CASE STUDY OF AN ANOMALOUS, LONG-LIVED CONVECTIVE SNOWSTORM

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On 23 March 1966, thundersnow was reported over a period of 9 hours (non-consecutive) at Eau Claire, Wisconsin. This event constitutes the longest period of thundersnow at a single station from a surface observation dataset of 226 stations spanning the years 1961-1990. For 30 years researchers have studied the atmospheric link between routine snowstorms and those with convection. We can assume that the ingredients are dynamic in nature and that thundersnow occurs around a surface cyclone, or near a lake area in association with surface-based instability, or in mountainous terrain.

In this study, the dynamic characteristics of a long-lived convective snowstorm are examined. Using objective analysis of rawinsonde data and model simulation output from the Workstation-Eta (WS-Eta), we have determined that the thermodynamic characteristics of the thundersnow event did not change with the evolution of the cyclone. For the duration of the event Eau Claire was north-northeast of the surface cyclone, with ample moisture, and forcing for ascent. Equivalent potential vorticity (EPV) and conditional symmetric instability (CSI) were present in cross-section analysis. The $\theta_v$ pattern at 700 mb indicates a trough of warm air aloft (TROWAL) upstream at 0000 UTC and then coinciding with Eau Claire at 1200 UTC. Elevated convective (potential) instability fails to develop.

The WS-Eta run provided an excellent meso-$\alpha$ simulation of the storm. The WS-Eta output was almost identical to the subjective surface analysis as well as the 48-hr precipitation field. This snow event resulted largely from the prolonged presence of frontogenesis in the presence of weak symmetric stability. That the event should remain convective after 0300 UTC when measurable SCAPE is not present is not altogether clear, even in the presence of a fine-scale model simulation.

The case presented here resulted from strong frontogenetical forcing in the presence of weak conditional symmetric stability northeast of a surface cyclone. This scenario was created and maintained by the presence of a TROWAL airstream over the EAU region for an extended period.
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Chapter 1

Introduction

Thundersnow, although rare, can profoundly affect the daily life of the average person. For the past 30 years, researchers have been studying the atmospheric link between routine snowstorms and those with convection. Researchers have assumed that the ingredients are dynamic in nature when they occur around a surface cyclone, although they are also frequent in lake areas in association with instability as well as in mountainous terrain.

Thundersnow events can bring blizzard-like conditions to an area: situations with snowfall, strong winds, and where visibility drops to zero rapidly. In the right situation, convective snowstorms have the potential to amplify heavy mesoscale banded precipitation that can drop 12-24 in. of snow in less than a day and leave even the best emergency management officials utterly helpless. When situations like this occur, not only is the average pedestrian restricted in travel, but emergency personnel are also affected.

It is important that this research be accomplished so that scientists and forecasters gain an understanding of the primary components of such events. Additionally, the general public will be better prepared for potentially dangerous situations.
1.1 Purpose & Objectives

1.1.1 Purpose

The purpose of this study is to reveal the dynamic characteristics of a particularly strong, long-lived convective snowstorm. Within this long-lived event, thundersnow persisted over a period of nine non-consecutive hours. Clearly, some quasi-persistent process acted to generate (and regenerate) the instability required for convection.

1.1.2 Objectives

Two objectives are identified to accomplish the aforementioned purpose. This research will seek to:

- Determine if the convection is slantwise or upright in nature, or if the environment evolved from one type to the other.
- Examine the stability tendency to determine the thermodynamic cause of such persistent convection.

1.2 Statement of Thesis

In recent years, great strides have been made to determine the dynamic situation in which convective snowstorms evolve. In this research, an investigation into the dynamic characteristics of a long-lived convective snowstorm will be undertaken. With this work, we expect to support the efforts of previous investigators (Market et al. 2002; Oravetz 2003). As a great deal of the thundersnow (TSSN) occurred northeast of the larger cyclone center, we expect that this event is driven largely by the release of conditional symmetric instability (CSI; Market et al. 2004). However, by the end of the TSSN period at Eau Claire, WI, the cyclone center is to the south southeast of the Eau Claire, WI airport. Consequently, we intend to show:
• Thundersnow at Eau Claire was primarily the result of CSI released north of a warm front, and

• The atmosphere became less stable for gravitational convection toward the end of the thundersnow period.
Chapter 2

Literature Review

With this chapter, recent studies on convective snowfall are examined, as well as forecasting methods for banded heavy precipitation. Previous studies have examined other cases of convective snow (e.g., Halcomb 2001), as well as a composite view of what convective snow events should look like on average (e.g., Oravetz 2003). Each of these studies includes an excellent literature review with that of Halcomb (2001) being the most thorough and expansive. Consequently, we dispense with needless repetition and focus here on those works that deal with banded precipitation and the nature of long-lived wintertime convection.

Curran and Pearson (1971) were among the first to write on convective wintertime snowfall. They were interested in correlating weather balloon soundings to thundersnow occurrences, and hoped to find some kind of inversion or characteristic that would shed light on how these situations occur. They chose to only include those situations where thundersnow occurred within 3 hours of the sounding and 90 nautical miles, limiting their study to only 13 cases. A mean proximity sounding was determined by averaging the temperature and dew point data for 850, 700, 500, and 400 millibar. When this procedure was applied to the 13 cases that Curran and Pearson were studying and then averaged, the mean soundings gave a Showalter index of $+12^\circ\text{C}$, a very stable value with respect to thunderstorm development. The SELS lifted index (Galway 1956) also gave a stable value of
+4°C from the top of the inversion to 500 millibars. Curran and Pearson (1971) deduced that the mean proximity sounding is essentially moist-adiabatic from 800-600 mb, without any significant changes in relative humidity, equivalent potential temperature ($\theta_e$), or wet-bulb potential temperature ($\theta_w$) with height. Although traditional stability measures suggest a statically stable environment, Oravetz’s (2003) re-analysis of their event did reveal appreciable convective available potential energy for an elevated parcel. This seeming conundrum was resolved with the refinement of the concepts of elevated convection and symmetric instability.

Moore and Lambert (1993) discussed the effect that CSI has in winter storms. It was noted that CSI is often a significant source for the elevated instability needed to produce mesoscale banded precipitation associated with winter storms. CSI-related precipitation is often oriented parallel to thickness contours, with a component of motion toward warm air, and with widths of less than 100 km. When a cross-section is constructed normal to the thermal wind, CSI occurs where lines of pseudo-angular geostrophic momentum ($M_g$) are more horizontal with respect to lines of $\theta_e$, and the relative humidity is greater than 80% (to allow an assumption of saturation). CSI tends to occurs in regions of large vertical shear, large anticyclonic shear, and low static stability. This profile occurs when large vertical shear “flattens” $M_g$ surfaces, anticyclonic shear creates weak inertial stability, and low static stability “pushes” $\theta_e$ surfaces toward the vertical. The equation for equivalent potential vorticity (EPV) includes all of these factors; it can be used to diagnose regions of potential CSI but does not necessarily mean CSI is being released. Simply, if EPV is negative in a convectively stable region, then CSI is present and absolute vorticity will be negative on a $\theta_e$ surface. In an environment where CSI and convective instability (CI) are occurring simultaneously, vertical motion as a result of CI will dominate over that resulting form of CSI. Therefore, the release of CI, and not CSI, will result in the elevated convection.
Research performed by Emanuel (1985) proposed a coupled dynamical relationship between frontogenesis and moist symmetric stability for producing mesoscale precipitation bands. With gridded model output available at National Weather Service offices, frontogenesis and CSI can be diagnosed operationally and incorporated into the forecast process (Weismuller and Zubirck 1998). Indeed the research performed by Nicosia and Grumm (1998) gave evidence that frontogenesis and CSI may work in concert to produce mesoscale precipitation bands in extratropical cyclones. They showed that both frontogenesis and CSI (and/or low moist symmetric stability) were present during major northeastern United States snowstorms that possessed mesoscale snowbands. They also found that frontogenesis leads to vertical motion by inducing a thermally direct ageostrophic circulation. Frontogenesis was calculated in GEMPAK 5.4 using total winds (geostrophic plus ageostrophic) instead of the geostrophic winds alone, because the ageostrophic winds are important near frontal zones, especially strong frontal zones.

EPV, also calculated from a built-in GEMPAK function for cross-sections normal to the isotherms, is defined as:

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

(2.1)

with the x direction representing the direction perpendicular to the thermal winds with x increasing in the direction of warmer air. In the baroclinic atmosphere CSI can be present when EPV is negative. The first product inside the brackets is the contribution to EPV from the vertical wind shear and the horizontal temperature gradient. When the vertical wind shear and associated horizontal temperature gradient are large, \( \frac{\partial M_g}{\partial p} \) will be more negative with the \( M_g \) surfaces attaining a more shallow slope in the x-p plane. The contribution to EPV from the first term, in this case, will be more negative and there will exist a chance that \( \theta_e \) will slope more steeply than the \( M_g \) surfaces a necessary condition for CSI. The second product inside the brackets of the EPV equation is a small, negative number, since absolute
vorticity is almost always positive in the northern hemisphere, the second product will be a smaller negative number than the first product and EPV will tend to be negative.

It is also possible that CI can be present in conjunction with CSI when EPV is negative. If a layer is moist statically unstable ($\theta_e$ decreasing with height) then the second product in the equation will be positive and contribute to negative EPV when CI is present. Upright convection tends to dominate upon the release of CI, whereas slantwise convection can be associated with the release of CSI. It is also possible that the release of CSI might trigger the release of CI resulting in upright convection. Thus, for slantwise convection to occur, the atmosphere should be statically stable within a saturated region of negative EPV.

In the cases studies performed by Nicosia and Grumm (1998), the cross-sections were taken north of each surface low perpendicular to the 700-hPa height contours at various forecast times moving northeast with each surface low. The 700-hPa level was chosen for the frontogenesis calculations (in the x-y plane) since a distinct maximum in frontogenesis formed near 700 hPa in the cross-sections with each case. Enhanced vertical motion from frontogenesis occurred parallel to the isotherms, with the enhanced low-level ageostrophic winds oriented perpendicular to the isotherms. In this way, it is postulated that negative EPV can develop in close association to a frontogenetic region during cyclogenesis. Nicosia and Grumm (1998) showed that frontogenetic forcing and negative EPV were important mechanisms for producing mesoscale bands of heavy snowfall in the case studies. The negative EPV was primarily associated with CSI; with regions of CI for short periods of time. EPV was found to be reduced in the zone where the dry tongue jet at midlevels overlaid the low-level easterly jet or cold conveyor belt (CCB), north of the surface cyclone, and to the warm side of the frontogenesis maximum. The synergy between frontogenesis and negative EPV is postulated to be
crucial for the development of long-lived mesoscale snowbands.

Novak and Horwood (2002), examined a New England snowstorm from 5-6 February 2001. The snowfall was banded in appearance with one primary band throughout the snowstorm. Data from the WSR-88D, as well as the 22-km Eta model analysis, were used to diagnose dynamic features during this event. On the synoptic scale, at 1800 UTC 5 February 2001, there was surface cyclone development off the Mid-Atlantic coast in response to a strong shortwave disturbance rounding the base of a 500-hPa trough. By 0000 UTC 6 February 2001 the storm system was deepening off the New England coast. Rapid cyclogenesis occurred as the 500-hPa trough became negatively tilted. A closed circulation existed at 700-hPa, which maximized deformation northwest of the surface cyclone. Mid-level deformation acting on the ambient temperature gradient contributed to intense mid-level frontogenesis, which served as forcing for the primary banded feature. The frontogenesis maximum also closely correlated to the primary band location. Examination of the cross-section taken perpendicular to the band at 2100 UTC 5 February 2001 revealed a sloping frontal structure in the saturated equivalent potential temperature (θe) field, with a deep layer of negative equivalent potential vorticity (EPV) near the middle of the cross-section. Also, CSI was possible because the environment is nearly saturated below 500 hPa. The θe field showed a transition from elevated CI in the far southeast where the fine-scale bands were initiating, to possible CSI near the middle of the cross-section where the primary band was located. Intense frontogenesis was found along the frontal zone, forcing the full wind and ω tangent to the cross-section, and leading to vertical motion.

In summary of this case study, synoptic flow configurations were favorable for mesoscale banding during cyclogenesis as the formation of a closed mid-level circulation maximized deformation northwest of the surface cyclone. The cross-section through the banded features showed the environment of the finescale bands
were characterized by elevated CI. The primary band was found in an environment characterized by weak symmetric stability. CI was the main cause of banding, and orographic effects were secondary in this case.

The main focus of the work by Novak et al. (2002) was how mesoscale bands can greatly affect the intensity, timing and subsequent accumulation of precipitation. As discussed by Nicosia and Grumm (1999), several case studies have established the occurrence of significant mesoscale banding associated with frontogenesis and small moist symmetric stability. A composite study of mesoscale bands was performed; these composites document the environments of the different types of bands through assessment of frontogenetical forcing, stability, and moisture. For the Novak et al. (2002) study, only systems exhibiting precipitation greater than 25 mm or 12.5 mm liquid equivalent in the case of frozen precipitation during a 24-hour period at a location within the study area were selected as cases for the study. Three dominant band categories are proposed (single, multi-band, and narrow cold frontal) with transitory structure serving as a bridge between the significantly banded structures and nonbanded cases. This method was applied to 88 cases for which composite radar data were available (Novak et al. 2002). The composite study focused on single-banded structures, with 70% of the single bands exhibiting some portion of their length in the northwest quadrant relative to the surface cyclone, occurring in the comma-head portion of the cyclone, while 30% were found ahead of the cyclone, a finding similar to that of Market et al. (2002). A strong jet is also present rounding the base of a trough, and a weaker jet is found in the confluent region over southeastern Canada. A large-scale deformation zone associated with frontogenesis is present northwest of the surface cyclone. Significant precipitation was associated with cases included in this composite, but the absence of a closed midlevel circulation precluded deformation and frontogenesis northwest of the surface cyclone, limiting band development. Fron-
togenetical forcing is readily evident in both cross-sections, with a sloping frontogenesis maximum found within each frontal zone. A sloping updraft is found on the equatorward flank of each frontal zone, consistent with a thermally direct circulation induced by frontogenesis. However, the northwest composite cross-section exhibits a stronger, narrower, and deeper updraft than the nonbanded composite cross-section. Northwest composite frontal zones exhibit a more upright configuration than the nonbanded composite frontal zone, consistent with smaller static stability. With this in mind it can be presumed that the closed midlevel circulation resulted in the development of a confluent asymptote northwest of the surface cyclone, leading to frontogenesis and subsequent band development. Also, the upright nature of the northwest composite frontal zone supported banded structures as well.

In a recent work on banded precipitation (Banacos 2004), the goal was to describe a “mode” of banded precipitation that occurs with a distinct col point in the horizontal flow aloft. An emphasis was placed on observational aspects of these events, which may aid in forecasting their timing and location, particularly in the short-range forecast period (0-12 hour). For cases with mesoscale banding, the following characteristics are present: horizontal deformation in the 850-500 mb layer, with a deformation zone often most pronounced near 700 mb. Large-scale deformation zones can be broken into two categories: (1) those occurring in conjunction with strong extratropical cyclogenesis, and (2) those associated with strong east-west oriented frontal zones with modest or minimal surface cyclone development. In the former case, the deformation develops as the cyclone deepens from an open wave to a closed low in the middle-troposphere. A diffluent flow region develops in the northwest quadrant of the cyclone, sometimes west of a surface inverted trough, where the low-level cold conveyor belt (CCB) splits.

In the frontal/weak cyclogenesis pattern, the 700-mb flow is often character-
ized by a positively tilted trough and downstream confluent flow several hundred kilometers to the north of a polar or arctic front. A weak low center is usually found along the surface boundary, with mesoscale banding occurring to the north and/or northeast of the cyclone. In cases of very strong low-level static stability, a surface low may not exist, making the situational awareness of the forecaster to heavy precipitation potential generally lower than in the strong cyclogenesis case. Deformation in the latter case is achieved through confluent 700-mb flow parallel to the baroclinic zone. Speed convergence at the nose of a southerly low-level jet, which increases the horizontal temperature gradient on its poleward edge, may also be present. The frontal/weak cyclogenesis pattern is common in continental areas, particularly the Great Plains and Great Lakes regions during the winter.

In order to identify horizontal deformation for the purpose of mesoscale banding potential, the two-dimensional frontogenesis function, is defined (Petterssen 1956) as follows:

\[ \mathcal{Z} = \frac{d}{dt} |\nabla \theta| \]  

(2.2)

where \( \mathcal{Z} \) is the scalar form of frontogenesis, defined as the total (lagrangian) tendency of the potential temperature gradient following the motion of the parcel. Observational experience suggests that the kinematics of the horizontal flow often play a dominant role in the component parallel to the isentropes. The component of frontogenesis normal to the isentropes \( F_n \) is likely more focused on smaller spatial scales because it can be enhanced by convergence (largely a mesoscale process), whereas \( F_s \) is associated with vorticity and deformation, which can be large on the synoptic-scale and associated with geostrophic motions.

Banacos (2004) noted that, since synoptic-scale divergence is small, deformation often initiates the banding process, with convergence increasing once thermal wind balance is disrupted and a secondary, ageostrophic circulation is established. Mesoscale bands are therefore well-approximated by horizontal deforma-
tion zones because they determine the approximate location of frontogenetic forcing that results in a narrow linear zone of ascending motion. There is evidence that suggests that mesoscale bands are most robust in association with frontal-scale circulations that are strong and well-defined through the 850-500 mb layer, but also when it occurs spatially separate from the forcing associated with \( F_s \). This can result in robust mesoscale bands with little or no precipitation other than that associated with the bands, a situation that can result in extreme precipitation accumulation gradients. According to Banacos (2004), if the deformation zone is colocated with a horizontal shear zone, then \( F_s \) can be large and the organization of precipitation will either include bands embedded in a larger field of precipitation or will be non-banded in nature. The best scenario from a banding perspective would appear to be strong deformation (and convergence) with minimal translation or vorticity, such that a col point exists in a system-relative and an absolute sense, creating a very favorable environment not only for banding, but also for long-lived bands affecting one specific area.

Indeed, mesoscale banding in cold season events is responsible for snow rates of up to 3-5 inches an hour and their limited scale can create extreme gradients in snow accumulation (Novak et al. 2004). In this paper a time-dependent strategy used to anticipate cold-season mesoscale band formation is discussed, along with the incorporation of emerging conceptual models of band development, model guidance, and observational tools available to forecasters. The authors examined the 25 December 2002 snowstorm for their case study. When examining a case of possible mesoscale banding approximately 24-48 hours before the event it is important to note the development of mid-level deformation, which is maximized north and west of the surface cyclone as the mid-level circulation develops and/or in diffluent flow ahead of the surface cyclone. Deformation is important since, in the presence of a temperature gradient, it can contribute to frontogenesis, which
serves as mesoscale forcing for banded precipitation (Novak et al. 2004).

12-24 hours before the event, when confidence in the synoptic-scale flow is established, the forecaster can examine the plan view as well as cross-sectional environment to identify deformation zones and frontogenesis maxima. Since frontogenesis in the presence of near-neutral moist gravitational or symmetric stability has been identified as the primary forcing mechanism for band development, assuming sufficient moisture over the location and timing of the coincidence of these parameters can outline the banded threat area. Observational studies have suggested that bands tend to form near the location where a layer of small moist gravitational or symmetric stability lies just above the mid-level frontogenesis maximum (Trapp et al. 2001; Novak et al. 2002). Operational experience has shown that no single threshold value can be correlated to band development, since the gravitational and symmetric stability modulates the frontogenetical response, and other factors, such as microphysics and moisture availability, can alter precipitation development (Novak et al. 2004). Similar challenges are noted when interpreting the moist symmetric stability as the calculation of saturation equivalent potential vorticity (EPV) which has been shown to be quite sensitive to the choice of representative wind (e.g., Schultz and Schumacher 1999; Clark et al. 2002) and model resolution.

In summary, each of these studies provides valuable support for the analysis which follows, although there has been a recent emphasis on banded winter precipitation northwest of a surface cyclone. Most of the precipitation in this case occurred to the northeast of a surface cyclone, without significant benefit from a deformation zone. Regardless, the recent literature has been quite useful for guiding much of the ensuing discussion.
Chapter 3

Methodology

This study relies upon both observed data and a special run of the Workstation Eta (hereafter the WS-Eta) mesoscale numerical weather forecasting system which is based upon the operational Eta model (Black 1994) at the National Centers for Environmental Prediction (NCEP). In this section, we explore in detail the observed data employed in this study, how it is treated when analyzed objectively, the architecture of the WS-Eta, and the diagnostics generated from both the objectively analyzed data and the model output fields.

3.1 Data

Surface data from 227 stations, as well as upper-air data from up to 82 North American radiosonde observation stations were used in the initial work done by Oravetz (2003). In this investigation, a component event of Oravetz (2003) is used, a case from Eau Claire, Wisconsin, 23 March 1966. With this case study, surface charts were created and hand analyzed every 3 h, along with upper-air analyses every 12 hours, although not all are shown.

Rawinsonde data were objectively analyzed (Barnes 1973) using the General Meteorological Package (GEMPAK) as discussed by Koch et al. (1983). The grid was centered on $39.8^\circ$N and $99.5^\circ$W using a polar stereographic grid situated over North America and possessing 32 points in the east-west direction and 25 points
in the north-south direction (Fig. 3.1). The horizontal grid-spacing was 150 km, with a vertical spacing of 50 mb, from 1000 to 300 mb, inclusive. Two passes are made through the data, and a search radius of \( 25 \Delta n \) was used, as suggested by Koch et al. (1983). The spectral response function, \( (\gamma) \), was set to 0.25 for those analyses. Fig. 3.2 shows the fraction, \( D_o \), of a given wavelength, \( \lambda \), that is resolved. The parameter \( \lambda^* \) is a ratio of \( \lambda \) to some prescribed wavelength \( L \), which is given as \( 2\Delta n \), the smallest theoretical distance over which a wave may be resolved. Here, \( \Delta n \) is 150 km, making \( L = 300 \) km. Figure 3.2 shows that for a \( \lambda^* \) of 3.0 about 76% of a 900-km wave is resolved.

### 3.2 Special Diagnostics

#### 3.2.1 Frontogenesis

Petterssen (1956) defined frontogenesis as the tendency toward the formation of a discontinuity, known as a front, or the intensification of such discontinuity. This
Figure 3.2: The spectral response curve (for $\gamma = 0.25$), showing what fraction of a wave is resolved ($D_o$) for a given wavelength ($\lambda^*$). For example, for a wave of $12\Delta n$ (1800 km when $\Delta n=150$ km), $\lambda^*$ has a value of 6.0, and 90% of the wave is resolved.

can be represented by (2.2), which may be rewritten as:

$$\mathcal{Z} = \frac{1}{2} \left| \nabla \theta \right| \left[ \text{def} \left[ \cos (2\beta) \right] - \text{div} \right]$$  
(3.1)

where

- $\text{def} = $ deformation of the horizontal wind field,
- $\text{div} = $ divergence of the horizontal wind field, and
- $\beta = $ the angle between the axis of dilatation and the isentropes.

According to Carlson (1991) frontogenesis is not merely the passive concentration of isotherms by an advecting wind, but a type of instability in which the differential advection of temperature and momentum feedback, via ageostrophic motions, produce a singularity known as a front. Thus, the effect of horizontal deformation alone is to promote frontogenesis when the axis of dilatation lies within $45^\circ$ of the isotherms, and to promote frontolysis when the axis of dilatation forms an angle of $45^\circ$ to $90^\circ$ with respect to the isotherms. Convergence acts frontogenetically, and divergence acts frontolytically (Bluestein 1986).
3.2.2 Equivalent Potential Vorticity

Equivalent potential vorticity (EPV) can provide quantitative values to assess the presence of CSI (Moore and Lambert 1993). The release of CSI is thought to culminate in the creation of slantwise convection. To illustrate how slantwise convection works, a cross section display of equivalent potential temperature ($\theta_e$) and the quantity $M_g$. Constant $M_g$ surfaces, may be viewed as surfaces of constant inertial stability. A saturated parcel displaced along an $M_g$ surface will maintain its $\theta_e$. If it encounters an environment with lower temperature, that is, when the slope of the $\theta_e$ lines are steeper than the $M_g$ lines, it becomes unstable and "convects" (McCann 1995). The Moore and Lambert (1993) expression for EPV, shown in this work as (2.1), is best applied with an analysis of relative humidity in a cross-section perpendicular to the thermal wind, and greater than 80%. Considering that this approach can be cumbersome in an actual forecast environment, McCann (1995) developed a three dimensional form of EPV that eliminates the need to compare the slopes of $M_g$ and $\theta_e$ on a cross-section.

\[
EPV = g \left[ \frac{\partial \theta_e}{\partial x} \frac{\partial v_g}{\partial p} - \frac{\partial \theta_e}{\partial y} \frac{\partial u_g}{\partial p} - \left( \frac{\partial v_g}{\partial x} - \frac{\partial u_g}{\partial y} + f_k \right) \frac{\partial \theta_e}{\partial p} \right] \quad (3.2)
\]

The three dimensional EPV allows for an easier determination of the atmospheric stability, either vertically or slantwise. If EPV is negative where $\theta_e$ decreases with height then the atmosphere is primarily potentially unstable. If EPV is negative and $\theta_e$ increases with height (potentially stable) then potential symmetric instability (PSI) is present; again, CSI can be assumed if the relative humidity approaches a value greater then 80%. In the situation where McCann’s three dimensional EPV shows the lapse rate to be slightly stable and the horizontal temperature gradient is strong, the large negative value of the first term more than compensates for the small positive value of the second term, and slantwise convection results (McCann 1995).
3.2.3 Static Stability Tendency Equation

The static stability tendency equation:

\[
\frac{\partial}{\partial t} \left( -\frac{\partial \theta}{\partial p} \right) = -\frac{\partial}{\partial p}(-\nabla \cdot \mathbf{v}_p \theta) - \omega \frac{\partial}{\partial p} \left( -\frac{\partial \theta}{\partial p} \right) - \frac{\partial}{\partial p} \left( \frac{\theta}{c_p T} \frac{dQ}{dt} \right) \tag{3.3}
\]

represents the local changes in static stability due to changes in four different terms: differential temperature advection (first on the right-hand side), vertical advection of static stability (second term on the right-hand side), vertical stretching or shrinking of an air column (third on the right-hand side), or differential diabatic heating (fourth on right-hand side; Bluestein 1992).

When the differential temperature advection term is considered, then it can be assumed that it is equivalent to geostrophic stability advection and differential ageostrophic temperature advection. In a quasigeostrophic atmosphere, ageostrophic temperature advection is neglected, and hence only geostrophic stability is considered (Bluestein 1992). We note that the geostrophic stability advection can only transport stability from one place to another. It can neither create or destroy stability. Thus, the ageostrophic wind is crucial in this term for affecting changes in stability. Also, the process of vertical advection cannot create or destroy stability. More importantly static stability can be created or destroyed through the stretching or shrinking of an air column or through differential diabatic heating (Bluestein 1992). For example, static stability is destroyed as cold, continental air flows over a relatively warm ocean, or if there is horizontal convergence. Static stability is created as a warm moist air mass flows over a relatively cold land mass, or if there is horizontal divergence (Bluestein 1992), thus making the level of non-divergence ineffective. Low-level convergence can be very effective at reducing the static stability associated with a capping inversion and eventually set the stage for cumulus convection (Bluestein 1992).

In this study, each term is calculated for the 700-600 mb layer using a 6-hr cen-
tered, time difference. All of the terms come from the output fields except for the last (differential diabatic heating) term, which is solved for as a residual. Perturbation analysis of the kind used by Moore (1985) revealed that error accounts for no more than 20% of the differential diabatic heating term.

### 3.2.4 Slantwise Convective Available Potential Energy

Slantwise convective available potential energy (SCAPE) is defined as the "positive area" between the environmental virtual potential temperature and a "lifted" parcel’s virtual potential temperature that follows a constant $M_g$ surface (Emanuel 1983; Snook 1992). The acceleration of a parcel in a positive SCAPE environment is identical to parcels in a more conventional positive convective available potential energy (CAPE) situation. The only difference between SCAPE and CAPE is the more "horizontal" extent of convective parcels with positive SCAPE (McCann 1995). In the real atmosphere, a value for CAPE can exceed 5000 J kg$^{-1}$, whereas a SCAPE of 500 J kg$^{-1}$ is significantly large (Snook 1992). This leads to the assumption that weather accompanying slantwise convection is less "intense" than weather with upright convection. However a forecaster may take advantage of EPV’s ability to detect multiple convective modes and use it as an all-purpose convection diagnostic tool in a saturated environment (McCann 1995).

### 3.3 Numerical Model Simulation

An example of the WS-Eta is a study by Jewett et al. (2002), wherein an ensemble forecasting strategy was used to account for cool-season forecasts that result from incorrect placement and timing of the mesoscale regions of mixed precipitation and of heavy snowfall and blizzard conditions. In this ensemble forecast; the Penn State/NCAR Mesoscale Model (MM5), the Weather Research and Forecasting Model (WRF), the NCEP WS-Eta and the NCEP Regional Spectral Model were
included. The MM5, WRF, and WS-Eta were initialized with NCEP operational Eta analysis (Jewett et al. 2002). Since the current study incorporates the WS-Eta only, its physics schemes will be discussed.

In the NCEP operational Eta, the schemes that are used include the Mellor-Yamada (1982) planetary boundary layer scheme, and the Betts-Miller-Janjic cumulus parameterization scheme (Betts and Miller 1986, Janjic 1994). The simulation presented here used NCEP/NCAR reanalysis data for initial and lateral boundary conditions. In this work, the WS-Eta is designed with a grid spacing of 32 km, and 45 vertical levels. Four soil levels are incorporated into the model. Along the west-east axis, there are 54 mass grid points along the first row, and 83 rows in the north-south direction. For a 32 km horizontal grid-spacing, a time step of 90 seconds is used. The grid spacing in the east-west direction is therefore $0.222^\circ$ and in the north-south direction $0.205^\circ$, with the top of the atmosphere set at 25 hPa.

### 3.4 Methodology Summary

The data used in this case study consist of subjectively analyzed surface and objectively analyzed upper air data for Eau Claire, Wisconsin, and the surrounding region during 22-23 March 1966. This study will also examine and discuss how frontogenesis, EPV, and the static stability equation can lead to determining the presence or nature of unstable layers in the atmosphere, and those mechanisms that might act to release instability. The WS-Eta model was run on this case not only to provide a model comparison of the synoptic conditions to that of the actual observed fields, but also to expand on the limited information afforded by the objectively analyzed upper-air data in 1966.
Chapter 4

Case Study: 23 March 1966

This chapter constitutes a case study of a blizzard in Wisconsin in late March of 1966. In particular, we shall focus our attention on a 12-hour period that encompasses the most significant weather at Eau Claire, Wisconsin. From 0300 UTC to 1200 UTC on 23 March 1966, snow with thunder was reported at Eau Claire for 9 non-consecutive hours, as shown by the meteorogram for the period 0000 UTC 23 March 1966 through 1500 UTC 23 March 1966 (Fig. 4.1). Although the accumulated snowfall totaled only 18.5 cm (7.3 inches), at Eau Claire over 27+ hours, the long-lived nature of convection with this event demands further scrutiny. In the evaluation of this case, the 48-hour snowfall accumulation has been calculated and is presented in Fig. 4.2. Notice the presence of a large band of snow from northern Wisconsin to western Iowa; within this large band, there are smaller mesoscale bands embedded causing a snow gradient to be distinctly identified.

4.1 0000 UTC 23 March 1966

The first time period that is analyzed for this case study is 0000 UTC 23 March 1966. This is three hours before the first report of thundersnow was issued by human observers at the Eau Claire Municipal Airport.
Figure 4.1: Meteorogram for Eau Claire, Wisconsin, for the period 0000 UTC to 1500 UTC on 23 March 1966. The top bar depicts the temperature (solid) and dew point (dashed) trends (°F); the second is a barograph of sea level pressure (mb); the middle bar shows the wind speed trend (solid line; knots) while the shafts with barbs show the wind direction and speed, respectively, as one might see in a standard station model; the fourth bar depicts the visibility trend (statute miles); the bottom bar shows standard symbols for weather type (above) and sky condition (below).
Figure 4.2: A map showing the 48-hour total snowfall accumulation (0000 UTC 22 March 1996 to 0000 UTC 24 March 1966) in inches from cooperative climate observation stations. Contours at 1, 5, 10, and 15 inches (2.5, 12.7, 25.4, and 38.1 cm).
Figure 4.3: The surface analysis valid at 0000 UTC 23 March 1966. Sea level pressure (solid; every 4 mb) and temperature (dashed; every 10°F) are analyzed subjectively. Eau Claire is identified by the orange dot in the station circle.

4.1.1 Synopsis

Examination of the surface analysis from 0000 UTC 23 March 1966 (Fig. 4.3) reveals a mature cyclone southwest of Eau Claire, Wisconsin, thereby placing the thunder-snow northeast of the surface cyclone. A warm front extends from northwestern Missouri through northern Indiana, putting Eau Claire north of the warm frontal boundary.

The cold front extends from northwest Missouri and continues to the south through southwestern Missouri and into Oklahoma. The central pressure of the cyclone is 997 hPa with a pressure of 1012 hPa at Eau Claire, with winds from the north-northeast at 10-15 kts, temperatures are well above freezing at 43°F with drizzle and thunderstorms reported in the area.

The 850-mb height and isotherm map for 0000 UTC 23 March 1966 (Fig. 4.4),
Figure 4.4: Analysis of the 850-mb geopotential height (solid; every 30 gpm) and temperature (dashed; every 5°C) valid at 0000 UTC 23 March 1966.

shows warm air advection (WAA) just south of the Eau Claire area, in a wide band that covers eastern Iowa, northern Illinois, and western Michigan. WAA just south of the event location signifies where the warm conveyor belt (WCB) is ascending into the upper Midwest around the cyclonic curvature of the upper-level low.

Fig. 4.5 shows the 700-mb \( \theta_e \) analysis; notice the slight presence of the trowal (Trough of Warm air Aloft) upstream (southwest) of our event location. The trowal can be identified as the axis of the mid-level \( \theta_e \) ridge. It defines the apex of the warm sector in the pre-occlusion extratropical cyclone (ETC), and the axis of the occluded front later in the life of an ETC. In the latter instance, the trowal is usually along, parallel to, and downstream from the axis of the surface occluded front. The trowal airstream (Martin 1999) is part of the WCB that turns cyclonically into the low pressure system.
Figure 4.5: Analysis of the 700-mb $\theta_e$ (solid; every 2 K) valid at 0000 UTC 23 March 1966. Notice the presence of the trough over eastern Nebraska.
Figure 4.6: Analysis of the 850-500 mb mean relative humidity, a linear average of 850, 700, and 500-mb relative humidities at 0000 UTC 23 March 1966.

The map of 850-500 mb mean layer relative humidity (Fig. 4.6), shows an area where the mean layer relative humidity is 70% or greater, showing that there is sufficient moisture throughout a deep enough layer for the production of precipitation. This map shows that not only Eau Claire, but the entire area along and north of the warm frontal boundary, has significant moisture available for precipitation.

500 mb heights and vorticity (Fig. 4.7) reveal the presence of two troughs. The first, located over Missouri, is a relatively weak (5520 gpm) negatively tilted trough that does not significantly affect this particular event. The second, deeper trough is positively tilted, and located through portions of Kansas, the panhandle of Ok-
Figure 4.7: Analysis of the 500-mb geopotential height (solid; every 60 gpm) and absolute vorticity (dashed; every 2 $\times 10^{-5} \text{s}^{-1}$) valid at 0000 UTC 23 March 1966.

lahoma and Texas and into New Mexico. This arrangement suggests a system that will continue to deepen as it translates. Indeed, the trough location evinces a favorable baroclinic structure.

Fig. 4.8 is the analysis of 400-700-mb layer Q-vector divergence. Q-vectors describe the changes that a parcel’s potential temperature gradient vector undergoes as the parcel moves with the geostrophic wind. Here, we employ the layer Q-vector which, based upon the GEMPAK code, we write here in a simplified form as

$$
\mathbf{\tilde{Q}}_{\text{layer}} = \frac{\partial p}{\partial \theta} \left( \left[ \frac{\partial \mathbf{V}_g}{\partial x} \cdot \nabla \theta \right] \hat{i} + \left[ \frac{\partial \mathbf{V}_g}{\partial y} \cdot \nabla \theta \right] \hat{j} \right) \tag{4.1}
$$

where $\mathbf{V}_g$ is the geostrophic wind, and $\theta$ is the potential temperature. The divergence of this vector is then formed simply as $\nabla \cdot \mathbf{\tilde{Q}}$, resulting in units of $\text{mb m}^{-2} \text{s}^{-1}$. With this formulation, when the Q-vector divergence is negative, then forcing
Figure 4.8: Analysis of the 400-700 mb layer mean Q-vector and its divergence (solid; every $5 \times 10^{-14}$ mb m$^{-2}$ s$^{-1}$) at 0000 UTC 23 March 1966

for upward vertical motion would result and Q-vector fields appear to converge. At 0000 UTC there is strong Q-vector convergence to the southwest of Eau Claire. This suggests strong forcing for upward motion in association with the surface cyclone. With the presence of synoptic-scale upward motion, and ample moisture in the layer, the ingredients needed for significant snow are becoming assembled.

The 300-mb map (Fig. 4.9), shows a meridional trough through western Kansas, the panhandle of Oklahoma and into Texas. This is in association with a curved 70-kt jet streak. Eau Claire is placed in the left exit region of this curved jet on the poleward side. This is typically an area of enhanced upward motion and convergence (Moore and Van Knowe 1992), which is verified by the 400-700 mb Q-vector divergence field (Fig. 4.9).

Isentropic surfaces are also used in order to diagnose storm-relative ($\tilde{V} - \tilde{C}$)
Figure 4.9: Analysis of 300 mb geopotential height (every 120 gpm) and isotachs (dashed; every 20 kts starting at 50 kts; shaded over 70 kts) for 0000 UTC 23 March 1966.
airstream structures and their associated descending and ascending motions; here \( \vec{V} \) is the actual wind, and \( \vec{C} \) is the storm motion. The first surface that was analyzed was the 280 K isentropic surface (Fig. 4.10) at 0000 UTC 23 March 1966. On this surface the range in height is from 900 mb through 700 mb. The streamlines on this surface denote where the cold conveyor belt (CCB) is descending from higher levels toward the ground, and beginning to swirl around the backside of the surface low pressure, before intersecting the surface. The 296 K isentropic surface (Fig. 4.11), from 0000 UTC 23 March 1966, reveals pressures from 950-400 mb. The WCB is apparent as warm air from the lower Ohio River Valley is being pulled up and around the surface low. Starting from the 950-mb level, warm air is ascending and wrapping around the surface low to a level of 550 mb. This image is a good representation of the vertical structure surrounding a cyclonic system on an isentropic surface. Finally, on the 312 K isentropic surface (Fig. 4.12) the levels start at 600 mb and continue to 300 mb. This image gives a representation of the dry descending airstream penetrating toward the surface over the southern plains, wrapping into the cyclone and providing contrast with the warm moist air on lower surfaces that are being advected and lifted into the cyclone. Yet, there is ascent diagnosed at this time over Eau Claire. While unsaturated, the air is not terribly dry, and the cooling from ascent in this flow only serves to drive up the relative humidity. While not in strict conformance with Nicosia and Grumm (1998), this situation certainly supports the production of precipitation.

4.1.2 Mesoscale Analyses

At 0000 UTC 23 March 1966 a cross-section taken from Kenora, Ontario, Canada (YQK), to Lacon, Illinois (C75), places Eau Claire, Wisconsin, just right of center, with the cross-section perpendicular to the 1000-500 mb thickness contours. The cross-section (Fig. 4.13) depicts \( M_g \) and \( \theta_e \). In an ideal thundersnow environment
Figure 4.10: An analysis of the 280 K isentropic surface with pressure (bold solid; every 50 mb) and storm-relative streamlines (where $\vec{C}$ has components of $u = +10.9 \text{ m s}^{-1}$, $v = +6.4 \text{ m s}^{-1}$), valid at 0000 UTC 23 March 1966.
Figure 4.11: An analysis of the 296 K isentropic surface with pressure (bold solid; every 50 mb) and storm-relative streamlines (where $\vec{C}$ has components of $u = +10.9 \text{ m s}^{-1}$, $v = +6.4 \text{ m s}^{-1}$), valid at 0000 UTC 23 March 1966.
Figure 4.12: An analysis of the 312 K isentropic surface with pressure (bold solid; every 50 mb) and storm-relative streamlines (where $\vec{C}$ has components of $u = +10.9 \, \text{m s}^{-1}$, $v = +6.4 \, \text{m s}^{-1}$), The thick red line represents the cross-section from Kenora, Ontario, Canada, to Lacon, IL, as seen in Fig. 4.13, valid at 0000 UTC 23 March 1966.
Figure 4.13: A cross-section from Kenora, Ontario (YQK) to Lacon, Illinois (C75) from 0000 UTC 23 March 1966. Dashed lines depict $M_g$ (every 5 kg m s$^{-1}$) and solid contours depict $\theta_e$ (every 2 K). The shaded region denotes where EPV is less than $1 \times 10^{-7}$ K m$^2$ kg$^{-1}$ s$^{-1}$. The black line represents the location of Eau Claire; notice how this line intersects the shaded area representing negative EPV.

The shaded area in the figure (4.13) represents negative EPV; when negative EPV is present conditional symmetric instability (CSI) is also possible.

In order to better diagnose the area of instability, a vertical profile through the objectively analyzed radiosonde data was prepared for Eau Claire, Wisconsin. The first profile taken at Eau Claire was of EPV (Fig. 4.14); in this profile EPV approaches zero and becomes slightly negative between 800-700 mb. This analysis shows that moist potential vorticity is present at Eau Claire three hours before the first report of thundersnow occurred. The presence of EPV around zero in a po-
tentially stable environment suggests the presence of weak symmetric instability in that layer. The next profile for Eau Claire is that of frontogenesis (Fig. 4.15), showing the vertical location where compression/stretching of the horizontal $\theta$ gradient is most likely to occur. In this profile, the frontogenesis maximum is located between 825-750 mb which resides beneath the area of negative EPV. This arrangement places the best frontogenesis ahead of the frontal zone, with the best forcing for ascent to the south and farther aloft of Eau Claire, in the presences of the negative EPV region.

The last profile that was taken at Eau Claire is that of $\theta_e$, (Fig. 4.16). In this vertical profile notice how $\theta_e$ almost doubles over with height. This indicates a moist neutral environment, and that the thundersnow development in the next three hours likely will be slantwise in nature. If the $\theta_e$ profile had doubled over with height this would lead us to believe that upright convection was the dominant cause of the thundersnow.

### 4.2 1200 UTC 23 March 1966

In the previous section a synopsis of the conditions three hours before the first thundersnow report was analyzed. In this section, conditions twelve hours later, when the last report of thundersnow was observed will be analyzed.

#### 4.2.1 Synopsis

In the 1200 UTC 23 March 1966 surface analysis (Fig. 4.17), the cyclone has tracked to the northeast and matured into an occluded cyclone. This evolution now places Eau Claire north-northwest of the surface cyclone. The warm front extends through southern Michigan, and the cold front is positioned south and through eastern Illinois into Kentucky and Tennessee. The occluded portion of the cyclone extends from extreme northern Illinois into southern Wisconsin. The central pressure has
Figure 4.14: The profile of 3-D EPV \(10^{-7} \text{Km}^2\text{kg}^{-1}\text{s}^{-1}\) valid for EAU at 0000 UTC 23 March 1966. This profile is interpolated from the objectively analyzed (OA) file, and valid for the location of and three hours prior to the beginning of thundersnow reports.
Figure 4.15: The profile of frontogenesis \((K \times 100 \frac{km}{hr} \times 3 \frac{hr}{hr})\) versus \(\log p\) for EAU at 0000 UTC 23 March 1966. This profile is interpolated from the OA file and valid for the location of and three hours prior to the beginning of thundersnow reports.
Figure 4.16: The profile of mean $\theta_e$ (K) versus log $\rho$ for EAU at 0000 UTC 23 March 1966. This profile is interpolated from the OA file and valid for the location of and three hours prior to the beginning of thundersnow reports.
Figure 4.17: The surface analysis valid at 1200 UTC 23 March 1966. Sea level pressure (solid; every 4 mb) and temperature (dashed; every 10°F) are analyzed subjectively.

dropped from 997 hPa at 0000 UTC to 994 hPa at 1200 UTC. The temperature at EAU has dropped from 43°F to 29°F. Thundersnow is being reported at the station, with winds backing to the north-northwest at 20 kts as the occluded portion of the cyclone begins to dominate the weather at the Eau Claire observation station.

Figure 4.18 represents the 850-hPa analysis from 1200 UTC 23 March 1966. In the twelve hours that have transpired from the last analysis the area of WAA has moved to the east of Eau Claire covering a region in extreme eastern Wisconsin, northern Michigan and into Ontario and Quebec, Canada. The WCB is continuing to ascend into Wisconsin, around the cyclonic curvature of the upper-level low present in the north central Plains. The area of WAA suggests that ample moisture and warming aloft are being provided to maintain instability in the temperature profile. We shall return to these points shortly. The geopotential height minimum
has decreased from 1350 gpm to 1320 gpm as the cyclone has deepened over the past twelve hours.

The 700-mb $\theta_e$ analysis (Fig. 4.19) shows that there is still a slight presence of the trowal that was there in the 0000 UTC $\theta_e$ analysis. The trowal has propagated to the north from the 0000 UTC location, and now approximates the axis of the occluded cyclone, and resides directly over the Eau Claire metropolitan area.

Figure 4.20 is the 1200 UTC analysis of the mean layer relative humidity from 850-500 hPa. A layer of moisture continues with a broad region over 70% from eastern Nebraska to Hudson Bay. Eau Claire as well as most of the state of Wisconsin, is in an area of at least 70% mean layer relative humidity, showing that there is still sufficient moisture throughout the layer to continue to support snowfall development.

500-mb heights and vorticity (Fig. 4.21) at 1200 UTC show that the positively
Figure 4.19: Analysis of the 700-mb $\theta_e$ (solid; every 2 K) valid at 0000 UTC 23 March 1966.
Figure 4.20: As in Fig. 4.6, but valid at 1200 UTC 23 March 1966.

A tilted trough from 0000 UTC has deepened and transitioned to become negatively tilted in the 1200 UTC observation. This trough axis is located in eastern Missouri into Tennessee and Alabama, and continues to denote a favorable baroclinic structure.

Figure 4.22 is the 400-700-mb layer Q-vector divergence analysis at 1200 UTC 23 March 1966. Here, the layer Q-vector and its divergence are defined as in (4.1) and its ensuing discussion. The area of Q-vector convergence has weakened over the past twelve hours, but is still present, and weakly favorable over the Eau Claire area. This continues to suggest forcing for upward motion into the upper-level environment. Again, sufficient synoptic-scale upward motion and moisture in the Eau Claire area continue to constitute the proper ingredients to support snowfall.

An analysis of 300-mb heights and isotachs (Fig. 4.23) was also created for 1200 UTC 23 March 1966. In this analysis, the jet streak has intensified, with a jet
Figure 4.21: As in Fig. 4.7, but valid at 1200 UTC 23 March 1966.

Figure 4.22: As in Fig. 4.8, but valid at 1200 UTC 23 March 1966.
Figure 4.23: As in Fig. 4.9, but valid at 1200 UTC 23 March 1966.

maximum of 90 kts near St. Louis, Missouri. The jet streak has developed east of the trough, and although still cyclonically curved, is oriented in a more north-south direction. Although the location of the trough has changed, Eau Claire is still securely positioned in the left exit region of the jet streak. This poleward side suggest an area of divergence aloft and upward motion which supports what was seen on the 400-700-mb Q-vector convergence field.

Isentropic analyses for 1200 UTC 23 March 1966 are performed on the same surfaces (280K, 296K, and 312K) that were used in the 0000 UTC analysis. On the 280 K surface (Fig. 4.24), with heights ranging from 950-700 mb, the CCB is more prominent than on the 0000 UTC analysis. On this analysis the ascending air is beginning to wrap around the location of the surface low. Ascent is suggested even at
this level over Eau Claire. On the 296 K analysis, (Fig. 4.25) the WCB is dominant over the Ohio River Valley, as ascending warm air from 850 mb rises to the 500 mb surface. The strong pressure gradient and nearly perpendicular storm-relative flow supports the idea of strong mid-level ascent over Eau Claire, as well as most of western Wisconsin. Note, too, that the 650-mb level, which is very near the pressure center of the 850-500-mb layer is positioned quite close to Eau Claire; clearly ample moisture is available to this flow (Fig. 4.20. Finally in the 312 K isentropic analysis (Fig. 4.26), descending, drying, air is being wrapped around in a cyclonic pattern between 500-350 mb. Clearly the cold air is descending and warming and capping the potentially colder air at lower surfaces, a situation expected with an occluded surface cyclone. This is also the situation sought by Nicosia and Grumm (1998).

4.2.2 Mesoscale Analyses

The cross-section at 1200 UTC 23 March 1966 taken from Fargo, North Dakota, to Tulip City, Michigan, places Eau Claire almost directly in the center of the cross-section and again perpendicular to 1000-500 mb thickness contours. The cross-section (Fig. 4.27) again depicts $M_g$ and $\theta_e$. Notice how, in the 1200 UTC cross-section (like that at 0000 UTC), the $M_g$ surface have begun to "lay down" when compared to the $\theta_e$ surfaces. The cross-section shows a large shallow pocket of weak negative EPV, with the strongest value occurring at Eau Claire around the 750 mb level. Since negative EPV is still present, CSI is probable, thus suggesting that slantwise convection is the cause of the thundersnow event.

In the updated examination of the vertical profiles of EPV, frontogenesis, and $\theta_e$ at Eau Claire, Wisconsin, the profile of EPV (Fig. 4.28) is negative between 700-600 mb, which suggests the presence of CSI. The vertical profile of frontogenesis (Fig. 4.29) at Eau Claire indicates again a frontogenetic maximum below the EPV
Figure 4.24: As in Fig 4.10, but valid at 1200 UTC 23 March 1966.
Figure 4.25: As in Fig 4.11, but valid at 1200 UTC 23 March 1966.
Figure 4.26: As in Fig 4.12, but bold line represents a cross-section line from Fargo, ND, to Tulip Lake, MI, shown in Fig. 4.27; analysis valid at 1200 UTC 23 March 1966.
Figure 4.27: A cross-section from Fargo, North Dakota (FAR), to Tulip City, Michigan (BIV), from 1200 UTC 23 March 1966. Dashed lines depict $M_g$ (every 5 kg m s$^{-1}$) and solid contours depict $\theta_e$ (every 2 K). The shaded region denotes where EPV is less than $1 \times 10^{-7} K m^2 kg^{-1} s^{-1}$. Notice the black line representing the location of Eau Claire and how this line intersects the shaded area representing negative EPV.
Figure 4.28: The profile of 3-D EPV ($10^{-7} K m^2 kg^{-1} s^{-1}$) valid for EAU at 1200 UTC 23 March 1966. This profile is interpolated from the OA file and valid for the location during the last time that thundersnow was reported.

maximum. Also, the frontogenesis is occurring in a more shallow layer closer to the surface as indicated by Oravetz (2003). The vertical profile of $\theta_e$ (Fig. 4.30) increases steadily, with height indicating a potential stable environment throughout. Near 650-600-mb layer, $\theta_e$ approaches an upright nature, but does not double over with height, further suggesting that slantwise convection has been the main driving force of this long-lived thundersnow.
Figure 4.29: The profile of mean frontogenesis (K 100 km$^{-1}$ 3 hr$^{-1}$) versus log $p$ for EAU at 1200 UTC 23 March 1966. This profile is interpolated from the OA file and valid for the location during the last time that thundersnow was reported.
Figure 4.30: The profile of mean $\theta_e$ (K) versus log $p$ for EAU at 1200 UTC 23 March 1966. This profile is interpolated from the OA file and valid for the location during the last time that thundersnow was reported.
4.3 WS-Eta results

An integral part of the research conducted in this case study, was to find a viable model to simulate the conditions in the upper Midwest for 22-23 March 1966. Investigators were also able to determine that this model could forecast the synoptic conditions necessary for convective snowfall development. In this analysis, the output generated by the WS-Eta not only provided a guidance model capable of forecasting synoptic conditions on 23 March 1966, but also provided insight into how the model could handle a snowfall event of this duration and magnitude.

4.3.1 Performance Assessment

The WS-Eta forecast of synoptic features for 0000 UTC 23 March 1966 (Fig. 4.31), placed the model cyclone over the northwestern section of Missouri, eastern Kansas, and southern Iowa. This simulation compares well to the actual observed features (Fig. 4.3) in terms of cyclone placement and depth. The model placement of the cyclone is too far north and east as compared to the hand analysis of observed conditions from that time period, although there is not a significant difference. The model also allowed the system to develop at a slightly faster rate than from what actually happened. Yet, the reproduction of the physical characteristics of the low are very accurate, with a cyclone strength of 1000 mb and the placement of the warm and cold front being almost exactly the same. The warm front cuts through extreme northern Missouri, into northern Illinois and Indiana. The cold front extends through western Missouri south into the Arkansas-Oklahoma border, although again a little too far to the north. The compactness of the isobars, and minor details occurring in the isobar pattern are very similar around the entire cyclone. To the northwest, the model has plotted isobars up to 1036 mb, whereas the actual event only has a maximum analysis of 1028 mb. Although the WS-Eta has over-analyzed the surface pattern in this location, it should be noted that the
Figure 4.31: The 12 hour WS-Eta forecast of sea level pressure (solid, every 4 mb) and 2 m air temperature (dashed, every 5°F) valid at 0000 UTC 23 March 1966.

1028-mb contour resides in almost the same location on the WS-Eta analysis and 0000 UTC hand-analyzed surface map. To the north the cyclone, both evaluations have included an inverted trough pattern through the Wisconsin area, although the WS-Eta analysis, is more to the east due to the over development of the surface cyclone. The importance of this analysis is how well the model handled the initial conditions of the surface features. Granted the WS-Eta over developed the surface cyclone somewhat, but the WS-Eta analysis is very accurate on most cyclone characteristics; the slight over development is to be expected in any model output.

The comparison of the 36 hour accumulated precipitation field from the WS-Eta (Fig. 4.32) to the actual precipitation Fig. 4.2 field for the study area on 23 March
1966 is also remarkable. Both analyses place a large mesoscale band of precipitation across lower Minnesota, and northern Wisconsin into central Iowa and Wisconsin. Throughout this large mesoscale band, there are smaller, heavier observed bands of snow that developed (Fig 4.2). The heaviest observed snowband that was produced during the event occurred on the eastern Minnesota-Wisconsin boarder where greater than 15 inches of snow fell. On the WS-Eta 36-hour output (Fig. 4.32), this snowfall was captured as one large band and shown as a liquid equivalent of 2.25 inches snow-to-liquid. The ratio of these values is nearly 7:1, which is a viable number in March when snowfall tends to be more moist and dense (Baxter 2003). In the Eau Claire area, a total of 8 inches was reported by observers and shown on the hand analysis. The WS-Eta also shows that Eau Claire received between 1.5 and 1.75 inches of liquid equivalent precipitation. If the rainfall from before the event started is taken into account then this is a valid analysis of the WS-Eta output of total accumulated precipitation. It can be stated that the total accumulated precipitation output by the WS-Eta captured the mesoscale banding that took place and the location where those bands formed, if not the fine-scale structures found in the observed snowfields.

It is important to note that the precipitation scheme obviously activated for this event, however the evaluation of the WS-Eta convective precipitation parameterization scheme did not indicate that convection occurred in the Eau Claire area. Since thunder, lightning, and heavy snowfall were observed in the Eau Claire area, we know that convection did take place. Thus, it can be assumed that a model with a grid spacing of 32 km, which is parameterizing convection over a large area, will not take into account isolated convective episodes. Furthermore, most parameterization schemes were written for tall, moist convection, typically 200-600 mb deep. It would not capture overturning in a 50-100 mb deep convective layer of the kind that was present in this case study. The convective parameterization scheme is
Figure 4.32: The WS-Eta accumulated liquid precipitation forecast (solid, every 0.25" (6.35 mm)); shaded over 1.0" (25.4 mm) for the 36-hour period from 1200 UTC 22 March 1966 to 0000 UTC 24 March 1966.
never turned on. Within the model, none of the precipitation produced is convective, but we understand that to mean that the convective parameterization scheme never was initialized because the convection was too shallow.

4.3.2 Assessing Static Stability

Output from the WS-Eta was used to assess the static stability and its tendency at Eau Claire, Wisconsin, between 0000 UTC and 1200 UTC 23 March 1966. We will begin with an examination of the modelled thermodynamic profiles and finish with calculations of static stability tendency for the crucial layer in the storm.

Skew-T Analysis

We begin our analysis with the 12 hour forecast (Fig. 4.33) valid at 0000 UTC 23 March 1966. Although quite moist through a significant depth, this sounding is also moist statically stable throughout the troposphere. The 700-500 mb lapse rate is 5.1°C km⁻¹. Note also the veering wind profile through the troposphere, maximized in the 2-5 km layer. This warm air advection signature is crucial for destabilizing the environment over Eau Claire, Wisconsin, over the ensuing 12 hours. Of course, this sounding also has a surface layer nearly 1 km deep that is above freezing. This matches quite closely the actual surface observation at Eau Claire at 0000 UTC (Figs. 4.1 and 4.3).

By 0300 UTC (Fig. 4.34) the surface layer is still above freezing, but is cold enough to support snow. Yet some characteristics have not changed. The same pre-warm frontal WAA signature persists, as does the level of origin of the most unstable parcel (650 mb). While the 700-500 mb lapse rate is essentially unchanged (5.2°C km⁻¹), the lifted index based on the 650 mb parcel is +3°C at this time. These facts together suggest that the 650 mb level is warming at a rate faster than the rest of the layer, which is shown by a careful inspection of the forecast sounding.
This trend continues through 0600 UTC (Fig. 4.35), with the most unstable parcel (originating at 650 mb) having a lifted index of 2°C, within a 700-500 mb layer that has a lapse rate of 5.8°C km⁻¹. What is of interest, however, is the development of a discontinuity in the wind profile near 5 km. This feature suggests a change to weak cold air advection near the 5-km level, which is also near the top of the 700-500 mb layer.

At 0900 UTC (Fig. 4.36), subtle changes occur in the deeper sounding, with the most unstable parcel persisting at 650 mb and a lifted index of +1°C. The 700-500 mb lapse rate is at its least stable value of 6.1°C km⁻¹. In addition to a wind profile continuing to feature a discontinuity near the top of the 700-500 mb layer (where weak cooling was simulated), the profile also reveals a strengthening northeasterly low-level jet. Note that, while the deep troposphere characteristics are largely unchanged, the lowest 1 km has cooled dramatically concurrent with the strengthening of the winds in that layer.

1200 UTC (Fig. 4.37) represents the end of convective snow in the real environment. Concurrently, the best sounding signatures for thundersnow (e.g. stability indices, moisture), are maintained through 1200 UTC before beginning to evolve into a post-storm environment. At this time, the sounding over Eau Claire still has its most unstable parcel at 650 mb, resulting in a lifted index of +1°C. The 700-500 mb lapse rate maintains its previous value of 6.1°C km⁻¹. The lower troposphere continues to cool in the presence of the strong northeasterly flow of the CCB.

In summary, the environment over Eau Claire, Wisconsin, during the period when TSSN was observed was certainly one of deep moisture, but only weak static stability. As we will see in a forthcoming section, potential instability was present during the latter part of the period we have studied, but it too was weak and confined to shallow layers. Indeed, these soundings also feature an inset entitled “Theta-e vs. Height,” which depicts the PI profile up to 500 mb. Careful inspection
Figure 4.33: A skew-T analysis valid at 0000 UTC 23 March 1966, from the 12-hour forecast fields produced by the WS-Eta.

of each sounding profile reveals moist neutral layers in the 650 to 500 mb layer at 0600 (Fig. 4.35), 0900 (Fig. 4.36) and 1200 UTC (Fig. 4.37).

**Static Stability Tendency**

Before exploring alternative stability diagnostics and convective modes, it is worthwhile to examine the time tendency of static stability. A plot of this value (Fig. 4.38), shows a marked trend toward destabilization in the model over Eau Claire, WI, beginning after 0300 UTC and lasting until after 0900 UTC.

Assessing the static stability tendency equation of Bluestein (1992) at 650 mb over Eau Claire, we see that the vertical advection of stability was negligible throughout the period, while the differential temperature advection and stability divergence terms then to offset one another (Fig. 4.39). Yet differential temperature advection was the dominant term. Indeed, a careful examination of the model sounding at 0600 and 0900 UTC reveals no change in the temperature at 700 mb, but several degrees of cooling at 600 mb.
Figure 4.34: A skew-T analysis valid at 0300 UTC 23 March 1966, from the 15-hour forecast fields produced by the WS-Eta.

Figure 4.35: A skew-T analysis valid at 0600 UTC 23 March 1966, from the 18-hour forecast fields produced by the WS-Eta.
Figure 4.36: A skew-T analysis valid at 0900 UTC 23 March 1966, from the 21-hour forecast fields produced by the WS-Eta.

| CURSOR DATA |
|--------------|---------------|
| PARCEL DATA  |
| *** MOST UNSTABLE PARCEL *** |
| LCL: 650mb | -8C/-8C 17F/16F |
| CAPE = 0 J/kg | L1 (500mb) = 1 C |
| LFC = 0 J/kg | L1min = N M |
| LCL = 0 J/kg | cap = M H |
| LEVEL | PRES | MSLT (ASL) | TEMP |
| LCL: 647mb | 9914kt |
| LFC | M | M |
| LCL | M | M |
| NWP | M | M |

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<td>500-500mb Lapse Rate = 17 C / 4.0 C/km</td>
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<td>ThetaE Diff = 80C</td>
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Figure 4.37: A skew-T analysis valid at 1200 UTC 23 March 1966, from the 24-hour forecast fields produced by the WS-Eta.

| CURSOR DATA |
|--------------|---------------|
| PARCEL DATA  |
| *** MOST UNSTABLE PARCEL *** |
| LCL: 650mb | -8C/-8C 17F/16F |
| CAPE = 0 J/kg | L1 (500mb) = 1 C |
| LFC = 0 J/kg | L1min = N M |
| LCL = 0 J/kg | cap = M H |
| LEVEL | PRES | MSLT (ASL) | TEMP |
| LCL: 647mb | 9914kt |
| LFC | M | M |
| LCL | M | M |
| NWP | M | M |

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<td>ThetaE Diff = 31C</td>
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62
Figure 4.38: A plot from the WS-Eta output of the static stability tendency (K mb$^{-1}$ s$^{-1}$) versus time for the period 0000 UTC to 1200 UTC, for the layer 700-600 mb over Eau Claire, WI.

The differential diabatic heating term (Fig. 4.40) contributed consistently to stabilization during the crucial period. Indeed, this indicates an increase in diabatic heating with height, which is due, in part, to the moister environment at the bottom of the layer (700 mb) than the top (600 mb). Yet, even within the confines of the established uncertainty, the differential advection of $\theta$ dominated over Eau Claire in this case.

In all, this exercise shows more clearly the various processes that lead to destabilization. The decrease with height of warm air advection and the increase with height of diabatic heating strongly suggest the presence of the trowal while paving the way for more elaborate stability analyses.
Figure 4.39: A plot from the WS-Eta output of the differential temperature advection term (represented by dots; dark green), vertical advection of stability (rectangles; magenta) and vertical stretching (triangles: blue) terms (all in K mb$^{-1}$s$^{-1}$) versus time for the period of 0000 UTC to 1500 UTC for the 700-600 mb layer over Eau Claire, WI.
Figure 4.40: A plot from the WS-Eta output of the differential diabatic heating (middle; red contour) term (K mb$^{-1}$ s$^{-1}$) versus time for the period 0300 to 1200 UTC for the 700-600 mb layer over Eau Claire, WI. The black “sidebars” above and below are a kind of error bar for each period, as the actual differential diabatic heating was solved for as a residual.
4.3.3 Assessing Symmetric Stability

To establish the WS-Eta cross-section analysis for 0300-0900 UTC 23 March 1966, a thickness in a layer from 400-700 mb was used. This layer was used because it placed the middle of the analysis around 550 mb, which is the area that better encompasses where the sounding profiles suggest convection should be, with the most unstable parcel beginning at 650-mb. The cross-section is taken perpendicular to the thickness pattern so that, in theory, the thermal wind is blowing normal to the thickness contours. Thus the only change in the wind is its speed not direction. In a geostrophic sense this has to be assumed.

A cross-section taken from International Falls, Minnesota (INL) to Davenport, Iowa (DVN) (Fig. 4.41), was evaluated for frontogenesis, EPV, and $\theta_e$ at 0300 UTC 23 March 1966. In the cross-section of frontogenesis (Fig. 4.42), notice a frontogenesis maximum of 6 K (100km)$^{-1}$ (3hr)$^{-1}$ directly above the location of Eau Claire, Wisconsin. This would indicate that over a 3 hour period the temperature gradient will compress by 6 K over 100 km, a considerable amount. The circulation vectors on this analysis are the scaled addition of the ageostrophic wind (horizontal component) and $\omega$ (vertical component). This pattern of relatively strong ascent on the warm side of a frontogenesis maximum is typical behavior in a well-developed frontal zone. In an evaluation of $\omega$ (Fig. 4.43) along this same cross-section there is a vertical velocity maximum of 20 $\mu$b s$^{-1}$ also directly above Eau Claire. The maximum vertical velocity is found towards the warm air from the frontogenesis maximum. This should be expected to occur as a response of the atmosphere to the frontogenesis maximum. Looking at the 0300 UTC cross-section analysis of $M_g$ and $\theta_e$ (Fig. 4.44) it is apparent that there is a location directly over Eau Claire at a level between 700-800 mb where the $\theta_e$ contours are more vertical then the $M_g$ contours. This signifies an area conducive to vertical motion. Since the $M_g$ are scooping concavely and the $\theta_e$ contours are scooping convexly, this indicates
an area of negative EPV. This is verified by the EPV analysis for this cross-section (Fig. 4.45), where EPV is $-0.5 \times 10^{-6} \text{Km}^2 \text{kg}^{-2} \text{s}^{-1}$.

In the analysis of SCAPE (Fig. 4.46), the $40 \text{m kg}^{-2} \text{s}^{-1}$ surface was used to make the calculation, resulting in a value of $13 \text{ Jkg}^{-1}$ and a vertical velocity of $5 \text{ m s}^{-1}$. Granted $5 \text{ m s}^{-1}$ is not a strong vertical velocity, but for wintertime convection in a shallow layer, it may be just enough of a push to initiate convection. It should also be noted that 0300 UTC is the time that convection started in the observed data.

The last cross-section taken at 0900 UTC from Park Rapid, Minnesota (PKD) to Milwaukee, Wisconsin (MKE) is shown in Fig. 4.41. Here in the cross-section
Figure 4.42: The cross-section analysis of frontogenesis, (K 100 km⁻¹hr⁻¹) along a line from International Falls, MN (INL) to Davenport, IA (DVN) from the WS-Eta output fields valid at 0300 UTC 23 March 1966, 15 hours into the simulation. The location of Eau Claire, WI, is approximated with the red 'EAU'.
Figure 4.43: The cross-section analysis of $\omega \ (\mu b \ s^{-1})$ from the WS-Eta output fields valid at 0300 UTC 23 March 1966, 15 hours into the simulation. The location of Eau Claire, WI, is approximated with the red 'EAU'.
Figure 4.44: The cross-section analysis of $M_g$ (dashed; every 10 kg m s$^{-1}$) and $\theta_e$ (solid; every 2 K), and relative humidity (shaded for 70%, 80%, and 90%) from the WS-Eta output fields valid at 0300 UTC 23 March 1966, 15 hours into the simulation. The location of Eau Claire, WI, is approximated with the red 'EAU'.
Figure 4.45: The cross-section analysis of 3-D Equivalent Potential Vorticity (solid; every $0.25 \times 10^{-6} \text{K m}^2 \text{kg}^{-1} \text{s}^{-1}$) from the WS-Eta output fields valid at 0300 UTC 23 March 1966, 15 hours into the simulation. The location of Eau Claire, WI, is approximated with the red 'EAU'.
of frontogenesis (Fig. 4.47), a frontogenesis/frontolysis couplet has developed. This means there should be rising motion on the warm side of the frontogenesis maximum and on the cold side of the frontolysis maximum. An $\omega$ of 16 $\mu$s$^{-1}$ (Fig. 4.48), lies directly in between the frontogenesis/frontolysis couplet directly over Eau Claire. The cross-section of $M_g$ and $\theta_e$ (Fig. 4.49) at 0900 UTC shows that the area where $\theta_e$ is more vertical than $M_g$ has become extremely shallow with only a small area of negative EPV present (Fig. 4.50). Again, there is no presence of SCAPE in the 0900 UTC calculation. For the time period of 0600 to 1200 UTC there is a steady state CAPE generation/consumption mode in the model over Eau Claire, which would explain persistent convection in a layer from 550-650 mb.

**WS-Eta Conclusions**

In the analysis of the WS-Eta it became apparent that the model guidance was extremely accurate in the placement of the surface characteristics and precipitation.
Figure 4.47: A cross-section analysis of frontogenesis from the WS-Eta output, as seen in Fig. 4.42, but from Park Rapid, MN, (PKD) to Milwaukee, WI, (MKE) and vaild for 0900 UTC 23 March 1966.
Figure 4.48: A cross-section analysis of $\omega$ from the WS-Eta output, as seen in Fig. 4.43, but from Park Rapid, MN, (PKD) to Milwaukee, WI, (MKE) and valid for 0900 UTC 23 March 1966.
Figure 4.49: A cross-section analysis of $M_p$ and $\theta_e$ from the WS-Eta output, as seen in Fig. 4.44, but from Park Rapid, MN, (PKD) to Milwaukee, WI, (MKE) and valid for 0900 UTC 23 March 1966.
Figure 4.50: A cross-section analysis of 3-D Equivalent Potential Vorticity from the WS-Eta output, as seen in Fig. 4.45, but from Park Rapid, MN, (PKD) to Milwaukee, WI, (MKE) and valid for 0900 UTC 23 March 1966.
The precipitation field is well-represented by the model, not only in its placement, but also its intensity. In the cross-sectional analysis, the warm front was well-defined in the frontogenesis and $\theta_e$ cross-sections, with the warm front laying in the $\theta_e$ pattern. Analysis of $\omega$ showed a sloped response with maximum vertical motion over the frontal surface directly over Eau Claire. Eau Claire’s 600-650 mb EPV quantifies the area where CSI is most prone. The SCAPE calculations revealed SCAPE present on the 40 m kg$^{-2}$s$^{-1}$ surface at 0300 UTC, the only time period where SCAPE was present, but assumed to be a generation/consumption mode for the next 6 hours.
Chapter 5
Conclusions

Of the thesis statements that we had hoped to establish, only the first has been confirmed. In this case, the thundersnow at Eau Claire, Wisconsin, was not the result of potential instability release and upright convection. Instead, the release of conditional symmetric instability and resulting slantwise convection dominated this event. Indeed, a detailed analysis of the system during the course of this work revealed that Eau Claire was essentially northeast of the surface cyclone throughout the entire event. Even the young occluded frontal zone apparent at 1200 UTC on 23 March 1966 (Fig. 4.17) has the appearance of a warm front type occluded system. Thus, the second thesis statement has been disproven, in that the convective type never evolves into a more upright mode. Even though the isentropic analysis shows that Eau Claire was under the influence of the westward extension of the warm conveyor belt (the trowal) especially late in the period, and the static stability tendency analysis shows that differential thermal advection due to the trowal airstream dominated destabilization in the critical layer, significant potential instability failed to materialize.

That thundersnow was so long-lived can be explained on the meso-\(\alpha\) scale through the tendencies of both the static stability, as well as the SCAPE. In the case of the former, the actual lapse rate tended toward instability especially at 0600 and 0900 UTC, the heart of the event. This is due to the lower portion of the 700-600
mb layer warming faster than the upper portion at those hours. Evidence of this behavior is still in evidence on the 296 K surface at 1200 UTC 23 March 1966 (Fig. 4.25). Slantwise ascent is strongly inferred over the Eau Claire region at this time. With regard to SCAPE, it is noteworthy that the value reaches its peak early in the event at 0300 UTC on 23 March 1966, and then hovers about 0 J kg$^{-1}$ along the same momentum surface (40 kg m s$^{-1}$) at both 0600 UTC and 0900 UTC. This behavior suggests that the event has reached a steady-state condition, wherein energy is being consumed at the same rate at which it is being produced. Yet, this outcome does not allow for additional energy to foster an updraft. The background ascent in the model (20 μb s$^{-1}$) is certainly sufficient to lift a parcel to the point where it could become unstable, if available potential energy (APE) existed in some form. However, it is not an updraft.

As one might expect, our primary conclusion on this point is that this run of the WS-Eta, while providing an excellent meso-$\alpha$ simulation of the storm, is still too coarse to resolve the mechanisms necessary to focus and maintain convective snowfall. Of course, this same conclusion may also be drawn back to the way in which the model treats convection as well as the cloud microphysics. In summary, this snow event resulted largely from the prolonged presence of frontogenesis in the presence of weak symmetric stability. That the event should remain convective after 0300 UTC when measurable SCAPE existed is not altogether clear, even in the presence of such a fine-scale model simulation.
REFERENCES


Rebecca Lynn Ebert was born on April 20, 1978 in St. Louis, Missouri to Anthony David Ebert and Ramona Christine (Gebhardt) Ebert. Along with an older brother, Robert Gregory, Becky lived in St. Peters, Missouri a suburb of St. Louis for most of her young life. The author has always been fascinated with the weather, whether it was severe storms in the summer or heavy snowfall in the winter, she has always had a passion for what drives these events. Becky experienced the phenomena of thundersnow for the first time when she was 3 years old on January 31, 1982 when St. Louis received 23 inches of snow in 24 hours. This was absolutely fascinating to her and was one of the reasons she wanted to study meteorology.

Becky graduated from Francis Howell North high school with honors in 1996, after which she attended St. Charles County Community College where in 1999 she was awarded an Associates of Art. The author continued her education at the University of Missouri-Columbia were she was an active member of the local and national chapter of the American Meteorological Society, National Weather Association and Sigma Xi scientific research society. As an undergraduate the author held the position of treasurer of the Meteorology Club for two years, as well as, a member of the funding raising committee and chair of the Mid-Missouri weather calendar. In December of 2002 the author received her Bachelor of Science in Atmospheric Science from the University. It was in that last semester of college that the author made the decision to continue her education and work towards a masters degree. Through good fortune and funding from the National Science Foundation she was able to continue at the University of Missouri where in July of 2004 she will complete her Masters of Science in Atmospheric Science at Mizzou.

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