
<td>119</td>
</tr>
<tr>
<td>References</td>
<td>120</td>
</tr>
</tbody>
</table>
**List of Figures**

<table>
<thead>
<tr>
<th>Figure</th>
<th>Page</th>
</tr>
</thead>
<tbody>
<tr>
<td>Figure 2.4.1 Schematic vertical cross section illustrating symmetric instability; x- and z-axes define horizontal and vertical distances, respectively. Solid contours represent absolute geostrophic momentum ($M_g$) of basic flow. Dashed contours represent equivalent potential temperature ($\theta_e$). Lettered points show sample displacements (dashed) and accelerations (double-shaft arrowheads; from Weismuller and Zubrick 1998)</td>
<td>8</td>
</tr>
<tr>
<td>Figure 2.4.2 A modified version of the Wiesmueller and Zubrick (1998) schematic vertical cross section of CSI by McCann (1999). Solid contours represent absolute geostrophic momentum ($M_g$). Dashed contours represent saturated equivalent potential temperature ($\theta_{es}$). Letters represent parcels of various displacements. Double-shaft arrows represent accelerations</td>
<td>9</td>
</tr>
<tr>
<td>Figure 2.5.1 Theoretical critical charge reversal layers (Saunders 1992). Y-axis is temperature in degrees Celsius</td>
<td>13</td>
</tr>
<tr>
<td>Figure 2.5.2 The proposed tri-pole cloud structure is similar to the charges acquired by graupel at different</td>
<td>14</td>
</tr>
<tr>
<td>Figure 3.3.1 Example sounding valid 0700 UTC 10 January 2003 for Joplin Missouri depicting the depth and location of the layer examined during analysis of the stability tendency and its individual terms</td>
<td>20</td>
</tr>
<tr>
<td>Figure 4.1.1 Average DADTP ($10^6$ K mb$^{-1}$ s$^{-1}$) occurring in thundersnow events. Asterisks represent the initial hour and final hour of reported lightning activity</td>
<td>23</td>
</tr>
<tr>
<td>Figure 4.1.2 RUC_20 initial hour valid 0800 UTC 10 December 2003. Typical DADTP pattern observed in TSSN events. Equivalent potential temperature (green contour) in Kelvin plotted with differential temperature advection in $10^6$ K mb$^{-1}$ s$^{-3}$ (red contour). Dashed line indicates TROWAL axis. Area of interest is Joplin (JLN)</td>
<td>24</td>
</tr>
<tr>
<td>Figure 4.1.3 Same as figure 4.1.2 except valid for 0900 UTC 10 December 2003</td>
<td>24</td>
</tr>
<tr>
<td>Figure 4.1.4 Same as figure 4.1.2 except valid for 1000 UTC 10 December 2003</td>
<td>25</td>
</tr>
</tbody>
</table>
Figure 4.1.5 Same as figure 4.1.2 except valid for 1100 UTC 10 December 2003

Figure 4.1.6 Average DADTP ($10^6$ K mb$^{-1}$ s$^{-1}$) occurring during non-thundering snow events. Asterisk indicates mid-event period where visibility ≥ ¼ statute mile

Figure 4.1.7 RUC_20 initial hour valid 1200 UTC 2005 January 06. Example of an average DADTP ($10^6$ K mb$^{-1}$ s$^{-1}$) pattern occurring during a non-thundering snow event. $\theta_e$ (green contour) in Kelvin plotted with differential temperature advection ($10^6$ K mb$^{-1}$ s$^{-1}$; red contour). Area of interest is Green Bay (GRB)

Figure 4.1.8 Comparison of average DADTP values ($10^6$ K mb$^{-1}$ s$^{-1}$) over the lifetime of TSSN (red contour) and non-thundering snow events (blue contour). Asterisks represent period of lightning observation and non-thundering mid-event period respectively

Figure 4.1.9 Average vertical advection of stability ($10^6$ K mb$^{-1}$ s$^{-1}$) occurring throughout the lifetime of a TSSN event. Asterisks indicate period of lightning activity

Figure 4.1.10 RUC_20 initial hour valid 0700 UTC 10 December 2003. A typical VADVS pattern observed during TSSN events. $\theta_e$ (green countour) in Kelvin plotted with vertical advection of stability ($10^6$ K mb$^{-1}$ s$^{-1}$; red contour). Dashed line is the TROWAL axis. Area of interest is Joplin (JLN)

Figure 4.1.11 Average vertical advection of stability ($10^6$ K mb$^{-1}$ s$^{-1}$) occurring throughout the lifetime of a non-thundering snow event. Asterisks indicate midpoint of event

Figure 4.1.12 RUC_20 initial hour valid 1200 UTC 16 January 2003. Example of a typical VADVS pattern ($10^6$ K mb$^{-1}$ s$^{-1}$) occurring during a non-thundering snow event. $\theta_e$ (green contour) in Kelvin plotted with differential temperature advection (red contour). Dashed line is the TROWAL axis. Area of interest is Springfield Missouri (SGF)

Figure 4.1.13 Comparison of average VADVS ($10^6$ K mb$^{-1}$ s$^{-1}$) values over the lifetime of TSSN (red contour) and non-thundering snow events (Blue contour). Asterisks represent period of lightning observation and mid-event period respectively
Figure 4.1.14 Average vertical stretching of the air column \( (10^6 \text{ K m}^{-1}\text{s}^{-3}) \) occurring throughout the lifetime of a TSSN event. Asterisks indicate period of lightning activity.  

Figure 4.1.15 RUC_20 initial hour valid 0700 UTC 10 December 2003. \( \theta e \) (green contour) in Kelvin plotted with vertical stretching \( (10^6 \text{ K m}^{-1}\text{s}^{-1}); \text{red contour}) \). Dashed line is the TROWAL axis. Area of interest is Joplin (JLN).  

Figure 4.1.16 RUC_20 initial hour valid 0400 UTC 23 November 2003. \( \theta e \) (green contour) in Kelvin plotted with vertical stretching \( (10^6 \text{ K m}^{-1}\text{s}^{-1}); \text{red contour}) \). Area of interest is Salina, Kansas (SLN).  

Figure 4.1.17 Average vertical stretching \( (10^6 \text{ K m}^{-1}\text{s}^{-3}) \) occurring throughout the lifetime of a non-thundering snow event. Asterisk indicate midpoint of event.  

Figure 4.1.18 RUC_20 initial hour valid 1000 UTC 22 January 2005. Example of an average STRCH pattern \( (10^6 \text{ K m}^{-1}\text{s}^{-3}) \) occurring along a warm front during a non-thundering snow event. \( \theta e \) (green contour) in Kelvin plotted with vertical stretching (red contour). Area of interest is Green Bay (GRB).  

Figure 4.1.19 Comparison of average STRCH \( (10^6 \text{ K m}^{-1}\text{s}^{-3}) \) values over the lifetime of TSSN (red contour) and non-thundering snow events (Blue contour). Asterisks represent period of lightning observation and mid-event period respectively.  

Figure 4.1.20 Average static stability tendency \( (10^6 \text{ K m}^{-1}\text{s}^{-1}) \) occurring throughout the lifetime of a TSSN event. Asterisks indicate period of lightning activity.  

Figure 4.1.21 Comparison of average static stability tendency \( (10^6 \text{ K m}^{-1}\text{s}^{-3}) \) components during a TSSN event. Differential temperature advection (blue contour), vertical advection of stability (red contour), vertical stretching (green contour) and Stability tendency (magenta contour).  

Figure 4.1.22 Average static stability tendency \( (10^6 \text{ K m}^{-1}\text{s}^{-3}) \) occurring throughout the lifetime of a non-thundering snow event. Asterisks indicate midpoint of event.
Figure 4.1.23 Comparison of average static stability tendency components. Differential temperature advection ($10^8$ K mb$^{-1}$ s$^{-1}$; blue contour), vertical advection of stability ($10^6$ K mb$^{-1}$ s$^{-1}$; red contour), vertical stretching ($10^5$ K mb$^{-1}$ s$^{-1}$; green contour) and stability tendency (magenta contour)………………………………………… 51

Figure 4.1.24 Comparison of average static stability tendency ($10^6$ K mb$^{-1}$ s$^{-1}$) values over the lifetime of TSSN (red contour) and non-thundering snow events (Blue contour). Asterisks represent period of lightning observation and mid-event period respectively……… 52

Figure 4.2.1 Average Total Totals occurring throughout the lifetime of a TSSN event. Asterisks indicate period of lightning activity……………………………… 55

Figure 4.2.2 RUC_20 initial hour valid 0800 UTC 10 December 2003. A typical pattern of TT observed during a TSSN event. $\theta_e$ (green contour) in Kelvin plotted with Total Totals (red contour). Area of interest is Joplin (JLN)………………………………… 57

Figure 4.2.3 Average Total Totals value occurring throughout the lifetime of a non-thundering snow event. Asterisks indicate midpoint of event……………………………… 59

Figure 4.2.4 RUC_20 initial hour valid 1200 UTC 14 February 2005. A typical TT pattern found in non-thundering snow events. $\theta_e$ (green contour) in Kelvin plotted with Total Totals (red contour). Area of interest is Green Bay (GRB)……………………………………………….. 60

Figure 4.2.5 Comparison of average Total Totals values over the lifetime of TSSN (red contour) and non-thundering snow events (blue contour). Asterisks represent period of lightning observation and mid-event period respectively…………………………… 61

Figure 4.3.1 Average K-index occurring throughout the lifetime of a TSSN event. Asterisks indicate period of lightning activity…………………………………………………… 63

Figure 4.3.2 RUC_20 initial hour valid 0600 UTC 10 January 2003. A typical K-index pattern observed during a TSSN event. $\theta_e$ (green contour) in Kelvin plotted with K-index (red contour). Dashed line is the TROWAL axis. Area of interest is Joplin (JLN)…………………... 65
Figure 4.3.3 Same as figure 4.3.2 except valid for 0700 UTC 10 January 2003

Figure 4.3.4 Same as 4.3.2 except valid for 0800 UTC 10 January 2003

Figure 4.3.5 Same as figure 4.3.2 except valid for 0900 UTC 10 January 2003

Figure 4.3.6 Average K-index value occurring throughout the lifetime of a non-thundering snow event. Asterisks indicate midpoint of event.

Figure 4.3.7 RUC_20 initial hour valid 1200 UTC 14 February 2005. A typical K-index pattern observed during non-thundering snow events. $\theta_e$ (green contour) in Kelvin plotted with K-index (red contour). Area of interest is Green Bay (GRB).

Figure 4.3.8 Comparison of average K-index values over the lifetime of TSSN (red contour) and non-thundering snow events (blue contour). Asterisks represent period of lightning observation and mid-event period respectively.

Figure 4.4.1 Average Mid-level lapse rate in K km$^{-1}$ occurring throughout the lifetime of a TSSN event. Asterisks indicate period of lightning activity.

Figure 4.4.2 RUC_20 initial hour valid 0500 UTC 10 December 2003. A typical pattern of mid-level lapse rates occurring during a TSSN event. $\theta_e$ (green contour) in Kelvin plotted with mid-level lapse rates in K km$^{-1}$ (red contour). Dashed line is the TROWAL axis, Solid black line is mid-level lapse rate axis. Area of interest is Joplin (JLN).

Figure 4.4.3 Same as figure 4.4.2 except valid for 0600 UTC 10 December 2003.

Figure 4.4.4 Same as figure 4.4.2 except valid for 0700 UTC 10 December 2003.

Figure 4.4.5 Same as figure 4.4.2 except valid for 0800 UTC 10 December 2003.
Figure 4.4.6 RUC_20 initial hour valid 0300 UTC 23 November 2003. A typical mid-level lapse rate pattern occurring during a TSSN event occurring along a warm front. $\theta_e$ (green contour) in Kelvin plotted with mid-level lapse rates in K km$^{-1}$ (red contour). Black solid line is the axis of mid-level lapse rate. Area of interest is Green Bay (GRB)…………………

Figure 4.4.7 Average mid-level lapse rate (K km$^{-1}$) occurring throughout the lifetime of a non-thundering snow event. Asterisk indicates the midpoint of the event……

Figure 4.4.8 RUC_20 initial hour valid 1200 UTC 22 January 2005. A typical pattern of mid-level lapse rate observed during a non-thundering snow event. $\theta_e$ (green contour) in Kelvin plotted with mid-level lapse rates in K km$^{-1}$ (red contour). Area of interest is Green Bay (GRB)………………………………………………

Figure 4.4.9 Comparison of average mid-level lapse rate (K km$^{-1}$) over the lifetime of TSSN (red contour) and non-thundering snow events (blue contour). Asterisks represent period of lightning observation and mid-event period respectively………………

Figure 4.6.1 Average most unstable lifted index occurring throughout the lifetime of a TSSN event. Asterisks indicate period of lightning activity…………

Figure 4.6.2 Average most unstable lifted index occurring throughout the lifetime of a non-thundering snow event. Blue contour represents average MULI lifted from 655 mb. Green contour represents true MULI lifted from 552 mb. The asterisks indicate midpoint of event……………………………………………………

Figure 4.6.3 Comparison of average most unstable lifted index, lifted from 655 mb, over the lifetime of TSSN (red contour) and the average most unstable lifted index, lifted from 667 mb, non-thundering snow events (blue contour). Asterisks represent period of lightning observation and mid-event period respectively…

Figure 4.7.1 Comparison of pressure level (mb) of -10°C isotherm (green contour) and lifted parcel pressure level (mb; red contour) of the average most-unstable parcel throughout the lifetime of a TSSN event……………………………………………………
Figure 4.7.2 Comparison of the pressure level (mb) of the -10°C isotherm (green contour) and the LPL (mb) at an average level of 655mb (red contour), average LPL (blue contour) of the actual most-unstable parcel throughout the lifetime of a non-thundering snow event……………………………………….. …

Figure 4.7.3 Average MSL height (m) of the -10°C isotherm occurring throughout the lifetime of a TSSN event. Asterisks indicate period of lightning activity……………………………………………… 91

Figure 4.7.4 Average MSL height (m) of the -10°C isotherm occurring throughout the lifetime of a non-thundering snow event. Asterisks indicate mid-point of event…………………………… 93

Figure 4.7.5 Comparison of the average MSL height (m) of -10°C isotherm throughout the lifetime of TSSN events (red contour) and non-thundering snow events (blue contour)…………………………… 95

Figure 4.8.1 Percentage of TSSN events by stability regime, valid at initial hour of observation. Curved flow (dark blue), potential instability (green), neutral symmetric instability (yellow) and stable environments (light blue)………………………………… 96

Figure 4.8.2 Percentage of TSSN events by stability regimes, valid for the hour of initial lightning observation. Curved flow (dark blue), potential instability (green), neutral symmetric instability (yellow), potential symmetric instability and stable environments (light blue)…………………………… 98

Figure 4.8.3 Percentage of non-thundering events per stability regime, valid at initial hour of observation. Curved flow (dark blue), potential instability (green), neutral symmetric instability (yellow) and stable environments (light blue)…………………………… 100
Figure 4.8.4 Percentage of non-thundering snow events by stability regimes, valid for the mid-point of the average non-thundering snow event. Curved flow (dark blue), potential instability (green), neutral symmetric instability (yellow), potential symmetric instability (red) and stable environments (light blue).
List of Tables

Table 4.1.1 Correlation of individual components and the total static stability tendency equation .......................... 48

Table 4.1.2 Correlation of individual components and the total static stability tendency equation .................. 52

Table 4.8.1 Represents the distribution of TSSN events per stability regime over the first three hours of observation prior to initial lightning activity ......................................................... 97

Table 4.8.2 Represents the distribution of TSSN events by stability regime valid during the hours of lightning observation .......................................................... 99

Table 4.8.3 Represents the distribution of TSSN events by stability regime over the final three hours of observation .......................................................... 99

Table 4.8.4 Represents the distribution of non-thundering snow events per stability regime over the first three hours of observation, prior to the mid-point of the event ......................................................... 101

Table 4.8.5 Represents the distribution of non-thundering snow events by stability regime over the final three hours of observation ......................................................... 102

Table 5.2.1 TSSN events selected for this study including: date and time of event, location, station identifier, hour of initial cloud-to-ground flash and duration of event ......................................................... 119

Table 5.2.2 Non-thundering snow events selected for this study including: date and time of event, location, station identifier, hour of the mid-event period and duration of event ......................................................... 119
Chapter 1
Introduction

Thundersnow is a challenging forecast problem which can result in large amounts of snow in a short time period with little warning. This heavy snow can cause reduced visibility and impassable road conditions in a relatively confined area. These two factors can be an inconvenience if not a danger to public welfare. Therefore it is imperative to define regions of the atmosphere that are conducive to thundersnow events in order to properly warn the public of a potentially crippling snow event.

Much is known about the environment supportive of convective banded snow events, which share similar synoptic profiles to thundersnow events. Market (et al. 2004) provided a set of guidelines (composite maps) that define the synoptic profile of thundersnow. Several authors (Schultz 1999, Jurewicz and Evans 2004, Moore and Lambert 1993) have also investigated banded snowstorms as well as snow events with which lightning occurred.

To compliment plan view composites offered by Market (et al. 2004) this study seeks to examine and eventually establish the stability profile of lightning-producing snow events. With this in mind two data sets were gathered from snow events featuring snow with lightning and snow events with no known lightning activity. These two data sets will be examined with special attention to stability and the thermodynamic profile found in each individual event.

It is hoped that this study will provide the operational forecaster a tool in which to diagnose the potential for a thundersnow event. This tool will enable
the forecaster to better warn the public of an upcoming snow event with the potential to deposit heavy snow in a short period of time. Additionally, the forecaster will be able to discriminate better between non-convective (non-thundering) snow events and thundersnow events with knowledge gathered from this study.

1.1 Purpose & Objectives

1.1.1 Purpose

The purpose of this study is to further the understanding of convective snow (thundersnow) by comparing this particular type of event to non-thundering snow events. It should be noted that for this study the terms thundersnow and convective snow are synonymous, therefore interchangeable. Specifically this study is concerned with the stability regime and evolution of stability associated with thundersnow events along with how they compare to non-thundering snow events. In order to determine if any thresholds exist, various stability indices are compared between these two types of snowstorms.

1.1.2 Objectives

Several objectives are identified to accomplish the aforementioned purpose.

This research will:

- Determine which stability regime is dominant in convective snow events.
• Determine if any thresholds exist in traditional stability indices to discriminate between thundersnow events and non-thundering snow events.
• Determine what component of the stability tendency equation dominates the total tendency in thundersnow events.

1.2 Statement of Thesis

In the past, numerous studies have been performed on banded convective snow events. However, little research has been done on thundersnow specifically. This thesis will aim to satisfy a few fundamental questions including:
• What stability regime is dominant in thundersnow events?
• Are there threshold values in traditional stability indices discriminating thundering events from non-thundering events?
• Are there recognizable patterns of these indices in plan view charts?
• How does stability evolve in time in thundersnow events and does it differ from the evolution of non-thundering snow events?
Chapter 2
Literature Review

Numerous articles are available on banded snowfall in the literature. Most of these studies ignore the topic of lightning or when lightning is addressed it is in literature involving lake effect snow. Few articles actually address thundersnow in the absence of topographic influences or lake enhancement. This review will look at several articles addressing climatology, synoptic setup, lightning with snow and a review of conditional symmetric instability. These topics will be addressed in this order.

2.1 Climatology

In a climatological study, Market et al. (2002) analyzed 30 years worth of thundersnow events. Several important factors were uncovered by Market during his investigation. First it was discovered that these types of events occur in three different geographical areas. These areas include the intermountain West where orographic lift would likely play a key role in enhancing vertical motion capable of creating both heavy snow and charge separation. A second region known for thundersnow is the Great Lake region. Undoubtedly, the Great Lakes provide abundant moisture to fuel these storms as well as providing a source of sensible and latent heating. Of interest in this study is the Midwest where instability (when present) is typically elevated; we see limited orographic effects and no lake enhancement.
2.2 Synoptic Scale

Of more significance to the operational forecaster are the following conclusions as proposed by Market et al. (2002). First of which is that thundersnow events often occur in association with a transient midlatitude cyclone, northwest or northeast of a cyclone’s center at a mean distance of approximately 440 km (Market et al. 2002). It is known that the deep surface layer thermal stratification during these events is very stable and near saturated.

2.3 Banded Snow

Along with being located both to the northwest and/or northeast of an extratropical cyclone, thundersnow often occurs in elevated bands. Banded precipitation events are well documented in the literature. One particular study by Jurewicz and Evans (2004) investigated two banded snow events with different background (synoptic) flows. The bulk of this investigation concentrated on the mesoscale frontogenetic circulations and the release of conditional symmetric instability (CSI).

Jurewicz and Evans (2004) found that in both cases CSI was present with a midtropospheric frontogenesis axis normal to the snowband. Jurewicz and Evans (2004) postulated that whenever it appears that frontogenetic forcing could be significant and potentially produce banded precipitation, viewing a cross section of frontogenesis, equivalent potential vorticity (EPV) or moist potential vorticity (MPV), omega, and relative humidity would assist forecasters in establishing a link between frontal scale forcing, instability, and upward motion.
Schultz (1999) examined lake-effect snowstorms both with and without lightning in two different geographic locations. He postulated that convective available potential energy (CAPE) is not a good indicator of lightning since it was absent during events that exhibited no lightning and those that did exhibit lightning. He found statistically significant results indicating the most useful parameters in forecasting lightning in lake-effect events are lower tropospheric temperatures and lifted index, whereas dewpoint depression and CAPE are not useful.

Two possibilities were given to explain Schultz’s (1999) results. First, at colder temperatures, as in non-lightning events, there may not be enough water vapor in the atmosphere to activate the riming process, which promotes the transfer of adequate electrical charge in the mixed phase region, even with sufficient vertical motion. Second, the warmer lower tropopause in the lightning events implies that the height of the -10°C isotherm (the region where charging is believed to occur most efficiently due to the coexistence of high concentrations of ice crystals and supercooled liquid water) is high enough off the ground that vertical motions in the lower troposphere during lake-effect events can become of sufficient intensity to separate charge.

2.4 CSI

Along with being primarily associated with banded precipitation systems, thundersnow is thought to be a product of low-topped elevated convection. Studies by Smith et al. (2005) found that cloud top temperatures averaged -45°C, in a study of seven thundersnow (TSSN) events. However, cloud top
temperatures were often only -20°C, indicating some form of low-topped instability. Due to the relatively low cloud top traditional warm season stability indices, such as the Total Totals (TT) index maybe unrepresentative of environments capable of supporting this phenomena. Traditional indices are more representative of deep instability found during summer time thunderstorms. Therefore, the forecaster must rely on alternative diagnostics of stability such as CSI.

CSI is an atmospheric condition in which combinations of gravitational (static stability) and horizontal forces (inertial stability) are acting simultaneously. When CSI is present in an otherwise insipid environment profound weather may occur in the form of banded precipitation. Studies by Wiesmueller and Zubrick (1998) reintroduce the theories of CSI, EPV and their affects on frontogenetic fields. McCann (1999) pointed out deficiencies that Wiesmueller and Zubrick’s (1998) work and offer his views on the topic, all of which we examine presently.
Weissmuller and Zubrick (1998) offer a graphic illustration (Figure 2.4.1), which depicts the basic theory of CSI. Figure 2.4.1 illustrates the resultant accelerations of a parcel when displaced in various manners. Parcel A is displaced horizontally, parcel B is displaced vertically. Both of these displacements result in the parcel undergoing a tendency of restoration to its initial condition.

However, Point C is displaced in a slantwise manner, which results in an acceleration in the direction of displacement, not a restoration force. Therefore, in a CSI environment, parcels resist vertical acceleration (due to vertical instability) and horizontal acceleration (due to horizontal instability), however, slantwise displacements result in motion parallel to the displacement (slantwise instability).
McCann (1999) modified Wiesmueller and Zubrick’s Figure 1 (Figure 2.4.1 in this study), in two ways (Figure 2.4.2). McCann (1999) modified the labeling of the dashed lines to make them lines of equal saturated equivalent potential temperature ($\theta_{es}$) instead of equivalent potential temperature ($\theta_e$). McCann (1999) points out that it is not necessary to assume a saturated environment just a saturated parcel. Secondly, McCann (1999) labeled the endpoints of the double arrows of each of their sample parcel displacements with a primed letter. McCann points out that since lines of both $\theta_{es}$ and absolute geostrophic momentum ($M_g$) are parallel, static stability is the same for each parcel as well as inertial stability. Therefore, all the parcels will be accelerated away from their initial positions (McCann 1999).
In Figure 2.4.1 parcel A has the same value of $M_g$ as $A'$ but since $\theta_e$ of the parcel is still warmer than the $\theta_e$ of the environment the parcel will continue to rise. As the parcel continues upward it will experience greater $M_g$ and be forced to the left. McCann (1999) specifies that since $\theta_es$ are sloped less than the $M_g$ surfaces the parcel will always be warmer than the environmental $\theta_es$, therefore the parcel will continue along a slantwise path.

Wiesmueller and Zubrick (1998) summarize the synoptic setting that is characteristic of CSI. CSI often occurs in regions experiencing increasing vertical wind shear, on the order of 10-20 ms$^{-1}$ in the lowest 1-2 km. Additionally, low static stability in the middle troposphere and a statically stable boundary layer are necessary. The thermodynamic profile should be near saturation and cool closely to the moist adiabatic lapse rate at the lower to mid-levels of the atmosphere. The lifted parcel must also be nearly saturated within the region of symmetric instability, typically occurring between 800 and 500hpa in the central United States (Weissmuller and Zubrick 1998). These conditions are often satisfied near a warm front and ahead of a synoptic scale trough.

Moore and Lambert (1993), showed that CSI can also be evaluated by examining fields of EPV in cross sections oriented normal to the geostrophic thermal winds.

EPV is defined as:

$$ EPV' = -\eta \cdot \nabla \theta_e $$

(2.4.1)

where $\eta$ is the three dimensional vorticity vector and $\nabla$ is the gradient operator in x,y and p coordinates. This form, derived by Martin et al. (1992), includes both
horizontal and vertical motions. By assuming geostrophic flow and neglecting vertical motion and terms with respect to $y$, Moore and Lambert (1993) arrived at:

$$EPV = g \left( \frac{\partial M_g}{\partial p} \frac{\partial \theta_e}{\partial x} - \left( \frac{\partial M_g}{\partial x} \frac{\partial \theta_e}{\partial p} \right) \right)$$

where term A is the product of the change of geostrophic momentum ($M_g$) with respect to pressure and the change of potential temperature ($\theta_e$) in the $x$ direction, where $x$ is the distance from the origin of the cross section. Term B is the product of the change of geostrophic momentum in the $x$ direction and the static stability. By multiplying equation 2.4.2 by gravity ($g$) EPV can be expressed in units of $10^{-6}$ m$^2$K s$^{-1}$ kg$^{-1}$, or potential vorticity units (PVU; Weissmuller and Zubrick 1998).

CSI may be inferred when the atmosphere is saturated and equation 2.4.3 is zero or negative. CSI occurs when term A dominates term B which is often the case since term A is usually negative. Term B can be negative or a small positive value. Negative EPV implies instability for parcel accelerations away from their original positions (McCann 1999).

McCann (1995) derived EPV in vector form:

$$EPV = g \left[ \hat{k} \cdot \left( \frac{\partial V_g}{\partial p} \times \nabla_p \theta_w \right) - \zeta_g \frac{\partial \theta_w}{\partial p} \right]$$

where term A represents the thermal wind crossed with the temperature gradient, term B is the product of the geostrophic absolute vorticity ($\zeta_g$) and static
stability. Once again negative EPV implies instability with parcel accelerations away from their original positions (McCann 1999).

McCann (1995) states that the first term of equation 2.4.3 is equivalent to the horizontal temperature gradient squared and so is always negative and "destabilizes" the environment. Therefore, the sign of EPV depends on the sign and relative magnitude of the second term (McCann 1999). A negative EPV can result from a negative term B (conditional static instability) or a small positive value, which results in CSI (McCann 1999). Without observing both terms A and B to assess why EPV is negative the forecaster cannot state whether the negative EPV represents conditional static instability or CSI. Therefore, EPV should also be used with traditional cross sections of $M_g$ and $\theta_{se}$ surfaces, or $M_g$ and $\theta_e$ with relative humidity.

### 2.5 Lightning

Cloud electrification is dependent upon the formation of ice crystals and the eventual formation of graupel at different temperature levels. Ice crystals formed below the minus 10° C isotherm acquire positive charges a negative charge regime exists in particles formed between the -15° C and -20° C isotherms (Figure 2.5.1). An area of positive charge is acquired by graupel formed above -20° C (Saunders 1992).

Further evidence of these charge acquisition levels was revealed by Takahashi, et al. (1999) who found that negatively charged graupel appeared near the -10° C level in a relatively strong updraft. Takahashi also observed many
positively charged ice crystals existing between the -30°C and -20°C isotherms.

Negative charges were again found at temperatures between -20°C and -10°C.

![Diagram](image)

Figure 2.5.1 Theoretical critical charge reversal layers (Saunders 1992). Y-axis is temperature in degrees Celsius.

Temperature is not the only factor that determines whether graupel acquires a positive or negative charge. Charge is also dependant on whether the growth of the particle is of a wet or dry mechanism (Jayaratyne 1993, Avila et al 1995). Wet growth occurs when liquid water content (LWC) is high at the graupel’s surface and evaporation occurs expanding the surface area of the graupel. When two particles interact, the surface that is growing by diffusion will acquire the positive charge (Jayaratyne 1993).

Dry growth occurs at lower LWC values, where super-cooled drops accrete to the graupel’s surface (riming). In a naturally occurring cloud, many different ice crystals and graupel particles coexist with different charges. Since smaller ice crystals fall at slower rates, than larger graupel, collisions are inevitable. Not only will the graupel develop a charge from aggregation but
graupel and ice may also be charged by frictional contact (Takahashi, et al. 1999). Thus, frictional contact can lead to charge separation.

Cloud electric field profiles have been observed to exhibit a tripole structure (Figure 2.5.2; Takashi, et al. 1999). A tripole structure is an expanded die-pole with an additional positive charge below the negatively charged region. These three charge regions, as proposed by Simpson and Scarce (1937) and Simpson and Robinson (1941), illustrate a positive charge at about -30° C levels capping a negative charged region to -10°C and a final positive field below -10° C (Takahashi, et al. 1999). This cloud electric profile closely resembles the Takahashi charge pattern for ice and graupel.

![Figure 2.5.2](image)

Figure 2.5.2 The proposed tri-pole cloud structure is similar to the charges acquired by graupel at different.

In natural clouds, ice crystals and graupel are not static; rather, they are very dynamic. Not only are these objects falling at different rates due to gravity and updrafts, but in-cloud circulations are also juggling them around. In a
natural thunderstorm updraft, positively charged ice crystals are carried upward colliding with negatively charged falling graupel (Takahashi et al. 1999). Therefore collisions between positive and negatively charged particles are inevitable.

These collisions generate a space charge. Takahashi found that “a space charge sufficient to initiate lightning was obtained with a graupel concentration greater than 1 liter$^{-1}$ (Takahashi et al. 1999). Ice crystal concentrations of 1 liter$^{-1}$ are average for a cloud with a temperature of less than -4°C. Theoretically clouds containing higher concentrations of ice would generate a greater space charge, thus producing more cloud-to-ground and cloud-to-cloud lightning.
Chapter 3
Methodology
3.1 Event Selection

This study will address the evolution of the atmospheric stability in two snowstorm collections in which significant snowfall was occurring. The first data set will represent environments that produced significant snow while lightning was occurring in the vicinity. A second data set represents environments, which produced significant snowfall, but lightning activity was absent.

Eight thundersnow (TSSN) events from the 2003 and 2004 winter season were chosen for this study. These eight events all occurred in the mid-western United States, following Market et al. (2002), and were not associated with any topographical influence or lake enhancement. Additionally, these eight events are all associated with a transient mid-latitude cyclone. An event was included when surface observations reporting snow or TSSN coincided with a lightning flash supported by the National Lightning Detection Network (NLDN). Finally, analysis of an event began three hours prior to the first lightning flash recorded by the NLDN and continued every hour until three hours after the final lightning flash was recorded.

Seven non-thundering snow events were chosen from winter seasons between the years of 2000 and 2004. Special care was taken to ensure that none of these events exhibited any lightning activity to the extent permitted by surface observations and the NLDN. All of these non-thundering events are associated
with a transient mid-latitude cyclone located in the central United States. None of these seven events were influenced by topographic or lake enhancement.

An event was considered when snow was falling in absence of thunder and/or lightning. Additionally, visibility during an event had to be \( \leq \frac{1}{4}\text{sm} \) due to snow falling. Analysis was performed three hours prior to the point at which visibility fell below \( \leq \frac{1}{4}\text{sm} \) and was carried on every hour for seven hours, capturing a majority of the snow event lifetime. The author acknowledges that this may not be the best method by which to approach non-thundering snow events, however both the author and his advisor agree that this is a sufficient method of event selection.

3.2 Data

This study utilizes Rapid Update Cycle (RUC) numerical model data in an isobaric coordinate system spaced on a 20km grid. The Atmospheric Radiation Measurement (ARM) group, sponsored by the U.S. Department of Energy, Office of Science, Office of Biological and Environmental Research, Environmental Sciences Division supplied these data. In preparation for analysis this data set was converted into GEMPAK form and various scripts were employed to calculate derived variables.

Ultimately this enabled synoptic scale plan view maps to be constructed and displayed in the General Meteorological Package (GEMPAK). Additionally, thermodynamic diagrams are also constructed through the aid of NSHARP, a sounding display program, which is a companion to GEMPAK. Ultimately,
these tools allow the meteorologist to develop diagnostics to analyze atmospheric phenomena.

We also acknowledge that the RUC may suffer from biases in certain environments (Thompson et al. 2002). Moreover, we are employing a model initial field, which differs from an objective analysis in that the initial field is a dynamically balanced state suitable for integration to a final solution 3 to 12 hours hence (in the case of the RUC). Nevertheless, the RUC represents the best assimilation of all available data (profiler, aircraft, etc.) each hour, and we proceed, mindful of its limitations.

3.3 Indices Examined

3.3.1 Traditional Indices

Data collected from these two data sets are analyzed with particular attention given to traditional stability indices. These traditional stability indices include; Total Totals, K-index, mid-level lapse rates, CAPE and the most-unstable Lifted Index. Each of the aforementioned indices is examined over the evolution of both TSSN and non-thundering snow events.

Once the specific indices for both types of events are documented, the data gathered are compared. This comparison will delineate whether or not a particular index is a good descriptor for a TSSN event. A simple two-tailed t-test is invoked to determine if the sample means are statistically different. If the sample mean is proven at 95% confidence level to be different then the index is said to be a good descriptor for TSSN.
The t-test used in this study is valid for n<30, which fits the small sample set of this study providing that the following assumptions are met. First the samples must be independent; the sample sets in this study are indeed independent. Second we must assume a normal distribution. One final assumption is that the sample variances are equal or close to being equal. Although the sample size is small, these assumptions are valid for this study.

Each of the traditional stability indices is also plotted in a plan view chart along with $\theta_e$ to determine if any spatial patterns exist further discriminating between TSSN and non-thundering snow events. $\theta_e$ is plotted at either 700mb or 500mb depending on which level best represents the TROWAL or frontal boundary.

### 3.3.2 Static Stability

The static stability tendency (STABTDY) equation:

$$
\frac{\partial}{\partial t} \left(-\frac{\partial \theta}{\partial p}\right) = -\frac{\partial}{\partial p} \left(-\mathbf{\nabla} \cdot \mathbf{\nabla} \theta\right) - \omega \frac{\partial}{\partial p} \left(-\frac{\partial \theta}{\partial p}\right) - \delta \frac{\partial \theta}{\partial p} - \frac{\partial}{\partial p} \left(\frac{\theta}{c_p T} \frac{dQ}{dt}\right)
$$

as defined by Bluestein (1992) represents the tendency of stability at a specific point and time. In this study the static stability tendency is partitioned into three of its four terms. These terms represent differential temperature advection (DADTP; term A), the vertical advection of stability (VADVS; term B) and stretching of the air column (STRCH; term C). Vertical change in diabatic heat (term D) is the fourth term, however at this time it is considered small and therefore is ignored in this study. This statement can be justified by assuming
that in a deep saturated environment little evaporation or condensation will occur. Additionally, it is felt that the change of diabatic heating will be even smaller since this study is dealing with a small, 100-mb layer.

This study will examine the stability tendency and each of its individual terms within a 100-mb layer. The center of this layer is located at the same level as the most unstable lifted parcel level (MULPL; figure 3.3.1). It is reasonable to evaluate the stability tendency at this particular level since the atmosphere begins to cool above this point; theoretically this point should also be the top of the TROWAL air stream. Additionally, by evaluating the stability tendency from the MULPL some level of consistency is established in the methodology of this study between the analysis of the most unstable lifted index, the evaluation of CAPE and of course the evaluation of the stability tendency itself.

Figure 3.3.1 Example sounding valid 0700 UTC 10 January 2003 for Joplin Missouri depicting the depth and location of the layer examined during analysis of the stability tendency and its individual terms.
Each of the three terms of the stability tendency equation are evaluated in the same manner as the traditional indices are subjected to. Again documenting the evolution of each individual term for both TSSN and non-thundering snow events. As with the traditional indices, plan view charts of each term are generated along with an elevated level of $\theta_e$.

Additionally, the magnitude of the total stability tendency is calculated for each event, further providing an idea of the stability tendency of both phenomena over time. The total stability tendency is then compared to each of the individual terms in order to determine which dominates the total tendency. Along with comparing the difference among the actual values of each component a simple correlation is performed in order to reveal which components correlate with increases and decreases of the total stability tendency.

### 3.3.3 CSI

Finally the stability regime of each type of event is examined with respect to CSI. CSI is subjectively analyzed through cross sectional analysis, which is carried out for each hour of each event. These cross sections aid in determining whether the atmosphere is unstable for a vertical displacement (release of potential or conditional instability) or a slantwise displacement (PSI or CSI).

Procedures for assessing CSI set by Moore and Lambert (1993) are followed in this study. These procedures diagnose CSI objectively through analysis of absolute geostrophic momentum ($M_g$) and equivalent potential temperature ($\theta_e$). An additional criterion for assessing CSI is that the observed layer be sufficiently saturated, therefore for CSI to be diagnosed the layer must
be of at least 80% relative humidity (Moore and Lambert 1993, Schultz and Schumacher 1999); thus \( \theta_e = \theta_{es} \) per Moore and Lambert (1993) these cross sections are taken normal to the thermal wind as represented by the 850 to 300mb thickness.

### 3.4 Methodology Summary

Through analysis of the traditional stability indices of both TSSN and non-thundering snow events certain discriminators are exposed in regards to TSSN. With the data gathered for each hour of each event a look at the evolution in stability of the atmosphere is also revealed and differs between TSSN and non-thundering snow events. Additionally, these procedures also give insight as to what mechanism contributes to the reduction of the stability tendency of the atmosphere in the vicinity of snow events. Finally, cross sectional analysis of the atmosphere lends credence to what regime of instability is responsible for TSSN and non-thundering snow events.
Chapter 4
Analysis
4.1 Stability Tendency

4.1.1 DADTP

The differential temperature advection (DADTP) term of the stability tendency equation (Eqn. 3.3.1) tends toward more positive values during the hours prior to lightning activity (Figure 4.1.1). When comparing TSSN events taking place with a TROWAL it is evident that DADTP values immediately left of a TROWAL air-stream tend towards positive values, or large negative values (Figure 4.1.2). Areas bordering the TROWAL and immediately right of the TROWAL air-stream are often negative.
Figure 4.1.2 RUC_20 initial hour valid 0800 UTC 10 December 2003. Typical DADTP pattern observed in TSSN events. Equivalent potential temperature (green contour) in Kelvin plotted with differential temperature advection in 10⁻⁶ K mb⁻¹ s⁻¹ (red contour). Dashed line indicates TROWAL axis. Area of interest is Joplin (JLN).

Figure 4.1.3 Same as figure 4.1.2 except valid for 0900 UTC 10 December 2003.
Figure 4.1.4 Same as figure 4.1.2 except valid for 1000 UTC 10 December 2003.

Figure 4.1.5 Same as figure 4.1.2 except valid for 1100 UTC 10 December 2003.
Upon further examination of the DADTP term it is observed that it represents the change with height of the advection of potential temperature. For example positive values of DADTP can mean warm air advection increasing with height, which when elevated is a stabilizing feature. Negative values can indicate cold air advection increasing with height, which of course when elevated the area below can become buoyant. In the case of a TROWAL, warm air is advected aloft, effectively stabilizing the column beneath and possibly destabilizing the column above. Therefore, when the TROWAL axis at a particular level moves into a region that region experiences warm air advection and DADTP tends towards positive values.

At the hour of initial lightning activity the average DADTP value tends towards even greater positive values. The average DADTP value occurring the hour of first reported cloud-to-ground activity is \(0.84 \times 10^{-6} \text{ K m}^{-1} \text{ s}^{-1}\) this average value has a standard deviation of \(1.24 \times 10^{-6} \text{ K m}^{-1} \text{ s}^{-1}\). The minimum recorded value at this time period is \(-4.1 \times 10^{-6} \text{ K m}^{-1} \text{ s}^{-1}\), with the greatest value obtained being \(6.6 \times 10^{-6} \text{ K m}^{-1} \text{ s}^{-1}\).

Once lightning activity begins average values of DADTP associated with lightning occurring while snow is falling tend toward negative values. The average DADTP value occurring during the hours of lightning activity is \(-0.5 \times 10^{-6} \text{ K m}^{-1} \text{ s}^{-1}\). The minimum value recorded while lightning is occurring is \(-0.9 \times 10^{-6} \text{ K m}^{-1} \text{ s}^{-1}\), with a maximum-recorded value of \(4.2 \times 10^{-6} \text{ K m}^{-1} \text{ s}^{-1}\).

An explanation for the tendency of DADTP towards more negative values could be that the TROWAL air-stream is departing the region, allowing for cold air to be advected into the region. Analyzing plan view charts of DADTP for this
time period reveals that the TROWAL axis is in fact departing the area of interest (Figure 4.1.2 see through Figure 4.1.5). Therefore it is reasonable that cold air advection is taking place and causing DADTP to tend towards more negative values.

During the first three hours of a non-thundering snow event the average DADTP value is $-1.2 \times 10^{-6}$ K mb$^{-1}$ s$^{-1}$ with a standard deviation of $2.2 \times 10^{-6}$ K mb$^{-1}$ s$^{-1}$. The minimum value occurring within the first three hours of a non-thundering event is $-6.8 \times 10^{-6}$ K mb$^{-1}$ s$^{-1}$ with the maximum value being $1.95 \times 10^{-6}$ K mb$^{-1}$ s$^{-1}$. It does appear that DADTP values do increase with time throughout the first three hours of an average non-thundering snow event (Figure 4.1.6). On average this value is quite small and remains a negative value.

Similar to the TSSN events this tendency towards positive values could be explained by an approaching warm front or TROWAL axis. Although this maybe
an explanation a majority of the non-thundering events occur behind the leading edge of a warm front or left of the TROWAL axis.

The average value for DADTP occurring at the midpoint of a non-thundering snow event is $-1.0 \times 10^{-6}$ K mb$^{-1}$ s$^{-1}$ with a standard deviation of $2.2 \times 10^{-6}$ K mb$^{-1}$ s$^{-1}$. The largest recorded value at this time is $1.5 \times 10^{-6}$ K mb$^{-1}$ s$^{-1}$ and the minimum value is $-4.7 \times 10^{-6}$ K mb$^{-1}$ s$^{-1}$. In the mid-point of a non-thundering event the average value is not a local maximum or local minimum in relation to its neighbors (Figure 4.1.7). The average value does fall somewhere in between the local maximum and minimum in its immediate area.

Similar to the hours after lightning is recorded in a TSSN event, DADTP values in non-thundering events tend towards greater negative values. The average DADTP value occurring during the final three hours of a non-
thundering event is \(-2.2 \times 10^{-6} \text{ K m}^{-1} \text{s}^{-1}\) with a standard deviation of \(3.2 \times 10^{-6} \text{ K mb}^{-1} \text{s}^{-1}\).

Figure 4.1.8 Comparison of average DADTP values \((10^{-6} \text{ K mb}^{-1} \text{s}^{-1})\) over the lifetime of TSSN (red contour) and non-thundering snow events (blue contour). Asterisks represent period of lightning observation and non-thundering mid-event period respectively.

Comparing DADTP values in TSSN events and non-thundering events reveals that TSSN events tend sharply towards greater positive DADTP values prior to the first report of lightning activity, where non-thundering events slowly tend toward greater negative values prior to the mid-point of the event (Figure 4.1.8). In fact, the average change from the beginning of the observation period to the first lightning report is \(2.9 \times 10^{-6} \text{ K mb}^{-1} \text{s}^{-1}\), where the difference between the beginning of a non-thundering snow event and its mid-point value is only \(0.3 \times 10^{-6} \text{ K mb}^{-1} \text{s}^{-1}\). From this we may infer that a greater amount of vertical change in warm air advection is occurring in TSSN events than in non-thundering events.
A t-test was performed on both the TSSN and non-thundering sample sets. The average value for DADTP occurring during the initial hour of a non-thundering event is $-1.3 \times 10^6 \text{ K mb}^{-1} \text{ s}^{-1}$, while the average value for DADTP for the same time period in a TSSN event is $-2.1 \times 10^6 \text{ K mb}^{-1} \text{ s}^{-1}$. With regard to the P value of a two-tail t-test it cannot be said that the DADTP sample mean is statistically different between non-thundering and TSSN events. The confidence level for this test is only 31% that the means are a non-random occurrence.

Of slightly greater confidence is the difference between DADTP values valid during the mid-point and those occurring during the initial hour of lightning activity in a TSSN event. With a 75% confidence level it can be said that the mean value of DADTP occurring at the initial hour of lightning activity is much greater, more positive, than that occur at the mid-point of a non-thundering event. Although, this is not statistically significant it is significant enough to note. Because of the low confidence level it cannot be stated that DADTP alone is a good indicator of a TSSN event.
4.1.2 VADVS

Figure 4.1.9 Average vertical advection of stability (10\(^6\) K mb\(^{-1}\) s\(^{-1}\)) occurring throughout the lifetime of a TSSN event. Asterisks indicate period of lightning activity.

On average the vertical advection of stability (VADVS) tends toward lesser positive values throughout the hours prior to lightning activity, with the exception of hour three which actually tends towards more positive values of VADVS (Figure 4.1.9). Over the first two hours prior to the indication of lightning activity the average value of VADVS is 2.4x10\(^6\) K mb\(^{-1}\) s\(^{-1}\) with a standard deviation of 0.2x10\(^6\) K mb\(^{-1}\) s\(^{-1}\). The best VADVS value for this two-hour period is -1.0x10\(^6\) K mb\(^{-1}\) s\(^{-1}\) and the largest positive value was 8.5x10\(^6\) K mb\(^{-1}\) s\(^{-1}\). Recall that ideally this component should be negative to optimize the tendency towards instability.
As previously mentioned vertical advection of stability tends towards lesser positive values during the hours prior to first indication of lightning activity, with the exception of hour three (Figure 4.1.9). There is an increase in the average value of VADVS occurring in hour three from the previous hour of $2.3 \times 10^{-6}$ K mb$^{-1}$ s$^{-1}$. This phenomenon is apparent in four of the eight sample sets with the remaining four sets tend slowly towards lesser positive values. The average VADVS value for this hour is $4.5 \times 10^{-6}$ K mb$^{-1}$ s$^{-1}$ with a standard deviation of $7.4 \times 10^{-6}$ K mb$^{-1}$ s$^{-1}$. As reflected in the standard deviation there is quite a large range in values at this hour. In fact the values range from $-0.2 \times 10^{-6}$ K mb$^{-1}$ s$^{-1}$ to $21.0 \times 10^{6}$.

Once again VADVS tends toward smaller positive values at the hour of initial lightning activity. In fact the average value of VADVS for this period is $1.7 \times 10^{-6}$ K mb$^{-1}$ s$^{-1}$ which is a decrease of $2.8 \times 10^{-6}$ K mb$^{-1}$ s$^{-1}$ from the previous hour. The standard deviation at this hour is $4.4 \times 10^{-6}$ K mb$^{-1}$ s$^{-3}$, which is quite large. The maximum value recorded at this hour is $8.7 \times 10^{-6}$ K mb$^{-1}$ s$^{-1}$ with a minimum value of $-6.1 \times 10^{-6}$ K mb$^{-1}$ s$^{-1}$.

During the hours while lightning is occurring VADVS tends towards smaller values. The average value occurring during this period is $1.4 \times 10^{-6}$ K mb$^{-1}$ s$^{-1}$ with a standard deviation of $0.5 \times 10^{-6}$ K mb$^{-1}$ s$^{-1}$. VADVS values ranged from $-6.1 \times 10^{-6}$ K mb$^{-1}$ s$^{-1}$ to $13.3 \times 10^{-6}$ K mb$^{-1}$ s$^{-1}$, once again revealing the great range of values present during thundersnow events. The cessation of lightning activity is marked by an increase towards greater positive values of VADVS.

The hour directly following lightning activity exhibits a marked rise in average VADVS value. The average value for this time period is $4.9 \times 10^{-6}$ K mb$^{-1}$ s$^{-1}$.
with a standard deviation of $6.7 \times 10^6$ K mb$^{-1}$ s$^{-1}$. This is an increase of $3.6 \times 10^6$ K mb$^{-1}$ s$^{-1}$ over the average value found while lightning is occurring. The hours leading up to the end of an average TSSN event tend once again towards lesser positive values of VADVS. The average VADVS for the final period of observation is $0.1 \times 10^6$ K mb$^{-1}$ s$^{-1}$, which is a difference of $5.7 \times 10^6$ K mb$^{-1}$ s$^{-1}$ over the hours after the cessation of lightning activity.

In plan view charts no definitive pattern in discernable specifically to TSSN events. Often the area of interest will be positioned along the apex of the TROWAL axis, were gradients of positive VADVS exist. Negative vertical advection of stability is usually present along the TROWAL axis closer to the cyclone center (Figure 4.1.10).

![Figure 4.1.10 RUC_20 initial hour valid 0700 UTC 10 December 2003. A typical VADVS pattern observed during TSSN events. $\theta$ (green countour) in Kelvin plotted with vertical advection of stability ($10^6$ K mb$^{-1}$ s$^{-1}$; red contour). Dashed line is the TROWAL axis. Area of interest is Joplin (JLN).]
Figure 4.1.11 Average vertical advection of stability ($10^6 \text{ K mb}^{-1} \text{s}^{-1}$) occurring throughout the lifetime of a non-thundering snow event. Asterisks indicate midpoint of event.

Similar to TSSN events VADVS tends towards smaller positive values in non-thundering snow events (Figure 4.1.11). But the trend towards greater positive values prior to the mid-point found in TSSN events is absent. The average VADVS value found during the first three hours of a non-thundering event is $4.1 \times 10^6 \text{ K mb}^{-1} \text{s}^{-1}$ with a standard deviation of $0.9 \times 10^6 \text{ K mb}^{-1} \text{s}^{-1}$. The greatest value found during this time period is $17.4 \times 10^6 \text{ K mb}^{-1} \text{s}^{-1}$ with the best value recorded being $0.9 \times 10^6 \text{ K mb}^{-1} \text{s}^{-1}$. The difference in VADVS from the initial hour of observation to the hour prior to the mid-point of the event is $1.8 \times 10^6 \text{ K mb}^{-1} \text{s}^{-1}$. This works out to be a decrease in VADVS of $0.60 \times 10^6 \text{ K mb}^{-1} \text{s}^{-1}$ from the initial period of observation and the midpoint of the non-thundering snow event.

At the midpoint of a non-thundering event VADVS ranged from $-0.3 \times 10^6 \text{ K mb}^{-1} \text{s}^{-1}$ to $3.8 \times 10^6 \text{ K mb}^{-1} \text{s}^{-1}$ with an average value of $1.2 \times 10^6 \text{ K mb}^{-1} \text{s}^{-1}$. From
the mid-point to the following hour a slight tendency toward a more positive value is noted in the average VADVS value. This increase is subtle $0.8\times10^6$ K mb$^{-1}$ s$^{-1}$ and is followed by a gradual tendency towards lesser positive values.

Few characteristic patterns are discernable in VADVS in relation to non-thundering snow events. However, a positive field of VADVS can often be found in non-thundering snow events occurring along the warm front (Figure 4.12). Along the trailing edge of the warm front negative values of vertical advection of stability can be found.

Figure 4.12 RUC_20 initial hour valid 1200 UTC 16 January 2003. Example of a typical VADVS pattern ($10^6$ K mb$^{-1}$ s$^{-1}$) occurring during a non-thundering snow event. $\theta_e$ (green contour) in Kelvin plotted with differential temperature advection (red contour). Dashed line is the TROWAL axis. Area of interest is Springfield Missouri (SGF).
In non-thundering snow events the average vertical advection of stability tends towards more positive values than it does in TSSN events (Figure 4.1.13). The average value for VADVS during the initial hour of a non-thundering event is $5.1 \times 10^6$ K mb$^{-1}$ s$^{-1}$, while TSSN events display an average value of $2.5 \times 10^6$ K mb$^{-1}$ s$^{-1}$. A t-test performed on this initial period yields a confidence level of 70% that these two sample means are in fact from different from one another.

At the hour of initial cloud-to-ground lightning and the corresponding mid-point of a non-thundering snow event the sample means are nearly identical. The average VADVS value occurring at this time period for a TSSN event is $1.7 \times 10^6$ K mb$^{-1}$ s$^{-1}$ similarly the average VADVS value for a non-thundering snow event valid at this time period is $1.2 \times 10^6$ K mb$^{-1}$ s$^{-1}$. The t-test
for these two sample means reveals only a 23% probability that the sample sets
are of a non-random nature.

A bimodal trend towards greater positive values is found in the average
model of TSSN events but is not found in the average model of non-thundering
events (Figure 4.1.13). However, a great deal of variation exists in the values of
VADVS during the periods encompassing these two extreme periods. Upon
individual analysis of the TSSN sample sets, it is obvious that this bimodal
period is not natural but rather a manifestation of statistical analysis. In fact this
bimodal phenomena only occurs in one of the eight TSSN sample sets.

4.1.3 STRCH

The stretching (STRCH) component of the stability tendency equation
tends to be positive with little change during the hours prior to lightning activity.
(Figure 4.1.14). Three hours prior to the indication of lightning activity the average value of STRCH is $0.5 \times 10^6 \text{ K mb}^{-1} \text{ s}^{-1}$ with a standard deviation of $1.7 \times 10^6 \text{ K mb}^{-1} \text{ s}^{-1}$. The best STRCH value for this period is $-1.8 \times 10^6 \text{ K mb}^{-1} \text{ s}^{-1}$ and the largest positive value is $3.1 \times 10^6 \text{ K mb}^{-1} \text{ s}^{-1}$.

A slight increase is evident at the second period in the average value of STRCH. At this time the average STRCH component is $0.7 \times 10^6 \text{ K mb}^{-1} \text{ s}^{-1}$ with a standard deviation of $2.1 \times 10^6 \text{ K mb}^{-1} \text{ s}^{-1}$. This increasing trend is evident in half of the sample sets with the other half illustrating little variation at all.

At the time of initial lightning activity the average STRCH value is similar to the previous hours. This average value is $0.6 \times 10^6 \text{ K mb}^{-1} \text{ s}^{-1}$ with a standard deviation of $1.7 \times 10^6 \text{ K mb}^{-1} \text{ s}^{-1}$. The initial hour of lightning activity exhibits the largest amount of variation associated with the stretching term. Values range from $-2.1 \times 10^6 \text{ K mb}^{-1} \text{ s}^{-1}$ to $2.7 \times 10^6 \text{ K mb}^{-1} \text{ s}^{-1}$. A slight tendency towards smaller positive values occurs between this hour and the following hour. This slight difference in average value is $0.1 \times 10^6 \text{ K mb}^{-1} \text{ s}^{-1}$.

A sharp trend towards negative stretching values occurs over the hours representative of lightning activity (Figure 4.1.14). The difference in average STRCH between the initial hour of lightning activity and the final hour of lightning activity is $-0.8 \times 10^6 \text{ K mb}^{-1} \text{ s}^{-1}$. A minimum value of $-4.5 \times 10^6 \text{ K mb}^{-1} \text{ s}^{-1}$ is recorded within the sample sets. However average values occurring throughout the period of lightning activity are only slightly negative. The average value for hours six and seven is $-0.2 \times 10^6 \text{ K mb}^{-1} \text{ s}^{-1}$ with the average STRCH value valid for the final period of lightning activity being $-0.6 \times 10^6 \text{ K mb}^{-1} \text{ s}^{-1}$.
After the cessation of lightning activity the stretching term tends towards positive values once again. The average value for stretching the hour after the final lightning strike is recorded is $0.1 \times 10^{-6}$ K mb$^{-1}$ s$^{-1}$ with a standard deviation of $2.4 \times 10^{-6}$ K mb$^{-1}$ s$^{-1}$. This trend towards positive values is present in five of the eight sample sets.

Speaking in terms of plan view charts of the stretching term, no consistent patterns are evident. With TROWAL events positive stretching values occur along the TROWAL axis. Ahead of the TROWAL axis negative STRCH values can usually be observed (Figure 4.15). Thundersnow events associated with warm fronts will be located along a positive STRCH gradient parallel to the frontal zone (Figure 4.16). Although, little impact is observed at the location of interest negative STRCH values can often be observed ahead of the warm front.

![Image](Figure 4.15 RUC_20 initial hour valid 0700 UTC 10 December 2003. θe (green contour) in Kelvin plotted with vertical stretching ($10^{-6}$ K mb$^{-1}$ s$^{-1}$; red contour). Dashed line is the TROWAL axis. Area of interest is Joplin (JLN).)
Figure 4.1.16 RUC_20 initial hour valid 0400 UTC 23 November 2003. \( \theta_e \) (green contour) in Kelvin plotted with vertical stretching \( \left( 10^{-6} \text{ K mb}^{-1} \text{s}^{-1} \right) \); red contour). Area of interest is Salina, Kansas (SLN).

Figure 4.1.17 Average vertical stretching \( \left( 10^{-6} \text{ K mb}^{-1} \text{s}^{-1} \right) \) occurring throughout the lifetime of a non-thundering snow event. Asterisk indicate midpoint of event.
Unlike the STRCH term in TSSN events the average STRCH term in non-thundering snow events tends toward negative values (Figure 4.1.17). In fact the average value recorded for the initial period of observation of a non-thundering event is \(-0.2 \times 10^6 \text{ K mb}^{-1} \text{ s}^{-1}\) with a standard deviation of \(1.1 \times 10^6 \text{ K mb}^{-1} \text{ s}^{-1}\). A tendency towards slightly more negative values occurs in the second and third hours of observation resulting in an average value of \(-0.6 \times 10^6 \text{ K mb}^{-1} \text{ s}^{-1}\) with a standard deviation of \(1.5 \times 10^6 \text{ K mb}^{-1} \text{ s}^{-1}\) for the hour preceding initial lightning activity.

Surprisingly the average STRCH value occurring at the mid-point of non-thundering snow events is identical to the previous hour. The average value for the mid-point of the event is \(-0.7 \times 10^6 \text{ K mb}^{-1} \text{ s}^{-1}\) with a standard deviation of \(1.7 \times 10^6 \text{ K mb}^{-1} \text{ s}^{-1}\). The most negative value at this time is \(-3.6 \times 10^6 \text{ K mb}^{-1} \text{ s}^{-1}\) and the greatest recorded value is \(1.5 \times 10^6 \text{ K mb}^{-1} \text{ s}^{-1}\). A continued tendency towards more negative average values occurs between the mid-point and the following hour.

The average value of STRCH occurring immediately after the mid-point is \(-1.7 \times 10^6 \text{ K mb}^{-1} \text{ s}^{-1}\) with a standard deviation of \(2.5 \times 10^6 \text{ K mb}^{-1} \text{ s}^{-1}\). Beyond this period the stretching component of the stability tendency equation tends towards zero but remains negative. The average value for the final period of observation of the STRCH term is \(-0.5 \times 10^6 \text{ K mb}^{-1} \text{ s}^{-1}\) with a standard deviation of \(2.2 \times 10^6 \text{ K mb}^{-1} \text{ s}^{-1}\).

Non-thundering snow events tend to be more negative in terms of the STRCH term of the stability tendency equation. This is also evident in plan view
charts of this term. The area of interest is often located along the TROWAL axis or ahead of a warm front in a region of negative STRCH values (Figure 4.1.18).

Figure 4.1.18 RUC_20 initial hour valid 1000 UTC 22 January 2005. Example of an average STRCH pattern \(10^6 \text{ K m}^{-1} \text{s}^{-1}\) occurring along a warm front during a non-thundering snow event. The (green contour) in Kelvin plotted with vertical stretching (red contour). Area of interest is Green Bay (GRB).
Figure 4.1.19 Comparison of average STRCH ($10^6$ K mb$^{-1}$ s$^{-1}$) values over the lifetime of TSSN (red contour) and non-thundering snow events (Blue contour). Asterisks represent period of lightning observation and mid-event period respectively.

It is apparent upon comparing the average STRCH values of TSSN events to non-thundering events that STRCH tends to be more positive in TSSN events than in non-thundering snow events (Figure 4.1.19). In fact the STRCH component in TSSN events remains positive through the period of initial lightning activity only becoming negative after lightning has been occurring for an hour. The STRCH term in non-thundering events remains less than zero through the entire event.

The average value for STRCH found in TSSN events for the initial hour of observation is $0.5 \times 10^6$ K mb$^{-1}$ s$^{-1}$ while the average value for non-thundering events for the same period is $-0.2 \times 10^6$ K mb$^{-1}$ s$^{-1}$. A two-tail t-test reveals a confidence interval of 64% that the STRCH component in non-thundering events is in fact less than the mean found in TSSN events. Therefore the STRCH term is
not statistically different between TSSN events and non-thundering snow events during the initial hour of observation.

Comparing the mid-point of the average non-thundering snow event and the hour of initial lightning observation associated with the average TSSN event reveals that the STRCH term is greater in TSSN events than the average value in non-thundering events (Figure 4.1.19). The average value of the STRCH term for the initial hour of lightning activity in a TSSN event is $0.7 \times 10^{-6} \text{K mb}^{-1} \text{s}^{-1}$ while the average value found in non-thundering events is $-0.7 \times 10^{-6} \text{K mb}^{-1} \text{s}^{-1}$. The two-tailed t-test confirms with an 85% confidence level that STRCH term is in fact greater in TSSN events for this period. However, this confidence level is still not statistically significant.

Little variation with time is seen in the STRCH term in either non-thundering or TSSN events. Both types of event do exhibit a tendency toward lesser positive values or lesser negative values the hour after the mid-point or after the initial hour of lightning activity. Also both cases tended toward more positive values leading towards the end of the observation period. Additionally, it is apparent that the STRCH term is more favorable in non-thundering events than in TSSN events since ideally STRCH should be a negative value to infer instability. Unfortunately, it cannot be shown statistically that this is in fact the case.
4.1.4 Total Tendency

The total stability tendency equation (STABTDY) tends towards more positive values over the first three hours of observation (Figure 4.1.20). Initially the average STABTDY value is 0.9x10^{-6} \text{ K} \text{ mb}^{-1} \text{ s}^{-1} with a standard deviation of 2.0x10^{-6} \text{ K} \text{ mb}^{-1} \text{ s}^{-1}. Over the course of three hours the average STABTDY value increases to 5.6x10^{-6} \text{ K} \text{ mb}^{-1} \text{ s}^{-1} with a standard deviation of 8.2x10^{-6} \text{ K} \text{ mb}^{-1} \text{ s}^{-1}. A great deal of variation exists among the sample sets the hour prior to initial lightning activity. The maximum value for this period is 21.2x10^{-6} \text{ K} \text{ mb}^{-1} \text{ s}^{-1} and the minimum recorded value is -1.4x10^{-6} \text{ K} \text{ mb}^{-1} \text{ s}^{-1}. After the third hour of observation the STABTDY tends towards lower positive values.

At the hour of initial lightning activity the average STABTDY is 3.2x10^{-6} \text{ K} \text{ mb}^{-1} \text{ s}^{-1} with a standard deviation of 6.6x10^{-6} \text{ K} \text{ mb}^{-1} \text{ s}^{-1}. The minimum value for this period is -9.9x10^{-6} \text{ K} \text{ mb}^{-1} \text{ s}^{-1} with a maximum value of 12x10^{-6} \text{ K} \text{ mb}^{-1} \text{ s}^{-1}. The
average stability tendency continues towards lesser positive values over the first two hours of lightning activity. The average STABTDY tends towards a more positive value the final hour of observed lightning.

This value is $2.1 \times 10^6$ K mb$^{-1}$ s$^{-1}$ with a standard deviation of $7.7 \times 10^6$ K mb$^{-1}$ s$^{-1}$. Once again the tendency becomes less positive the hour after the final lightning observations are recorded. During the hours after the cessation of lightning activity STABTDY tends toward greater positive values until the end of the observation period.

Overall the stability tendency equation yields a propensity towards stability throughout the period of observation. Two periods of increased tendency towards stability are apparent when the average stability tendency is plotted (Figure 4.1.20). The first node of increased stability tendency occurs the hour prior to initial observation of lightning activity. A tendency towards greater stability, but at a slower pace occurs throughout the hours of lightning observation. After the final hour of lightning observation the second mode of increased stability tendency occurs.

To further support the stability tendency's propensity towards stability a second experiment was performed. This second experiment was executed in the same manner as the previous experiment (see Fig. 3.3.1) but at a layer, which is 50 millibars higher in the atmosphere than the previous experiment. It was hoped that by raising the level of analysis by 50 millibars that some trend toward instability may have been revealed by looking solely in the warm air above the top of the inversion.
However, this second experiment revealed the same results as our initial experiment: a tendency towards stability. Yet, the rates of stabilization were cut roughly in half. Nevertheless, this conclusion contrasts with Ebert (2004) who performed similar experiments on a single TSSN event and found a tendency towards instability in the TROWAL air stream above the frontal inversion. However, it should be noted that the Ebert (2004) study also considered the diabatic heating term that is ignored in this particular study. Additionally Ebert's (2004) study examined only one TSSN event that was associated with a TROWAL, while this study examined 8 separate events, which took into consideration both events occurring northwest and northeast of an extratropical cyclone.

![Figure 4.1.21 Comparison of average static stability tendency ($10^6$ K mb$^{-1}$ s$^{-1}$) components during a TSSN event. Differential temperature advection (blue contour), vertical advection of stability (red contour), vertical stretching (green contour) and Stability tendency (magenta contour).](image-url)
In the hours preceding initial lightning activity DADTP is the only component tending to be negative (Figure 4.1.21). This component does however become less negative as time approaches the hour of initial lightning activity. The VADVS component tends to be a small positive number and STRCH component tends to be a greater positive number. Although DADTP maybe the component that forces STABTDY to take on smaller positive values, it appears that the total stability tendency equation follows the tendency of the VADVS term (Figure 4.1.21). As VADVS increases so does the stability tendency; as STABTDY decreases so does VADVS.

During the hours of lightning activity both STRCH and DADTP tend towards smaller values, as does the total stability tendency equation. It appears that the STRCH component only decreases marginally, $1.3 \times 10^{-6}$ K mb$^{-1}$ s$^{-1}$ through the hours of lightning activity. While, DADTP changes by $2.5 \times 10^{-6}$ K mb$^{-1}$ s$^{-1}$ over the course of four hours. Referring to the above figure this decrease in DADTP seems to impact the STABTDY significantly. It should also be noted that VADVS also decreases appreciably.

Table 4.1.1 Correlation of individual components and the total static stability tendency equation.

<table>
<thead>
<tr>
<th></th>
<th>DADTP</th>
<th>VADVS</th>
<th>STRCH</th>
<th>STABTDY</th>
</tr>
</thead>
<tbody>
<tr>
<td>DADTP</td>
<td>1</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>VADVS</td>
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<td>1</td>
<td></td>
<td></td>
</tr>
<tr>
<td>STRCH</td>
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<td>-0.2</td>
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<td></td>
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<tr>
<td>STABTDY</td>
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<td>0.4</td>
<td>0.3</td>
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</tbody>
</table>

According to the Pearson correlation DADTP exhibits the best positive correlation with STABTDY (Table 4.1.1). This correlation coefficient is however not that great and leans closer toward zero than one, inferring only a weak correlation exists between STABTDY and DADTP. An almost equal correlation to
that found between STABTDY and DADTP exists between VADVS and STABTDY. This correlation coefficient is 0.397 illustrating that VADVS holds almost the same weight in influencing the total stability tendency, as does DADTP.

Also referring to Table 4.1.1 an inverse correlation exists between VADVS and DADTP. This inverse relation is greater than the correlation found between either components in relation to the total stability tendency. This correlation may be partially due to the appearance, in both VADVS and DADTP, of the change in potential temperature with respect to the change in pressure.

Figure 4.1.22 Average static stability tendency \((10^6 \text{ K mb}^{-1} \text{ s}^{-1})\) occurring throughout the lifetime of a non-thundering snow event. Asterisks indicate midpoint of event.

At the initial hour of observation the average stability tendency is \(3.5\times10^6\) K mb\(^{-1}\) s\(^{-1}\) with a standard deviation of \(3.7\times10^6\) K mb\(^{-1}\) s\(^{-1}\). A trend towards lesser values, towards zero, is evident over the first three hours of observation in the model representation of the STABTDY in non-thundering snow events (Figure
This trend results in a net change of $2.2 \times 10^6$ K mb$^{-1}$ s$^{-1}$ over this initial three-hour period. Six of the seven sample sets displayed this trend towards lower values during this particular three-hour period of observation. It should also be noted that a great deal of variation exists within the sample sets valid for this period. The standard deviation ranges from $3.3 \times 10^6$ K mb$^{-1}$ s$^{-1}$ valid for the first hour of observation to $5.3 \times 10^6$ K mb$^{-1}$ s$^{-1}$ in the second hour of observation.

At the mid-point of a non-thundering snow event the average stability tendency is $-0.5 \times 10^6$ K mb$^{-1}$ s$^{-1}$ with a standard deviation of $2.3 \times 10^6$ K mb$^{-1}$ s$^{-1}$. From this point on, the average stability tendency remains negative reaching a minimum average value of $-2.1 \times 10^6$ K mb$^{-1}$ s$^{-1}$ at the fifth hour of observation. A slow tendency towards higher values (i.e. closer to zero) of stability occurs after this period until the end of the observation period (Figure 4.1.22). Again six of the seven sample sets follow this trend reinforcing the validity of this stability tendency model for non-thundering snow events.
As mentioned in the previous section, the average stability tendency occurring in non-thundering snow events trends toward smaller values over the course of the first five hours of observation, at which point the tendency shifts towards more positive values (Figure 4.1.22). Comparing the individual components to the total stability tendency reveals that DADTP is the dominating factor in the stability tendency of non-thundering events (Figure 4.1.23). Whereas VADVS and STRCH show little variation in time over the first four hours of observation, both exhibit change at hour five.

To support this hypothesis a correlation was performed on the three terms of the stability tendency equation with respect to the total stability tendency of a non-thundering event. It is evident (Table 4.1.2) that DADTP is indeed the dominating component in the stability tendency of non-thundering cases with a
correlation coefficient of 0.7. Like the results found with TSSN events this is only a weak correlation at best.

In regards to the correlation of VADVS and STRCH with the total stability tendency only a weak inverse relationship is evident. In fact almost no correlation is inferred by the correlation coefficient -0.2 resulting from comparing VADVS with STABTDY and even less correlation is found between STRCH and STABTDY. This evidence further supports the dominating nature of the DADTP component of STABTDY.

Table 4.1.2 Correlation of individual components and the total static stability tendency equation.

<table>
<thead>
<tr>
<th></th>
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<th>VADVS</th>
<th>STRCH</th>
<th>STABTDY</th>
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<tbody>
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<td>-0.01</td>
<td>1</td>
</tr>
</tbody>
</table>

Figure 4.1.24 Comparison of average static stability tendency (10^4 K mb^{-1} s^{-1}) values over the lifetime of TSSN (red contour) and non-thundering snow events (Blue contour). Asterisks represent period of lightning observation and mid-event period respectively.
Initially STABTDY is greater in non-thundering snow events than in TSSN events (Figure 4.1.24). A t-test supports this statement with as 92% confidence level that STABTDY is greater in non-thundering events than in TSSN events. Over the initial three hours of observation STABTDY decreases in non-thundering snow events where as STABTDY increases in TSSN events for the same time period.

Comparing the mid-point of a non-thundering snow event to the hour of initial lightning activity in TSSN events it is apparent that STABTDY is greater in TSSN events than in non-thundering events (Figure 4.1.24). This statement can only be made with 82% confidence therefore it cannot be claimed to be statistically significant. Both types of events exhibit a tendency towards smaller values over the next hour.

The stability tendency of non-thundering events slowly tends towards zero from the fifth-hour of observation till the end of the observation period. While STABTDY associated with TSSN events remains steady for one additional hour before tending toward more positive values once again (Figure 4.1.24). Again it should be noted that STABTDY tends to be less than zero in non-thundering events from the mid-point through the end of the observation period. At no time does the average value of STABTDY associated with TSSN events ever reach zero or fall below zero.

It was found that in both TSSN events and non-thundering events DADTP correlated the best with changes in STABTDY. Statistical analysis shows that this is the case in TSSN events with a confidence level of 0.4, while in non-thundering cases this confidence level increased to 0.7. This suggests not only does DADTP
dominate in both events but also it dominates to a greater level in non-thundering events.

Unlike the tendency of VADVS found in non-thundering events VADVS is very dynamic in TSSN events. The vertical advection of stability also weakly correlates with STABTDY in TSSN events, falling only 0.02 behind the correlation between STABTDY and DADTP. Additionally, VADVS associated with TSSN events shares an inverse correlation with DADTP, which is not found in non-thundering events.

It is also evident that in both TSSN and non-thundering events STRCH remains fairly constant with time. Also in both cases STRCH tends toward smaller values after the mid-point of and event for a short period prior to tending once again to greater values. In reference to TSSN events STRCH correlates only weakly with STABTDY while in regards to non-thundering events little to no correlation is evident.
4.2 Total Totals

The Total Totals index (hereafter TT):

\[ TT = T_{A850} + T_{d850} - 2T_{500} \]

is a measure of atmospheric stability which takes into account the temperature and dewpoint of the boundary layer and the midtropospheric temperature. Term A in equation 4.2.1 is the temperature at 850mb, term B is the dewpoint at 850mb and term C is the temperature at 500mb.

During the three hours prior to cloud-to-ground lightning activity average values for the TT index ranged from 22 to 49. The average TT value during this period was 36 with a standard deviation of 8.1. Little variation is noted in the average values of this index over the first three hours of the event (Figure 4.2.1).
Comparing the TT values prior to the first indication of lightning activity to the period of initial lightning activity resulted in no statistical significance. However, half of the individual sample sets exhibited a decrease in stability prior to lightning activity.

At the hour of initial lightning activity the average TT value is 36 with a standard deviation of 8.9. The maximum TT value seen at the time of initial lightning activity is 48 and the lowest recorded value is 24. Examination of the average TT values occurring through the period when lightning is taking place reveals a slow decrease with time of about 1 TT unit per hour (Figure 4.2.1). Individual analysis of the data set over this time period offers little insight. Four of the sample sets exhibit a decrease in value over the time period, two actually exhibit an increase in value and two of the sample sets remain steady.

The hours following lightning activity yield well-behaved TT values; little variation is seen among the sample sets over time. Average values for these hours range from 34 to 32. The average standard deviation for TT values after lightning has ceased is 7.1.
Figure 4.2.2 RUC_20 initial hour valid 0800 UTC 10 December 2003. A typical pattern of TT observed during a TSSN event. $\theta_e$ (green contour) in Kelvin plotted with Total Totals (red contour). Area of interest is Joplin (JLN).

In a plan view perspective cases, involving a TROWAL appeared to occur along a gradient of TT values (Figure 4.2.2). Most often the location of lightning activity would occur along the gradient of TT and the TROWAL axis. Additionally, it should be noted that in most cases the location of interest is not a TT maximum but rather falls along a gradient towards lower values. Higher values of TT occur closer to the TROWAL. Curiously, as the TROWAL approaches our area of interest an increase of TT values is not always the case.

Considering the equation representing the TT (equation 4.2.1) it is apparent that the resulting value is dependent on the 850-mb temperature and dewpoint as well as the temperature at 500 mb. When holding the 850-mb temperature and dewpoint constant and warming the 500-mb temperature the TT value will decrease. If the temperature at 500 mb is held constant and either the 850-mb temperature or dewpoint is decreased the final value for TT also
decreases. However, warming the 500-mb temperature is a more efficient way to decrease the TT values. In a series of experiments performed on the TT equation it was observed that warming the 500-mb temperature while holding the other two terms constant is much more efficient (by a factor of two) at lowering the TT value than warming either the 850-mb dew point or temperature by the same amount (while holding the 500-mb temperature constant).

With this in mind, consider the level of the TROWAL during an elevated convective event. If the TROWAL is located at 500 mb above the area of interest, the TT value for that particular area will decrease as the TROWAL axis approaches. However, if the TROWAL is located at a lower level, which is most often the case little change will be noted since the TT equation does not take directly into account the layer between 850 mb to 500 mb. Therefore, only TSSN events associated with a TROWAL around 500 mb will see a decrease in TT values. If the location of the event is located closer to and northeast of the cyclone the TROWAL may actually increase the TT values.

Plan view charts of TT in TSSN events associated with a warm front once again place the area of interest along a gradient of TT values. Similar to the TSSN events associated with a TROWAL the area of interests is not always the location of the greatest TT value. Additionally it appears that the gradient of TT values parallels the warm frontal boundary.
Throughout the first three hours of a non-thundering snow event the TT values ranged from 16 to 42. The average value during this period is 31 with a standard deviation of 9.5. Similar to the TSSN events little variation is noted in the average values of this index over the first three hours of the event (Figure 4.2.3). When comparing the non-thundering TT sample sets individually for the first three-hour period, six of the seven cases also exhibit little to no variation in TT value with time.

Midway through a non-thundering snow event TT values range from 26 to 43. The average value for this time period is 31 with a standard deviation of 9.3. No statistical significance is evident in TT values, between the first three hours of a non-thundering event as it transitions to the mid-event period. In fact, the difference between the average value of the hour prior to the midway value and the mid-way value itself is only 0.6.
Again little difference is noted between the average mid-event TT value and the average value occurring over the final three hours of observation (Figure 4.2.3). The average value occurring during the final three hours of non-thundering events is 31. The standard deviation of this average is 7.8. Total totals values occurring for this period range from 17 to 43. Again, little variation with time is seen when comparing the individual sample sets of TT over the final three hours of a non-thundering event.

In plan view charts non-thundering heavy snow events also can occur along gradients of TT. However, the gradients in TT associated with non-thundering snow events are not as intense as those found in TSSN events. Additionally, when a TROWAL is involved it is not always as close in proximity to the area of interest as in TSSN events.

Figure 4.2.4 RUC_20 initial hour valid 1200 UTC 14 February 2005. A typical TT pattern found in non-thundering snow events. θ (green contour) in Kelvin plotted with Total Totals (red contour). Area of interest is Green Bay (GRB).
In the case of warm frontal non-thundering snow events the line of equal TT values like found TSSN events also parallels the frontal boundary (Figure 4.2.4). But once again the gradient of TT is not quite as intense as that found in TSSN events. Additionally, TT values do not run parallel along the warm frontal spatially as long as found in TSSN events.

No real plan view features in TT values distinguish TSSN events from non-thundering snow events. The tighter gradient in TT values related to TSSN events could readily occur in non-thundering events if a significant boundary exists. A recurring axis of instability was not found in either TSSN or non-thundering events.

![Figure 4.2.5 Comparison of average Total Totals values over the lifetime of TSSN (red contour) and non-thundering snow events (blue contour). Asterisks represent period of lightning observation and mid-event period respectively.](image)

At first glance it appears that the TT index maybe a good discriminator for TSSN events (Figure 4.2.5). Average TT values at the beginning of observation of
TSSN events are 36 and average TT values found at the beginning of non-thundering events only approach 31. However, after performing the t-test on the sample means only a 66% confidence interval is obtained stating that there is only a slim chance that the sample sets are in fact different at this time period.

Turning our attention to the mid-point of a non-thundering snow event and the hour of initial lightning activity in a TSSN event it is apparent that the average TT value at this time period is greater in TSSN events than in non-thundering events (Figure 4.2.5). But once again the t-test reveals only a 67% chance that the sample means are in fact different from one another.

Although, at first glance average TT values appear greater in TSSN events than in non-thundering events for all periods this hypothesis cannot be supported statistically. Therefore, the TT index alone is not a good indicator of whether or not a snow event will feature lightning or merely yield snow.
### 4.3 K-index

The likelihood for showers and thunderstorms can be assessed through the K-index:

\[
K = (T_{850} - T_{500}) + D_{850} - (T_{700} - D_{700})
\]

where term A is the difference between the 850-mb temperature and the 500-mb temperature, term B is the 850-mb dewpoint and term C is the difference between the 700-mb temperature and 700-mb dewpoint. As the K-index value increases the potential is potential for thunderstorm development becomes greater.

Figure 4.3.1 Average K-index occurring throughout the lifetime of a TSSN event. Asterisks indicate period of lightning activity.
During the three hours prior to cloud-to-ground lightning activity values for the K-index range from -1 to 18. The average value for this period is 11 with a standard deviation of 1.6. A gradual rise in the average K-index value is noted over the first three hours of the event (Figure 4.3.1). This negligible increase is about 1°C/hr.

Comparing the K-index values for the initial period of observation to the period of initial lightning activity resulted in a 5.4°C increase inferring a more favorable environment for precipitation. Five of the eight individual sample sets exhibited this increase in value prior to lightning activity.

At the hour of initial lightning activity the average K-index is 14.5 with a standard deviation of 7.0. The maximum K-index value recorded at this time is 26 and the minimum value is 6. Examination of the average K-index values occurring through the period when lightning is taking place slowly drop by a value of .5°C/hr. Individual analysis of the data set over this time period offers little insight, as the individual sets exhibited a variety of behaviors beyond this point.

The hours subsequent to lightning activity yield a slow decrease in average K-index values (Figure 4.3.1). K-index values for these hours range from 2 to 22. The average value for this time period is 11 with a standard deviation of 8.0. Although a great deal of variation exists during this period all of the TSSN sample sets exhibited a gradual and sometimes sharp decrease in the K-index through the period.

Plan view analysis of the K-index often depict a gradient of the K-index aligned with the θ_e ridge or TROWAL (Figure 4.3.2). An increase of the K-index is
apparent with the approaching TROWAL, which coincides with the period of initial lightning activity (Figure 4.3.2 thru Figure 4.3.5). When the TROWAL axis is overhead the K-index decreases and this tendency continues as the TROWAL air-stream departs.

Figure 4.3.2 RUC_20 initial hour valid 0600 UTC 10 January 2003. A typical K-index pattern observed during a TSSN event. $\theta_e$ (green contour) in Kelvin plotted with K-index (red contour). Dashed line is the TROWAL axis. Area of interest is Joplin (JLN).
Figure 4.3.3 Same as figure 4.3.2 except valid for 0700 UTC 10 January 2003.

Figure 4.3.4 Same as 4.3.2 except valid for 0800 UTC 10 January 2003.
At no time is the area of interest a local maximum or minimum. Often the greater K-index values are observed equatorward of the TROWAL axis or $\theta_e$ ridge. The gradient of K-index will often follow the isotherms of $\theta_e$ in both TROWAL and warm frontal events. Curiously events associated with the TROWAL do exhibit an axis of instability with the area of interest located in the vicinity of the TROWAL axis and the K-index axis.
During the first three hours of a non-thundering snow event the K-index values ranged from -6 to 17. The average value during this period is 9 with a standard deviation of 0.2. Little variation is noted in the average K-index over the first three hours of a non-thundering snow event (Figure 4.3.6). In fact, average values varied less than a half of a degree during this time.

A slight decrease is noted in the average K-index value between the hours preceding the mid-event hour and the mid-event hour itself (Figure 4.3.6). This average decrease is 1.4°C and this decrease continues at approximately the same rate leading up to the end of the event. Although some variation does exist in the individual sample sets this decrease is evident in all but one of the non-thundering cases.

Similar features exist in the plan view charts of K-index in non-thundering snow events as seen in plan view charts of TSSN events. The area of interest is...
often along a K-index gradient, which is aligned with the $\theta_e$ ridge or TROWAL (Figure 4.3.7). However, the gradient is often not as tight as those found TSSN events and non-thundering events lack the axis of instability present in TSSN events. Additionally the equal lines of the K-index are not always congruent with the isotherms related to the TROWAL or $\theta_e$ ridge as found in TSSN events.

![Figure 4.3.7 RUC_20 initial hour valid 1200 UTC 14 February 2005. A typical K-index pattern observed during non-thundering snow events. $\theta_e$ (green contour) in Kelvin plotted with K-index (red contour). Area of interest is Green Bay (GRB).](image)

According to Figure 4.3.8 both the TSSN and non-thundering snow events share similar average K-index values for the initial hour of observation. In fact the average K-index value for TSSN events at the initial hour of observation is 8 and the average value at this time period for non-thundering events is 7. A two tailed t-test reveals only a 1% probability that the sample means are not identical between non-thundering and TSSN events for the initial hour of observation.
A noteworthy phenomenon associated with TSSN events and not with non-thundering events is the trend towards greater values of the K-index prior to the first recorded lightning activity or mid-point of the event (Figure 4.3.8). Non-thundering cases lack this feature in all but one of the sample sets. Whereas five of the eight TSSN sample sets include this rise in the K-index prior to lightning activity.

![Graph showing comparison of average K-index values over the lifetime of TSSN (red contour) and non-thundering snow events (blue contour). Asterisks represent period of lightning observation and mid-event period respectively.](image)

Figure 4.3.8 Comparison of average K-index values over the lifetime of TSSN (red contour) and non-thundering snow events (blue contour). Asterisks represent period of lightning observation and mid-event period respectively.

Also apparent (Figure 4.3.8) is that the average value of the K-index in TSSN events at the hour of initial lightning activity when compared to the mid-point value of a non-thundering event is much greater. In fact the average value at the time of initial lightning activity is 14.5, whereas the average mid-point value on non-thundering event is only 7.5. This seems like a significant difference.
however the two-tailed t-test only suggests a 85% probability that the sample means are not random in nature.

Once again, at first glance it appears that the K-index would be a good discriminator between TSSN events and non-thundering events (Figure 4.3.8). However, statistical analysis indicates that the sample means are not quite different enough to support this hypothesis. Trends in the K-index may be a discriminator between the two phenomena as the K-index increases in TSSN events up to the time of initial lightning activity, whereas the K-index in non-thundering events varies only slightly with time.

Plan view patterns appear similar between non-thundering and TSSN events. Both types of events occur along a gradient of the K-index, which often parallels the isotherm of $\theta_e$. However TSSN events occur along a tighter $\theta_e$ gradient, which intersects the TROWAL air stream. Additionally, TSSN event often exhibit an axis of K-index values in the vicinity of the TROWAL axis.
4.4 Mid-level Lapse Rate

The mid-level lapse is the change of temperature over the depth of a defined atmospheric layer. This study examines the mid-level lapse as defined:

$$LR = \frac{(T_{700} - T_{500})}{(H_{700} - H_{500})}$$

(4.4.1)

where the numerator represents the difference of the 700-mb temperature and the 500-mb temperature and the denominator represents the difference in height of the 700-mb level and the 500-mb level. The greater the lapse rate, more negative, is the less stable the greater the potential is for convective development.

During the three hours prior to cloud-to-ground lightning activity mid-level lapse rates ranged from -4.1 K km$^{-1}$ to -7.4 K km$^{-1}$. The average mid-level lapse rate during this period is -6.4 K km$^{-1}$ with a standard deviation of 1.0 K km$^{-1}$.
Little variation is noted in the average mid-level lapse rate over the first three hours of the event (Figure 4.4.1).

Comparing the mid-level lapse rates between the hours prior to the first indication of lightning activity to the time of initial activity illustrates that a slight increase in stability occurs prior to the onset of lightning (Figure 4.4.1). This increase in stability is reflected in a three-tenths of a degree decrease in mid-level lapse rate. Six of the TSSN cases exhibited this decrease in mid-level lapse rate to some degree.

At the hour of initial activity the average mid-level lapse rate is -6.3 K km\(^{-1}\) with a standard deviation of 1.0 K km\(^{-1}\). The maximum lapse rate seen at the time of initial lightning activity is -6.8 K km\(^{-1}\) and the least recorded value -4.1 K km\(^{-1}\). On average the mid-level lapse rate decreases at a rate of 0.2 K hr\(^{-1}\) during the hours that lightning activity is recorded. The standard deviation tends to increase with TSSN events during the periods in which lightning is occurring; therefore this stabilizing trend is not always apparent in a case-to-case comparison.

The hours subsequent to lightning activity are not as well behaved as the Total Totals index, with a standard deviation approaching -2.2 K km\(^{-1}\) by the end of the period. Typical values for these hours range from -3.5 K km\(^{-1}\) to 8.1 K km\(^{-1}\). Part of the variation in mid-level lapse rates occurring during the final hours of a TSSN event could be due to noise generated by the determination of the observation termination. Some events are longer in duration than others and some of the noise could be a second weather system beginning.
Looking at the plan view charts of mid-level lapse rates in TSSN events featuring a TROWAL it is apparent that an axis of greater mid-level lapse rates penetrates the occluded quadrant of the associated cyclone (Figure 4.4.2, 4.4.3, 4.4.4 and 4.4.5). The area of interest will most often be located at the intersection of the TROWAL axis and the axis of the instability. A second scenario occurs when the axis of instability intersects the northern periphery of the TROWAL airstream.

Figure 4.4.2 RUC_20 initial hour valid 0500 UTC 10 December 2003. A typical pattern of mid-level lapse rates occurring during a TSSN event. θ (green contour) in Kelvin plotted with mid-level lapse rates in K km\(^{-1}\) (red contour). Dashed line is the TROWAL axis, Solid black line is mid-level lapse rate axis. Area of interest is Joplin (JLN).
Figure 4.4.3 Same as figure 4.4.2 except valid for 0600 UTC 10 December 2003.

Figure 4.4.4 Same as figure 4.4.2 except valid for 0700 UTC 10 December 2003.
In warm frontal TSSN events, no reoccurring axis of stability is present. However, a closed isoline of mid-level lapse rates along and ahead of the warm front is usually present. Additionally, the axis of instability will be positioned southwest of the area of interest, penetrating into the cyclone (Figure 4.4.6). The mid-level lapse rates are also not as great as those found in TROWAL TSSN events.
In either case the area of interest is not a local maximum value nor is it a local minimum value. Similar to the hypothesis presented with the Total Totals index when the 500-mb temperature is increased the mid-level lapse rate decreases. Likewise when the 700-mb temperature is increased and the 500-mb temperature is held constant, disregarding any change in the thickness of the 500 to 700-mb layer, the mid-level lapse rate increases. Therefore depending on the proximity of an event to the center of an occluded cyclone the height at which the TROWAL is located may dictate the tendency of the mid-level lapse rate.

If the location of interest intersects the TROWAL air-stream at 500 mb the mid-level lapse rate will possibly be more stable than the surrounding areas. However, if the area of interest intersects the TROWAL air-stream at 700 mb a local increase in mid-level lapse rate (decrease in stability) maybe seen.
Additionally, as the TROWAL air-stream departs from a location of interest the axis of instability shifts as well (Figure 4.4.2 thru Figure 4.4.5). The previously mentioned figures illustrate that the instability axis shifts as the TROWAL air stream arrives to and departs from the area of interest. As the TROWAL air stream exits to the east of the area of interest, that area may experience additional warm air advection or cold air advection resulting in a change of the mid-level lapse rate at the point of interest. Whether this is a positive or negative result, is dictated by the level at which the TROWAL is initially found.

![Graph showing mid-level lapse rate](image)

Figure 4.4.7 Average mid-level lapse rate (K km\(^{-1}\)) occurring throughout the lifetime of a non-thundering snow event. Asterisk indicates the midpoint of the event.

During the first three hours of a non-thundering event mid-level lapse rates ranged from -2.2 K km\(^{-1}\) to -6.5 K km\(^{-1}\). The average mid-level lapse rate during this period is -4.6 K km\(^{-1}\) with a standard deviation of ± 0.5 K km\(^{-1}\). Little
variations is noted in the average mid-level lapse rate over the first three hours of a non-thundering snow event (Figure 4.4.7).

Comparing mid-level lapse rates occurring during the first three hours of a non-thundering event to the average mid-level lapse rate occurring mid-event reveals little change. Although, Figure 4.4.7 shows a rise in mid-level lapse rate this rise on average is only -0.2 K km\(^{-1}\), which is less than one standard deviation from average values occurring prior to the mid-point. Four of the non-thundering cases exhibited this increase in mid-level lapse rate to some degree.

Little change in the average mid-level lapse rate is noted in the hours subsequent to the mid-point of an event (Figure 4.4.7). The maximum mid-level lapse rate for this period is -6.2 K km\(^{-1}\) and the minimum value recorded for this period is -1.4°C/km. The average value occurring within the hours leading up to the end of the observation period is -4.7 K km\(^{-1}\) with a standard deviation of 1.6 K km\(^{-1}\).

Much like the mid-level lapse rate plan view fields associated with TSSN events, certain features reoccur with non-thundering events. For instance the area of interest is not a local maximum or minimum. However, unlike TSSN events no axis of instability is recorded in the non-thundering events in this study. Rather non-thundering events occurring both with a TROWAL or a frontal boundary feature closed isolines of mid-level lapse rate values often ahead of the boundary or TROWAL air-stream (Figure 4.4.8).
Figure 4.4.8 RUC_20 initial hour valid 1200 UTC 22 January 2005. A typical pattern of mid-level lapse rate observed during a non-thundering snow event, $\theta_e$ (green contour) in Kelvin plotted with mid-level lapse rates in K km$^{-1}$ (red contour). Area of interest is Green Bay (GRB).

Figure 4.4.9 Comparison of average mid-level lapse rate (K km$^{-1}$) over the lifetime of TSSN (red contour) and non-thundering snow events (blue contour). Asterisks represent period of lightning observation and mid-event period respectively.
Directly comparing mid-level lapse rates between TSSN events and non-thundering events reveals that mid-level lapse rates in TSSN events are much greater than in non-thundering events (Figure 4.4.9). In fact the average mid-level lapse rate occurring 3 hours prior to initial reports of lighting is -6.5 K km\(^{-1}\) where the average mid-level lapse rate at the beginning of a non-thundering event is only -4.7 K km\(^{-1}\). A t-test was performed on the sample means resulting in a p-value of .1 supporting with over a 95% confidence level that this difference is statistically significant.

As stated earlier the average mid-level lapse rate occurring at the hour of initial lightning activity is -6.0 K km\(^{-1}\) and the average mid-level lapse rate occurring at the midpoint of a non-thundering case is -4.9 K km\(^{-1}\). A statistically significant difference occurs when comparing these two periods as well. Again, a t-test was performed resulting in a confidence level of 95% that the mean values do not belong to the same sample set.

It appears that the mid-level lapse rate is a good discriminator between TSSN events and non-thundering events. The critical value for a midlevel lapse rate would be -6.0 K km\(^{-1}\) ± 0.98 K km\(^{-1}\). It is important to note that achieving this threshold value alone does not guarantee that a snow event will exhibit lightning activity. For a TSSN event to occur this threshold should be met with the proper background flow and moisture content of the air column.
4.5 CAPE

Convective available potential energy (CAPE) is a controversial subject when referring to convective snow. Generally this is due to the theory that convective bands form in environments that are primarily supported by slantwise rather than purely upright ascent. However, some of the more robust TSSN events are associated with moderate values of CAPE. In terms of this study CAPE is present at various time periods in six of the eight observed TSSN events.

Although CAPE is present in more than half of the observed events there is insufficient data to support a pattern to its occurrence. In some instances CAPE is present prior to lightning activity while in other instances it maybe present the hour of initial lightning observation or even after lightning activity has ceased. When CAPE is present in a TSSN event it is generally less than 100 J kg\(^{-1}\). In rather robust cases values of up to 311 J kg\(^{-1}\) have been observed. Due to the small sample size and discontinuous, sporadic nature of CAPE this author has little confidence in any statistical analysis of this phenomena.

However, CAPE is even more scarce in non-thundering snow events than in TSSN events. Only two of the seven non-thundering events displayed any evidence of CAPE in this study. When CAPE is observed, it is generated by lifting a parcel from a level of 550mb or greater. At this level any potential energy generated is of little use in producing convective elements.

In this study CAPE is not considered a good discriminator of TSSN against non-thundering events. The following two reasons are offered to support this statement. First and foremost, CAPE is present in both non-thundering and TSSN sample sets. Secondly, due to the random nature of CAPE in TSSN events
it cannot be stated that CAPE is present in any particular time period. Perhaps, analysis of a larger more statistically sound data set may elude better insight to this phenomena's role in convective snowfall.

### 4.6 MU Lifted Index

![Graph showing the average most unstable lifted index occurring throughout the lifetime of a TSSN event. Asterisks indicate period of lightning activity.](image)

The average most unstable lifted index (MULI) occurring during the initial hour of observation is 2 with a standard deviation of 1.1. In the world of summer time convection this is a very marginal value inferring little chance of convective development. A slight tendency towards more unstable LI is evident over the first three hours of observation (Figure 4.6.1). The minimum LI observed in the sample sets for this time period is -1, which is considered unstable, and the greatest observed value is 5, which is considered stable.
At the time of initial lightning activity the average LI is reduced to 1.4 with a standard deviation of 1.3. The minimum LI recorded in the sample sets at the time of initial lightning activity is 0, considered neutral, and the maximum-recorded LI at this time is 4. A continued trend towards zero is evident between the LI of this hour and the beginning observation period (Figure 4.6.1). This trend of decreasing average LI values continues into the next hour before again rebounding to greater LI values.

Throughout the hours of lightning activity an initial decrease is noted over the first two hours of activity. This decrease in average LI is 0.4 resulting in an average LI for the fifth hour of observation of 1. The minimum value found in the sample sets for the corresponding hour is 0 and the maximum value observed is 2. Beyond hour five the LI begins to return toward more stable values at a rate of 0.3 per hour through the end of the period representing hours of lightning activity. This trend towards increased stability continues until the end of the observation period (Figure 4.6.1).
The methods for analyzing the LI associated with non-thundering snow events are slightly different than that used in examining TSSN events. This is because the author found it reasonable to compare the LI at a similar average lifted parcel level (LPL) as was used to define the LI in TSSN events although the most unstable LPL (MULPL) in non-thundering events is obtained much higher in the atmosphere. The average MU LPL found in this study for TSSN events is 655mb while the average MU LPL in non-thundering events is 552mb. Therefore the depth of instability is of great importance in identifying TSSN events over non-thundering events. This topic will be covered in later sections.

When considering the LI at an average LPL of 667mb it is evident that TSSN events are less stable than non-thundering events. At the hour of initial observation the average LI in non-thundering events is 5.9 with a standard
deviation of 3.6. Unlike TSSN events the LI tends towards stability over the first three hours of observation (Figure 4.6.2). It should also be noted that the smallest LI value in the non-thundering sample set for this time period is 2 and the maximum value is 13.

At the mid-point of the average non-thundering event the LI value is nearly 7 at the average LPL of 667mb. Again the best LI found among the sample set for the mid-point is 2 with the maximum value of 15 recorded. Recall that the average value at the time of initial lightning activity is 1.38 which is much less stable than the average value found for this time period.

Beyond the mid-point of the non-thundering event the LI tends to increase in value slightly until the end of the observation period (Figure 4.6.2). The average value occurring two-hours after the mid-point is 6.9. At the final hour of observation the average LI actually decreased to 6.1 with a standard deviation of 4.7.

Recall that the methodology for analysis of LI in non-thundering snow events required the examination of the LI at an average LPL of 667mb in order to be consistent with TSSN events. Now we will examine the LI at an average LPL of 552mb, which is where the average LPL is actually found in non-thundering events. Bear in mind that this level is considered too high in the atmosphere to support convective elements.

With the average LPL of 552mb LI values ranging from 0.6 to 0.9 are obtained over the first three hours of observation with reasonable standard deviations of 0.8. These LI values fall far below those found in TSSN events however the average LPL associated with TSSN is geometrically much lower in
the atmosphere. Lifting a parcel from such a high level results in little to no variation in the LI over time. The average LI valid for the mid-event period is 0.9, which is identical to the previous hours value. Some variation can be observed in the final three hours of observation at this level.

Now that the average LI for non-thundering snow events has been presented for both parcels lifted from 667mb and 552mb it is the author’s intention to compare just the LI values resulting from the 667mb level to the average TSSN LI values. This methodology will serve two purposes. First by comparing only the LI values resulting from a non-thundering LPL of 667mb more consistency will be obtained with the average LI of 655mb LPL found in TSSN events. Secondly, the hypothesis that TSSN events are less stable in a deeper layer than non-thundering events can be substantiated.

![LI Comparison Graph](image)

Figure 4.6.3 Comparison of average most unstable lifted index, lifted from 655 mb, over the lifetime of TSSN (red contour) and the average most unstable lifted index, lifted from 667 mb, non-thundering snow events (blue contour). Asterisks represent period of lightning observation and mid-event period respectively.
It is apparent (Figure 4.6.3) that the average LI achieved by lifting an air parcel from 655mb associated with TSSN events is more favorable than the average LI found by lifting an air parcel from 667mb associated with non-thundering events. Comparing the average LI of both phenomena for the initial period of observation reveal that the average LI for non-thundering snow events is 5.9 while the average LI in TSSN events for the same period is 2. A two-tailed t-test suggests with a confidence interval of 98% the LI associated with TSSN events is indeed less than non-thundering events.

Similarly, the LI valid for the mid-point of a non-thundering snow event is also greater than the LI valid for the hour of initial lightning activity in the average TSSN event (Figure 4.6.3). The two-tailed t-test supports this hypothesis at a 99% confidence level. This evidence suggests that this parameter is more desirable in TSSN events than in non-thundering snow events.

A secondary observation relevant to this study is the trend towards lesser LI values found in TSSN events. Although, this trend is not a strong transition to more favorable LI values, it is absent in non-thundering events. In fact non-thundering events exhibit a tendency towards more stable LI values as time tends toward the mid-point of the event. This phenomenon further supports the hypothesis that LI is in fact a good discriminator for TSSN event identification.
4.7 The -10°C Isotherm

![Graph showing comparison of pressure levels](image)

Figure 4.7.1 Comparison of pressure level (mb) of -10°C isotherm (green contour) and lifted parcel pressure level (mb; red contour) of the average most-unstable parcel throughout the lifetime of a TSSN event.

In order for lighting activity to occur, graupel is thought necessary. Graupel occurring at various temperature levels will acquire either a positive or negative charge. In turn, we must have sufficient vertical motion through the charge reversal zones in order for graupel to form. Therefore the hypothesis for this section is that in TSSN events the most unstable lifted parcel level must be below the -10°C isotherm level, thus suggesting instability supportive of vertical motion at or around the -10°C isotherm level.

Referring to Figure 4.7.1 the average LPL at the first hour of observation is 674 mb with a standard deviation of 95mb. Meanwhile the average height of the -10°C isotherm for this time period is 621 mb with a standard deviation of only
33mb. Indeed the -10°C isotherm is in fact located above the average MU LPL by 53 mb.

Although the average MU LPL does rise suggesting a shallower layer of instability with time, the -10°C isotherm remains above the MU LPL over the first three-hours of observation (Figure 4.7.1). At the time of initial lightning activity the average MU LPL is 660mb with a standard deviation of 75mb. The -10°C isotherm at this time is 619mb with a standard deviation of 34mb, still 41mb further aloft.

Over the hours of lighting activity the average MU LPL retreats further aloft resulting a net change of 24mb over the four hours observation period (Figure 4.7.1). The level of the -10°C isotherm migrates closer to the surface over this period resulting in a 5 mb decrease in height over the first three hours of lightning observation with an abrupt decrease of 12 mb between the final hour of observation and the previous hour.

The average MU LPL remains below the -10°C isotherm through the final hour of lightning activity (Figure 4.7.1). At the hour following the cessation of lightning activity the atmosphere begins to cool and the -10°C isotherm retreats towards the surface at an average rate of 12 mb per hour.

Simultaneously the atmosphere stabilizes and the MU LPL begins to rise to 646 mb at the hour after the cessation of lightning activity. At this point the -10°C isotherm is located at 642 mb, nearly at the same level of the MU LPL. During the following hours the -10°C isotherm is located at 663 mb and the average MU LPL retreats to 642 mb. The atmosphere is now stable at the -10°C level which theory suggests will result in a cessation of graupel formation.
Referring to Figure 4.7.2 the average -10°C isotherm occurring during a non-thundering snow event is located at 729 mb. However, the average MU LPL happening during a non-thundering snow event is 554 mb which is located far above the -10°C level. In fact the average temperatures found at the MU LPL range from -18°C at the beginning of an event to -20°C at the end of the observation period. From this we may infer that the atmosphere is stable at the -10°C isotherm throughout the period of observation, theoretically creating a condition in which vertical motion is inhibited thus resulting in restricted graupel formation at this important charge reversal level.
Figure 4.7.3 Average MSL height (m) of the -10°C isotherm occurring throughout the lifetime of a TSSN event. Asterisks indicate period of lightning activity.

Michimoto (1989) investigated the altitude of the -10°C isotherm level of winter thunderstorms and found that the altitude that forms the threshold between strong and weak or no lightning activity is 1.8km AGL (Michimoto 1991). The average mean sea level (MSL) height of the -10°C isotherm varied less than 55m over the first three hours of observation during the average TSSN event (Figure 4.7.3). This average height is 3882m (3.9km) above MSL with a standard deviation of 450m. The lowest level the -10°C isotherm, is located at during the first three hours of observation is 3171m (3.17km) occurring and the greatest level at which the -10°C isotherm level valid for the same observation period is 4693m (4.7km).

At the initial hour of lightning activity the average height of the -10°C isotherm is 3898m (3.89km) with a standard deviation of 422m. The minimum height at this time is 3065m (3.07km) and the greatest level is 4443m (4.44km).
Over the hours of lightning activity the height of -10°C isotherm initially increases on average to 3950m (3.60km) before trending towards lower heights at the final hour of lightning observation (Figure 4.7.3).

The average height of the -10°C isotherm drops to 3606m (3.61km) at the hour after the cessation of lightning activity. This downward trend continues into the next hour resulting in an additional decrease of 182m. Beyond this point the -10°C isotherm retreats upward into the atmosphere, reaching a height of 3835m (3.84km).

![Figure 4.7.4](image_url) Average MSL height (m) of the -10°C isotherm occurring throughout the lifetime of a non-thundering snow event. Asterisks indicate mid-point of event.

Referring to Figure 4.7.4 the average height of the -10°C isotherm associated with non-thundering events is found much lower in the atmosphere than in TSSN events. In fact the average level of the -10°C isotherm at the initial
hour of observation is 2704m (2.7km) with a large standard deviation of 1325m (1.3km). Over the first three hours of observation the average height of the -10°C isotherm decreases to 2614m (2.61km) (Figure 4.7.4). The lowest level at which the -10°C isotherm was observed for this period is 602m and the highest level found in the sample sets is 4199m (4.20km).

At the mid-point of the event the average height of the -10°C isotherm is 2597m (2.60km) with a standard deviation of 1485m (1.5km). The lowest level observed for this period is 480m and the greatest level observed in the sample sets is 4285m (4.29km). From the mid-point through the end of observation the average height of the -10°C isotherm continues to decrease. By the end of observation the average level that the -10°C isotherm is found is 2353m (2.35km).

Initial analysis reveals that the -10°C isotherm is located much higher in the atmosphere during TSSN events than in non-thundering events (Figure 4.7.5). At the initial period of observation the average height of the -10°C isotherm associated with TSSN events is 3882m (3.88km) above the surface while it is found on average at 2704m (2.70Km) in association with non-thundering snow events. A two-tailed t-test reveals with a confidence level of 97% that the average height of the -10°C isotherm associated with TSSN events is indeed higher in the atmosphere than in non-thundering events.
Figure 4.7.5 Comparison of the average MSL height (m) of -10°C isotherm throughout the lifetime of TSSN events (red contour) and non-thundering snow events (blue contour).

Again the -10°C isotherm is found lower in the atmosphere during the mid-point of non-thundering events than in TSSN events at the point at which lightning is first observed. The average height of the -10°C isotherm at the mid-point of non-thundering events is 2597m (2.60km), while the -10°C isotherm is found at 3898m (3.90km) at the hour of initial lightning activity. The two-tailed t-test supports this statement with a 97% confidence interval. Again suggesting that the -10°C isotherm is located higher in the atmosphere in TSSN events than in non-thundering snow events.

Several observations can be gathered from the above results. First off it can be inferred that along with being more stable in a deeper layer, non-thundering snow events occur in a much cooler environment than TSSN events do. This can be inferred from the height of the -10°C isotherm. A slight increase in the depth of the lower atmosphere can be inferred during the period of
lightning activity, whereas the lower atmosphere associated with non-thundering events tends to experience a net cooling as noted by the decrease in height of the -10°C isotherm. These trends along with the t-test results suggest that the level of the -10°C isotherm is a good discriminator for TSSN events.

4.8 Symetric instability

Assessing potential symmetric instability (PSI) is a rather subjective procedure. This study follows methodology by Moore and Lambert (1993), which involves cross-sectional analysis of $\theta$e, geostrophic momentum and relative humidity. This methodology fails once the synoptic flow becomes curved, which violates the geostrophic flow criterion. Unfortunately many of the cases examined in this study exhibit some form of curvature during one period.
or another. This problem may introduce error in determining the type of stability regime responsible for convection in TSSN events.

Initially, 50% of the TSSN cases investigated in this study are stable (Figure 4.8.1). Neutral symmetric instability (NSI) is present in 25% of the cases examined; 12.5% exhibited potential instability (PI). Curvature inhibits the assessment of stability mode in 12.5% of the TSSN cases during the initial hour of observation.

Table 4.8.1 Represents the distribution of TSSN events per stability regime over the first three hours of observation prior to initial lightning activity.

<table>
<thead>
<tr>
<th></th>
<th>Initial hour</th>
<th>2nd hour</th>
<th>3rd hour (prior to initial CG flash)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Stable</td>
<td>50</td>
<td>13</td>
<td>13</td>
</tr>
<tr>
<td>PSI</td>
<td>0</td>
<td>25</td>
<td>13</td>
</tr>
<tr>
<td>NSI</td>
<td>25</td>
<td>13</td>
<td>13</td>
</tr>
<tr>
<td>PI</td>
<td>13</td>
<td>38</td>
<td>50</td>
</tr>
<tr>
<td>Curved</td>
<td>13</td>
<td>13</td>
<td>13</td>
</tr>
</tbody>
</table>

In the hours leading up to the initial hour of lightning activity it appears that PI dominates the atmosphere (Table 4.8.1). Potential instability accounts for 37.5% of the events during the second hour of observation and 50% of the events the hour prior to initial lightning activity. Potential symmetric instability is responsible for only 25% of the observed environments during the second hour of observation and 12.5% the hour prior to initial lightning activity.
At the initial hour of lightning observation 25% of the cases investigated are dominated by PSI (Figure 4.8.2). In 50% of the cases investigated PI is the dominating mode of instability present in TSSN events. This finding would allow us to infer that CAPE would be present in at least 50% of TSSN events, which is not always the case. Potential instability does not demand the presence of CAPE. Only 25% if the observed TSSN events displayed CAPE at the hour of initial lightning activity.

The inflated value of PI events maybe partially attributed to the fact that 12.5% of the cases evaluated at this time period exhibit some form of curvature, which may mask potential PSI events. Another response for this inflated value of PI is the subjective nature of this analysis, perhaps the authors cross section was not taken exactly perpendicular to the thermal wind. Yet another possibility is
that not all of the PSI is accounted; for instance PI events could also contain layers of PSI, which are ignored due to the dominating nature of PI over PSI.

Table 4.8.2 Represents the distribution of TSSN events by stability regime valid during the hours of lightning observation.

<table>
<thead>
<tr>
<th></th>
<th>Initial CG</th>
<th>2nd Hour</th>
<th>3rd Hour</th>
<th>Final CG</th>
</tr>
</thead>
<tbody>
<tr>
<td>Stable</td>
<td>0</td>
<td>0</td>
<td>0</td>
<td>0</td>
</tr>
<tr>
<td>PSI</td>
<td>25</td>
<td>0</td>
<td>13</td>
<td>14</td>
</tr>
<tr>
<td>NSI</td>
<td>13</td>
<td>25</td>
<td>25</td>
<td>43</td>
</tr>
<tr>
<td>PI</td>
<td>50</td>
<td>63</td>
<td>50</td>
<td>43</td>
</tr>
<tr>
<td>Curved</td>
<td>13</td>
<td>13</td>
<td>13</td>
<td>0</td>
</tr>
</tbody>
</table>

During the first 4 hours of lightning activity PI continues to dominate the atmosphere, accounting for 63% of the observations occurring during the second hour of lightning activity (Table 4.8.2). Environments neutral to symmetric instability begin to dominate over PSI environments, accounting for 25% of the observed environments occurring for this period. At the final hour of observed lightning activity, PI was present in 43% of the cases with NSI also representing 43% of the observed environments.

Table 4.8.3 Represents the distribution of TSSN events by stability regime over the final three hours of observation.

<table>
<thead>
<tr>
<th></th>
<th>Hr after CG</th>
<th>Final Period</th>
</tr>
</thead>
<tbody>
<tr>
<td>Stable</td>
<td>0</td>
<td>33</td>
</tr>
<tr>
<td>PSI</td>
<td>0</td>
<td>0</td>
</tr>
<tr>
<td>NSI</td>
<td>43</td>
<td>33</td>
</tr>
<tr>
<td>PI</td>
<td>57</td>
<td>17</td>
</tr>
<tr>
<td>Curved</td>
<td>0</td>
<td>17</td>
</tr>
</tbody>
</table>

The time periods following the cessation of lightning activity seem to favor PI environments. Referring to table 4.8.3, 57% of the observed cases occur in a PI environment the hour after the final lightning strike is observed. Interestingly, NSI environments are evident in 43% of the observed cases valid for this time period. This trend toward NSI continues into the next hours with
33% of the cases exhibiting NSI. The final hour of observation PI, once again dominates 67% of the environments, however this is due in part to the curved flow, which restricts the evaluation of the stability regime in 33% of the cases.

![Figure 4.8.3 Percentage of non-thundering events per stability regime, valid at initial hour of observation. Curved flow (dark blue), potential instability (green), neutral symmetric instability(yellow) and stable environments (light blue).](image)

Initially the environment of a non-thundering snow event appears to be dominated by NSI, which is responsible for 57% of the initial observations (Figure 4.8.3). Non-thundering snow events often initiate in stable environments as reflected by Figure 4.8.3, 43% of the observed events occurred in stable environments. Note that the absence of PSI, PI or curved flow occurring in this time period.
Table 4.8.4 Represents the distribution of non-thundering snow events per stability regime over the first three hours of observation, prior to the mid-point of the event.

<table>
<thead>
<tr>
<th></th>
<th>Initial hour</th>
<th>2nd hour</th>
<th>3rd hour</th>
</tr>
</thead>
<tbody>
<tr>
<td>Stable</td>
<td>43</td>
<td>29</td>
<td>57</td>
</tr>
<tr>
<td>PSI</td>
<td>0</td>
<td>29</td>
<td>0</td>
</tr>
<tr>
<td>NSI</td>
<td>57</td>
<td>29</td>
<td>29</td>
</tr>
<tr>
<td>PI</td>
<td>0</td>
<td>14</td>
<td>0</td>
</tr>
<tr>
<td>Curved</td>
<td>0</td>
<td>0</td>
<td>14</td>
</tr>
</tbody>
</table>

The hours leading up to the mid-point of non-thundering snow events reflect a mixed bag of stability regimes. Referring to table 4.8.4 it is observed that 29% of the observed environments valid for the second hour of observation were of the PI regime, 29% were NSI and 29% exhibited characteristics of PSI. Potential symmetric instability is absent the third hour of observation, while stable atmospheric regimes account for 57% of the cases.

Figure 4.8.4 Percentage of non-thundering snow events by stability regimes, valid for the mid-point of the average non-thundering snow event. Curved flow (dark blue), potential instability (green), neutral symmetric instability (yellow), potential symmetric instability (red) and stable environments (light blue).
An equal number of NSI and stable atmospheric regimes are observed at the mid-event period (Figure 4.8.4). These two stability regimes account for 86% of the non-thundering sample sets. Potential symmetric instability is observed in 14% of the non-thundering cases. At no time was the atmosphere observed to be PI or have curved flow during this time period.

Table 4.8.5 Represents the distribution of non-thundering snow events by stability regime over the final three hours of observation.

<table>
<thead>
<tr>
<th></th>
<th>5th hour</th>
<th>6th hour</th>
<th>Final hour</th>
</tr>
</thead>
<tbody>
<tr>
<td>Stable</td>
<td>29</td>
<td>57</td>
<td>71</td>
</tr>
<tr>
<td>PSI</td>
<td>0</td>
<td>0</td>
<td>0</td>
</tr>
<tr>
<td>NSI</td>
<td>57</td>
<td>43</td>
<td>14</td>
</tr>
<tr>
<td>PI</td>
<td>0</td>
<td>0</td>
<td>14</td>
</tr>
<tr>
<td>Curved</td>
<td>14</td>
<td>0</td>
<td>0</td>
</tr>
</tbody>
</table>

At the fifth hour of observation NSI appears to be the dominant stability regime (Table 4.8.5) comprising 57% of the sample set. The hours leading to the end of the observation period become more dominated by stable regimes, which are responsible for 71% of the sample set by the final hour of observation. Potential symmetric instability is not present in any of the sample sets leading to the end of the observation period.

A comparison of the mode values of stability regime occurring over the first three hours of observation during both TSSN and non-thundering reveals some differences between these two phenomena. The dominating mode for this time period in the observed TSSN events is potential instability. While non-thundering snow events are initial neutrally symmetric evolving in to PSI environments and finally become stable by the third hour of observation.

The mode value valid at the mid-point of the non-thundering snow events illustrates that NSI environments dominate the sample sets. While the mode
environment of TSSN events at the period of initial lightning observation is potential instability. This evidence suggests that stability associated with TSSN events tends to be more upright where as slantwise instability is more dominate in non-thundering events.

As time progresses towards the end of the observation period, NSI dominates the environment of non-thundering events and eventually the environment becomes stable (Table 4.8.5). However, in TSSN events the stability mode of the atmosphere remains PI through the end of the observation period (Table 4.8.3). Once again this evidence suggests that TSSN environment a prone to upright instability while stability regimes of non-thundering snow events are more slantwise in nature.
Chapter 5
Summary

5.1 Discussion

This study set out to answer several fundamental questions:

- How does the stability tendency change with time and how do the individual components of the governing equation affect the whole?
- Are there threshold values in traditional stability indices discriminating thundersnow from non-thundering events?
- Are there recognizable patterns of these indices in plan view charts?
- How does stability evolve in time in thundersnow events and does it differ from the evolution of non-thundering snow events?
- What stability regime is dominant in thundersnow events?

Several outcomes pertaining to these questions will now be presented. First the question to be covered is how does the stability tendency change with time and how does its individual components affect the outcome. Positive values of differential temperature advection infer warm air advection (Term A in equation 3.3.1). Warm air advection lower in the atmosphere acts as a destabilizing feature, however when warm air advection is elevated, the region beneath tends to stabilize. Likewise the region above tends to destabilize. Whereas elevated cold air advection tends to destabilize an air column.

Differential temperature advection tends towards more positive values prior to lightning activity (Figure 4.1.1). Recall that the lifted parcel level for a TSSN event is elevated in nature therefore the region beneath is tending towards stability with time. At the time of initial lightning activity, the DADTP
component begins to tend towards negative values and this trend continues through the period of observed lightning activity. This statement could suggest that cold air advection is taking place aloft and destabilizing the atmosphere below. Plan view charts reveal that DADTP values left of the TROWAL axis tend towards positive values (Figure 4.1.2), while the areas right of TROWAL axis and closer to the cyclone tend to be more negative.

Differential temperature advection associated with non-thundering snow events is presumably more negative throughout the lifetime of observation than what is observed in TSSN events. To further investigate this statement a two-tailed t-test is invoked on the initial hour of observation of both types of events. For the initial hour of observation the t-test resulted in no statistical difference between the two sample means. Therefore it cannot be stated the DADTP valid for the initial hour of observation is in fact different in TSSN and non-thundering snow events.

Similarly a two-tailed t-test was performed on the means of both types of events valid for the hour of initial lightning activity and the mid-point of a non-thundering snow event. The test resulted in a confidence interval of 75% that the two sample sets are in fact different. Unfortunately, this confidence interval is not considered significant and it cannot be stated that DADTP valid at the time of initial lightning activity is greater than that found at the mid-point of a non-thundering snow event.

Since we cannot state that the sample means are not different between TSSN and non-thundersnow events the DADTP value alone is not a good discriminator for TSSN identification. However, referring back to figure 4.1.8 we
see that monitoring DADTP over time may show some indication of a possible TSSN event. Primarily this is due to the tendency of DADTP to become more positive prior to initial lightning activity, this trend towards a more positive value is absent in non-thundering snow events.

According to Bluestein (1992) the process of vertical stability advection cannot create nor destroy stability it can merely relocate it. Positive values of VADVS imply upward advection of greater stability while negative values indicate advection of lower stability. Referring to Figure 4.1.9 the vertical advection of stability tends towards more positive values as time approaches the period of lightning observation. This suggests that greater stability is being advected upward into the atmosphere. Stability advection tends toward lesser positive values during the period representative of lightning activity (Figure 4.1.9). A second mode of more positive VADVS occurs at the final hours of lightning observation, at which point VADVS begins a gradual tendency towards zero once again. The average VADVS component never reaches zero until the final hours of observation.

Similar to tendencies observed in TSSN events VADVS tends towards smaller positive values over the first few hours of observation. However, unlike VADVS tendencies in TSSN events, the tendency in non-thundering events continues towards zero through the mid-point of the event. This trend towards smaller values continues through the end of observation and never reaches zero.

The vertical advection of stability tends to be greater in association with non-thundering snow events than in TSSN events the first hour of observation. However, a two-tailed t-test reveals only a 70% confidence level that the sample
means are not of random origin therefore this may not be the case. This, result of course is not statistically significant, therefore it can not be stated that at the initial hour of observation non-thundering snow events tend towards more positive values of VADVS than TSSN do.

At the hour of initial lightning activity and corresponding mid-event the sample means are nearly identical with only a 23% confidence interval that the sample means are in fact random (Figure 4.1.13). These results suggest that little difference exists in the values of VADVS between the two types of snowstorms. However, like the results found in DADTP the trend in VADVS over time does differ between TSSN and non-thundering snow events.

In the hours prior to lightning activity the tendency of VADVS trends towards more positive values in TSSN events. This trend is absent in non-thundering snow events as time approaches the corresponding mid-point of the event. Additionally, a second mode of more positive VADVS values is evident over the final hours of lightning activity. Once again this second mode of more positive VADVS values is absent in non-thundering snow events.

Static stability can be created or destroyed locally through the vertical stretching and shrinking of an air column (Bluestein 1992). Static stability is destroyed when there is horizontal convergence, this would be represented by negative values of STRCH. Horizontal convergence at the surface is especially affective in creating instability near regions under a capping inversion (Bluestein 1992). On the other hand positive STRCH values indicate horizontal divergence, which creates stability in the air column.
In terms of TSSN events STRCH is often positive, suggesting a tendency towards creating stability. Little change with time is noted in the hours leading up to lightning activity (Figure 4.1.14). However, once lightning activity is noted a tendency towards negative values of STRCH is apparent, inferring the destruction of stability. This tendency towards negative STRCH values lasts through the duration of lightning activity once again the tendency towards positive values returns by the final hour of lightning activity.

At no time is the average value of STRCH above zero in association with non-thundering snow events (Figure 4.1.17). This would infer a tendency for stability to be destroyed over the lifetime of this type of event. Referring to Figure 4.1.19 it is apparent that this term is smaller in non-thundering snow events than in TSSN events.

Initially no statistical difference exists between the sample means of TSSN and non-thundering snow events. However, by the mid-point of the non-thundering snow event and the corresponding period of initial lightning activity in TSSN events sample means do differ. Although, only an 85% confidence interval is obtained supporting the statement that STRCH is in deed greater than that found in non-thundering snow events, this level is unacceptable. The author feel that with a larger data set this confidence level may increase.

The total average stability tendency (STABTDY) of TSSN events never falls below zero (Figure 4.1.20). However, a great deal of variability exists among the sample sets. Two modes of more positive stability tendency appear throughout the lifespan of an average TSSN event (Figure 4.1.20). These two modes correlate temporally with the initial and final periods of lightning activity.
A tendency of lesser STABTDY values occurs at the time period of lightning activity separating these two positive modes.

The stability tendency found in the average non-thundering snow event is initially positive (stable) and slowly tends towards destabilization (values less than zero) with time (Figure 4.122). The STABTDY actually becomes less than zero by the mid-event period and tends toward further destabilization for the remainder of the observation period.

Comparing the STABTDY for both TSSN and non-thundering snow events suggests that initially non-thundering snow events have a tendency to be more stable than TSSN events (Figure 4.124). The two-tailed t-test suggests this to be true with a confidence interval of 92%. This confidence interval however is not a great enough to support the hypothesis that non-thundering snow events are in-fact more stable initially than TSSN events.

Again referring to Figure 4.124, non-thundering snow events tend towards zero over the first few hours of observation and do in fact, reach zero by the mid-point of the event. While TSSN events approach zero towards the hour of initial lightning activity, the average STABTDY never reaches zero. STABTDY is more positive in TSSN events than in non-thundering snow events. Again the two-tailed t-test disposes of this statement since the resulting confidence interval is only 82% that the sample means are in fact different.

This study suggests that STABTDY is not a good discriminator for TSSN based on the values alone. However, this study has established that the temporal evolution of STABTDY is quite different in TSSN events than in non-thundering snow events. STABTDY in the average TSSN event never falls below zero while
non-thundering snow events trend towards negative values over the lifetime of the event. Additionally a bimodal pattern towards greater stability is evident (Figure 4.1.24) in TSSN events and absent in non-thundering snow events.

Now that it has been established that the average STABTDY over the lifetime of a TSSN event differs from that found in non-thundering snow events we examine what component drives the total stability tendency of TSSN events. In the hours preceding initial lightning activity DADTP is the only component tending to be negative. This component does however become less negative as time approaches the hour of initial lightning activity. The VADVS component tends to be a small positive number and STRCH component tends to be a greater positive number.

Although DADTP may be the component that forces STABTDY to take on smaller positive values, it appears that the total stability tendency equation follows the tendency of the VADVS term. As VADVS increases so does the stability tendency as STABTDY decreases so does VADVS. Only weak correlations are found linking DADTP directly to the STABTDY (Table 4.1.1). Actually nearly identical correlations are observed between VADVS and STABTDY, suggesting that VADVS and DADTP have an equal share in dominating the sign of STABTDY.

Comparing the individual components to the total stability tendency reveals that DADTP is the dominating factor in the stability tendency of non-thundering events (Figure 4.1.23). Whereas VADVS and STRCH show little variation in time over the first four hours of observation, only exhibiting change at hour five. It is evident (Table 4.1.2) that DADTP is indeed the dominating
component in the stability tendency of non-thundering cases with a correlation coefficient of 0.7.

These results suggest that DADTP plays a dominant role in both TSSN and non-thundering snow events. Differential temperature advection correlates slightly better with the total stability tendency equation in regards to non-thundering snow events than in TSSN events. The weaker correlation found with TSSN events maybe due to the bigger role that the vertical advection of stability plays in influencing the stability tendency of TSSN events.

Some insight has been gained in the role of the stability tendency in regards to both TSSN and non-thundering snow events. Lets now turn our attention to the traditional stability indices and how they differ from TSSN to non-thundering snow events.

The total totals (TT) index appears to be greater in TSSN than in non-thundering snow events (Figure 4.2.5). Unfortunately the t-test reveals a low confidence interval of only 66% that this is the case during the initial hour of observation. Comparing the TT value valid for the initial hour of lightning activity and the corresponding mid-point of the non-thundering snow event also suggests that TT associated with TSSN is greater than in non-thundering snow events (Figure 4.2.5). Again the two-tailed t-test results in only a 67% confidence interval suggesting that the sample means are not different. Therefore the TT value alone is not a good discriminator of TSSN events.

In plan view charts TSSN events occur in a gradient of TT left of the TROWAL axis and along the $\theta_e$ ridge associated with the warm frontal boundary. At no time is the event location a local maximum or minimum.
Although, the average TSSN event seems to fall closer to the minimum TT value within the gradient.

Non-thundering snow events tend to fall along a similar gradient of TT along the leading edge of the $\theta_e$ gradient associated with the TROWAL axis. It is this similarity in TT gradients to those found in TSSN events that suggests that plan view charts of TT are not a good discriminator for TSSN events. However, the observed gradient may not be as great as that found in TSSN events.

Likewise the K-index appears greater in association with TSSN events than in non-thundering snow events (Figure 4.3.8). Initially the K-index value is similar in TSSN and non-thundering snow events. In fact the two-tailed t-test suggests that the sample means are identical with only a 1% probability that the sample means are random. After the initial hour of observation the K-index tends towards much greater values in TSSN events, while the K-index in non-thundering snow events decrease.

At the mid-point of the non-thundering event and the corresponding initial hour of lighting activity in TSSN events the K-index is much greater in association with TSSN events. Unfortunately, the two-tailed t-test suggests that the sample means are different with only an 85% confidence interval. Therefore, the K-index may not be a good discriminator for TSSN events.

Similar to plan views of TT associated with TSSN, the K-index also falls along a gradient of $\theta_e$ associated with the TROWAL axis (Figure 4.3.5). The location of the TSSN event is never a local maximum or minimum. In cases with a TROWAL the gradient of K decreases as the TROWAL moves over-head. The
K-index gradient will follow the equivalent potential temperature iso-lines as the TROWAL axis shifts.

Again similar to plan views of TT associated with non-thundering snow events, plan views of the K-index appear as a gradient that also falls along gradients of $\theta_e$ associated with the TROWAL axis or warm frontal boundary. Since this plan view is so similar to results found in TSSN events no distinctive characteristics separate fields of K-index associated with non-thundering snow events from K-index fields associated with TSSN events.

The average mid-level lapse rates in TSSN events are much greater than those found in non-thundering snow events (Figure 4.4.9). At the initial hour of observation the average mid-level lapse in TSSN events is 6.5 K km$^{-1}$ while the average midlevel lapse rate found in non-thundering snow events is 4.8 K km$^{-1}$. The two-tailed t-test suggests that the sample means valid for this period are in fact different with a 99% confidence interval.

At the hour of initial lightning activity the average mid-level lapse rate in association with TSSN is 6.0 K km$^{-1}$ compared to the average mid-level lapse rate of 4.9 K km$^{-1}$ found with non-thundering snow events is much greater. Again, confirmed by a 95% confidence interval the sample sets are different. Therefore, it can be stated that mid-level lapse rates are good discriminators for TSSN events.

From plan view charts of mid-level lapse rates in TSSN events featuring a TROWAL, it is apparent that an axis of greater mid-level lapse rates penetrates the occluded quadrant of the associated cyclone (Figure 4.4.3). The area of interest
will most often be located left of the TROWAL axis and in the vicinity of the mid-level lapse rate axis.

To further support the authenticity of the mid-level lapse rate being a good discriminator of TSSN events the following should be noted. None of the non-thundering snow events illustrated an axis of instability in mid-level lapse rates. Also, non-thundering snow events occur within closed iso-lines of mid-level lapse rates along the TROWAL axis or ahead of a frontal boundary.

Convective available potential energy (CAPE) appears in both TSSN and non-thundering snow events. The appearance of CAPE in both types of events is sporadic and does not regularly occur at any particular point in time. Due to its sporadic temporal placements and the fact that CAPE occurs in both TSSN and non-thundering events leads this author to support Schultz’s (1999) hypothesis that CAPE is not a discriminator of snow events featuring lightning.

The most-unstable lifted index appears more favorable in TSSN events than in non-thundering snow events (Figure 4.6.2). Comparing the average LI of both phenomena for the initial period of observation reveals that the average LI for non-thundering snow events is 5.9 while the average LI in TSSN events for the same period is 2. A two-tailed t-test suggests with a confidence level of 98% the LI associated with TSSN events is indeed less than non-thundering events.

Similarly, the LI valid for the mid-point of a non-thundering snow event is also greater than the LI valid for the hour of initial lightning activity in the average TSSN event (Figure 4.6.2). The two-tailed t-test supports this hypothesis at nearly a 100% confidence level. This evidence suggests that this parameter is more desirable in defining TSSN events than in non-thundering snow events.
In regards to the level of the -10°C isotherm in relation to the most-unstable lifted parcel (MULPL) of an average TSSN event, it is found that the -10°C isotherm is indeed higher in the atmosphere than the MULPL. This evidence suggests that the environment around the -10°C isotherm is buoyant and will enhance vertical motion through this critical charge reversal level.

Unlike the level at which the -10°C isotherm is found in regards to TSSN, the -10°C isotherm level is below the MULPL in non-thundering snow events (Figure 4.7.2). Thus, suppressing vertical motion through this critical charge reversal level and providing an environment less conducive to lightning activity.

Another discovery that can be inferred from this data is that the environment of a non-thundering snow event is much more stable in a deeper layer than TSSN events. Additionally, the environment of a non-thundering snow event is much cooler as well. This inference can be stated since the -10°C isotherm is located closer to the surface in non-thundering snow events than in TSSN events.

Finally, it was found that the stability regime responsible for the promotion of TSSN is that of potential instability rather than potential symmetric instability. This suggests that TSSN events maybe more gravitationally unstable than previously thought. However, PSI may have been present in more cases than noted since in area of mixed PI and PSI, PI will dominate.

Another important realization is that non-thundering snowstorms are prone to neutrally symmetric environments. Non-thundering events do evolve into stable environments towards the end of the observation period. This
suggests that non-thundering snow events are prone to be stable to gravitational stability and less stable with regards to symmetric instability.

5.2 Conclusions

This study set out to answer several fundamental questions:

- How does the stability tendency change with time and how do the individual components of the equation affect the whole?

  It was found that the stability tendency of thundersnow events tends towards greater stability prior to lightning activity (Figure 4.1.20). During the hours of lightning activity the stability tendency was reduced and began to increase towards the final hour of lightning activity. The dominating component(s) appears to be differential advection, which may be the key player to the reduction of the stability tendency during the hours of lightning activity. However, a nearly identical relationship was also exposed between the vertical advection of stability and the stability tendency equation.

- Are there threshold values in traditional stability indices discriminating TSSN events from non-thundering snow events.

  At first glance traditional stability indices appear more favorable in TSSN events than in non-thundering snow events. However, only three statistically supported parameters can be considered good discriminators for TSSN. The first is the mid-level lapse rate, with a threshold value of 6.0 ± 1.0. A second statistically sound indicator is the most-unstable lifted index (MULI). The threshold value for the MULI is 1.4 ± 1.3. Finally the most significant threshold is the depth of the -10° C isotherm. It was found
that the critical depth at which the -10° C isotherm should be found is 3898 m ± 418 m.

• Are there recognizable patterns of these indices in plan view charts?

   No plan view patterns specific to TSSN events were documented in this study. Although, it was found that several of the stability indices tend to align themselves along the θ_e ridge and or the TROWAL air-stream, this phenomena is not exclusive to TSSN events and is often associated with non-thundering snow events.

• What stability regime is dominant in thundersnow events?

   This study has established that the environment of TSSN events is dominated by potential instability (PI). While non-thundering snow events are dominated initially by neutral symmetric instability and evolve into a stable regime with time. These results may change with a larger data set since a number of the TSSN events exhibited a large amount of curved flow, which violates the geostrophic criterion of the conditional symmetric instability theory.

   In conclusion it has been established that the thermodynamic environment is quite different in TSSN events than in non-thundering snow events. This is the case specifically in terms of the evolution of stability throughout the lifetime of the event. Additionally, the stability regimes differ between the two types of snow events. An unexpected discovery unveiled by this study is that the depth of the warm layer is much greater in TSSN events than in non-thundering snow events.
This study should be continued in the future since the data set was rather small. However it is felt that these results do have merit since similar results are attained to what Market et al (2006) published. It is however felt that more substantial results would be obtained by expanding the data set especially with regards to symmetric instability.
Appendix A: Events

Table 5.2.1 TSSN events selected for this study including: date and time of event, location, station identifier, hour of initial cloud-to-ground flash and duration of event.

<table>
<thead>
<tr>
<th>DATE</th>
<th>LOCATION</th>
<th>STATION</th>
<th>FIRST CG (UTC)</th>
<th>DURATION (UTC)</th>
</tr>
</thead>
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<tr>
<td>11-23-2003</td>
<td>Salina KS</td>
<td>SLN</td>
<td>0300</td>
<td>0000-0800</td>
</tr>
<tr>
<td>12-09-2003</td>
<td>Beatrice NE</td>
<td>BIE</td>
<td>1400</td>
<td>1100-1800</td>
</tr>
<tr>
<td>12-10-2003</td>
<td>Tulsa OK</td>
<td>RVS</td>
<td>0300</td>
<td>0000-0600</td>
</tr>
<tr>
<td>12-10-2003</td>
<td>Joplin MO</td>
<td>JLN</td>
<td>0800</td>
<td>0500-1400</td>
</tr>
<tr>
<td>01-27-2004</td>
<td>Marion IL</td>
<td>MWA</td>
<td>0400</td>
<td>0100-0900</td>
</tr>
<tr>
<td>02-05-2004</td>
<td>Mountain Home</td>
<td>BPK</td>
<td>0300</td>
<td>0000-0800</td>
</tr>
<tr>
<td>03-05-2004</td>
<td>Eau Claire WI</td>
<td>EAU</td>
<td>0700</td>
<td>0400-1600</td>
</tr>
<tr>
<td>03-13-2004</td>
<td>Hutchinson WI</td>
<td>HCD</td>
<td>1200</td>
<td>0900-1800</td>
</tr>
</tbody>
</table>

Table 5.2.2 Non-thundering snow events selected for this study including: date and time of event, location, station identifier, hour of the mid-event period and duration of event.

<table>
<thead>
<tr>
<th>DATE</th>
<th>LOCATION</th>
<th>STATION</th>
<th>MID-EVENT (UTC)</th>
<th>DURATION (UTC)</th>
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<td>0900-1500</td>
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<td>SGF</td>
<td>1200</td>
<td>0900-1500</td>
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<td>OMA</td>
<td>1200</td>
<td>0900-1500</td>
</tr>
<tr>
<td>03-15-2004</td>
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<td>OMA</td>
<td>1200</td>
<td>0900-1500</td>
</tr>
<tr>
<td>01-06-2005</td>
<td>Green Bay WI</td>
<td>GRB</td>
<td>1200</td>
<td>0900-1500</td>
</tr>
<tr>
<td>01-22-2005</td>
<td>Green Bay WI</td>
<td>GRB</td>
<td>1200</td>
<td>0900-1500</td>
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<tr>
<td>02-14-2005</td>
<td>Green Bay WI</td>
<td>GRB</td>
<td>1200</td>
<td>0900-1500</td>
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REFERENCES


