

A Process-Oriented Methodology Toward Understanding the Organization of an Extensive Mesoscale Snowband: A Diagnostic Case Study of 4–5 December 1999

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ABSTRACT

A case study of a long, narrow band of heavy snowfall is presented that illustrates those processes that force and focus the precipitation in a unique linear fashion. System-relative flow on isentropic surfaces shows how the trough of warm air aloft (trowal) formed to the north-northwest of a weak synoptic-scale surface cyclone. To the north of the trowal, midtropospheric frontogenesis formed as the warm, moist, high- θ_e air in the trowal canyon became confluent with cold, dry air to the northwest of a closed midlevel circulation. Within the trowal airstream, isentropic upslope is shown to contribute to vertical motion, while transverse to this flow, mesoscale lift is enhanced on the warm side of a frontogenetical zone in the presence of weak symmetric stability and conditional symmetric instability. Further, it is shown that a sloping zone of small positive to negative equivalent potential vorticity forms to the southeast of the midtropospheric system-relative closed circulation as low- θ_e air associated with the dry conveyor belt, seen in water vapor imagery, overruns warm, moist high- θ_e air associated with the warm conveyor belt. In this way cold season instability forms due to differential moisture advection on the warm side of the frontogenesis axis. Finally, a conceptual model is shown that encapsulates the key processes that contributed to the extensive, narrow band of heavy snow in the presence of a weak synoptic-scale surface cyclone.

1. Introduction

During the 24-h period ending 1200 UTC 5 December 1999, a long (~1000 km) swath of heavy snow (greater than 15 cm or ~6 in.) fell from the Texas Panhandle to the northwest corner of Missouri (Fig. 1). The half-width of this snowband (the distance from the maximum snowfall to half that value) was approximately 50–60 km, although the band's width was impacted by precipitation type as warmer temperatures resulted in rain on the southern edge of the band. Within this narrow band, maximum snowfall amounts of 30 cm (12 in.) were recorded in Texas, Oklahoma, and Kansas, along with reports of convection [i.e., thundersnow; NCDC (1999)], especially in portions of southeast Kansas. The snow fell several hundred kilometers northwest of a weak surface cyclone, having a minimum pressure of only 1008 hPa, which formed along a northeast–southwest-oriented quasi-stationary frontal boundary that stretched from Michigan to north-central Texas. Although unusual, long, banded

heavy snowfall has been previously documented in the plains (e.g., Cissell and Marwitz 1998). Banacos (2003) has noted that the spatial location and duration of these heavy snowbands is often difficult to predict accurately.

The challenge of this heavy snow event is to explain the length and breadth of the heavy snowfall as well as thundersnow in the presence of a weak surface cyclone. Toward that end, this paper examines the processes that help to force and focus this narrow region of heavy snowfall, while also acting to destabilize the atmosphere producing convective-scale vertical motions in the cold air.

Section 2 will describe the relevant literature addressing the issues of cold season atmospheric airstreams, system-relative (SR) flow, frontogenesis, and destabilization processes that often occur during cyclogenesis. Section 3 describes the datasets used and methodology applied to this diagnostic study. Section 4 illustrates the processes believed to play a critical role in producing the extensive, yet relatively narrow, snowband and destabilizing the atmosphere for the present storm. Finally, section 5 summarizes the results and presents a schematic diagram that highlights the major processes acting in the storm in the context of three-dimensional airstreams.

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24 Hour Snowfall Isohyet Analysis Valid 5 December 1999

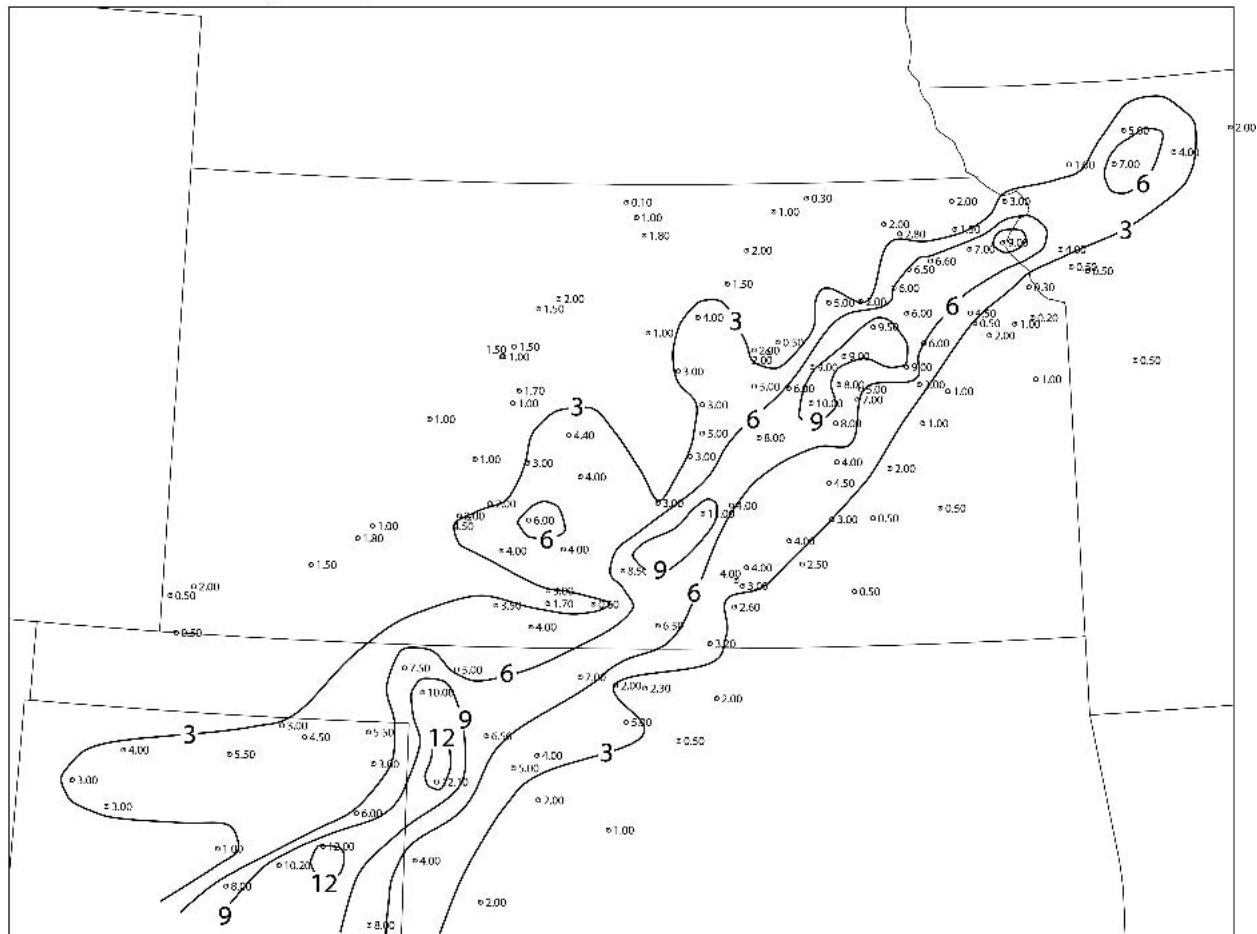


FIG. 1. Isohyets of snowfall (in.) for the 24-h period ending 1200 UTC 5 Dec 1998, based upon cooperative station data from NCDC.

2. Processes important to the production of narrow snowbands

a. Atmospheric airstreams and conveyor belts

Analysis of an animated loop of water vapor channel imagery readily reveals that atmospheric flow tends to be concentrated within airstreams having unique thermodynamic characteristics. Browning (1990) and Carlson (1991) note that airstreams contain relatively narrow ranges of potential temperature (θ ; for dry air) or wet-bulb potential temperature (θ_w ; for moist air) according to the airstream's origin. To a first approximation airstreams can be identified using an SR wind analysis on an isentropic surface, as air parcels conserve their potential temperature under adiabatic flow. The SR flow (also referred to in the literature as relative wind) on an isentropic surface defines the flow *relative* to the movement of the large-scale disturbance (Green et al. 1966; Carlson 1980), under the assumption that the system translates steadily without change of speed or shape. Thus, subtracting the system's phase speed

from the ground-relative winds results in SR flow. In this way the motion of the isentropic surface is accounted for and airstreams can be identified more easily in this quasi-Lagrangian framework. Carlson (1991) has noted that for atmospheric circulations that are neither deepening or weakening significantly with time, "relative wind streamlines are identical to trajectories relative to the moving system." Carlson (1991) and Moore et al. (1998) have shown that for slowly developing weather systems using SR flow essentially accounts for that component of the local time tendency of a scalar property (e.g., pressure) on the isentropic surface that is due to the motion of the isentropic surface. It does not, however, account for any diabatic processes that describe vertical flux in an isentropic coordinate system. Assessing the motion, C , of the isentropic surface for a well-defined system can be done by measuring the motion of a vorticity maximum on the isentropic surface, or tracking the speed of a trough in the Montgomery streamfunction field. It should be stressed, however, that the SR flow field thus obtained is only

valid for the region within which the system speed is applicable. Locations at a considerable distance from the immediate area will likely have a different system speed.

Research by Harold (1973), Carlson (1980), and Danielsen (1964) has identified three major airstreams associated with cyclogenesis, using relative-wind analysis. These airstreams, termed the warm, cold, and dry conveyor belts, have been shown to directly influence and dictate the organization of precipitation attending an extratropical cyclone (ETC). A conveyor belt can be thought of as an ensemble of air parcels, originating from a common source region and layer, tracked over synoptic-scale time periods (on the order of 24–36 h). The three major conveyor belts that have been shown to play important roles in the distribution of precipitation within and near an ETC are depicted in Fig. 2. The warm conveyor belt (WCB), originating at low levels within the warm sector, to a large degree is associated with the convective precipitation within the warm sector of an ETC, as well as the broad shield of typically stratiform precipitation poleward of the warm front. The cold conveyor belt (CCB) consists of dominantly low-level easterly flow ahead of the warm front and organizes precipitation within the comma head of an ETC. Recently, the CCB portrayed by Carlson (1980),

in which the airstream moves upward while abruptly turning anticyclonically as it approaches the ETC comma head, has come under review. Schultz (2001) has shown that the CCB has at least one branch that turns cyclonically as it rises, but stays within the lower troposphere. Thus, the deeper clouds in the comma head are more likely the result of an ascending WCB that has “branched off” to the northwest of the ETC. The dry conveyor belt (DCB) has been described by Danielsen (1964) as originating in the mid- to upper troposphere and consisting of high potential vorticity. The DCB is usually seen in water vapor imagery as a dry slot entering the ETC from the southwest of the cyclone. A second branch of the DCB typically turns anticyclonically to the south forming a low-level deformation zone that contributes to frontogenesis in the vicinity of the cold front (Bluestein 1993). It is the three-dimensional interaction of these three airstreams in the vicinity of the ETC that in many cases leads to an environment favorable for heavy banded precipitation (Nicosia and Grumm 1999).

b. The trowal airstream

Iskenderian (1988) used isentropic analysis and trajectories to diagnose a bifurcation of the WCB north of a warm front associated with a nondeveloping ETC that

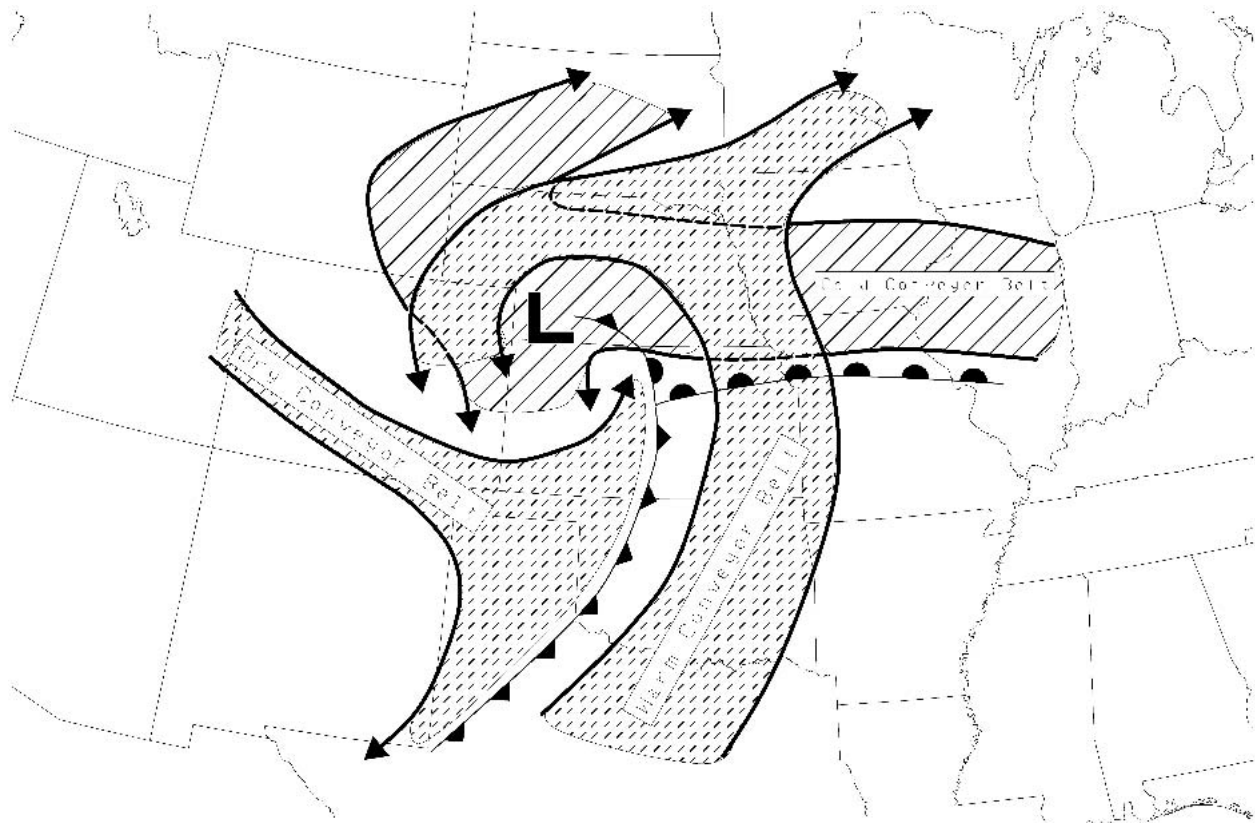


FIG. 2. Conveyor belts associated with extratropical cyclogenesis. (A color version of this schematic diagram is available online: <http://www.eas.slu.edu/CIPS/Publications.html>.)

produced copious amounts of precipitation as it moved to the northeast. He showed that part of the WCB airstream moved to the northeast as it rose north of the warm front, contributing to precipitation there. However, another branch of the WCB airstream was observed to move to the northwest of the center of low pressure, ascending above the CCB in the region of the comma head (see Fig. 2). This cyclonically curved branch of the WCB contributed to extensive precipitation to the northwest of the ETC [also noted by Halcomb (2001)].

Martin (1998a,b) identified this cyclonically curved branch of the WCB as the “trowal” airstream, where trowal denotes the trough of warm air aloft. Originally described by Canadian meteorologists Crocker et al. (1947), Godson (1951), and Penner (1955), the trowal is essentially defined as the sloping tongue of warm air that wraps cyclonically around an ETC. Martin (1998a) defined it within the context of a warm occlusion as the “3D sloping intersection of the upper cold-frontal portion of the warm occlusion with the warm frontal zone.” It is important to point out, though, that in the Canadian model, the trowal was identified in nonoccluded systems. This point will be borne out by the present study.

In the case of the warm occluded structure, Martin (1999) presented convincing evidence that the trowal structure can be explained by the differential rotation of the thermal gradient implied by the opposing $\mathbf{Q}_s$ vectors [see Keyser et al. (1992) for a description of $\mathbf{Q}_s$ and $\mathbf{Q}_n$ vectors]. The convergence of the $\mathbf{Q}_s$ vectors also results in quasigeostrophic forcing of upward vertical motion in the vicinity of the trowal. However, this $\mathbf{Q}_s$ forcing is a synoptic-scale dynamical process and does not explain the presence of frontal-scale banded precipitation that often is found in the vicinity of some trowal structures.

c. Cold season instability and frontogenesis

Common measures of warm season instability including the convective available potential energy (CAPE) and the lifted index (LI) are typically ineffective at describing the instability within a cold air mass that supports heavy snow rates. Over the last decade a great deal of research has been done attempting to address those processes that contribute to either upright or slantwise convection within relatively cold air masses. Bennetts and Hoskins (1979) were among the first researchers to describe how frontal rainbands might be explained by the presence of conditional symmetric instability (CSI), a condition wherein the atmosphere is convectively stable (i.e., equivalent potential temperature, θ_e , increasing with height) and inertially stable (geostrophic absolute vorticity, η_g , greater than zero), yet is unstable to slantwise ascent. Emanuel (1983, 1985) has shown that the upward vertical motion branch of a frontal-scale circulation in the presence of CSI or even weak symmetric stability (WSS) is both

enhanced and contracted. Moore and Lambert (1993) developed a two-dimensional form of equivalent potential vorticity (EPV) that could be used to diagnose regions of CSI within a cross-sectional plane. This concept was later generalized by McCann (1995) to a three-dimensional form of EPV that could be used in a plan or cross-sectional view. He basically showed that CSI can be diagnosed in a region of negative EPV that tends to form in a saturated environment within which the vertical wind shear is strong and the convective stability is weak.

In an attempt to clarify the distinction between potential symmetric instability (PSI) and CSI, Schultz and Schumacher (1999) have noted that PSI is evaluated using θ_e , while CSI is evaluated using θ_{es} , the saturated equivalent potential temperature. In either case, the instability is not realized unless the atmosphere is saturated in the presence of large-scale lifting, in which case θ_{es} and θ_e are identical. Operationally, it is more convenient to use θ_e and require the relative humidity to be at least 80% to diagnose regions susceptible to CSI (Market and Cissell 2002). In addition, Schultz and Schumacher (1999) illustrate that within the cold season environment near frontal zones a *spectrum of instability regimes* is often found ranging from convectively unstable air (i.e., $\partial\theta_e/\partial z$ less than zero) in the warm sector, to CSI north of the frontal boundary, to WSS even farther to the north.

Many studies (e.g., Sanders and Bosart 1985; Moore and Blakley 1988; Nicosia and Grumm 1999) have noted the presence of frontogenesis in or near the region of CSI. As Nicosia and Grumm (1999) illustrate, this is not merely an interesting coincidence. In a region of frontogenesis, the gradient of potential temperature increases with time, which, through the thermal wind relationship, requires an increase in the geostrophic wind shear. In a properly configured environment this geostrophic shear results in differential moisture advection that steepens the vertical slope of θ_e isentropes. The subsequent weakening of the convective stability, together with a strengthening of the vertical wind shear, results in a reduction of EPV, often leading to CSI. Their conceptual model (see their Fig. 17) illustrates that it is the unique juxtaposition of the cold, warm, and dry conveyor belts associated with a typical ETC that favors an EPV reduction zone in the vicinity of a region of midlevel frontogenesis. What is most important with respect to the current study is the fact that these processes can take place in or near a region of weak cyclogenesis, often quite a distance away from the surface cyclonic circulation.

3. Dataset and methodology

The case under examination covers the period 1200 UTC 4 December–1200 UTC 5 December 1999. Standard surface and upper-air rawinsonde data from the

National Weather Service (NWS) observational network were collected via Unidata's Internet Data Distribution (IDD) network, as well as Rapid Update Cycle version II (RUC II) initialized model data. Infrared (IR) and water vapor *Geostationary Operational Environmental Satellite-8 (GOES-8)* satellite imagery were obtained from the University of Wisconsin—Madison Space Sciences and Engineering Center. Derived fields of observed surface and upper-air parameters were computed using Saint Louis University's SLUbrew diagnostic software. This software objectively analyzes surface (upper air) fields using the Barnes (1973) scheme on a 95.25-km (190.5 km) grid that resolves approximately 50% of the two-delta wavelength of a given parameter, providing a synoptic-scale analysis of the data. RUC II (Benjamin et al. 2004) gridded model data on an 80-km grid were analyzed using the General Meteorological Package (GEMPAK; Koch et al. 1983), available to academic institutions through Unidata. Saturated EPV_g (geostrophic form of EPV) was computed using the McCann (1995) expression:

$$\text{EPV}_g = g \left[\frac{\partial \theta_{es}}{\partial x} \frac{\partial v_g}{\partial p} - \frac{\partial \theta_{es}}{\partial y} \frac{\partial u_g}{\partial p} - \left(\frac{\partial v_g}{\partial x} - \frac{\partial u_g}{\partial y} + f \right) \frac{\partial \theta_{es}}{\partial p} \right], \quad (1)$$

where u_g and v_g are the east–west and north–south components of the geostrophic wind, respectively; f is the Coriolis parameter; and θ_{es} is the saturated equivalent potential temperature. Petterssen's (1956) two-dimensional scalar frontogenetic function was computed as

$$F = \frac{1}{2} |\nabla \theta| [\text{Def}_r \cos(2\beta) - \text{Div}], \quad (2)$$

where $\nabla \theta$ is the potential temperature gradient, Def_r is the resultant deformation, β is the angle between the isentropes and the axis of dilatation, and Div is the divergence. To compute frontogenesis, the resultant deformation, defined as the square root of the sum of the squares of the stretching and shearing deformation, and β were computed following Bluestein (1993).

4. Diagnostic analysis of the 4–5 December 1999 snowband

a. Surface analyses and GOES-8 satellite imagery

Surfaces analyses for the period 1800 UTC 4 December–1200 UTC 5 December 1999 (Figs. 3a–d) document a weak surface cyclone with a central pressure of 1009 hPa centered over southeast Oklahoma that moves slowly northeastward with little or no intensity change to a position over St. Louis, Missouri. Throughout the period, moderate to heavy rain fell to the north-northeast of the surface low, with elevated thunder-

storm activity noted on the cold side of the stationary front north of the cyclone center. However, the focus of this paper is on the separate area of snow that developed to the northwest of the surface low, initially (see Fig. 3a) over the panhandles of Oklahoma and Texas as well as northwestern New Mexico, and translated over western Oklahoma into northeastern Kansas by 1200 UTC 5 December. This snowfall was on the northwestern periphery of a precipitation region that was distinctly separate from the rain that fell to the northeast and, in later periods, to the southeast, of the cyclone center. *GOES-8* enhanced IR satellite imagery during this period (Figs. 4a–d) reveals how the linear region of clouds about 400 km northwest of the surface low is detached from the primary area of clouds located over the Mississippi River valley. Over the time period shown the former region propagates from the Texas–Oklahoma panhandles into south-central Iowa. Light to moderate rain fell along and south of the major axis of this cloud band, while snow fell along and north of the same axis. It is likely that if the latter area of precipitation had been snow, the width of the snowfall band would have been wider. Thus, the width of this snowfall event was impacted to some extent by precipitation type as well as intensity. However, this distinct area of precipitation would still result in an extended track of snowfall.

The *GOES-8* enhanced water vapor imagery (Figs. 4e–h) reveals an area of midlevel dry air, noted by the dark slot, that wraps around the closed low from the southwest from north-central Texas into eastern portions of Oklahoma and southeast Kansas. As will be shown later, this midlevel dry-air advection helped to destabilize the atmosphere to the southeast of the heavy snowband, essentially creating a convectively unstable region (i.e., $\partial \theta_e / \partial z$ less than zero). Thus, differential moisture advection in the low to midtroposphere played a very important role in changing the vertical slope of the θ_e surfaces, quantified in the last term of (1). In the cold season, in the absence of strong diabatic heating through insolation, differential moisture advection in the low to midtroposphere creates either weak convectively stable or convectively neutral profiles that result in a greater response to forcing associated with either frontogenesis or jet streak–induced vertical circulations.

b. Upper-level flow

Initially, at 1200 UTC 4 December at 850 hPa (not shown) a closed circulation was located over southwest Oklahoma with a strong thermal gradient to the northwest of the low. Over the next 24 h (Figs. 5a,d) this low-level circulation moves to the northeast to a position over west-central Missouri. Initially the 500-hPa closed low (not shown) is centered over northeastern New Mexico with a neutrally tilted short-wave trough to the south. In subsequent time periods the 500-hPa circulation (Figs. 5b,e) lags to the west-southwest of the

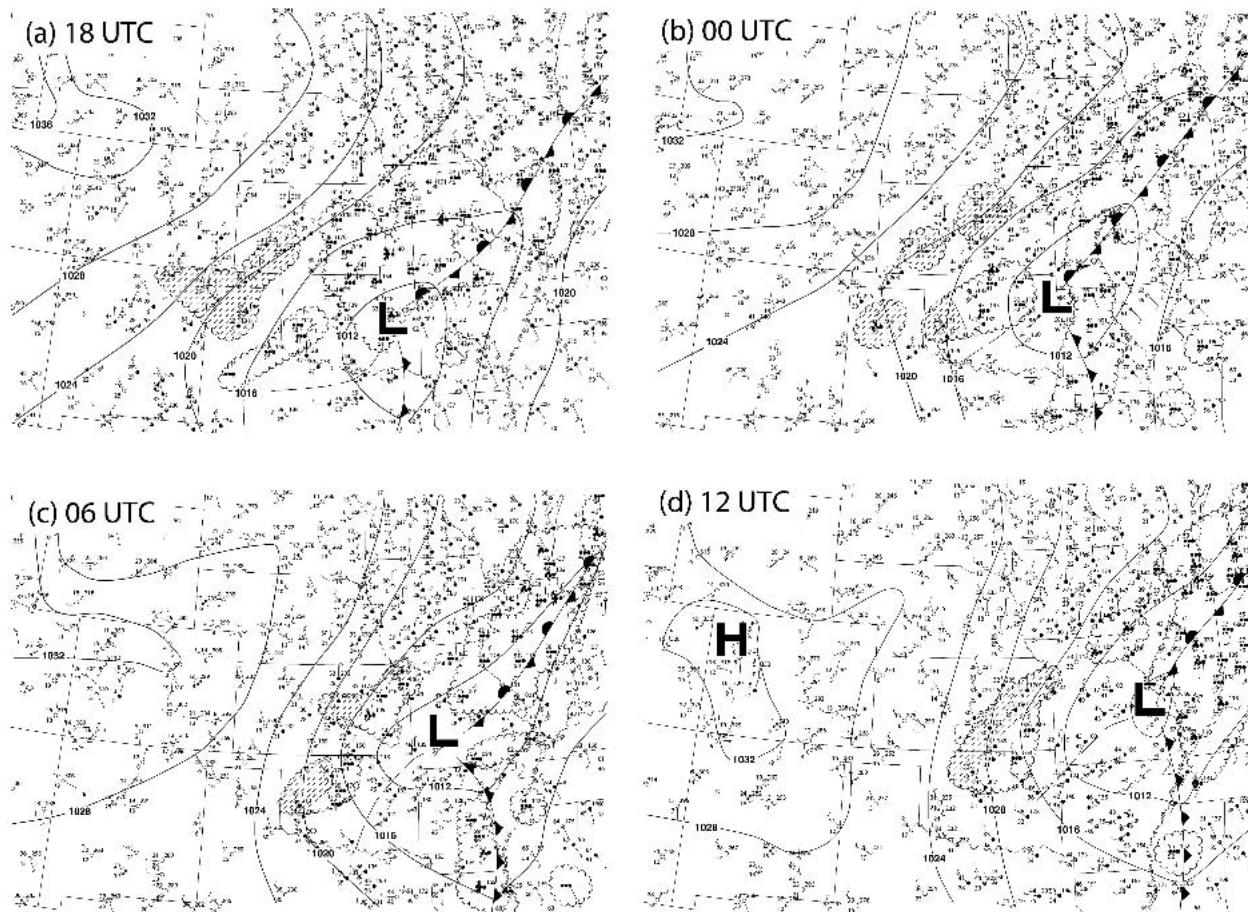


FIG. 3. Subjective analysis of surface weather at (a) 1800 UTC 4 Dec, (b) 0000 UTC 5 Dec, (c) 0600 UTC 5 Dec, and (d) 1200 UTC 5 Dec 1999. Solid lines are mean sea level (MSL) isobars (hPa), scalloped regions indicate precipitation (hatched area indicates snow), and fronts are indicated using standard notation. Note: subjective analysis is based upon more station data than indicated in plot. Thick dashed line indicates a surface trough.

850-hPa low but becomes quasi barotropic by the end of the period. A 300-hPa isotach analysis for the last two time periods (Figs. 5c,f) reveals the presence of a highly amplified flow with a strong 65 m s^{-1} jet streak propagating around the base of the trough at 0000 UTC 5 December, positioned along the eastern side of the trough axis by 1200 UTC 5 December. Thus, at 0000 UTC 5 December, most of the precipitation associated with the second cluster of clouds noted by the satellite imagery was located on the cyclonically sheared side of the upper-level jet (ULJ) in central Kansas. Although the exit region of the jet streak over northern Texas is not particularly well defined, the amplitude of the long-wave trough and sharp curvature of the cyclonic flow undoubtedly enhanced the upper-level divergence over Oklahoma and Kansas.

c. Evolution of the trowal

Generally, the trowal develops as the WCB begins to bifurcate with the westernmost extension of the WCB

turning toward the west-northwest. This evolution is best depicted through SR flow on an isentropic surface. In the present case we chose the 298-K isentropic surface as it best displays the isentropic upslope associated with the trowal airstream. The SR flow was computed by tracking the circulation center on the 298-K isentropic surface for successive 3-h time periods.

The SR streamlines and isobars on the 298-K isentropic surface for the period 0300–0900 UTC 5 December (Fig. 6) depict the trowal airstream as the branch of the WCB that turns cyclonically southwestward across Kansas. Note how the isobars on the 298-K surface form a ridge of high pressure, first in southern Kansas (Fig. 6a) and later along northern Missouri and Kansas (Fig. 6c) along the cyclonically turned SR streamlines of the WCB. As the trowal airstream ascends along an isentropic surface, it transports moisture to the west of the surface cyclone. System-relative moisture transport vectors on the 298-K surface for the period 0300–0900 UTC 5 December (Fig. 7) illustrate how moisture from the WCB airstream is advected as it ascends to the

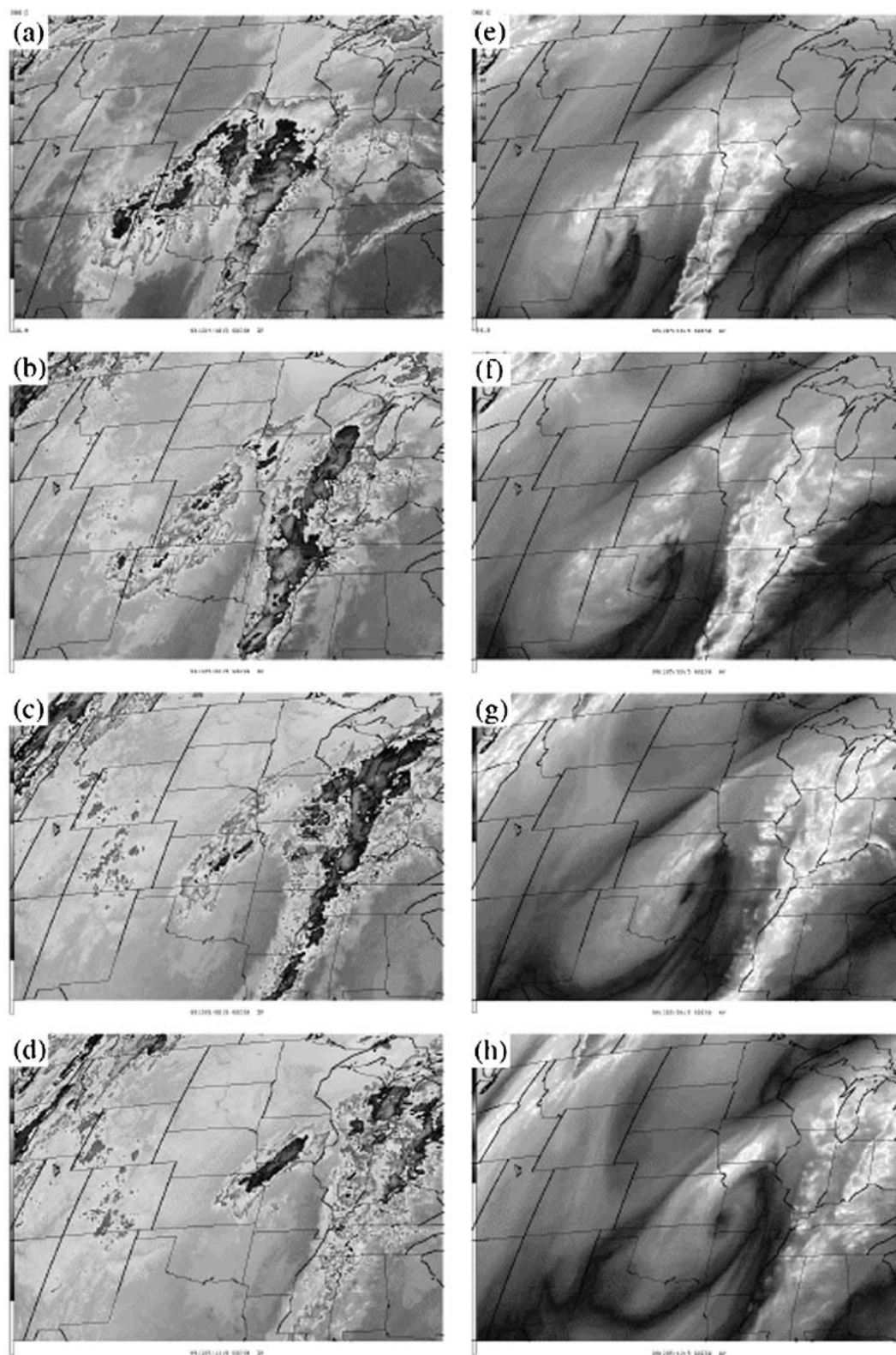


FIG. 4. *GOES-12* IR imagery for (a) 1800 UTC 4 Dec, (b) 0000 UTC 5 Dec, (c) 0600 UTC 5 Dec, and (d) 1200 UTC 5 Dec 1999. Water vapor imagery for (e) 1800 UTC 4 Dec, (f) 0000 UTC 5 Dec, (g) 0600 UTC 5 Dec, and (h) 1200 UTC 5 Dec 1999.

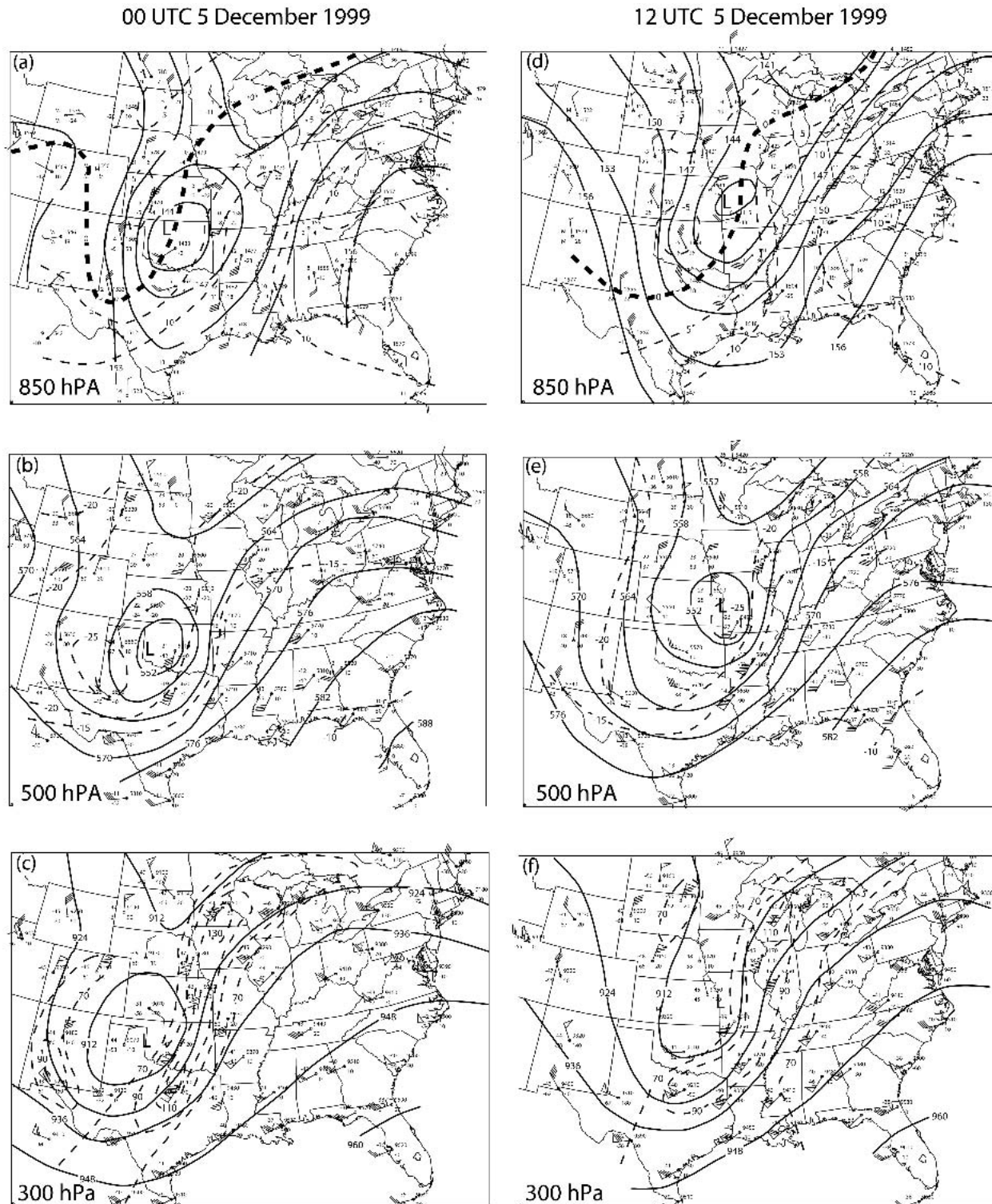


FIG. 5. Subjective analysis of upper-level pressure surfaces: (a) 850 hPa for 0000 UTC 5 Dec 1999; solid lines are isohypses in dm, dashed lines are isotherms in °C, and thick dashed line indicates the 0°C isotherm; (b) 500 hPa for 0000 UTC 5 Dec 1999; solid lines are isohypses in dm, and dashed lines are isotherms in °C; (c) 300 hPa for 0000 UTC 5 Dec 1999; solid lines are isohypses in dm, and dashed lines are isotachs in kt; (d) same as in (a) except for 1200 UTC 5 Dec 1999; (e) same as in (b) except for 1200 UTC 5 Dec 1999; and (f) same as in (c) except for 1200 UTC 5 Dec 1999.

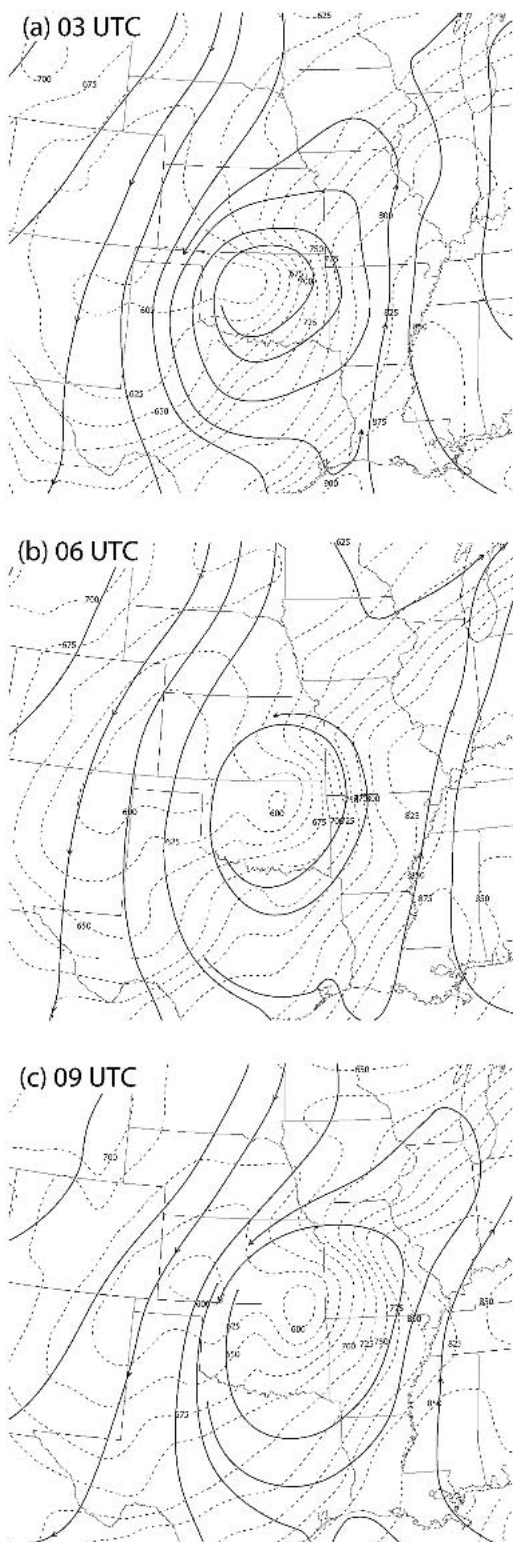


FIG. 6. System-relative streamlines (solid lines) and isobars (dashed lines, hPa) for the 298-K isentropic surface based upon RUC II initial analysis data for (a) 0300, (b) 0600, and (c) 0900 UTC 5 Dec 1999: for (a) $C = 238.2^\circ$ at 9.8 m s^{-1} , (b) $C = 226.9^\circ$ at 9.0 m s^{-1} , and (c) $C = 247.3^\circ$ at 10.7 m s^{-1} .

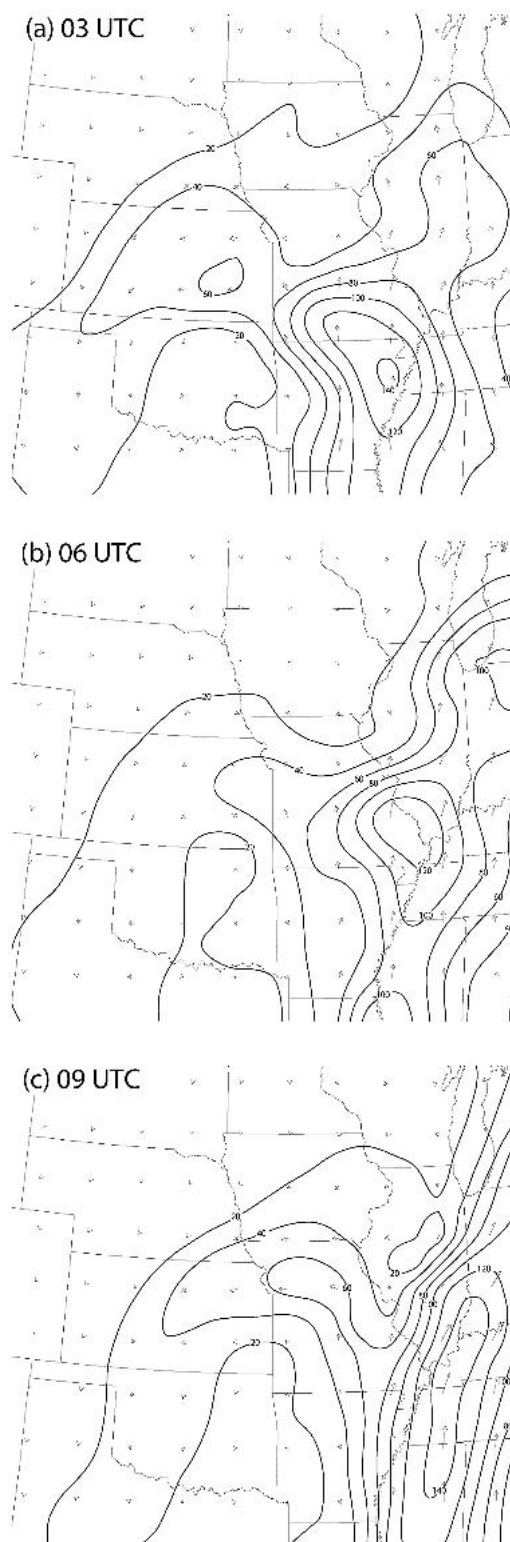


FIG. 7. System-relative moisture transport vectors (arrows) and magnitudes (solid lines, $\text{g kg}^{-1} \text{ m s}^{-1}$) for the 298-K isentropic surface based upon RUC II initial analysis data for (a) 0300, (b) 0600, and (c) 0900 UTC 5 Dec 1999. Values used for the system speed were the same as noted in Fig. 6.

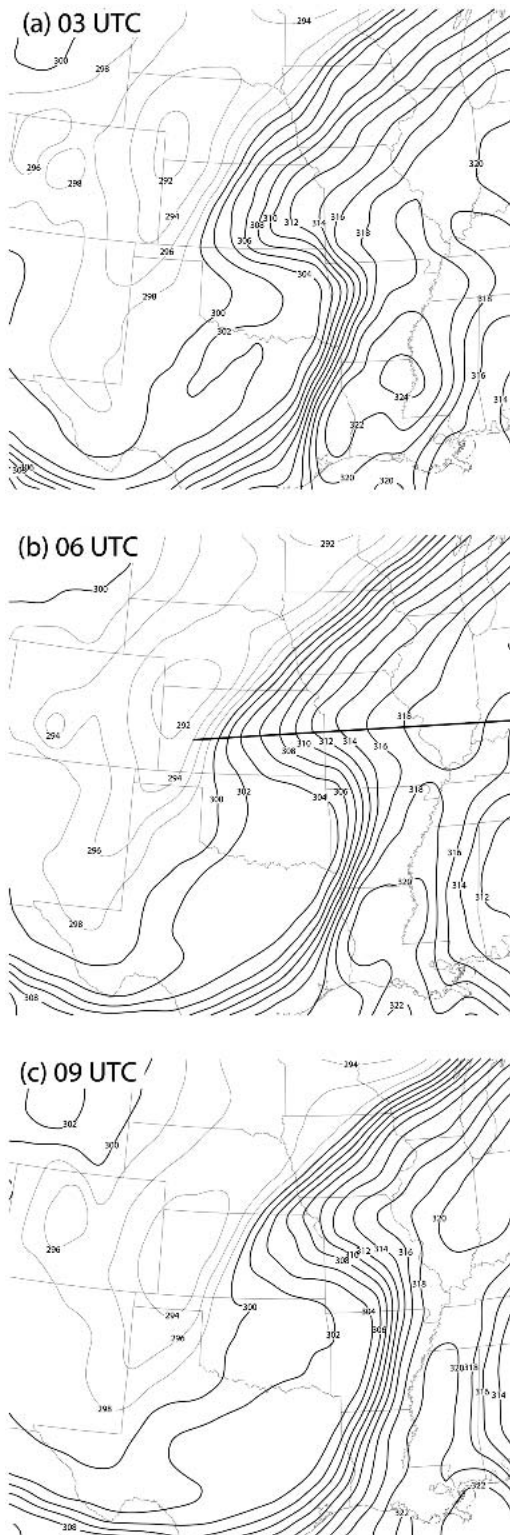


FIG. 8. Isentropes of equivalent potential temperature (θ_e , K) on the 700-hPa surface based upon RUC II initial analyses for (a) 0300, (b) 0600, and (c) 0900 UTC 5 Dec 1999. Solid line (trowal axis) in (b) indicates line along which cross section was taken in Fig. 9.

west-southwest along the trowal airstream. Consequently, to the northwest of the surface cyclone the depth and amount of moisture is increased substantially.

The trowal can also be diagnosed through an analysis of the 700-hPa θ_e isentropes. Initially (Fig. 8a) the trowal axis can be seen branching westward from the main area of high- θ_e values (greater than 320 K) located over the mid-Mississippi River valley to central and southern Kansas. Over the next 6 h (Figs. 8b,c) the trowal axis becomes sharper as it gradually moves to a position over northern Missouri and northeastern Kansas. Also note that during this 9-h period the θ_e gradient to the north of the trowal axis increases in magnitude, graphically illustrating the frontogenetical forcing north of the trowal axis.

The three-dimensional shape of the trowal has been described as a sloping canyon of high θ_e (Martin 1998a,b). By examining a cross section of θ_e along the trowal axis at 0600 UTC 5 December (Fig. 9), one can evaluate the atmospheric stability in the vicinity of the trowal. In this figure the highest- θ_e air in the trowal is seen in the eastern part of the cross section as a broad valley of high- θ_e air (values exceeding 318 K) with a maximum value of 324 K at its center. In the western periphery of the trowal, vertically folded isentropes of θ_e can be seen between 800 and 600 hPa, convincingly depicting a region of convective instability. It is the low- θ_e values associated with the midtropospheric DCB, noted earlier in the water vapor imagery, overrunning the warm, moist air of the trowal airstream, that leads to this region of convective instability.

d. Midlevel frontogenesis and equivalent potential vorticity

Plan views of layer-averaged frontogenesis for 800–650 hPa (Fig. 10) reveal the presence of a frontogenetic–frontolytic couplet that consistently moves slowly to

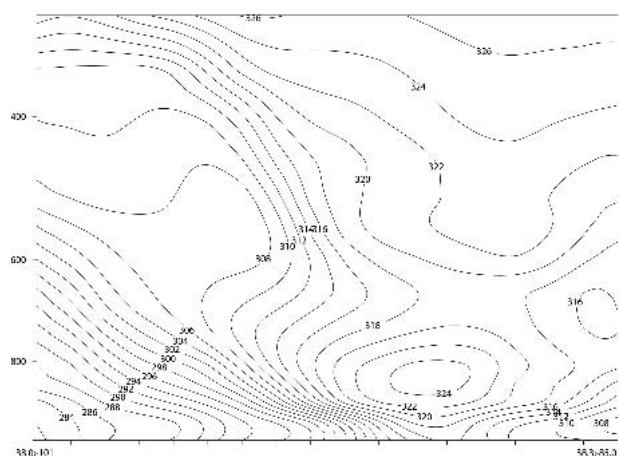


FIG. 9. Cross section of θ_e (K) along line shown in Fig. 8b based upon RUC II initial analysis for 0600 UTC 5 Dec 1999.

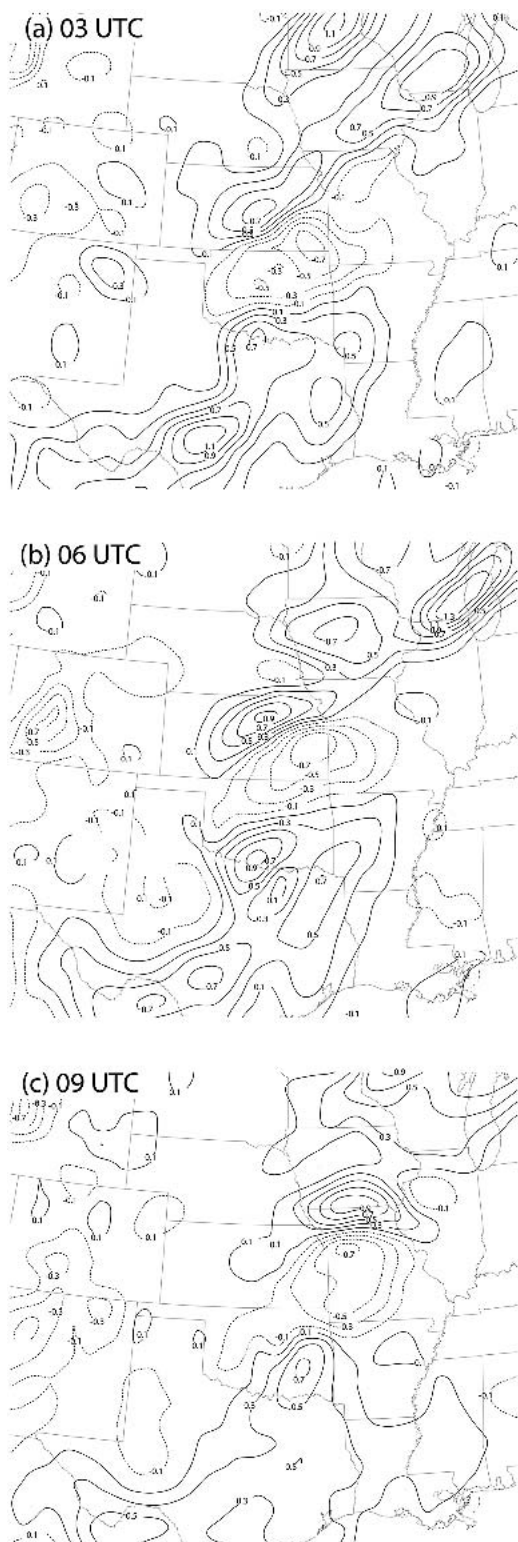


FIG. 10. Average frontogenesis for the 800–650-hPa layer based upon RUC II initial analyses for (a) 0300, (b) 0600, and (c) 0900 UTC 5 Dec 1999. Solid lines indicate frontogenesis while dashed lines indicate frontolysis. Units are $\text{K} (3 \text{ h} - 100 \text{ km})^{-1}$.

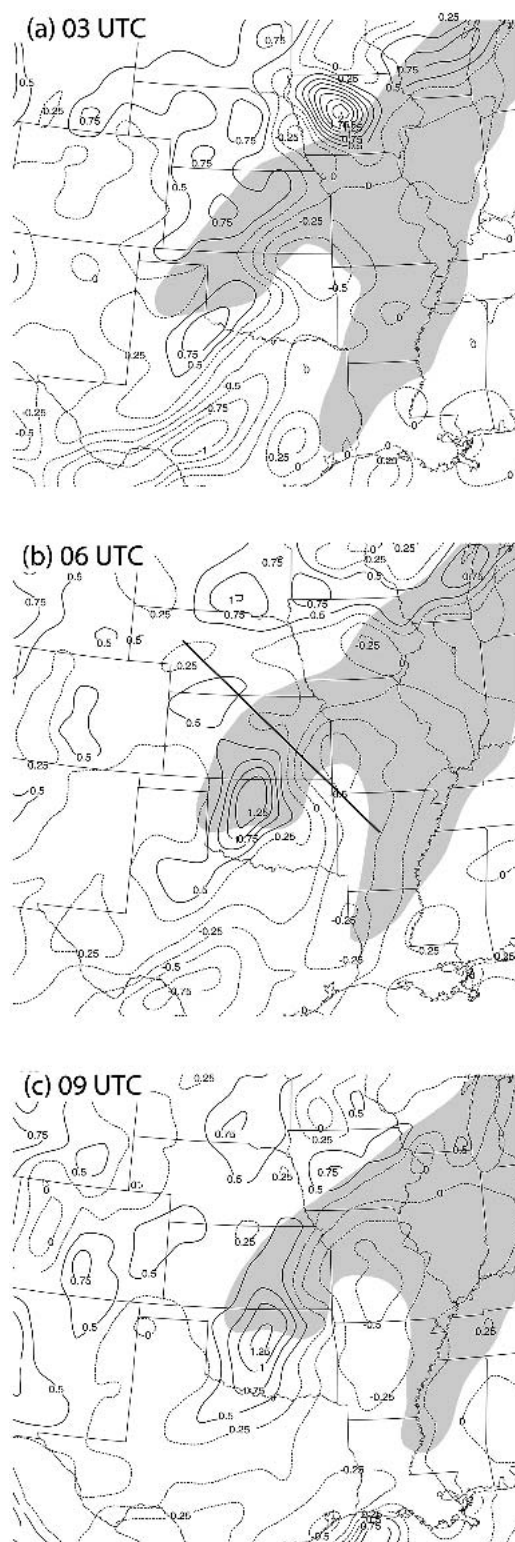
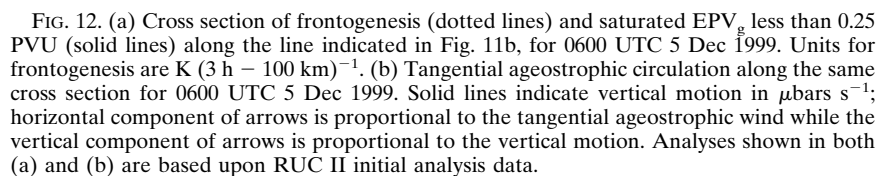


FIG. 11. Average saturated EPV_g in units of PVU for the layer 750–600 hPa based upon RUC II initial analyses for (a) 0300, (b) 0600, and (c) 0900 UTC 5 Dec 1999. Dashed lines indicate values of EPV_g less than or equal to 0.25 PVU. Shaded region indicates mean relative humidity for the 750–600-hPa layer equal to or greater than 80%. Solid line in (b) indicates line along which cross section was taken in Fig. 12.



the northeast from south-central Kansas into southern Iowa, paralleling the heavy snowband. This couplet, with the frontogenetic maximum slightly larger in magnitude than the frontolytic maximum, is located well to the north of the stationary front on the cold side of the surface boundary to the northwest of the surface cyclone. While substantial rain is falling in the region of frontolysis, light to moderate snow, associated with the frontogenesis region, fell as it moved through Kansas.

As noted earlier, the upward vertical motion associated with a frontogenetic circulation is both contracted and enhanced in the presence of weak-neutral static stability (i.e., $\partial\theta/\partial z$ is close to zero). Cold season instability is best measured through saturated EPV_g, which, when negative, indicates either CSI or convective instability. Plan-view plots of layer-averaged saturated EPV_g for 750–600 hPa for the 0300–0900 UTC 5 December time period (Fig. 11) consistently reveal a region of EPV_g less than 0.25 potential vorticity units ($1 \text{ PVU} = 10^{-6} \text{ m}^2 \text{ s}^{-1} \text{ K kg}^{-1}$) over eastern Kansas. In addition, this region is nearly saturated since average relative humidities exceed 80%; thus, the distinction between θ_e and θ_{es} is small. The vertical cross section of θ_e shown earlier (see Fig. 9) documented that negative EPV_g values are dominantly due to vertically folded θ_e surfaces. However, values of EPV_g between 0 and 0.25 PVU, in central Kansas at 0600 UTC (Fig. 11b), where θ_e surfaces are not folded, depict regions of WSS. This description is consistent with the spatial continuum of stability described by Schultz and Schumacher (1999) in which the stability regime begins with surface-based convective instability, progressing to elevated convective instability to CSI, and finally WSS as one moves to the north.

A cross section of both frontogenesis and EPV_g at 0600 UTC 5 December taken from southwest Nebraska to central Arkansas (Fig. 12a) displays an axis of frontogenesis sloping toward the cold air with maxima at about 925 and 700 hPa of over $1.0^\circ\text{C} (100 \text{ km} - 3 \text{ h})^{-1}$. At the same time there is an elevated layer of weak to negative EPV_g that is parallel to the frontogenesis axis but located about 200 km to the southeast along the cross-sectional axis. Thus, the rising branch of the direct thermal circulation is located in a region of weak-negative EPV_g, which acts to both contract the rising zone into a meso- β -scale band and enhance the magnitude of the upward vertical motion. Emanuel (1985) notes that this strong sloping updraft is typically found between 50 and 200 km away from the frontogenetical forcing on the warm side of the axis. Figure 12b, which displays the tangential ageostrophic circulation along the same cross-sectional path, reveals substantial upward vertical motion in the center of the cross section located in central Kansas. There is also an indication of an indirect (direct) thermal circulation on the south (north) side of this cross section that corresponds to the frontolytic (frontogenetic) forcing noted earlier.

Banacos (2003) has noted that “the best scenario

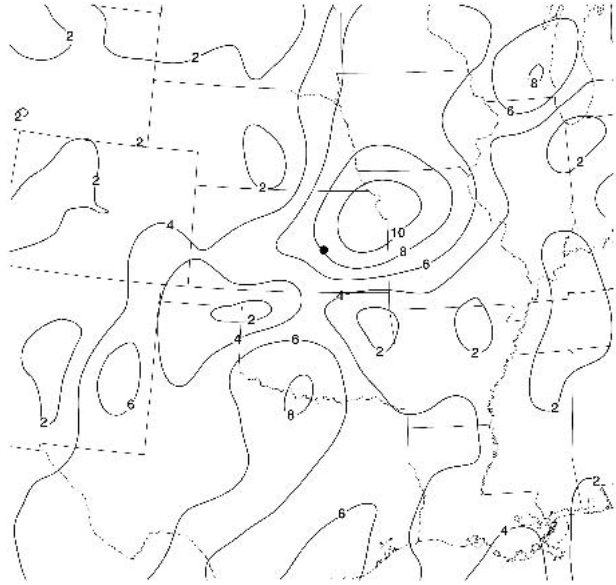


FIG. 13. Average resultant deformation for the 800–650-hPa layer ($\times 10^{-5} \text{ s}^{-1}$) for 0600 UTC 5 Dec 1999 based upon RUC II initial analysis data. Black dot indicates location of RUC II sounding shown in Fig. 14.

from a banding perspective would appear to be strong deformation (and convergence) with minimum 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.” Figure 13 shows that there was a strong area of resultant deformation

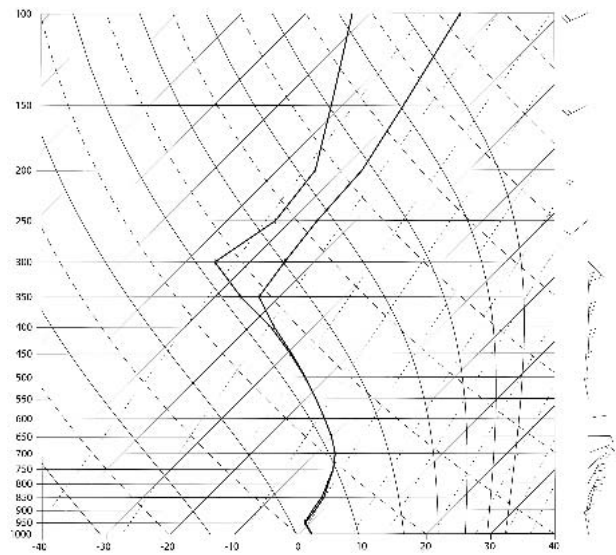


FIG. 14. RUC II skew T -log p sounding for the point shown in Fig. 13 in Kansas at 0600 UTC 5 Dec 1999. Solid lines indicate temperature and dewpoint ($^\circ\text{C}$) while wind barbs are indicated on the right-hand side in kt.

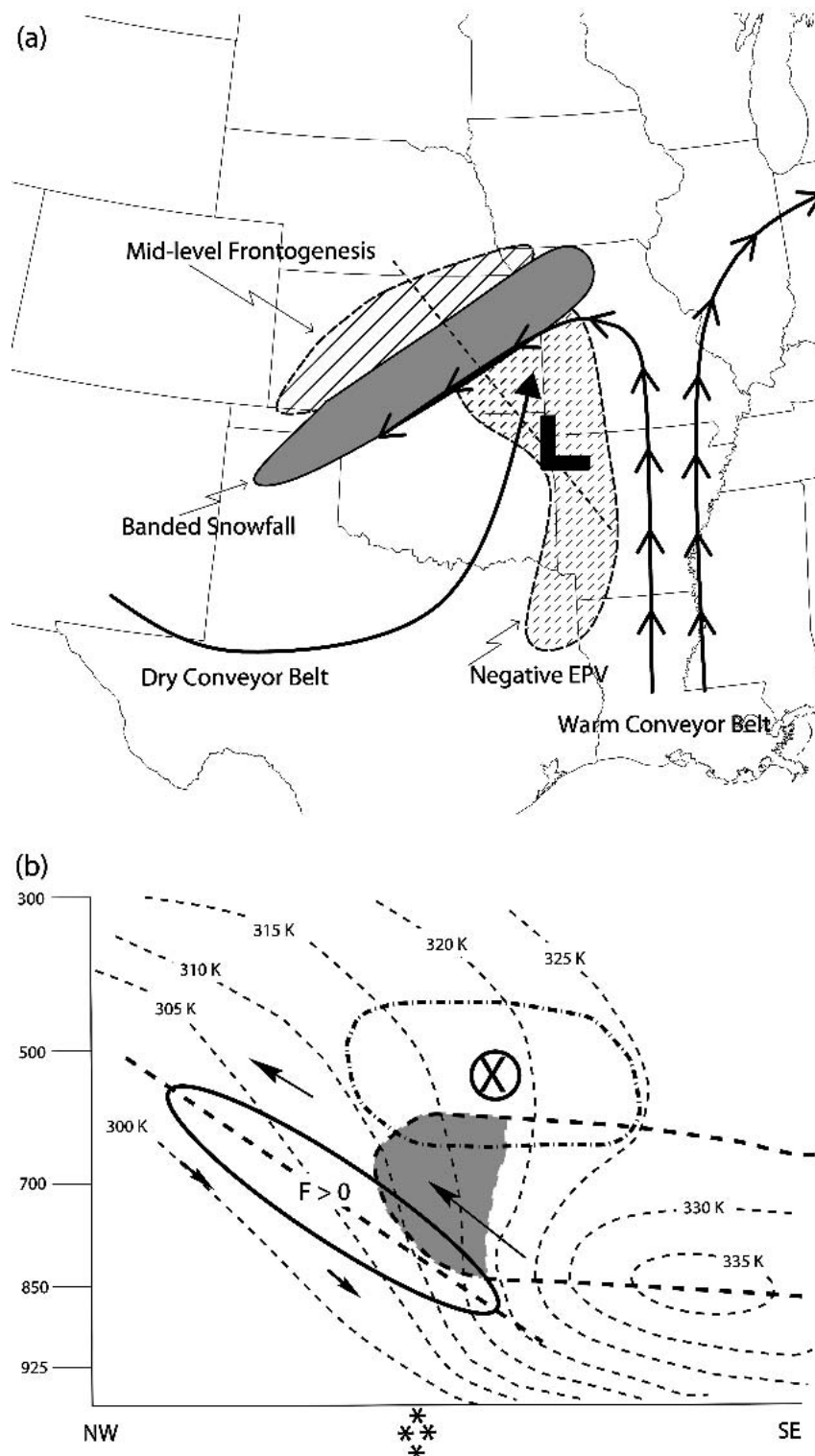


FIG. 15. Conceptual model of those physical processes contributing to heavy banded snow formation for this case study. (a) Plan view indicating the warm conveyor belt, dry slot, areas of negative EPV_g , midlevel frontogenesis, and banded snowfall. Here, L indicates position of surface low, while dashed line indicates axis along which the cross section shown in (b) is taken. (b) Cross section taken northwest to southeast across heavy banded snow, frontogenesis zone, and into negative EPV_g region. Dashed lines are θ_e (K). Elliptical region indicates zone of positive frontogenesis, with thick dashed line indicating the major axis of the zone of frontogenesis. Thick dashed line on right-hand side of figure denotes region of elevated convective instability. Dash-dot line above 600 hPa denotes region of dry-air advection in mid- to upper levels with circled X indicating the dry conveyor belt entering the cross section. Arrows indicate sense of slantwise circulation with length of arrows indicating the strength of the vertical motion. (Color versions of these schematic diagrams are available online at <http://www.eas.slu.edu/CIPS/Publications.html>.)

mation (greater than 10^{-4} s^{-1}), contributing to the frontogenesis maximum in north-central Kansas (Fig. 10b), over the majority of eastern Kansas. This region of deformation dominated eastern Kansas during both the 3-h period before and after 0600 UTC (not shown). Thus, there was a strong deformation zone over the area of interest for an extended period of time, along and north of the trowal axis. Further, a RUC II sounding taken at a point in central Kansas at 0600 UTC (Fig. 14) diagnoses several features that would point to heavy snow. First, there is a deep, moist, nearly isothermal lapse rate from approximately 950 to 750 hPa. In the actual atmosphere such isothermal lapse rates are typically due to atmospheric cooling as ice crystals aloft fall into a warm layer with temperatures over 0°C . This cooling can result in a relatively deep (often up to 1 km) isothermal layer at or below 0°C . This saturated isothermal layer is often associated with a mesoscale indirect thermal circulation, which enhances precipitation amounts near rain–snow boundaries (Szeto et al. 1988). Second, above this layer there is a deep, moist-adiabatic lapse rate extending to about 400 hPa. This shows that there was a deep layer of moisture with cloud temperatures well below -10°C , so that ice crystals were plentiful to seed the warmer, supercooled layer below. Finally, the vertical profile of winds reveals a distinct col region above the strong north-northeasterly winds, beginning at about 650 hPa. These facts document the conditions noted by Banacos (2003), which would produce not only banded precipitation, but a long-lasting banded event, which indeed did occur.

5. Discussion and conclusions

A comprehensive case study has been presented to document those processes that contributed to the formation of an extended, narrow band of heavy snowfall throughout parts of Texas, Oklahoma, and Kansas during 4–5 December 1999. These processes are illustrated in the conceptual model shown in Figs. 15a and 15b. In Fig. 15a, the plan view, one can see an area of negative EPV_g to the southeast of the surface cyclone center that is basically an area of surface-based or possibly elevated convective instability. To the north of this region EPV_g is also negative, which, in the absence of convective instability, indicates a region of CSI (if the layer is nearly saturated). Note that the latter area is near the vertical superposition of warm moist air, associated with the trowal airstream, and slightly cooler, dry air associated with the dry conveyor belt, which is seen as a dry slot in water vapor imagery. To the northwest of the negative EPV_g region a narrow zone of mid-troposphere frontogenesis develops as the warm, moist trowal airstream becomes confluent with the in situ cold air that lays to the north of the cyclone. The heavy snowband formed south of the midtropospheric zone of frontogenesis and north of the negative EPV_g zone as is seen in this plan view.

A vertical cross section taken normal to the major axis of frontogenesis (Fig. 15b) reveals the sloped, direct thermal circulation with a contracted, enhanced zone of upward vertical motion (noted by the longer upward-sloping arrows) on the warm side of the frontogenesis axis in a region of WSS (EPV_g between 0 and 0.25 PVU) or the negative region of EPV_g with θ_e surfaces nearly vertically sloped. Since this region is nearly saturated the difference between θ_{es} and θ_e is inconsequential (Halcomb and Market 2003). In this vertical depiction of θ_e surfaces one can see the southeast-to-northeast spectrum of instabilities noted earlier. There is a gradual transition from surface-based convective instability to elevated convective instability to CSI to WSS.

It is interesting that, in this particular case, the CCB played a relatively modest role in organizing the heavy snowband. However, Nicosia and Grumm (1999) have noted that for heavy banded precipitation in the northeast United States, the CCB plays a major role in transporting relatively warm, moist air from the Atlantic inland to provide deep moisture. This is in contrast to what was observed in the present case in which the WCB provided deep moisture along the trowal airstream. However, what is of paramount importance is that the underlying physical processes are basically the same in the central United States as are those in the northeast United States, namely the synergistic interaction of the major conveyor belts to create a mesoscale region of enhanced lift, instability, and deep moisture supportive of heavy snow. Furthermore, these same processes were also identified by Novak and Horwood (2002) and Novak et al. (2002) as key components to banded heavy snowfall in the northeast United States.

What is perhaps most significant about this particular case is that we have found these important processes associated with a weak surface cyclone as opposed to the more classic deep, occluding cyclones often described in the literature (Martin 1999; Novak and Horwood 2002). Second, this case underscores the importance of looking at SR flow on isentropic surfaces to better understand the development of the trowal, deep moisture advection, and isentropic lift to the northwest of the cyclone. Finally, this case highlights how important it is to focus on key processes that can interact under a special set of environmental conditions to force and focus precipitation along a narrow band over a long time period, often covering several states.

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