THE HISTORIC CHRISTMAS 2004 SOUTH TEXAS SNOW EVENT: DIAGNOSIS OF THE HEAVY SNOW BAND

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Abstract

On 24 and 25 December 2004, a rare, banded heavy snow event occurred over portions of south Texas. Maximum snow amounts of 20-33 cm (8-13 inches) were reported in the band region. Such amounts had not been observed in south Texas since the late 1800s. Moisture, lift, stability, and thermal profiles for the event were examined. On the synoptic scale, the event was associated with a strong low-latitude upper- level trough (26-28°N) and jet stream maximum of 72-77 m s^{-1} across northern Mexico and southern Texas. This trough and jet maximum combined to produce large scale lift over an already sufficiently moist south Texas region. At lower levels, a deep freezing/sub-freezing air mass was present as far south as northern Mexico. Confluent upper-level flow located northeast of the snow region, helped to maintain a deep cold air mass by producing northerly flow and cold advection at lower levels.

It was surmised that the rarity of the event was due to the combination of the very low- latitude upper-level trough, and the deep cold air mass. Anomaly calculations confirmed that both the 500 hPa heights and 850 hPa temperatures within the trough over north central Mexico were around four standard deviations below normal during the event.

To diagnose the snow band region, cross sections of saturated equivalent potential vorticity (EPV*), Petterssen frontogenesis, saturated equivalent potential temperature (θ_{es}), and relative humidity (RH) were constructed perpendicular to the snow band. During the entire heavy snow period, the cross sections indicated the presence of negative EPV* located just above an axis of mid-level frontogenesis. However, the θ_{es} surfaces in the cross sections showed a transition from the release of conditional instability (CI) and upright ascent at the beginning of the event, to the release of conditional symmetric instability (CSI) and enhanced slantwise ascent about half way through the event, as the θ_{es} surfaces became more horizontal.

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1. Introduction

On 24 and 25 December 2004, much of south Texas received a record snowfall. Measurable snow was reported from the Galveston/Houston region, southward to Corpus Christi, Brownsville and into Northeastern Mexico (Fig. 1). Embedded within the larger snow field was a narrow, heavier snow band. This band produced maximum snow depths of 33 cm (13 inches), a half width (distance between maximum snowfall and half that amount) of approximately 48-64 km (30-40 miles), and a total length of more than 320 km (200 miles). Although Fig. 1 reveals only one heavy snow band, a composite of the radar data from Weather Surveillance Radar-1988 Doppler (WSR-88D) sites across south Texas indicated that there were at least two distinct bands that formed during this event. The entire snow event occurred mainly between the hours of 0000 and 1200 UTC 25 December 2004. The heaviest snowfall of 5 to 10 cm h^{-1} (2 to 4 inches h^{-1}) occurred from 0200 UTC to 0800 UTC within the band region.

To put this Christmas 2004 snow event into historical perspective, the last time Corpus Christi or Victoria, TX, received similar 24 hour snowfall totals was February 1895. The February 1895 snow event also produced a narrow band of heavier snow across south and southeast Texas, but the snow totals were nearly double (25-50 cm, 10-20 inches) those of the current case (Griffiths and Ainsworth 1981). Snowfall amounts were likely higher with the 1895 storm since it had a longer duration (three days) than the 2004 event, which persisted for approximately 10 to 12 hours. Tables 1 and 2 show all of the recorded snow events for Corpus Christi and Victoria, TX respectively. These two
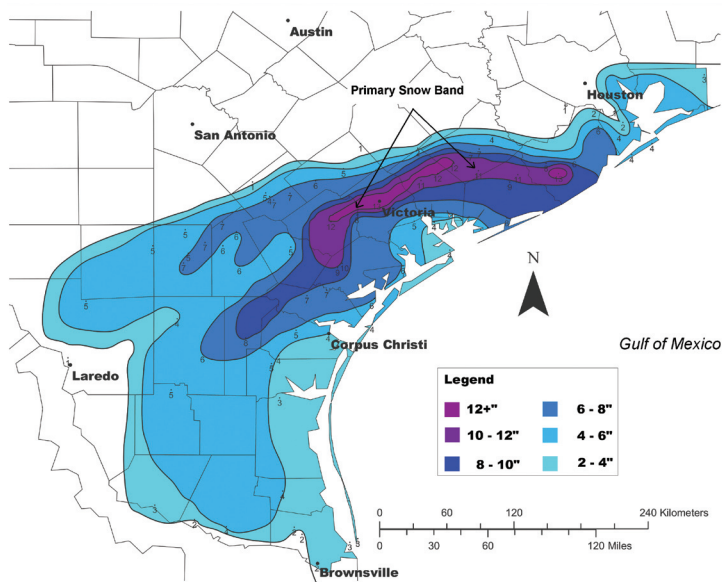
stations were chosen because they were the closest official National Climatic Data Center (NCDC) climatological sites within/near the snow band region. Table 1 shows that 72% of Corpus Christi's snow events had daily totals of one inch or less. Similarly, Table 2 reveals that 83% of Victoria's snow events had daily totals of three inches or less. For this event, both Corpus Christi and Victoria broke their records for 24 hour snowfall totals, 4.4 inches and 12.5 inches respectively. In addition, many locations from the Houston/Galveston area, southward to the lower Rio Grande Plains and northeast Mexico, recorded their first white Christmas since local records began (National Climatic Data Center 2004).

Year	Date	Amount
1888	January 15	1.0
1888	January 16	0.2
1895	February 14	4.3
1897	January 25	1.0
1897	January 28	2.0
1897	January 29	3.0
1906	February 7	0.1
1912	January 12	0.1
1914	February 25	0.5
1918	January 11	0.4
1918	December 25	0.1
1924	December 20	0.1
1924	December 21	0.2
1926	January 23	1.5
1932	March 10	0.2
1940	January 19	1.6
1940	January 21	0.1
1949	January 29	0.2
1958	February 12	0.5
1967	January 9	0.1
1973	January 11	0.2
1973	February 8	1.0
1973	February 9	0.1
2004	December 24	2.3
2004	December 25	2.1

Table 1. Snow events for Corpus Christi, TX from 1888 to 2004. Snowfall amounts are daily totals (inches), listed in chronological order. The December 2004 event was a record 24 hour snowfall amount (4.4 inches), 2.3 inches on December 24, and 2.1 inches on December 25.

Fig. 1. Analysis highlighting snowfall totals over south Texas, based on measurements obtained at 1200 UTC 25 December 2004. Snowfall is contoured every 2 inches, with a maximum contour of 12 inches within the primary band region.

The primary motivation for this study was to diagnose the potential cause(s) for the narrow snow band, which was not predicted by operational numerical weather prediction (NWP) models, and to document the occurrence of a seemingly rare, low- latitude heavy snow event. The key meteorological features present during this case were: a strong, low-latitude upper-level trough (26-28°N), and a deep, saturated, sufficiently cold air mass. Given the significance of a deep cold air mass for this case, a brief investigation into its origin and maintenance will be provided. In addition, it will be shown that it was possibly the simultaneous occurrence of a vigorous low-latitude upper- level trough and a deep sub-freezing air mass that signified the true rarity of this event. Overall, this study will reveal that, despite the rare snow totals and deep cold air mass over low-latitudes, the associated synoptic and mesoscale dynamics were not necessarily unique when compared to other more northern snow band events.

Section 2 will provide a general background on some methods used to diagnose the stability within heavy snow bands, as well as a possible mesoscale mechanism to release the instability. The data and methodology used to diagnose the features associated with the snow band will be discussed in section 3. Section 4 includes a brief synoptic and mesoscale overview of the event (including satellite and radar trends), a discussion on the origin and maintenance of the deep cold air mass, and an attempt to quantify the rarity of the low-latitude upper low and cold air mass across northern Mexico and southern Texas. Section 5 focuses on diagnosing the cause(s) of the heavy snow band, and section 6 provides a summary and conclusions.

2. Assessing the Stability and Forcing Mechanisms for a Heavy Snow Band

a. Diagnosing stability in the banded region

Many studies in the refereed literature have focused on the relationship between instability (upright or slantwise) and the existence of heavy banded snow (e.g., Seltzer et al. 1985; Moore and Lambert 1993; Martin 1998; Wiesmueller and Zubrick1998; Nicosia and Grumm1999; Market and Cissell 2002; Jurewicz and Evans 2004; Moore et al. 2005). Schultz and Schumacher (1999) emphasized that any instability, including conditional symmetric instability (CSI), will not be realized until ascent occurs within a saturated region where the instability exists.

Cross sections of equivalent potential vorticity (EPV) or saturated equivalent potential vorticity (EPV*) and equivalent potential temperature (θ_e) or saturated equivalent potential temperature (θ_{es}), taken perpendicular to the heavy snow band, can be used to help diagnose the stability. EPV/EPV* has also been referred to as moist potential vorticity (MPV) (e.g., Bennetts and Hoskins 1979; Shields et al. 1991; Martin et al.1992; Moore and Lambert 1993; McCann 1995; Nicosia and Grumm 1999).

To better describe both the mathematical and physical relationship between EPV and stability, Moore and Lambert (1993) derived a two dimensional form of EPV as shown in (1). Similarly, EPV* can be calculated in (1) by replacing θ_e with θ_{es} in (1). For simplicity, the equation was left in the form for EPV, where $M_g = v_g + f_x$ is the geostrophic absolute momentum (v_g is the geostrophic wind normal to the cross section, f is the Coriolis parameter, with the x-direction perpendicular to the thermal wind vector and increasing toward the warmer air), and g is the gravitational acceleration.

Year	Date	Amount
1895	February 14	12.0
1906	February 6	0.2
1918	January 11	1.0
1923	February 4	3.0
1926	January 23	2.6
1932	March 10	0.7
1932	March 11	1.7
1940	January 20	2.0
1940	January 22	3.0
1949	January 30	1.0
1958	February 12	3.4
1960	February 12	4.0
1973	January 11	1.2
1973	February 9	1.0
1976	November 28	0.2
1985	January 12	2.1
2004	December 24	10.0
2004	December 25	2.5

Table 2. Snow events for Victoria, TX from 1893 to 2004. Snowfall amounts are daily totals (inches), listed in chronological order. The December 2004 event was a record 24 hour snowfall amount (12.5 inches), 10.0 inches on December 24, and 2.5 inches on December 25.

$$EPV = g\left[\left(\frac{\partial M_g}{\partial p}\frac{\partial \theta_e}{\partial x}\right)_A - \left(\frac{\partial M_g}{\partial x}\frac{\partial \theta_e}{\partial p}\right)_B\right] \quad (1)$$

Term A includes the vertical wind shear ($\partial M_g/\partial p$), and the horizontal equivalent potential temperature gradient ($\partial \theta_e/\partial x$). Term B represents contributions from absolute vorticity ($\partial M_g/\partial x$), and the stability ($\partial \theta_e/\partial p$). The contribution to EPV at a point from term A will typically be negative, assuming a baroclinic environment in which the wind increases with decreasing pressure, and the temperature increases in the positive x-direction. In a baroclinic environment, EPV < 0 equates to the presence of CSI. In term B, the absolute vorticity is almost always positive, which will act to contribute to larger negative values for EPV. The key to term B is to properly assess the stability. Negative EPV could be diagnosed, but potential or conditional instability (CI) could still be present (i.e., $\partial \theta_e/\partial p$ or $\partial \theta_{es}/\partial p$ > 0 respectively), which would allow upright convection to be dominant over slant-wise convection.

In practice, heavy banded snow events have been observed with small positive EPV values between 0 and 0.25 potential vorticity units (1 PVU =10^{-6} m^2 K s^{-1} kg^{-1}) (Schumacher 2005, personal communication). Therefore, it can be very difficult to forecast well in advance (e.g., >24 to 36 hours) whether banding will be caused by enhanced frontogenetical circulations due to reduced EPV/weak symmetric stability (WSS), or due to negative EPV and the release of CSI (Schultz and Schumacher 1999). The relationship between EPV/stability and frontogenetical circulations will be discussed in the next section. Operationally, the EPV and stability cross section technique mentioned above is generally best performed within 24 hours of the banded event, when model output is likely to be more skillful.

b. Frontogenesis as a possible release of the instability

If sufficient moisture and instability (i.e., CI or CSI) are determined to be present, the next step would be to assess what mechanism(s) could force the ascent that would release the instability, and potentially form a band of heavy precipitation/snow. Schultz and Schumacher (1999) and Market and Cissell (2002) noted that heavy banded snow events may be forced by various mechanisms such as: frontogenesis, boundary layer instabilities, ducted gravity waves, orography, and Kelvin-Helmholtz instability. Out of the mechanisms listed above, mid-level frontogenesis (e.g., between 800 and 600 hPa) was frequently shown to be a significant factor in the existence, placement and strength of mesoscale heavy snow/precipitation bands (e.g., Novak et al. 2006; Schumacher 2003; Schultz and Schumacher 1999).

However, when diagnosing the potential for a heavy precipitation band, one should not focus exclusively on mesoscale lifting mechanisms, such as frontal circulations. Schumacher (2003) argued the importance for diagnosing the presence of larger scale lift in conjunction with mesoscale mechanisms when forecasting the potential for narrow banded precipitation. For example, the ageostrophic frontal circulation associated with a frontogenetical region may be enhanced through a deeper atmospheric layer if it is in phase with the synoptic scale lift produced by an upper trough/potential vorticity (PV) anomaly.

If frontogenesis is present, then the resulting thermally direct ageostrophic circulation could be the trigger to release the instability within the ascent region on the warm side of the frontal zone. As EPV becomes smaller on the warm side of the frontal circulation, the updraft becomes constricted and enhanced, leading to the production of a narrow heavy precipitation band (Martin 1998; Sanders and Bosart 1985). Modeling studies from Xu (1989) showed that in the presence of frontogenetical forcing, a single band can intensify and contract as EPV reduces toward zero or becomes negative. Numerical simulations by Knight and Hobbs (1988) also showed that as precipitation bands moved into areas of CSI, the vertical motions within the bands increased exponentially with time.

3. Data and Methodology

The entire model data used to diagnose this event was taken from archives produced from the National Weather Service's (NWS) Advanced Weather Interactive Processing System (AWIPS) at the NWS Weather Forecast Office (WFO), Corpus Christi, TX. Since this study was intended to benefit operational forecasters, only data that were operationally available to the affected south Texas WFOs were utilized to diagnose this event. Analyses from the 80 km North American Mesoscale Eta (NAM-Eta) model, from 0000 and 0600 UTC 25 December, were used to diagnose the synoptic environment.

To diagnose the possible cause(s) for the heavy band(s) of snow, cross sections of Petterssen frontogenesis from AWIPS (Petterssen 1956), saturated equivalent potential vorticity (EPV*), θ_{es}, and relative humidity (RH) were constructed perpendicular to the heavy snow band. This method was similar to that used by Nicosia and Grumm (1999), Novak and Wiley (2002), Schumacher (2003), and Jurewicz and Evans (2004). The cross sections were generated using the Local Analysis and Prediction System (LAPS) from data archived on AWIPS (Albers et al. 1996). The LAPS output was desirable since it was available hourly over a regional domain, and with a horizontal grid resolution of 10 km. For this case study, the LAPS analyses

were centered on the Corpus Christi WFO's County Warning Area (CWA) of responsibility across south Texas, which coincided with the region of the primary heavy snow band.

In order to track the snow band evolution, reflectivity data were utilized mainly from composites from four NWS NEXRAD WSR-88D Doppler radars at WFOs across south Texas, including: Corpus Christi (KCRP), Brownsville (KBRO), Austin/San Antonio (KEWX), and Houston/ Galveston (KHGX). These composited reflectivity products made it possible to view a much larger geographic region than what would be seen using a single radar.

The snowfall data were compiled and hand analyzed from cooperative observers, Automated Surface Observing Systems (ASOS), and from reports by public and law enforcement officials. Most of the snowfall data appeared to be accurate. However, when discrepancies existed, particularly within the suspected regions of heavier snowfall, the data were either discarded or estimated using the ratio of liquid water equivalent to snowfall as reported by the Victoria cooperative observer, who was trained by National Weather Service personnel to properly measure snowfall and the liquid equivalent.

In an attempt to better quantify the combined rarity of a strong, low-latitude upper-level trough, and a deep sub-freezing air mass, 500 hPa height, 850 hPa and 925 hPa temperatures, and surface pressure standardized anomaly graphics were produced. These graphics were generated using the National Centers for Environmental Prediction (NCEP) reanalysis gridded dataset, and 30 year normals from the National Climatic Data Center (NCDC) from 1961-1990. The anomalies were standardized using 21-day running mean values. A detailed explanation of the methodology used to compute these anomaly graphics can be found in Grumm and Hart (2001).

4. Synoptic and Mesoscale Overview

a. Large scale pattern

The discussion of the significant synoptic features associated with this storm will focus on the period between 0000 and 0600 UTC, 25 December 2004, which were the synoptic times closest to the heavy snow period (the heaviest snowfall occurred between 0200 UTC and 0800 UTC 25 December 2004). Additional details on the synoptic-scale features out to 48 hours prior to the beginning of the banded snow event can be found in Wilk et al. (2007).

At 0000 UTC, 25 December 2004, the upper-level pattern exhibited a deep layer ridge axis over the Pacific Northwest of the United States, a broad trough over the

Upper Midwest and Central Plains states, and a separate short wave trough over north central Mexico (Figs. 2a-b). The upper trough over north central Mexico began to separate from the Upper Midwest trough when it was over the central Baja California region, approximately 24 hours prior to the event (not shown). This more southern upper trough then moved east across north-central Mexico by 0000 UTC 25 December, as seen in Fig. 2. At 250 hPa, there were two distinct jet maxima along the eastern sides of the two upper troughs (Fig. 2a). One jet maximum was located between the Middle Atlantic Coast and the southern Mississippi River Valley, and the other was located over northern Mexico. The strongest jet core was over the Tennessee River Valley and the Mid Atlantic states, with maximum speeds of 93-103 m s^{-1} (180-200 kt). A secondary jet core of 72-77 m s^{-1} (140-150 kt) ran across northern Mexico and southern Texas. This configuration placed much of southern and eastern Texas (the region of the eventual heavy snow band) in the exit region of the southern-most upper jet streak, and the entrance region of the northern-most one. At 500 hPa, there was a strong cyclonic vorticity center over north central Mexico, with cyclonic vorticity advection moving into northeastern Mexico and southern Texas.

At the lower and middle levels of the atmosphere, the noteworthy features were the depth of freezing/ sub-freezing air, and the lack of any significantly strong surface cyclone or anti-cyclone (Fig. 2c-d). The 850 hPa analyses (Fig. 2c) showed that the 0°C isotherm had pushed down to the border of Texas and Mexico. At the surface, where temperatures (not shown) were in the 0 to 3°C range (lower to mid 30s°F), there was a weak ridge of high pressure, with a long east-west axis from northern Mexico, northeast into the Ohio River Valley (Fig. 2d). This ridge axis was the remnant high pressure area left behind by the Arctic air mass that initiated the deep cold push of air into south Texas early on 23 December. There was a weak surface cyclone (1009 hPa) over the southwest Gulf of Mexico, which was approximately 835 to 925 km (450 to 500 nautical miles) southeast of the eventual heavy snow band. Between the surface cyclone, well offshore to the east, and a ridge of higher pressure over northern and central Texas, the entire south Texas region was under the influence of low-level northerly flow.

By 0600 UTC 25 December 2004, upper-level analyses indicated that the base of the trough continued to move eastward toward south Texas (Figs. 3a-b). The split between the southern and northern upper troughs, noted six hours earlier, was still evident The strongest upper-level jet at 250 hPa (82-93 m s^{-1}, 160-180 knots) was starting to concentrate on the east side of the northern portion of the upper trough, over the southeast U.S. and

Middle Atlantic regions with a secondary jet over the western Gulf and northeastern Mexico (Fig. 3a). At 500 hPa, strong cyclonic vorticity advection was now occurring across all of south Texas, with the center of the vorticity maximum still over north central Mexico. Deep cold air, surface and aloft, remained over south Texas and the snow band region. Temperatures at 850 hPa ranged between 0 to -4°C (Fig. 3c), and surface temperatures (not shown) were a bit lower than the previous 6 hours, ranging from -2 to +1°C (upper 20s to lower 30s °F). A weak surface low continued moving eastward over the western Gulf, with a nearly steady central pressure. A broad surface high pressure ridge axis also continued to stretch from the Ohio River Valley, southwest to central Texas and north central Mexico.

b. Satellite features

Beginning at 0000 UTC 25 December 2004 (Figs. 4a-b), the upper low was clearly evident in the Geostationary Operational Environmental Satellite (GOES) water vapor imagery over northern Mexico and southern Texas (Fig. 4a). A dry slot was seen around the south and east side of the upper low, which included southern Texas. The infrared (IR) imagery indicated that the brightest/coldest cloud top temperatures were over the western and northern Gulf of Mexico. Of particular interest was the orientation of the cloud and moisture features in both the water vapor and IR data. For example, the clouds east of the upper low over south Texas were oriented east/west, while closer to the upper low, they were more northeast/
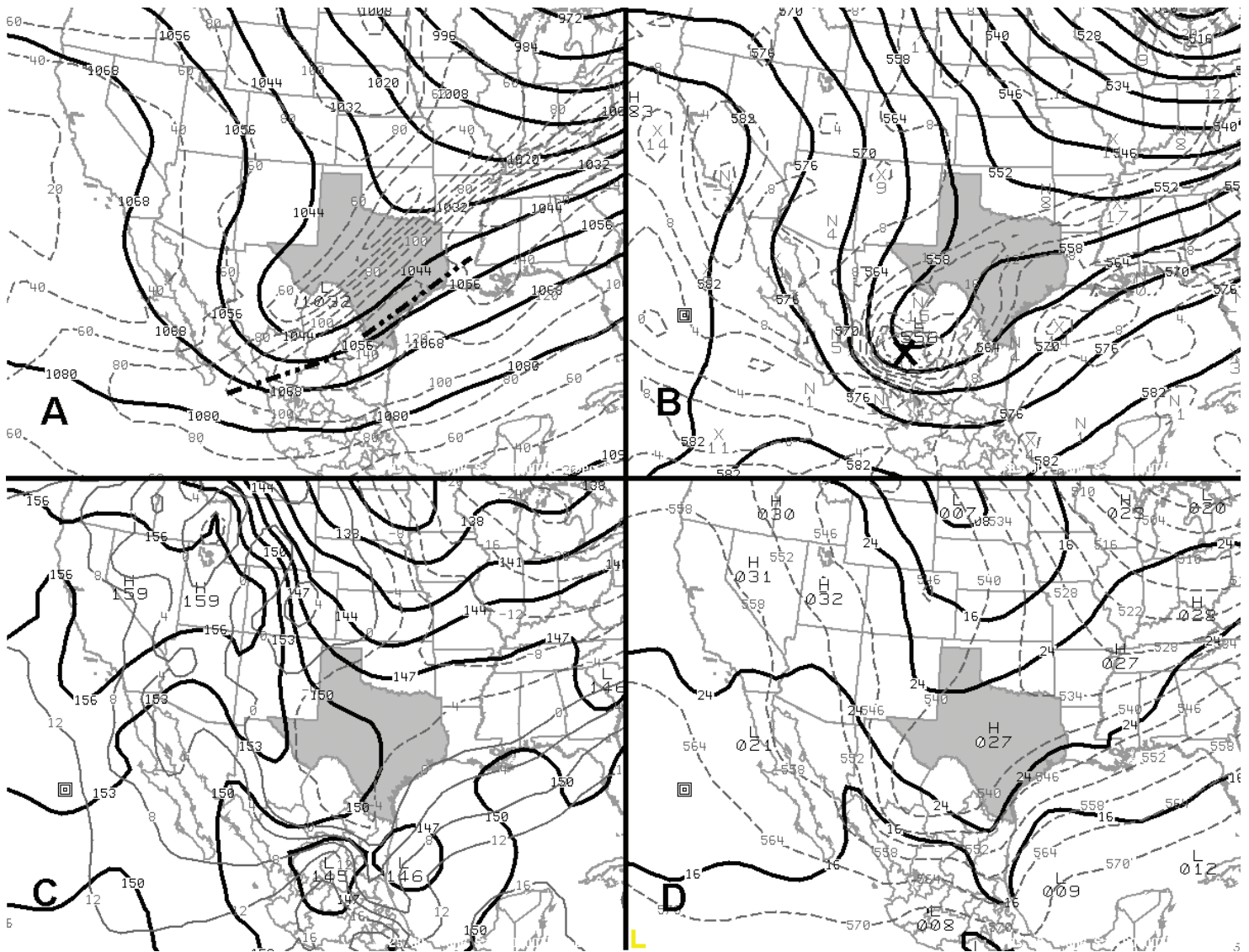

Fig. 2(a-d). 80 km NAM-Eta analysis for 000 UTC 25 December 2004: (a) 250 hPa geopotential height (solid, every 12 dm), isotachs (dashed, every 20 knots, with the jet core marked by dot-dash line); (b) 500 hPa geopotential height (solid, every 6 dm) and absolute vorticity (dashed, every 4 x 10^{-5} s^{-1}, with center of cyclonic vorticity max marked with bold X); (c) 850 hPa geopotential height (solid every 3 dm) and temperature (positive values are solid lines, negative values are dashed lines every 4°C); (d) Mean sea level pressure (solid, every 8 hPa), and 1000-500 hPa thickness (dashed, every 6 dm).

southwest. The east/west oriented cloud features appeared to be associated with more convective activity (discussed below in section 4c.), and the northeast/ southwest oriented clouds seemed more stratiform, and aligned parallel to the 1000-500 hPa thickness (thermal wind) field (compare to Fig. 2d).

As the upper low continued to move eastward toward south Texas, the cloud features became oriented more northeast/southwest by 0600 UTC (Figs. 4c-d), aligning parallel to the 1000-500 hPa thickness field (compare to Fig. 3d). Cloud top temperatures, as depicted by the IR imagery, changed very little over south Texas during the period between 0000 and 0600 UTC, with the coldest cloud tops well offshore over the Gulf of Mexico, east-southeast of the main upper jet core discussed above.

c. Radar trends

The radar data around 0000 UTC 25 December 2004 indicated that the precipitation was dominated by convection over the eastern portions of south Texas (Fig. 5a). At this time, convective-looking bands were oriented east/west (as discussed above in the section 4b.), with embedded higher reflectivity values of 35-45 dBZ, especially over the coastal counties of south Texas and the western Gulf of Mexico. The National Lightning Detection Network (NLDN) data even indicated a few cloud to ground (CG) strikes within the convective regions, particularly around Corpus Christi Bay and the western Gulf of Mexico. Despite the very few CG strikes over the land areas throughout this event, reports from the
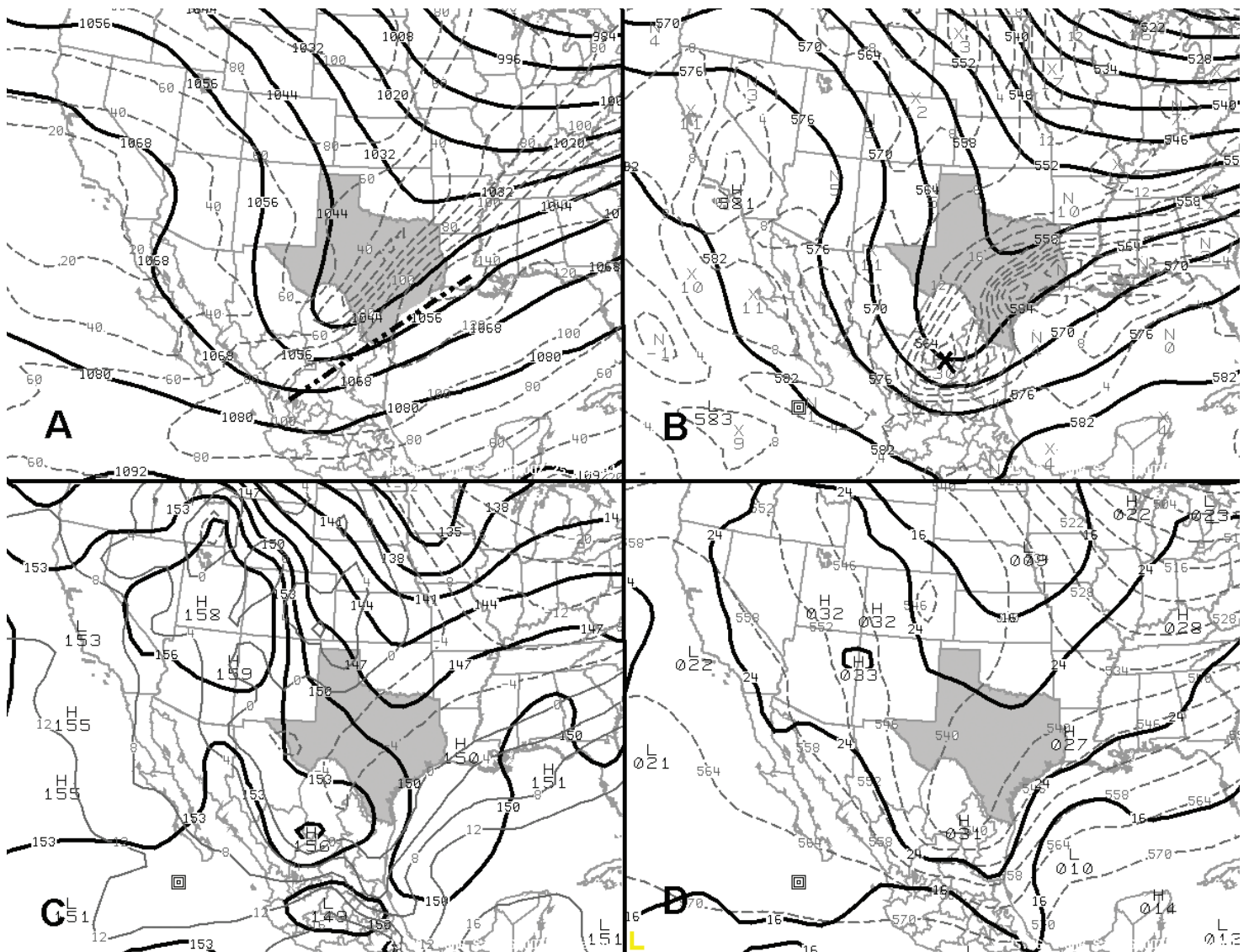

Fig. 3. Same as Fig. 2, except from 0600 UTC 25 December 2004.

Corpus Christi surface observation and residents across the coastal region of south Texas confirmed the presence of periodic "rumbles" of thunder, especially early in the event between 0000 and 0200 UTC 25 December 2004.

The transition from a convective to a more stratiform precipitation event over south Texas seemed to occur around 0200 UTC (Fig. 5b), with the dominant convective regions moving eastward into the western Gulf waters. Also by this time, at least two distinct bands of moderate to heavy snow were developing within the regions labeled as "developing primary" and "developing secondary" bands. The secondary band, located about 48 km (30 miles) northwest of the center of the primary band, was not clearly evident in the final snowfall analysis (Fig. 1) due to its shorter duration as compared to the primary band. This resulted in snowfall totals of approximately 40% lower in the secondary band.
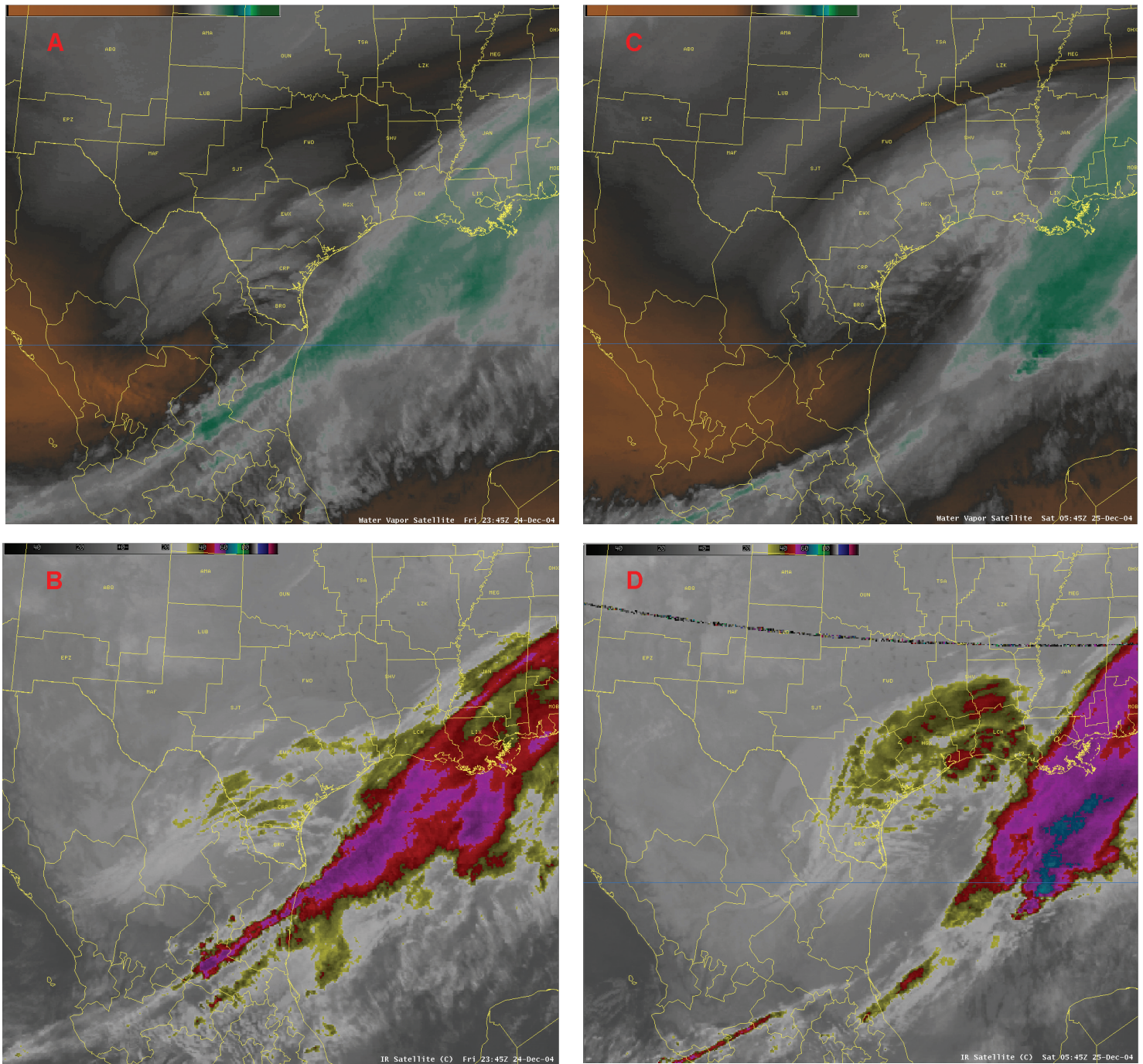
Fig. 4(a-d). GOES Water Vapor (WV) imagery (a,c) and Infrared (IR) imagery (b,d) valid at 0000 UTC (a, b) and 0600 UTC (c, d) 25 December 2004. Darker, more orange shades on the WV imagery signify less moisture in the middle troposphere, while whiter, greener shades indicate higher moisture. Pink, red and blue shaded regions on the IR imagery indicate colder/higher cloud tops.

After 0400 UTC 25 December, the secondary (northern) heavy snow band weakened and/or merged into the location of the primary band, which became the dominant band through the end of the event (Figs. 5c-d). The primary band weakened and began to move eastward offshore between 0800 and 1000 UTC (Fig. 5e-f). The heaviest snow in the banded region occurred between 0200 and 0800 UTC 25 December, with maximum observed snowfall rates of 5 to 10 cm h^{-1} (2 to 4 inches h^{-1}). The time scale for the heaviest snowfall (4 to 6 hours) was similar to those observed by Novak et al. (2004) in many banded precipitation events over the Northeast U.S.

d. The presence and maintenance of deep cold air

During the late afternoon hours of 24 December 2004, just prior to the onset of the heavy snow, cloudy and breezy conditions prevailed, with surface temperatures holding around 1-2°C (34 to 36° F) across all of south Texas. A moderately strong surface ridge was likely held in place over northern Texas and the southern Mississippi River Valley by confluent upper level flow (Figs. 2 a-b, and 3a-b). This surface ridge helped to maintain northerly low-level flow, which continuously advected colder, drier air across the region. Figure 6 shows an example of the surface conditions at 2100 UTC 24 December, which was a few hours prior to the onset of the event. The magnitude of the low-level dry air was evident from observations of surface dew point temperatures, which ranged from 12°C (10°F) over central Texas to -1°C (30°F) near the Texas/Mexico line. The 1200 UTC 24 December 2004 Corpus Christi (CRP) sounding (the sounding closest to the eventual heavy snow band region) revealed that the low-level drier air was confined to below 950 hPa, with saturated, sub-freezing conditions up to 650 hPa (Fig. 7). Even though the 1200 UTC 24 December 2004 sounding indicated deep sub-freezing temperatures, the flow veered quickly from northerly at low levels, to southwesterly around 850 hPa and above, which indicated the presence of warm advection from 850 hPa and above. However, despite strong southwest flow of 15-20 m s^{-1} above 850 hPa at this time, there was only weak warm advection (1-3°C 12 h^{-1}) in the layer between 850 and 500 hPa (not shown). At lower levels, a mix of light (non-accumulating) precipitation fell into a relatively dry boundary layer late in the afternoon of the 24th, causing evaporative cooling that resulted in wet bulb zero temperatures falling to 0°C or less near the surface (implied from Fig. 7).

By 0000 UTC 25 December 2004, the precipitation increased in coverage and intensity. The CRP sounding at this time continued to show dry low-level conditions, and a deep freezing/sub-freezing saturated layer up to around 600 hPa, which was 100 hPa higher than the previous 12 hours (Fig. 8). There was a slight warming of 1-2°C between 700 hPa and 800 hPa, which was likely the result of weak warm advection observed during the day of 24 December 2004, as discussed above. However, this elevated warm layer was relatively shallow (less than 700 m), and the northerly flow had now deepened another 100 hPa to around 800 hPa (compare Figs. 7 and 8).

By 0600 UTC (around the time of peak snow intensity), there was generally neutral to slightly negative (cold) temperature advection over all of south Texas between 850 hPa and 500 hPa (see Figs. 3c-d). Therefore, it is surmised that the combination of evaporative cooling due to precipitation falling into the drier air in the lowest 50 to 100 hPa, the implied adiabatic cooling due to the intensifying ascent associated with the approaching upper trough and associated jet streak circulations, and possible cooling from melting precipitation at mid-levels were likely enough to offset the weak mid level warm advection. Although likely not to be the case for this study, the cooling effects due to melting have been shown to be significant enough to change liquid precipitation to snow (Kain et al. 2000). Additionally, as discussed in section 4a above, the presence of a persistent surface/low-level anticyclone helped to maintain low-level freezing/sub-freezing air across the area.

e. Quantifying the rarity of the event

When this study first began, it was thought that the exceptionally low-latitude upper-level low that moved across northeast Mexico and southern Texas during the event was equally as rare as the deep, low latitude sub-freezing air mass. However, additional personal forecast and observational experience across south Texas suggested that upper level troughs/lows moving across southern Texas and northern Mexico, although not common, could not be considered very rare. Therefore, it was then suspected that it was the combination of the very low-latitude upper-level trough and deep cold air mass that embodied the true rarity of the case.

To investigate this theory, anomaly plots for upper level heights and lower to mid-level temperatures were produced for 0000 UTC 25 December, the approximate onset time of the event. At 500 hPa, these graphics revealed that the upper-level low that was moving across north central Mexico, just prior to the onset of the event, was approaching four standard deviations below normal for that date (Fig. 9a). Similarly, temperature anomalies at both 850 and 925 hPa were around four standard deviations below normal over north central Mexico,
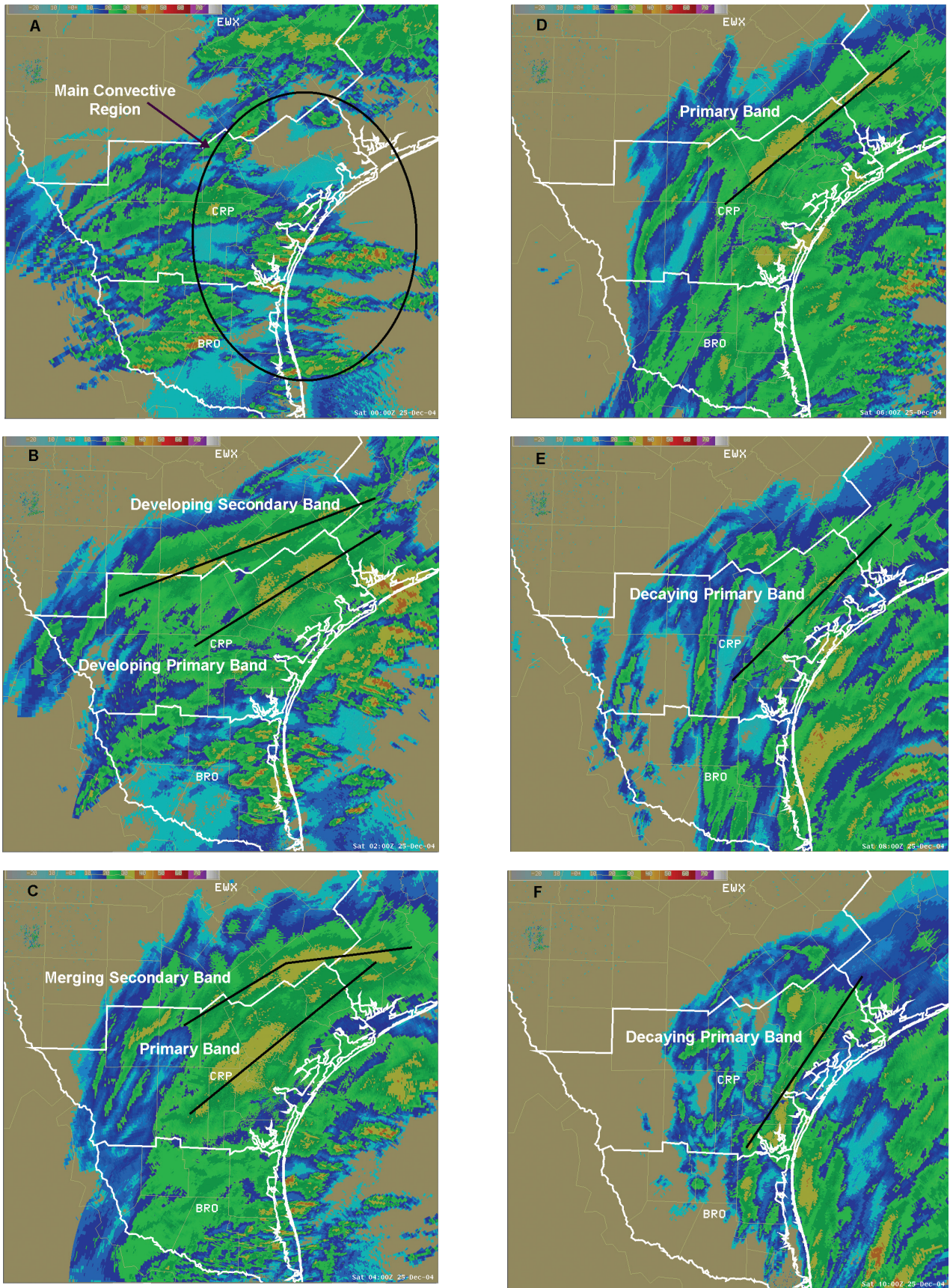
Fig. 5(a-f). Reflectivity composites from the CRP, BRO, EWX and HGX radars at 0000 UTC (a), 0200 UTC (b), 0400 UTC (c), 0600 UTC (d), 0800 UTC (e), and 1000 UTC (f) 25 December 2004. The heavy snow bands are depicted by solid black lines. County warning Areas of each NWS office outlined in solid white lines. CRP =Corpus Christi, BRO = Brownsville, EWX= Austin/San Antonio, and HGX=Houston/Galveston.
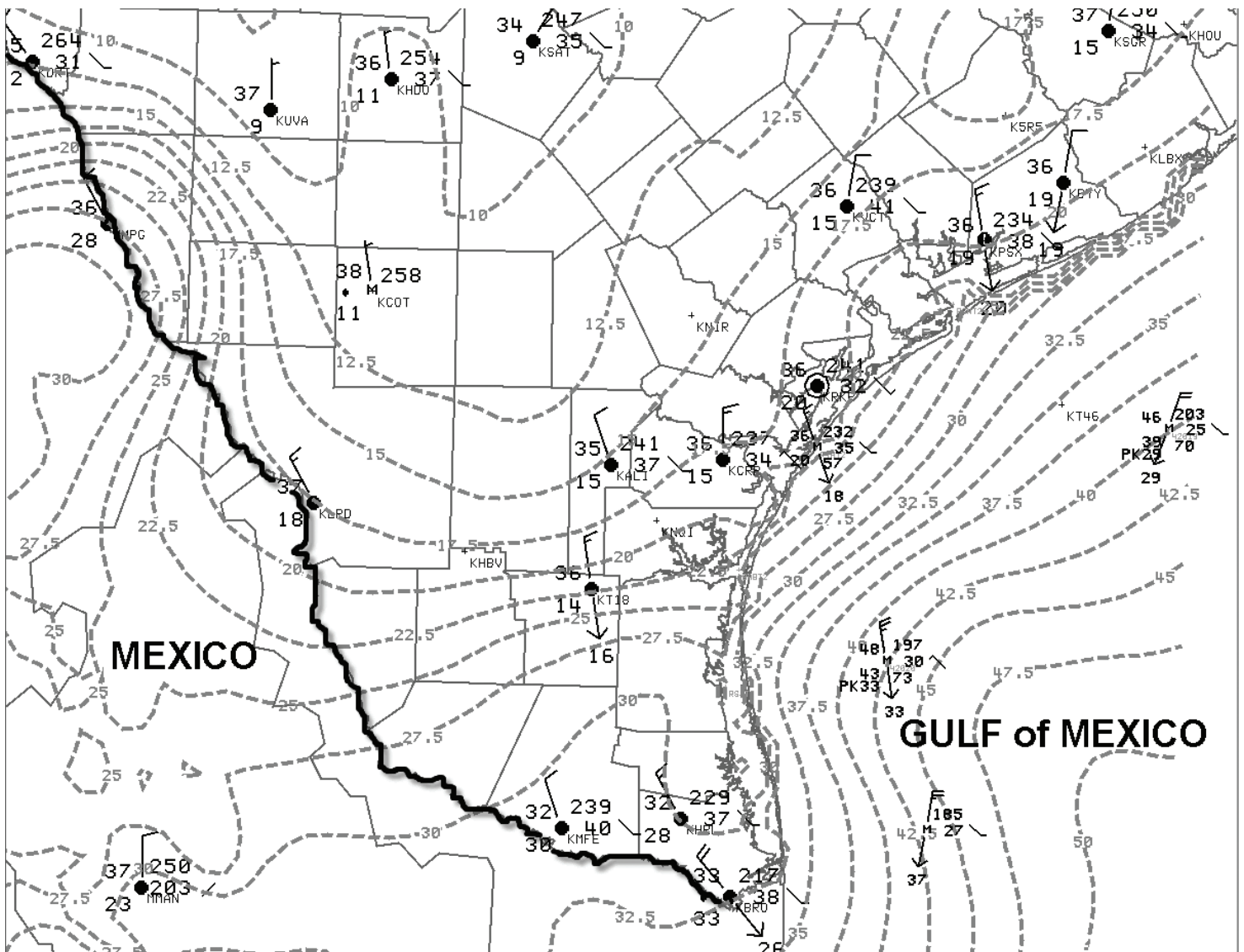
Fig. 6. Surface METAR and buoy data plotted in the standard surface station model, and LAPS analysis of surface dew point temperatures (dashed lines, every 2.5° F) valid at 2100 UTC 24 December 2004.

under the upper trough region, and two to three standard deviations below normal over south Texas and the snow band region (Figs.9b-c). Temperature anomalies over the snow band region remained two to three standard deviations below normal through the end of the snow event (not shown).

Closer to the surface, the main features of note were the anomalously high surface pressures over northern Mexico and northern Texas (Fig. 9d), and the lower pressures over the central Gulf of Mexico. The center of the high pressure anomaly ranged from two to three standard deviations above normal over northern Mexico, to one to two standard deviations over northern Texas. It is interesting to note that the core of higher surface pressure over north central Mexico was not well analyzed by the NAM at this time (Fig. 2d). Conversely, the low pressure

anomalies were only one to two standard deviations below normal over the central and southern Gulf. Thus, due to its closer proximity to the snow band region, the surface high pressure anomalies likely had more influence on the event than did the low pressure anomalies well to the east of the region. These higher pressures helped to maintain the cold northerly low-level flow previously discussed.

Grumm and Hart (2001) defined anomaly values of > 2.5 standard deviations from the 30 year means as being anomalous. Both the 500 hPa trough and lower-level temperature anomalies easily exceeded this threshold. The above analysis provides evidence that the combination of a very low-latitude upper–level trough and deep cold air mass contributed to the rarity of the snowfall. Additional investigation quantifying the rarity of the joint occurrence of such anomalies should be conducted in the future.
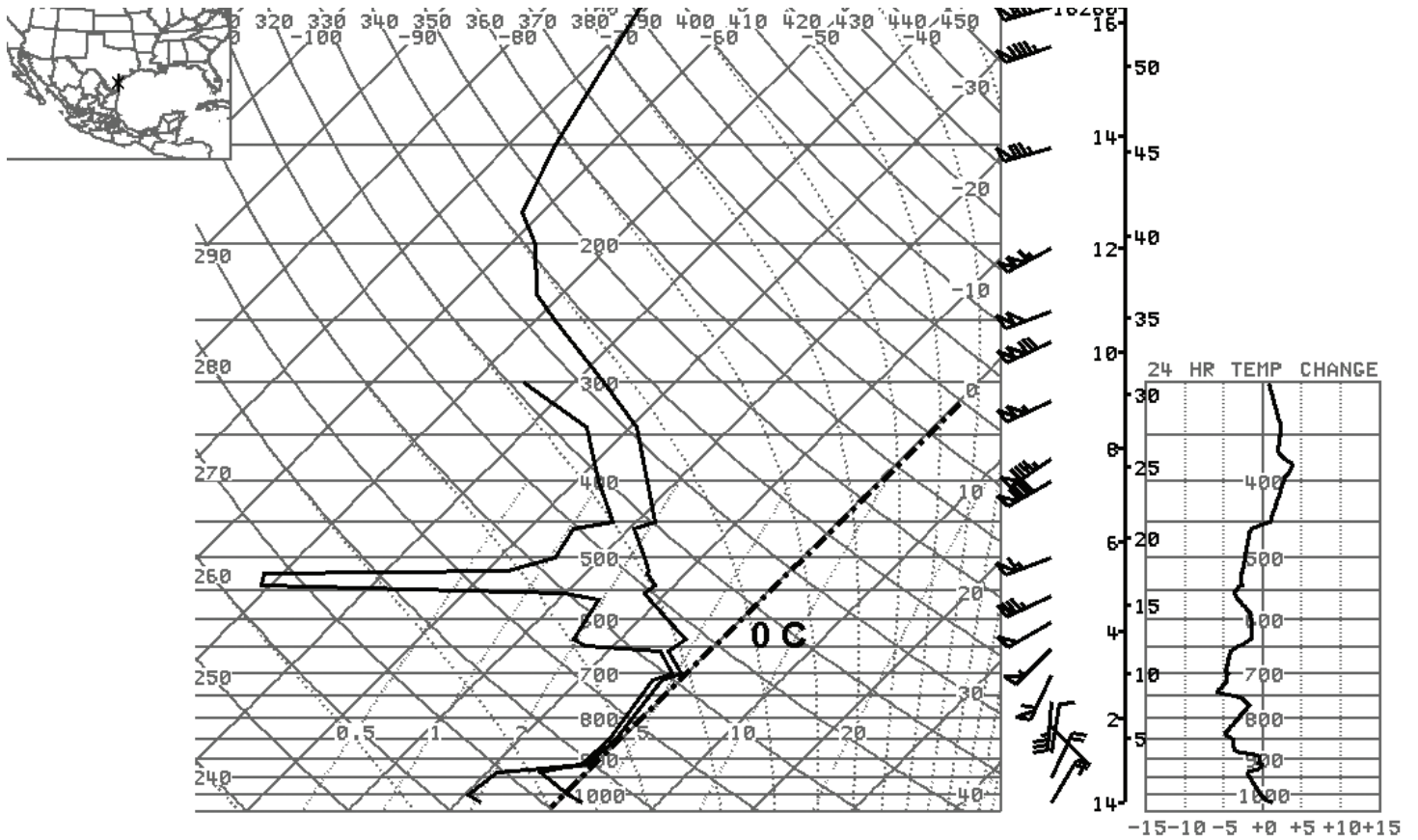
Fig. 7. Observed skew T-log p sounding from the WFO Corpus Christi, Texas (CRP) for 1200 UTC 24 December 2004. The dot-dashed line denotes the 0 °C isotherm.
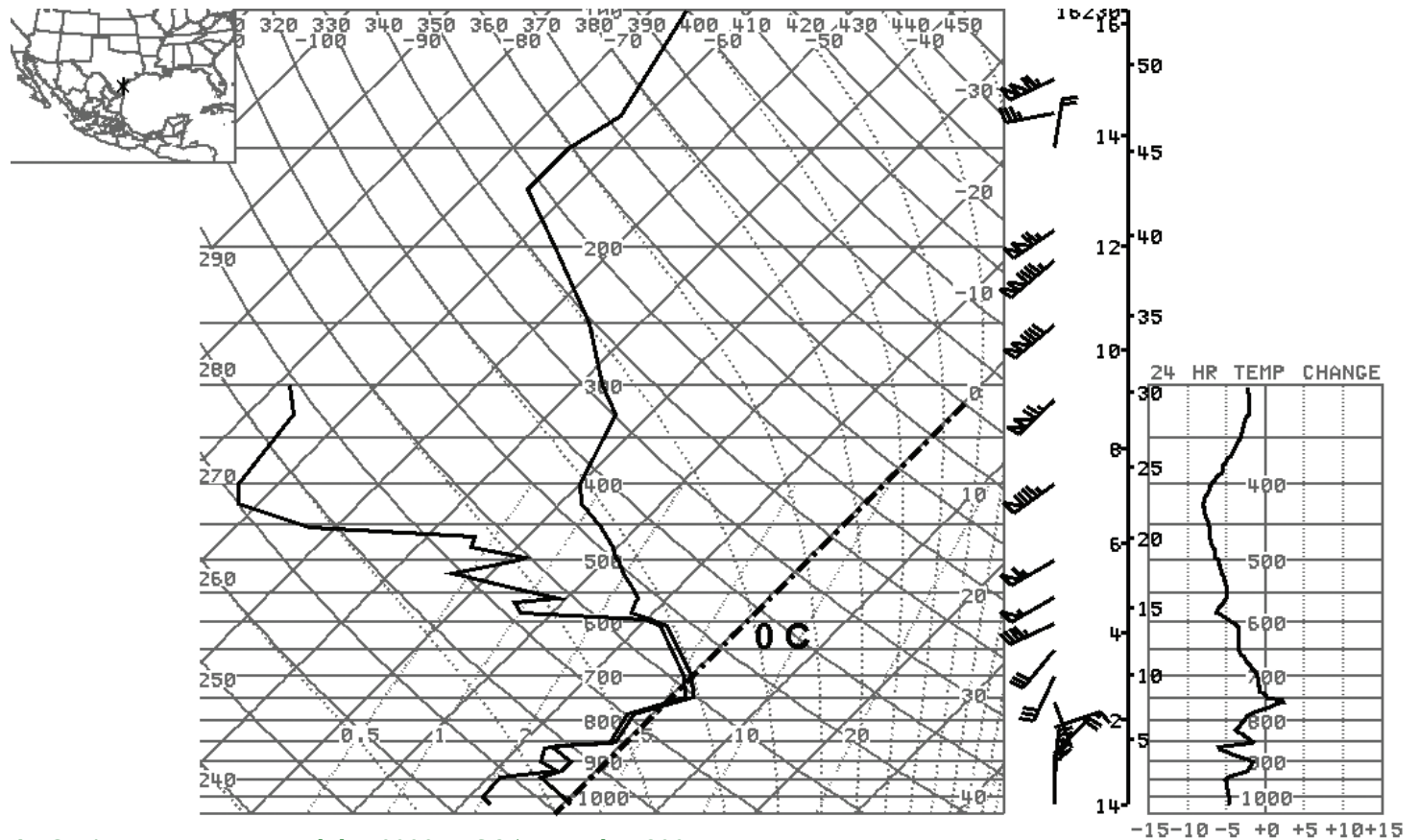

Fig. 8. As in Fig. 7, except valid at 0000 UTC 25 December 2004.

5. Diagnosis of the Heavy Snow Band

The time frame chosen to diagnose the heavy snow band was 0000 to 0600 UTC 25 December 2004. As mentioned in section 4, this was the synoptic time that was closest to the time period of the heaviest snowfall, which occurred between 0200 and 0800 UTC. The cross sections were constructed every two hours (beginning at 0000 UTC 25 December 2004) perpendicular to the heavy snow band region.

Beginning at 0000 UTC 25 December 2004, the cross section revealed an axis of frontogenesis gently sloping from east to west in the layer between 850 and 700 hPa

(Fig. 10). The ageostrophic vertical circulation (not shown) produced by this low to mid-level axis of frontogenesis appeared to be the main mesoscale source of ascent within the banded region. In addition, this frontogenetic region was also embedded within a saturated (RH > 80%) environment of deep synoptic-scale ascent due to the approaching upper-level trough and embedded jet streaks discussed in section 4. For simplicity, the RH data was omitted from the cross sections. In general, RH values of >80% were observed from around 400 hPa down to the surface for the entire event. This saturation was necessary for the existence of either CI or CSI.
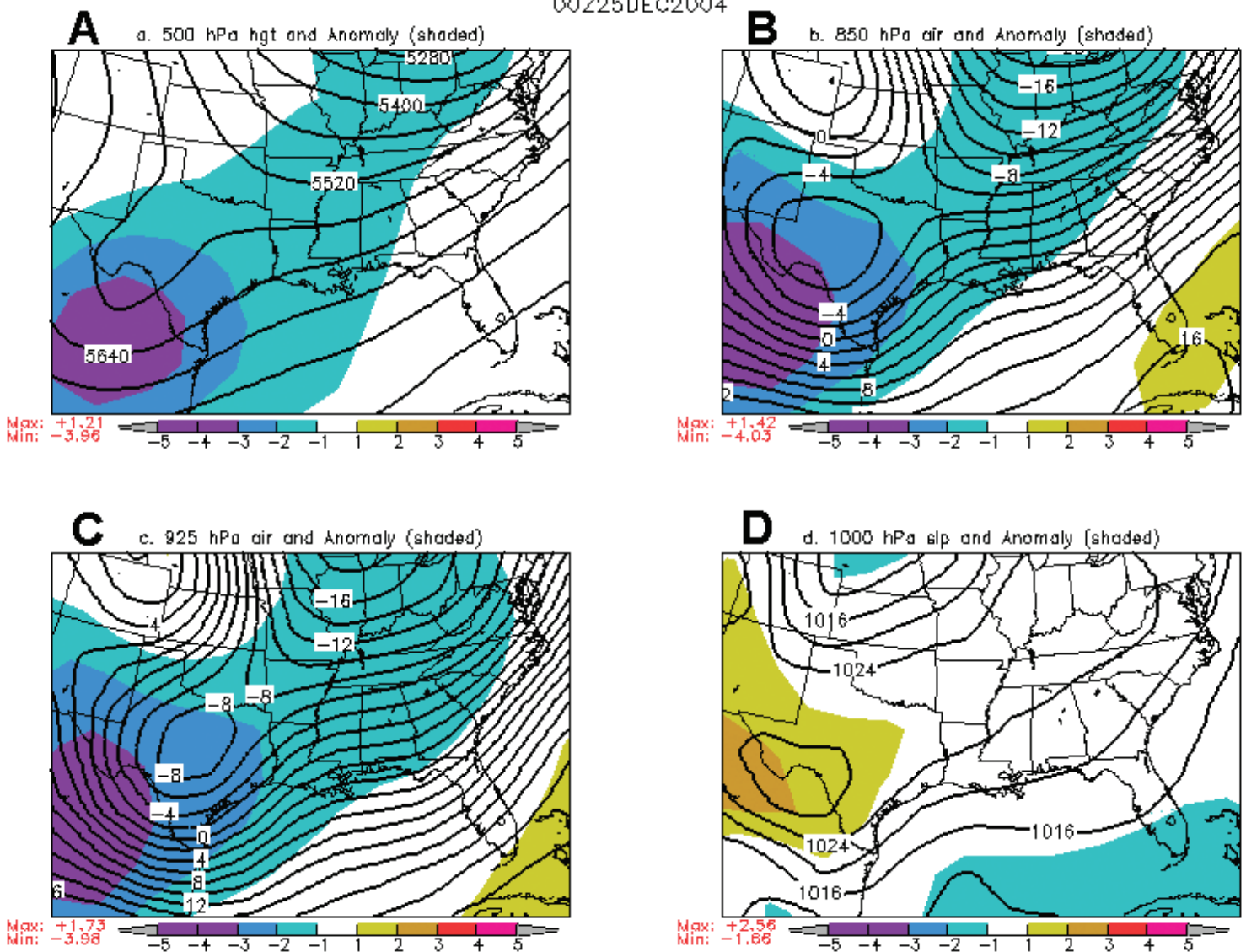

Fig. 9 (a-d). Observed meteorological fields and standardized anomalies at 0000 UTC 25 December 2004 including: (a) 500 hPa heights (contoured every 60 m) and standardized anomalies, (b) 850 hPa temperatures (contoured every 2°C) and standardized anomalies, (c) 925 hPa temperatures (contoured every 2°C) and standardized anomalies, and (d) surface MSLP (contoured every 4 hPa). Standardized anomalies are color shaded every 1 standard deviation.

Just above the axis of maximum frontogenesis, there was a large region of negative saturated equivalent potential vorticity (EPV*) of -0.5 to -1.0 PVU (Fig. 10). The θ_{es} surfaces just above the frontogenesis axis were showing a decrease with height, indicating the presence of CI. Therefore, the stability favored upright convection and the release of CI, which likely helped to enhance the snowfall through this time. A preliminary study of this case by Becker et al. (2005) also supported this conclusion.

As the dry conveyor belt wrapped around the south and east side of the upper low around 0000 UTC 25 December (Fig. 4a), this likely helped to produced a potentially unstable layer (θ_e decreasing with height) above the frontogenetic zone. In turn, the vertical motion associated with the upper-level trough and jet, as well as the frontogenesis zone, could then lift the potentially unstable layer to become conditionally unstable (θ_{es} decreasing with height). This scenario is typically observed on the eastern side of a developing extra-tropical cyclone (Carlson 1980). Therefore, the decreasing mid and upper-level moisture during this time likely helped to enhance upright convection (CI) within the heavy snow band region. Reports of thunder did seem to fit well with this conclusion. However, thunder alone may not necessarily preclude the existence of CSI and slantwise convection (Holle and Watson 1996).

The other significant contributor to the negative values of EPV* observed above the heavy snow band region was the strong vertical wind shear. Winds above the frontal inversion from the 0000 UTC 25 December 2004 CRP sounding (Fig. 8) were generally from the southwest, increasing from around 15 m s^{-1} at 750 hPa, to more than 70 m s^{-1} at 300 hPa. As was seen in term A in equation (1) in section 2a of this paper, a strong increase in wind speeds with height would help to contribute to negative values of EPV*.

Between 0200 and 0400 UTC 25 December 2004, the cross sections continued to reveal a deep region of negative EPV* near and above the mid-level frontogenetic zone (Figs. 11-12). By 0400 UTC, the θ_{es} surfaces were indicating a more neutral, to possibly slightly conditionally unstable stratification, which indicated the decreasing potential for upright convection to be dominant (Fig. 12). In addition, this layer of CI was higher and shallower by 0400 UTC, with flatter θ_{es} surfaces near the top of the frontal zone.

By 0600 UTC 25 December 2004 (Fig. 13), the cross section showed that a region of negative EPV* near and above the mid-level frontogenetic zone remained in place, but the magnitude and extent (horizontal and vertical) of the negative EPV* region had decreased. Also, the θ_{es} surfaces were much more horizontal above the frontal
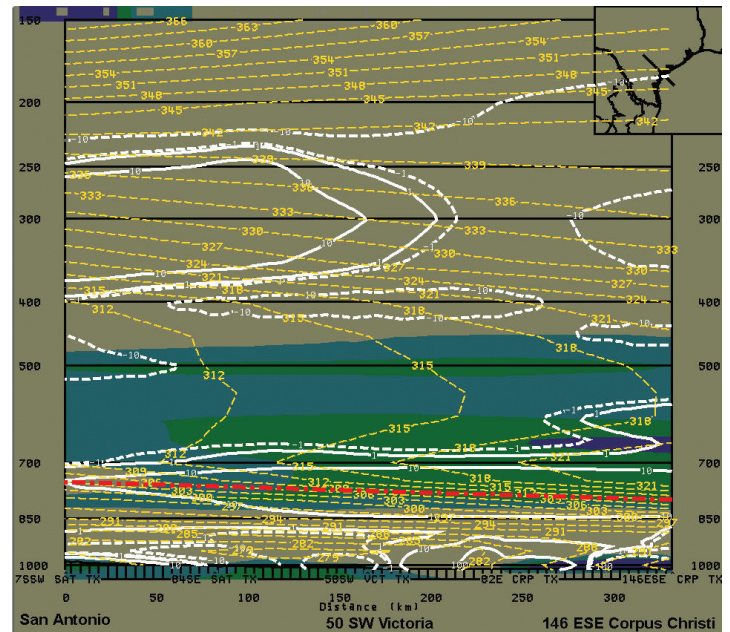

Fig. 10. LAPS cross section, along line seen in the upper right inset, valid at 0000 UTC 25 December 2004 and perpendicular to the primary snow band. Cross section includes: Petterssen frontogenesis (solid white contours represents positive frontogenesis, x 10^{-10} K m^{-1}s^{-1}; the red dot-dashed line marks axis of maximum frontogenesis at mid levels), saturated equivalent potential temperature (yellow dashed line, degrees kelvin), and saturated equivalent potential vorticity (negative values with dark shading, every 0.25 PVU, potential vorticity units, where 1 PVU =10^{-6} m^2 K s^{-1} kg^{-1}) Black text labels denote locations along the cross section. The center of the primary heavy snow band was near Victoria, TX. Note: The relative humidity (not shown) was > 80% from 400 hPa down to the surface.

zone, and no longer decreasing with height. Given the evolution to flatter θ_{es} surfaces in the region above the frontogenesis zone at this time, the heavy snow banding was likely the result of enhanced slantwise ascent from the release of CSI on the warm side of the frontal zone. Therefore, the banded heavy snow region appeared to transition from the release of CI through upright convection, to the release of CSI through enhanced slantwise upward motion, sometime between 0400 and 0600 UTC. Therefore, the first half of the heavy snow period seemed to be dominated by the release of CI and upright convection, and the later half became dominated by the release of CSI and enhanced slantwise ascent. The coverage and intensity of snowfall began to steadily decrease after 0800 UTC 25 December 2004, as the large-scale upper trough and upper-level jet began to move east of the region.

6. Summary and Conclusions

On 24-25 December 2004, a rare banded heavy snow event occurred over portions of south Texas. On the synoptic scale, the event was attended by a very low-latitude upper-level low that moved across north central Mexico and south Texas, a strong upper jet around 77 m s^{-1}, and a deep sub-freezing, saturated air mass. It was hypothesized that it was the combination of the very low-latitude upper trough and a deep sub-freezing air mass that contributed to this rare event. To better quantify this suspected rarity, anomaly calculations were made for the 500 hPa heights, 850 and 925 hPa

temperatures, and surface pressures for the onset time of the storm. The anomaly calculations revealed 500 hPa heights and 850 and 925 hPa temperatures around 4 standard deviations below the 30 year normal within the upper trough region over north central Mexico. These same fields had anomalies of 2 to 3 standard deviations below normal over the banded snow region across south Texas. Therefore, it is concluded that the combination of a very low-latitude upper-level trough and deep cold air mass contributed to the rarity of the snowfall.

In order to diagnose the cause of the heavy banded snowfall, cross sections of saturated Equivalent Potential Vorticity (EPV*), Petterssen frontogenesis, θ_{es}, and relative
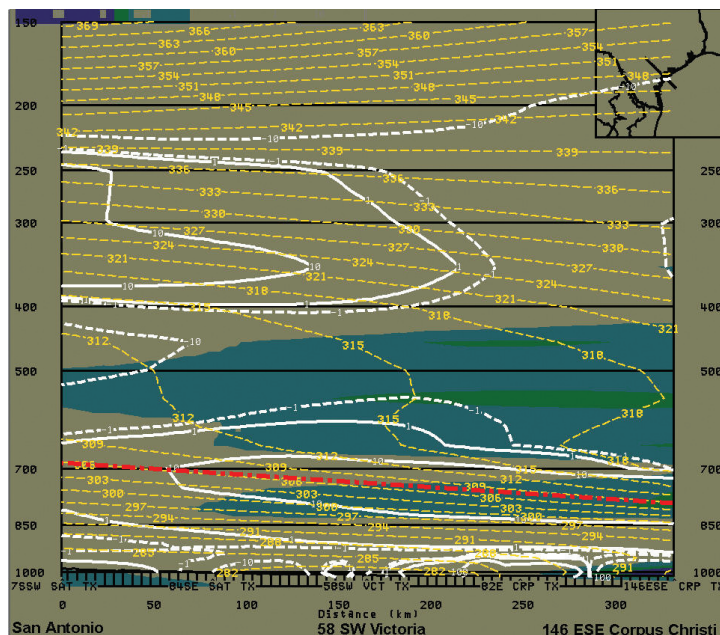

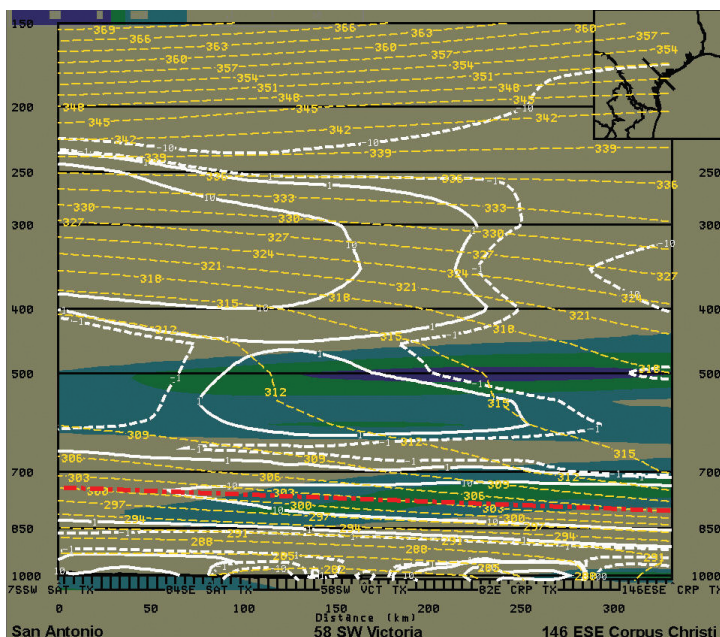Fig. 11. As in Fig. 10, except for 0200 UTC 25 December 2004.

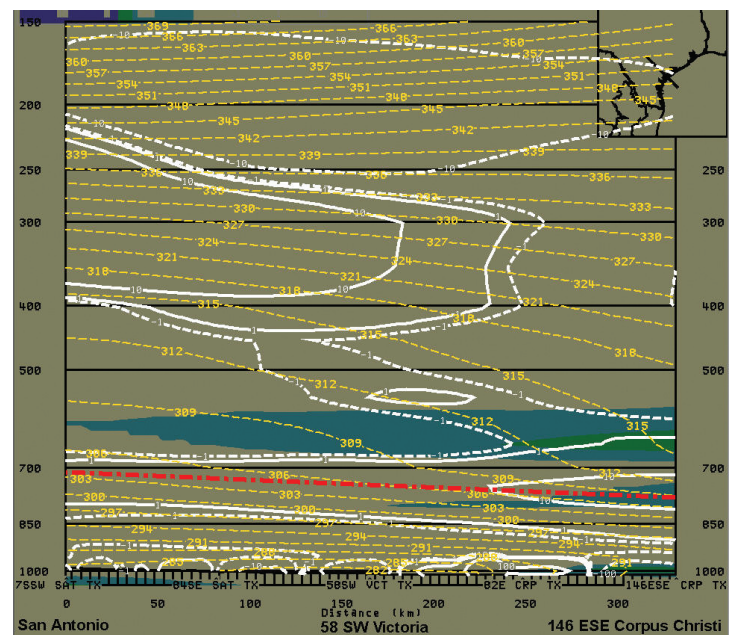Fig. 13. As in Fig. 10, except for 0600 UTC 25 December 2004.

Fig. 12. As in Fig. 10, except for 0400 UTC 25 December 2004.

humidity (RH) were constructed perpendicular to the heavy snow band. This method was similar to that used by several other published studies on heavy banded snow events. The general features that were observed in the cross sections throughout the entire event were: a mid-level (650-850 hPa) frontogenesis axis that sloped gently from east to west, negative values of EPV* just above the axis of frontogenesis, and saturated (RH > 80%) conditions within the banded region.

It was shown in section 2 that CSI exists when EPV or EPV* is negative in a baroclinic atmosphere that is convectively and inertially stable. In addition, as small positive values of EPV decrease or become negative, frontal circulations can be enhanced and constricted horizontally, which could by itself explain the formation of narrow, intense band(s) of precipitation. However, even if the EPV (EPV* in this case) was negative within the region of heavy banded precipitation, which it was for this event,

both CI and CSI could co-exist. Therefore, it was necessary to assess the conditional stability to determine whether CI or CSI would dominate. If CI and CSI coexisted, CI (upright convection) was expected to dominate, due to its much shorter time scale for growth (Moore and Lambert, 1993).

To determine the conditional stability, θ_{es} surfaces were analyzed within the cross sections. For the first few hours of the event (centered around 0000 UTC 25 December), θ_{es} was decreasing with height just above the frontogenesis axis, indicating the presence of CI, and therefore the likelihood that upright convection was dominant within the heavy snow band region. By 0600 UTC, the θ_{es} surfaces were no longer folding back, and showed a more gradual/horizontal slope upward from east to west. This indicated more conditionally stable conditions, which have been observed to be more typical within regions of CSI. In addition, the bands were aligned parallel to the 1000-500-hPa thickness gradient (compare the thickness in Figs. 2d and 3d to band orientation in Fig. 5), which is typically observed within banded regions that result from the release of CSI. Therefore, by 0600 UTC, there appeared to be a transition from upright to slantwise convection through the release of CSI within the heavy snow band. A formal explanation of this stability evolution is beyond the scope of this paper.

In general, studies showing a transition from CI to CSI within a heavy banded snow region were difficult to find in the refereed literature. However, recently Chamberlain and Hamilton (2007), and Novak et al. (2008) have documented a similar stability transition of CI to CSI, to small conditional stability in banded snow events. In contrast, Emanuel (1983) proposed the opposite type of stability evolution than found in this study within a frontal band. He gave an example of how a two-dimensional circulation resulting from CSI could cause θ_e surfaces to become over-turned (i.e., become convectively unstable). This type of scenario could occur within the upper branch of frontal circulation along the warm side of a frontal zone. The top branch of this circulation could act to fold the θ_e surfaces back on themselves, creating a more convectively unstable environment.

In conclusion, this study not only documented that heavy banded snow can occur at low-latitudes (i.e., south of 30°N), but also demonstrated that techniques developed to diagnose banded events at higher latitudes could be applied for southern latitude cases. This study also revealed that the synoptic and mesoscale forcing mechanisms (e.g., strong upper-level trough, jet maxima, and mid-level frontogenesis) were very similar to those present for more northern heavy banded snow events. Therefore, forecast strategies such as those documented by Novak et al. (2006) and Schumacher (2003) could be applied to more southern banded snow events. Finally, although heavy banded snowfall events are rare across south Texas, it is theorized that banded heavy rainfall events may be more common. Additional case studies are needed to determine if the techniques utilized in this study could be used to forecast and diagnose southern latitude banded heavy rainfall events.

Author

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