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AN ANALYSIS OF AN EARLY SEASON HEAVY SNOW EVENT OVER NORTH DAKOTA AND THE POSSIBLE INVOLVEMENT OF CONDITIONAL SYMMETRIC INSTABILITY

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1. Introduction

Forecasting winter storms in the Great Plains is often a difficult task to accomplish accurately. There can be many factors acting in concert to bring about the formation of an intense weather system, capable of producing heavy precipitation. This was the case on 4-5 October 2005 over the Northern Plains. A low pressure system brought heavy snow amounts (10 to 20 in.) from southwestern North Dakota into southwestern Manitoba. The snow storm was record breaking in that no storm of this magnitude had occurred so early in the autumn season, dating back to 1874. The heaviest snow occurred over North Dakota in a narrow band approximately 500 km long by 150 km wide (Figure 1). Numerous banded precipitation features were observed during the event. It is often observed that heavy precipitation areas tend to occur in relatively narrow bands (1000 km long by 60-100 km wide) to the north and west of a surface cyclone (e.g. Cissell and Marwitz 1998, Novak and Horwood 2002, Sanders and Bosart 1985, Moore and Blakeley 1988). This mesoscale banding occurs quite frequently, not only with winter storms, but also in the presence of frontal areas (Bennets and Hoskins 1979, Neiman and Fedor 1993). The formation of these bands and their subsequent persistence can many times be attributed to the release of conditional symmetric instability (CSI) through frontogenetical forcing (Bennets and Hoskins 1979, Moore et al. 2005, Nicosia and Grumm 1999, Xu 1986). This was believed to be the case with the 4-5 October 2005 event over North Dakota. It will be shown, through various analyses, that the release of CSI via intense frontogenetical forcing may have been involved in the intensification and maintenance of a mesoscale precipitation band early on in the event.

Section 2 briefly reviews some of the important processes involved in creating mesoscale snow bands. Section 3 describes the methodology used to perform this case study. Section 4 details the synoptic scale and mesoscale setting for the event. Section 5 contains further mesoscale analysis of the event that includes investigations of CSI and frontogenesis. Finally, Section 6 will summarize important conclusions reached.

2. General Theory

The main objective of this paper is not to analyze or critique the copious amount of research that has been performed relating to banded precipitation. We want to utilize these concepts to diagnose and better understand the processes which occurred during the heavy snow event of 4-5 October 2005.

Moore et al. (2005) provided an excellent overview of the important dynamical features involved in winter storm systems. They pointed out that research done by Harold (1973), Carlson (1980), and Danielson (1964) led to the identification of three airstreams, the cold conveyor belt (CCB), warm conveyor belt (WCB), and dry conveyor belt (DCB). It is the interaction of these belts that produces an environment favoring heavy banded precipitation (Nicosia and Grumm 1999). Typically, the WCB moves north of the warm front, splitting and creating two branches. One branch moving northwest around the surface low, and the other moving north and east with respect to the surface low, rising upward [see Moore et al. 2005 Fig. 2]. The branch moving northwest and wrapping cyclonically around the low has been referred to as the "TROWAL airstream" (Martin 1998a, b). A TROWAL (trough of warm air aloft) is crucial to increasing frontogenetical forcing leading to further intensification of vertical motion and possible banded precipitation (Moore et. al. 2005).

Nicosia and Grumm (1999) further demonstrated that the interaction of the cold conveyor belt (CCB) with the dry conveyor belt (DCB) can lead to further destabilization of the environment. This process is evident through an analysis of saturated equivalent potential temperature (θ_{es}). As the DCB and CCB interact, the slope of the θ_{es} surfaces becomes steeper. This is the result of cold dry air aloft overrunning cold moist air at low levels, decreasing the static stability of the atmosphere. This is very important as the net result is destabilization. Hence, the environment can eventually become conditionally or potentially unstable with respect to slantwise (conditional/potential symmetric instability) and/or gravitational convection (conditional/potential instability). However, in the plains states it is the interaction of the moist WCB and the DCB that results in a reduction of static stability (Moore et al. 2005).

Stability is a very important parameter to diagnosis in both winter and summer storms, much more in the former than many forecasters realize. Forecasters mainly use parameters such as convective available potential energy (CAPE), the Showalter index, and the lifted index to assess the stability of the atmosphere. However, symmetric instability can also be paramount in the production of precipitation, especially in the winter.

The theory of symmetric instability has been around for quite some time. Bennetts and Hoskins (1979) were some of the first authors to apply CSI to frontal rainbands. In general, symmetric instability occurs in an atmosphere favored by the following: low static stability, strong vertical geostrophic shear, and weak or anticyclonic geostrophic vorticity (Bluestein 1993). CSI is an atmospheric state in which the atmosphere is stable with respect to vertical displacements (i.e., statically stable with equivalent potential temperature (θ_e) increasing with height), and stable with respect to horizontal displacements (i.e., inertially stable with geostrophic absolute vorticity (η_g) greater than zero), but conditionally unstable to slantwise displacements. Schultz and Schumacher (1999) as well as Nicosia and Grumm (1999) discussed why cross sections of θ_{es} as well as saturated geostrophic potential vorticity (MPV_g^*) can be used in assessing instability in the atmosphere through the geostrophic absolute momentum $M_g - \bar{\theta}$ relationship ($\bar{\theta}$ can be replaced by θ, generally a valid approximation) in two and three dimensions. Hence, one can use θ_{es} and MPV_g^* to diagnose areas that might possess CSI. Symmetric instability occurs in regions where θ_{es} surfaces are steeper in slope than M_g surfaces (or areas where MPV_g^* is negative). One must use MPV_g^* instead of MPV_g (unsaturated) to detect CSI as indicated by Schultz and Schumacher. Areas of potential symmetric instability (PSI) can be detected with MPV_g. If the atmosphere is not saturated then CSI cannot be released, thus saturation is the key. Many authors (i.e.

Market and Cissel 2002) used a relative humidity of 80% or more to assume saturation has occurred in a forecasting scenario. This way one could use θ_e and MPV_g to detect areas of possible CSI. Schultz and Schumacher (1999) provided an excellent review of CSI theory, and we highly recommend the reader refer to their paper for the technical aspects of CSI. Schultz and Schumacher (1999) stressed that one must first determine if instability is present and then whether it can be released. Just because CSI or conditional instability (CI) may be present does not mean an intense precipitation band will develop. There must be forcing to allow the release of the instability. They also emphasize that a mixture of both types of instability may be present and that checks for CI must be made in order to try and differentiate the two.

Another important and related parameter is frontogenesis. Not only is it important in the production of vertical motion necessary for the formation of precipitation, it can also force the release of any slantwise instability present as long as saturation is present. Increasing frontogenesis disrupts the thermal wind balance as warm air is advected into a region. In response to the warming, ageostrophic flow develops creating a thermally direct vertical circulation to restore the thermal wind balance. This can lead to a positive feedback mechanism, which can intensify low level convergence leading to stronger vertical motion and heavier precipitation (Nicosia and Grumm 1999). Moore et al. (2005) found that the heavy snow tended to fall on the warm side of the frontogenetical maximum aloft, in a region of weak symmetric stability (WSS) (0 to 0.25 potential vorticity units, 1 PVU = 10^{-6} m^2 s^{-1} K kg^{-1}). This indicates that frontogenetical forcing allows the release of the instability that was present, which can lead to a heavy precipitation band similar to Nicosia and Grumm (1999).

3. Data and methodology

Most of the data used in this research study were archived through the Advance Weather Interactive Processing System (AWIPS) archiving package. Satellite, surface, and upper air data were obtained for 4-5 October 2005 as well as GFS, RUC and NAM forecast model runs and analyses. A detailed upper air and surface analysis was performed to evaluate the state of the atmosphere. All cross sections and most computed fields were created by AWIPS through the volume browser tool. Surface and upper-air analyses were also obtained from the Storm Prediction Center and the Hydrometeorological Prediction Center as indicated. Cross sections of MPV_g^*, θ_{es}, Lifted Index, and frontogenesis were created to diagnose areas that potentially contained CSI and/or CI. These fields were chosen based on the Moore et. al (2005) and Nicosia and Grumm (1998) analyses. Isentropic analyses were performed to indicate positions of the TROWAL and other system relative features similar to Moore et al. (2005).

4. Synoptic Analysis

At 1200 UTC on 4 October 2005 cold surface high pressure was centered over central Saskatchewan with northeast surface flow extending south into the Northern Plains (Fig. 2a). Low pressure was located over the southern Rockies, with a stationary low pressure trough extending from the Central Plains, northeast to the Great Lakes. A 500-mb trough with embedded shortwaves was located over the northwest U.S. (Fig. 2b) with broad southwest flow extending from the central Rockies, into the Northern Plains. An upper-level ridge extended from Texas into the western Atlantic. During the morning and early afternoon of October 4th, precipitation ranged from showers and thunderstorms over the eastern Dakotas and Minnesota to light rain over western North Dakota and a mixture of light rain and light snow over eastern Montana (not shown). Surface temperatures in western and central North Dakota remained nearly steady from the mid 30s to lower 40s ºF through the afternoon of the 4th.

By late afternoon (2100 UTC) deep convection had developed over central and eastern Nebraska and moved northeast along the stationary front (not shown). Also, at this time an area of light to moderate rain developed over eastern Wyoming and western South Dakota (Fig. 3c). Through analysis of water vapor imagery and the RUC model (Figs. 3a, b) it appears that precipitation was developing over eastern Wyoming ahead of a mid level shortwave moving across the central Rockies. The precipitation intensified as it entered western South Dakota and developed north into southwest North Dakota.

At 0000 UTC on 5 October 2005 the surface trough remained nearly stationary from the central Rockies to the Great Lakes. Surface low pressure was located over eastern Colorado, with another area of developing low pressure over northwest Iowa (Fig. 4a). The 500-mb trough (Fig 4b) had progressed eastward to the vicinity of western Wyoming and central Montana. An upper jet streak was observed over southern Canada at 250 mb (Fig. 4c). North Dakota was located under the entrance region of this jet streak, a favorable region for the release of CSI (Bluestein 1993). Radar imagery at 0000 UTC showed an area of moderate precipitation occurring from western South Dakota into southwest North Dakota (Fig. 5). Other showers and thunderstorms were occurring over eastern North Dakota and parts of Minnesota at this time. However, it was the intense precipitation occurring over western North Dakota that is the primary interest of the authors. The precipitation persisted for several hours as it translated from western North Dakota into southern Manitoba. It will later be shown that the maintenance and intensification of the precipitation may have been due, in part, to the release of CSI through frontogenetical forcing.

By 0600 UTC on 5 October 2005, the weak surface low over northwest Iowa had moved north to near Sioux Falls, South Dakota (Fig. 6a). The 500-mb low (Fig. 6b) was positioned over northern Wyoming with a vorticity maximum (not shown) over the western Dakotas. Banded precipitation over southwest North Dakota had propagated northeast with the system, into north central North Dakota. To the south and east of the banded precipitation, numerous thunderstorms had developed over central and eastern South Dakota (Fig. 7).

At 1200 UTC on 5 October 2005 a 1004-mb surface low was centered over extreme southeast North Dakota with an occlusion extending southeast into south central Minnesota (Fig. 8a). A stationary front extended from the occlusion, east to the Great Lakes and the cold front extended south through Iowa and then southwest into the southern Rockies. The widespread convection over central and eastern North Dakota had lifted northeast into far northeast North Dakota and northern Minnesota. The 500-mb low (Fig. 8b) had closed off and was located over extreme northwest South Dakota. The dry air intrusion, associated with the 500-mb shortwave, had progressed northward through central North Dakota, into southwest Manitoba (not shown). Satellite imagery from 1200 UTC 5 October 2005 denoted a well defined comma head signature over eastern Montana and far western North Dakota, typical of a mature cyclone (Fig. 9). Widespread light to moderate snow was still occurring over western and north central North Dakota at this time.

The surface low was centered over eastern North Dakota at 1800 UTC 5 October 2005 (Fig. 10a) while the 500-mb low pressure center was located over central North Dakota (Fig, 10b). Light to moderate precipitation was still occurring over western and central North Dakota as the event transitioned into a “wrap around moisture” precipitation event by this time. The precipitation continued to diminish after 1800 UTC, as the system slowly moved east out of the region.

5. Mesoanalysis

As mentioned earlier, instability is very important to the development of an intense precipitation band. Three ingredients must be present in order for deep convection to be present: lift, moisture, and instability. Schultz and Schumacher (1999) reiterate that these three ingredients must also be present for moist slantwise convection (the release of CSI). Through various analyses, we will now demonstrate in this section that all three of these components were present.

Instability began to build into the region throughout the day of 4 October 2005. By 0000 UTC 5 October 2005, it was apparent that the stability over western North Dakota had weakened. Cross sections were made through the region of the precipitation band over western and central North Dakota. The cross sections were made perpendicular to the thermal wind at or near the TROWAL axis, a requirement that must be satisfied when examining symmetric stability, using θ_{es} and MPV_g^*. Figure 11 is a cross-section of θ_{es} and MPV_g^* made through western North Dakota at 0000 UTC 5 October 2005. The cross-section was made when the precipitation band was at its most intense point. In examining Figure 11, it is evident that MPV_g^* was negative over western North Dakota from approximately 700 mb to 400 mb. Farther southeast along the cross-section in eastern North Dakota and central Minnesota, the θ_{es} surfaces actually fold over, such that $\partial\theta_{es}/\partial z < 0$, indicating that the atmosphere was unstable with respect to vertical displacements. This area coincided with a region where thunderstorms (gravitational convection) were ongoing. The θ_{es} surfaces over western North Dakota were not folded over, but were nearly vertical in orientation, indicating the possible presence of CSI. Moving from southeast to northwest, the precipitation regime changes from a convective mode to a stratiform mode.

Progressing to 0300 UTC 5 October (Fig. 12), the area of possible CSI had lessened over western and central North Dakota from approximately 700 mb to 400 mb. The RUC analysis indicated a small region of negative values over central areas of the state. However, it is important to note that the majority of the region exhibited values very close to zero and only slightly positive. With model limitations, it is possible that the atmosphere still could have been exhibiting CSI or WSS since these values were so small. As one progresses east into eastern North Dakota, the nearly vertical isentropes transition into folded isentropes. In examining the radar, deep convective type precipitation is visible over eastern North Dakota into South Dakota and Minnesota (Fig. 13).

Advancing to 0600 UTC October 5 (Fig. 14), the RUC analysis indicated a CI regime over western Minnesota and eastern North Dakota, transitioning very quickly into possible CSI then WSS in central North Dakota, as one moves west. This correlates well with the radar at 0600 UTC (Fig. 7), which indicates this same regime (deep convective transitioning into stratiform). However, at this time a strong vorticity maximum had moved over the region. Areas that are prone to possessing symmetric instability occur where the absolute geostrophic vorticity of the environment is small (as mentioned earlier), but we still feel that CSI might still have been present.

The heaviest precipitation had moved out of North Dakota by 1200 UTC on 5 October 2005. Light to moderate precipitation continued over western and central North Dakota and additional banded precipitation developed from northwest North Dakota into southwest Manitoba (Fig. 15). This entire area of precipitation is associated with the cyclonically turning portion of the WCB (Schultz 2001). A cross section taken at this time indicates a stable environment with respect to slantwise convection in western North Dakota with a possible WSS regime in central areas (Fig. 16). Eastern North Dakota still exhibited possible CSI with negative values of MPV_g^* present between 700 and 400 mb, quickly transitioning into a CI environment (folded isentropes). It is important to note that the majority of possible CSI, as indicated by the RUC analysis, had pushed northeast into northwestern Minnesota and southern Manitoba (not shown), coinciding with the primary TROWAL axis. In conclusion, the atmosphere was exhibiting various forms of instability during the time that a heavy precipitation band was observed over western North Dakota. The precipitation area transitioned into mostly a stratiform event with embedded banding by 1200 UTC 5 October, as the storm system lifted northeast into

Canada. It was from this time on, that the intensity of the precipitation diminished to light snow and persisted for several more hours, as the mid level vorticity maximum was centered over the state.

Next, the presence of mid level forcing will be shown. The forcing was responsible for helping to develop the banded precipitation, releasing any instability that may have been present. A strong mid-level shortwave (mentioned earlier) had entered the western Dakotas by 0000 UTC 5 October, helping to ignite a band of showers and thunderstorms in eastern Wyoming and Colorado. This shortwave was responsible for advecting cold dry air down and over the western Dakotas. This cold dry air was advected over warm moist air (WCB) streaming northwestward as a result of the developing surface low in eastern South Dakota, thus decreasing the static stability of the environment over the Dakotas and eastern Wyoming. The WCB and the development of the TROWAL are apparent in Figures 17a-f. As time progressed, warm moist air was transported north into the Northern Plains and by 0000 UTC the TROWAL axis is evident along with a large area of deformation. The TROWAL lifted northeast with the system and by 1200 UTC 5 October it was located in extreme eastern North Dakota (Fig. 17e). The deformation zone expanded in area and intensified while moving northeast during the period 0000-1500 UTC. At 0000 UTC, the deformation zone was between 700 and 600 mb and was visible in the Bismarck sounding (Fig. 18). This deformation zone developed where warm moist air advected north and interacted with cold dry air present, leading to frontogenesis and forced vertical ascent.

An interesting feature can be observed at 1200 UTC. A secondary TROWAL had developed over southern North Dakota, moving north across the state (Figs. 17e-f).

An area of weak deformation can be observed to the north of the secondary TROWAL. This weak deformation zone may have enhanced and maintained the precipitation over central portions of the state through the morning hours.

The deformation zone that developed as a result of the WCB streaming north and west led to strong frontogenesis aloft. Figures 19a-d depict the 850 to 500-mb layer frontogenesis that was occurring over western North Dakota from 0000 UTC through 1200 UTC on 5 October. The frontogenetical area in western North Dakota indicated in Figure 19 would have been responsible for releasing any CSI present in the early period of the event. As noted in Moore et al. (2005), the maximum upward vertical motion is always located on the warm side of the frontogenetical maximum. This would place the maximum vertical velocity in western and central North Dakota. Cross-sections of frontogenesis and vertical motion (Figs. 20a-d) reaffirm the fact that an associated upward directed vertical motion maximum was located over western and central North Dakota. This upward directed vertical motion maximum occurred on the warm side (thermally direct circulation) of the frontogenetical maximum, and is consistent with frontogenetic circulations. Much weaker upward vertical motion was observed over eastern North Dakota and western Minnesota. In examining Figs 20a-d., it is clear that the frontogenetical maximum moved northeast into southern Manitoba and Saskatchewan as it strengthened and spread out in area. This is important because as noted earlier, the atmosphere over western and central North Dakota transitioned from one that contained possible CSI to one in which was stable with respect to convection (according to model analyses). Since the precipitation band maintained its strength while expanding in area, it is possible that the frontogenetical forcing, via the secondary TROWAL, may have been strong enough to directly force banded precipitation--precipitation that may have been enhanced by the release of instability in eastern portions of North Dakota. Other important forcing, via the 500-mb low and associated vorticity maximum, most likely aided in the maintenance of the precipitation area over western and central North Dakota in time periods after 0900 UTC on 5 October.

Another important feature was the location of the upper level jet. The entrance region of the upper level jet streak located over southern Canada was positioned over North Dakota at 0000 UTC 5 October (Fig. 4c). It is important because this is a favored region for the release of CSI. The direct thermal circulation that exists below the jet streak in the entrance region allows for the development of a forced ascent on the warm side of the jet streak. It is also a region where the atmosphere exhibits weak positive to slightly negative vorticity, all attributes found in favored regions of CSI (Bluestein 1993).

6. Conclusion

Using the methodology of previous authors, a case study was performed to assess the atmospheric state over the Northern Plains on 4-5 October 2005. The object of the research was to investigate the processes involved in the production of heavy snow over western and central North Dakota. Specifically, the authors wanted to examine if CSI played a role in the development of the heavy precipitation.

It was determined that the development of surface low pressure over northwestern Iowa (Fig. 4a) allowed for the advection of warm moist air northward over the Dakotas. This process is visualized in the development of the TROWAL axes (Figs. 17, 19). The aforementioned process, coupled with a vigorous shortwave approaching from the Rocky

Mountains, led to the development of an unstable air-mass over much of the Northern Plains (as determined from previous analysis). This instability gradually transitioned from one of a vertically convective nature (eastern Dakotas and Minnesota), to one of a slantwise convective nature (western North Dakota). The instability over North Dakota was at its greatest intensity at approximately 0000 UTC, and diminished through 1200 UTC.

Various forcing mechanisms were responsible for the production of precipitation. Warm moist air streaming northward as a result of the surface low led to an area of strong frontogenesis/deformation over North Dakota, hence producing a large area of significant upward vertical motion early on in the event. As the system progressed, the forcing mechanisms changed. The majority of the frontogenesis and instability had moved out of the region by 1200 UTC. However, by this time, a closed 500-mb low had moved into North Dakota. The forcing produced from this low (through vorticity advection coupled with deformation from the cyclonically curved portion of the WCB) allowed for the persistence of light to moderate snow over much of North Dakota. A combination of these factors above led to nearly twenty four hours of continuous light to moderate snowfall over western and central North Dakota. Areas from southwest North Dakota into north central North Dakota received the greatest amounts (Fig. 1).

One interesting finding from this research was the discovery of a secondary TROWAL axis. This type of feature may be more common than one would suspect. With better model resolution these types of small scale features may be easier to uncover now than in the past. It appears that the secondary TROWAL was important in the maintenance of precipitation by the added small scale forcing it produced. A small area of frontogenesis was observed over western North Dakota (Fig. 19d) ahead of the secondary TROWAL. It is possible that this secondary forcing could have helped to produce the greater snow amounts observed from the southwest to north central region of North Dakota.

Once again, the atmosphere can be very complicated as is evident in this case. Many features typically act together in concert to produce an end result, heavy snow in this case. Earlier on in the developmental stages of the research project, the authors believed the release of CSI was the main contributor to the production of large snowfall totals. After examining various features, it was determined that the persistence of various forms of forcing (frontogenesis aided by instability early on, weak deformation, and positive vorticity advection) over North Dakota and slow system speed, were the major factors that resulted in the heavy snowfall totals observed in Figure 1. However, the possible release of CSI may have been one of several factors that contributed to this record breaking snowfall event.

7. References

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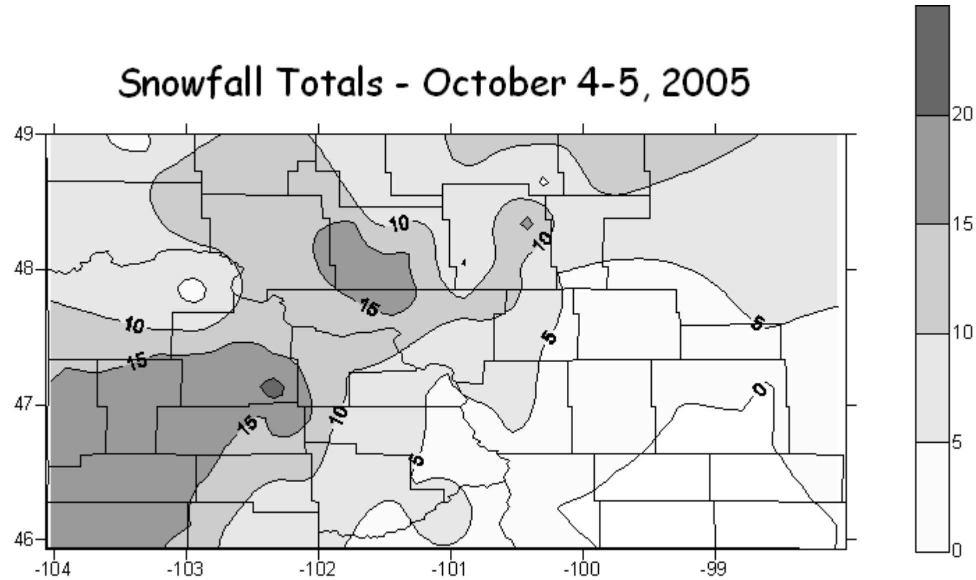

Figure 1. Total snowfall (inches) over western and central North Dakota on 4-5 October 2005.

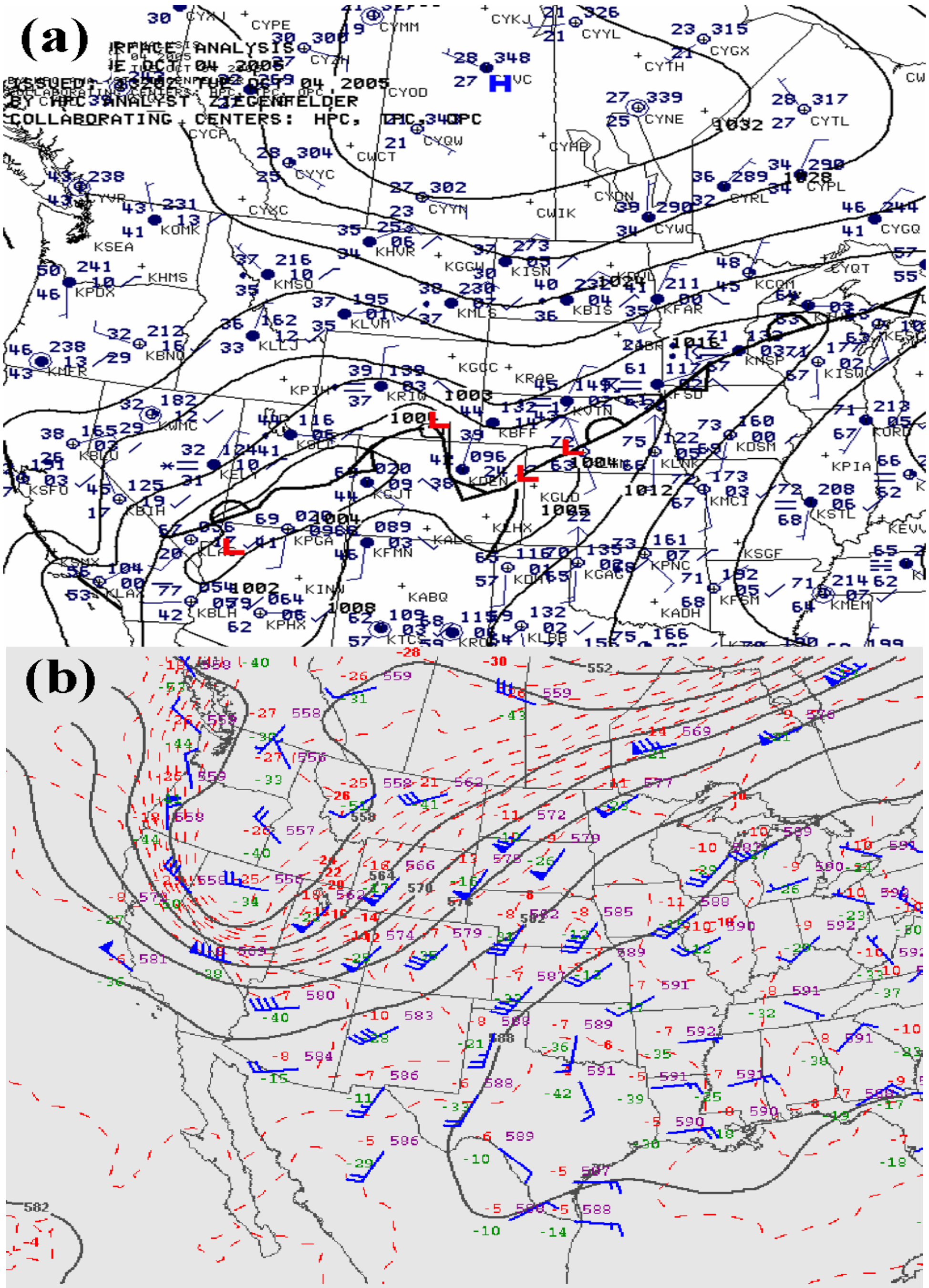

Figure 2. (a) 00-h Surface analysis (*from Hydrometeorological Prediction Center*) and observed station plot for 1200 UTC 4 October 2005. Solid lines are isobars in mb. (b) 00-h 500-mb objective analysis (*from Storm Prediction Center*) for 1200 UTC 4 October 2005. Solid lines are heights in geopotential decameters and dashed lines are temperature in degrees Celsius.

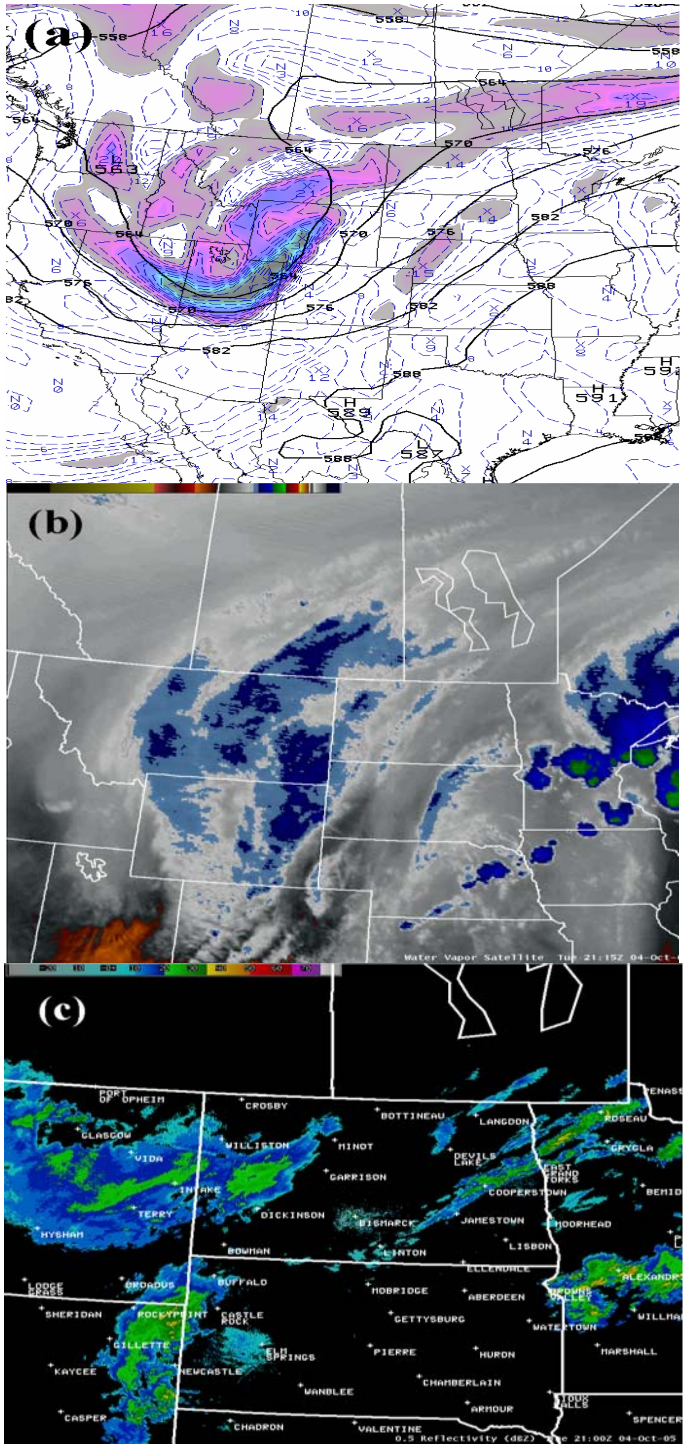

Figure 3. (a) 500-mb Rapid Update Cycle (RUC) 00-h analysis valid 2100 UTC 4 October 2005. Height (solid) contours are in geopotential decameters at 60-m intervals. Absolute vorticity in 10^{-5} s^{-1} (dashed) contours and shading above 12 (X represents maximum; N, minimum). (b) GOES water vapor image at 2115 UTC 4 October 2005. (c) WSR-88D base reflectivity (0.5^{o} elevation) mosaic at 2100 UTC 4 October 2005.

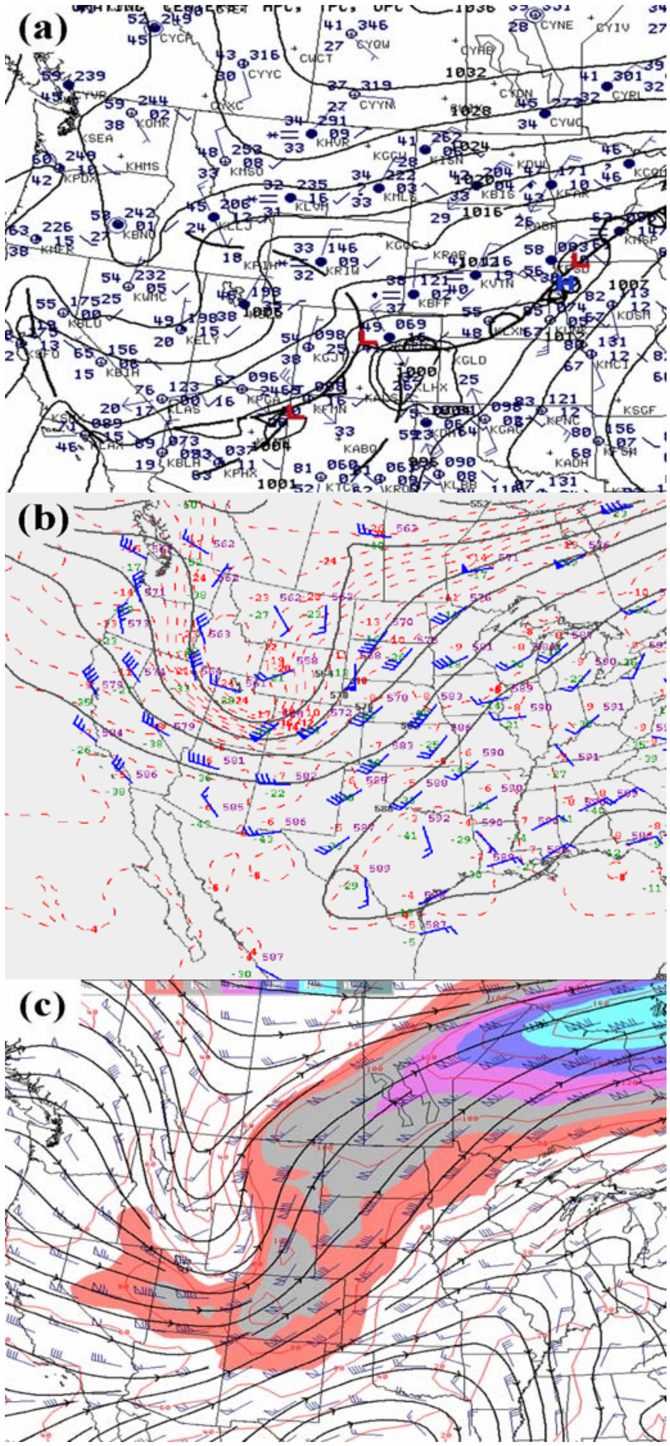

Figure 4. (a) Same as Figure 2a, except valid 0000 UTC 5 October 2005. (b) Same as Figure 2b, except valid 0000 UTC 5 October 2005. (c) 250-mb RUC 03-h analysis valid 0000 UTC 5 October 2005. Thick dark lines are streamlines. Contours (lines and shading) are winds in knots at a 20 knot interval. Wind speeds in shaded area begin at 70 knots.

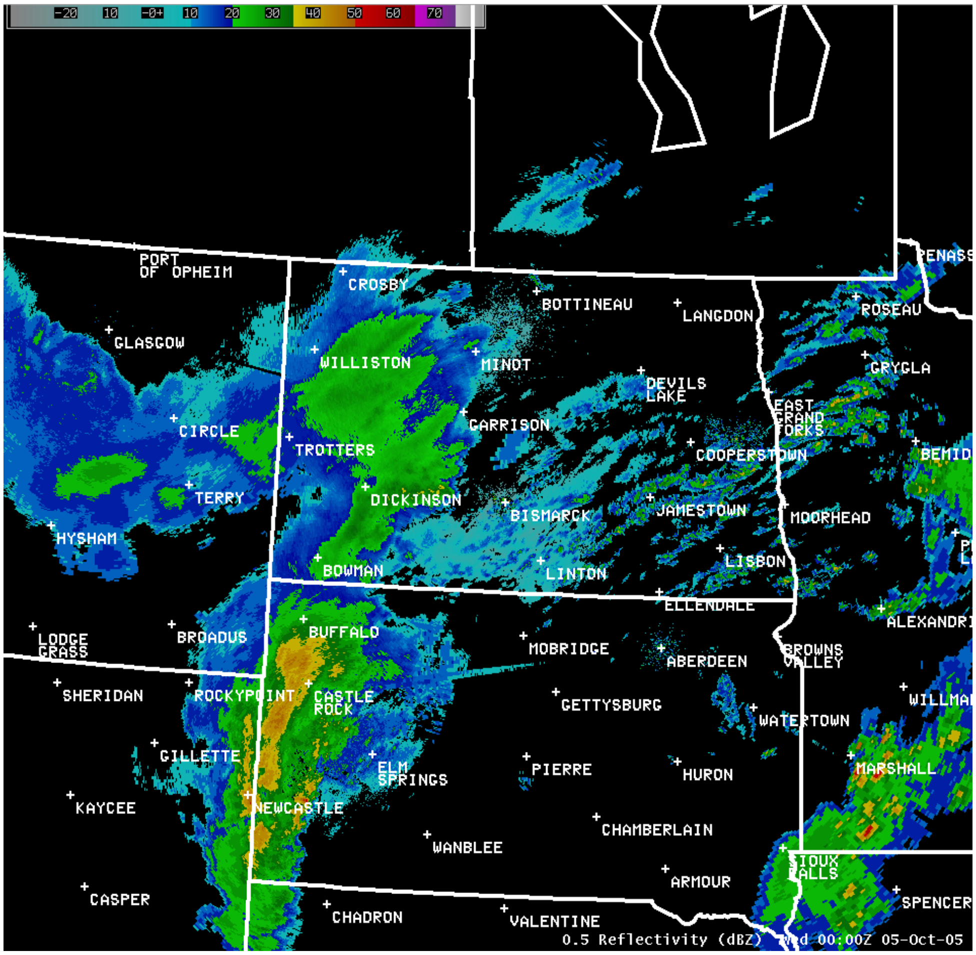

Figure 5. WSR-88D base reflectivity (0.5^{o} elevation) mosaic valid at 0000 UTC 5 October 2005.

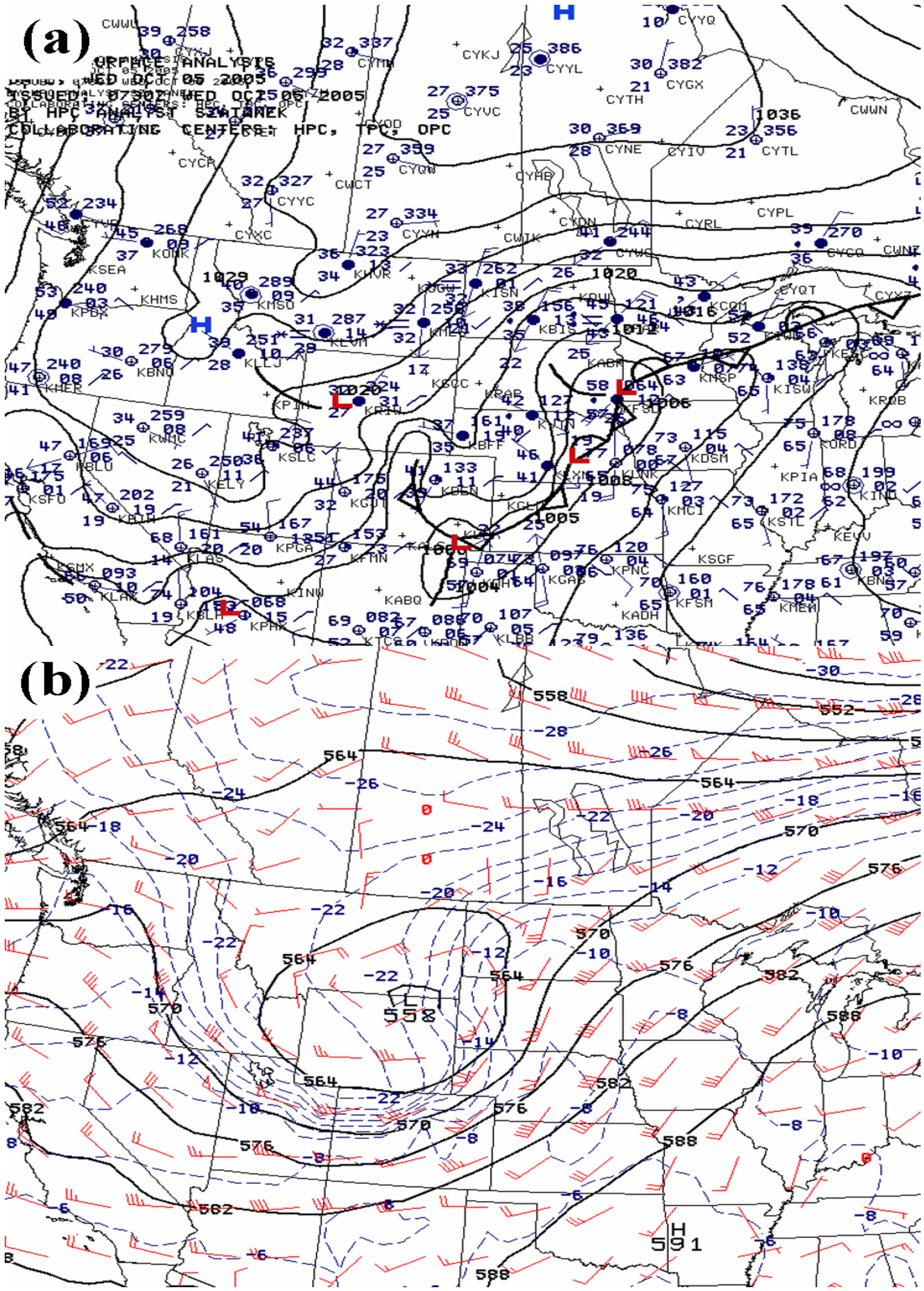

Figure 6. (a) Same as Figure 2a, except valid 0600 UTC 5 October 2005. (b) 500-mb RUC 00-h analysis valid 0600 UTC 5 October 2005. Height (solid) contours in geopotential decameters at 60-m intervals, temperatures (dashed) contours in degrees Celsius in 2 degree intervals and winds in knots.

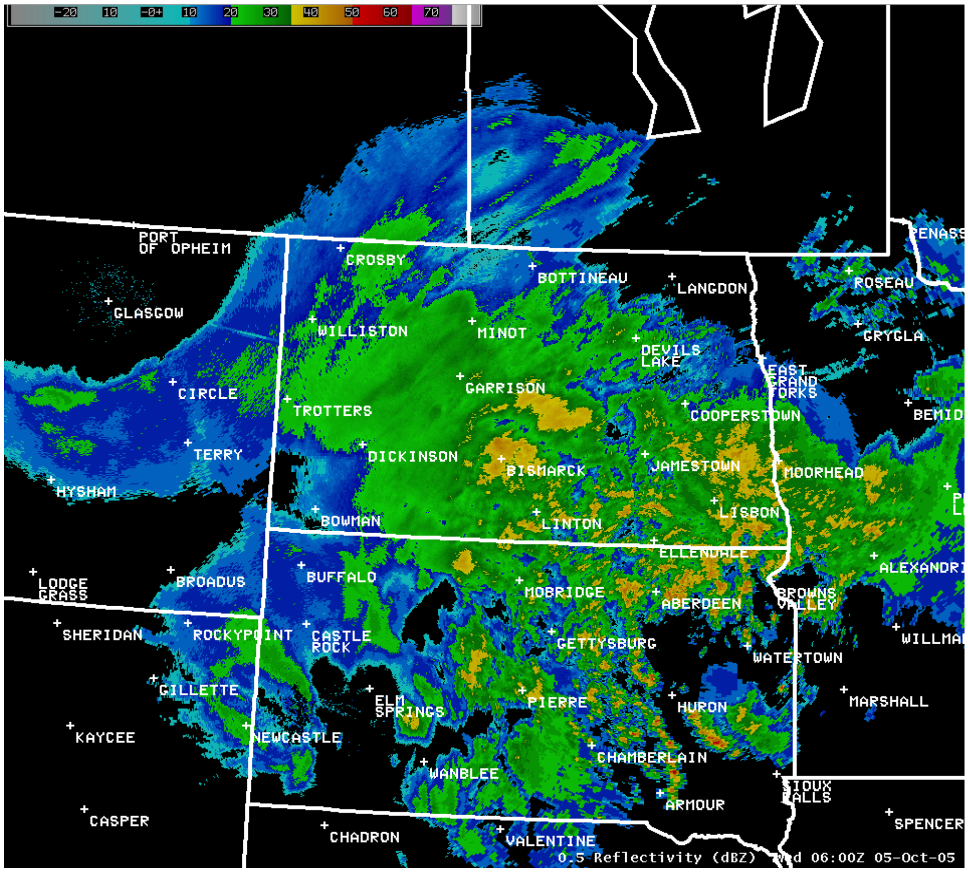

Figure 7. WSR-88D base reflectivity (0.5° elevation) mosaic valid at 0600 UTC 5 October 2005.

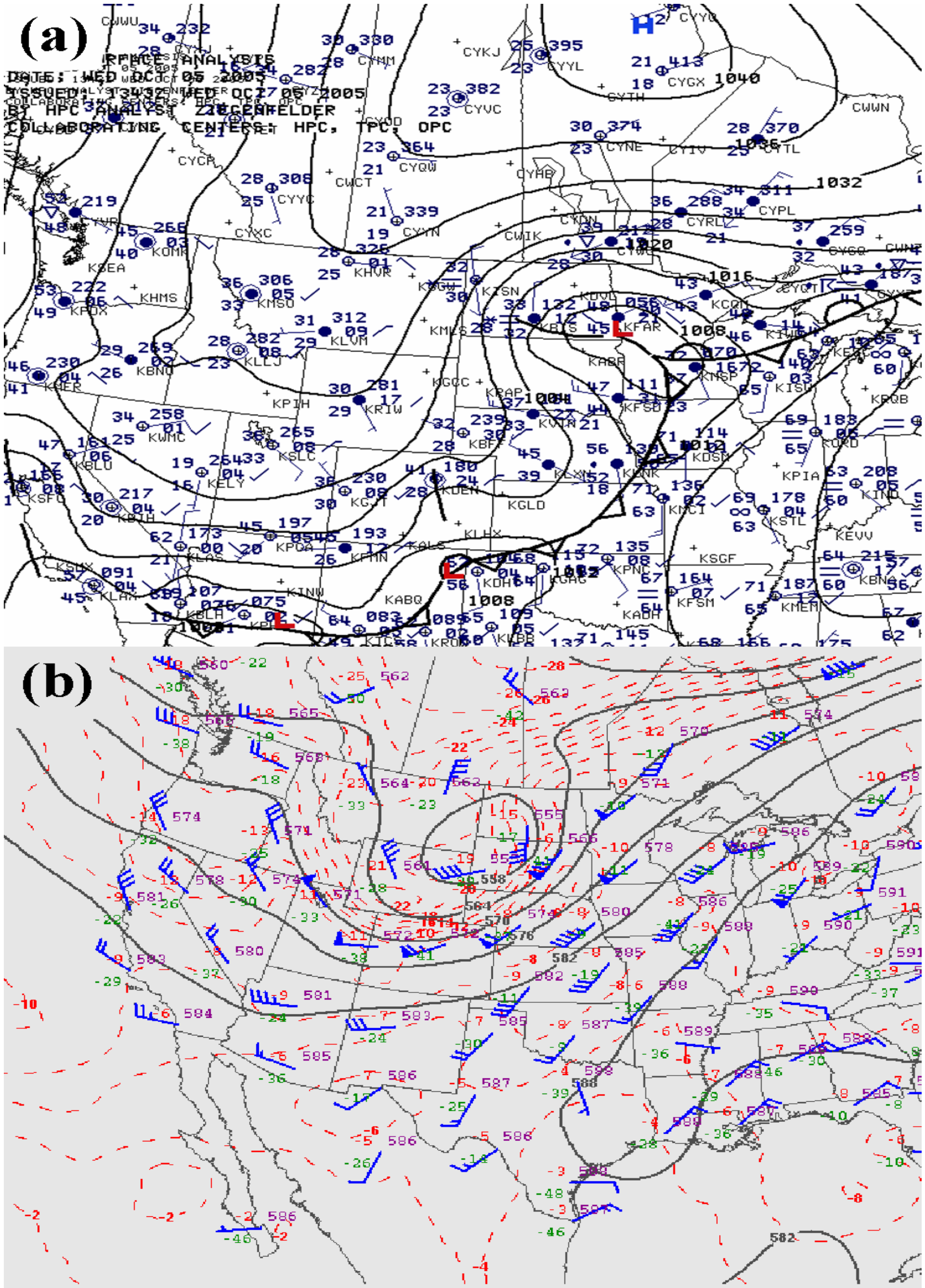

Figure 8. (a) Same as Figure 2a, except valid 1200 UTC 5 October 2005. (b) Same as Figure 2b, except valid 1200 UTC 5 October 2005.

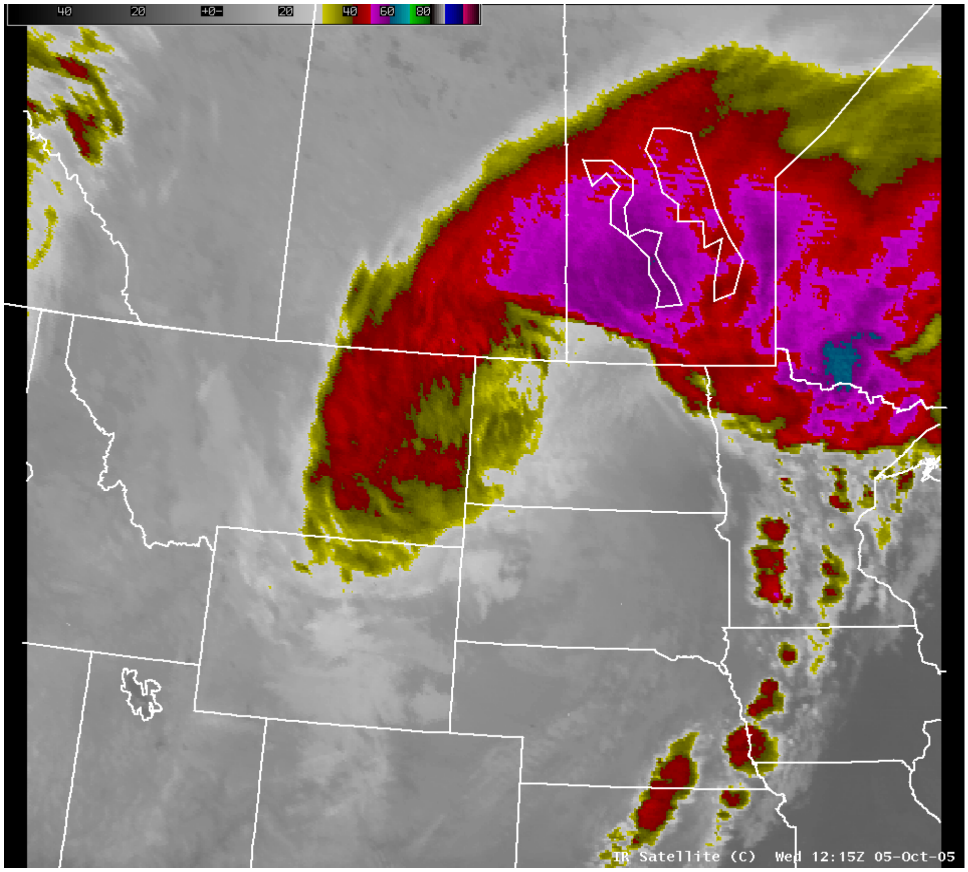

Figure 9. GOES Infrared Image valid at 1215 UTC 5 October 2005.

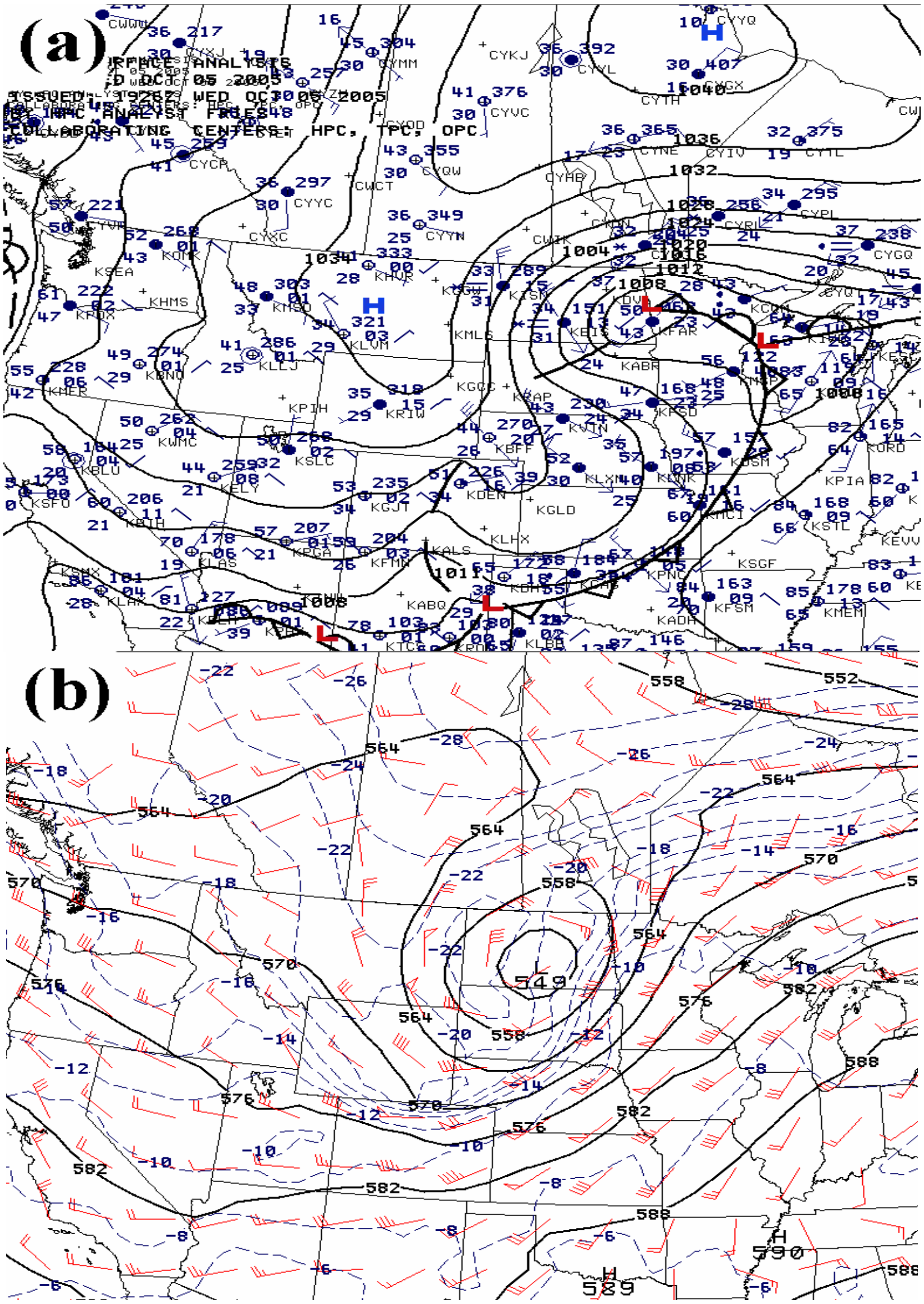

Figure 10. (a) Same as Figure 2a, except valid 1800 UTC 5 October 2005. (b) Same as figure 6b, except valid 1800 UTC 5 October 2005.

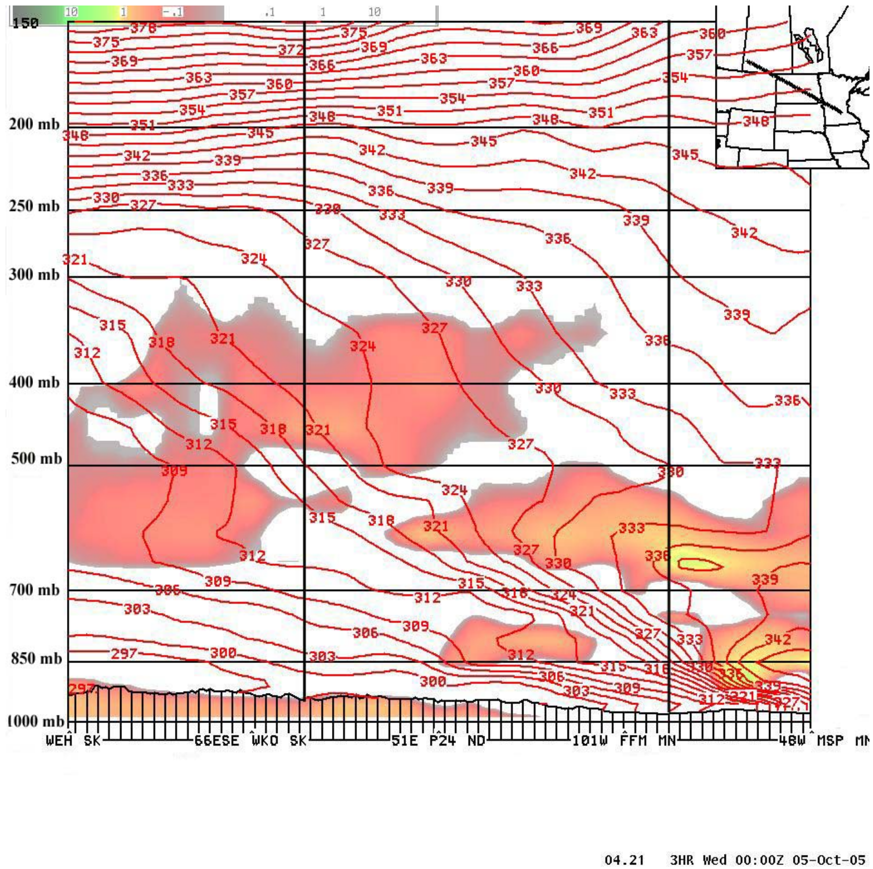

Figure 11. Vertical cross section of θ_{es} (contours) and MPV_g^* (shaded) from the RUC (Rapid Update Cycle) 3 hour forecast valid 00 UTC 5 October 2005. The cross section is along a line from southwestern Saskatchewan to near the Minneapolis, MN (see above insert). Salmon shaded areas indicate negative regions of MPV_g^*. Space between the two black vertical lines is the area over North Dakota.

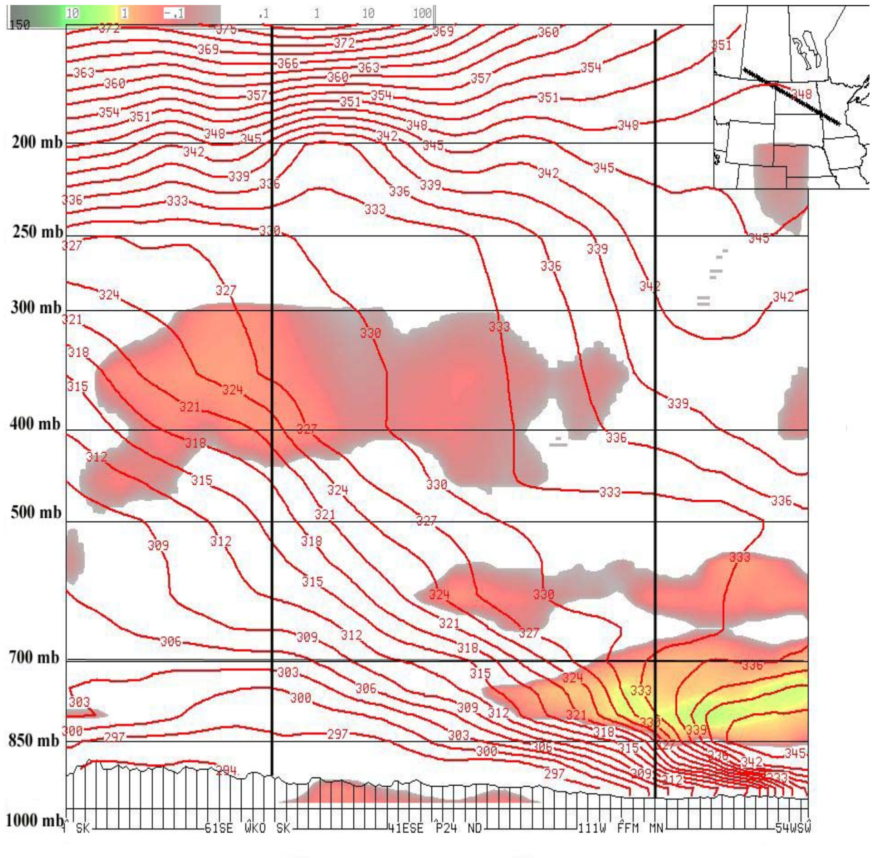

Figure 12. Same as Figure 11, except valid 03 UTC 5 October 2005.

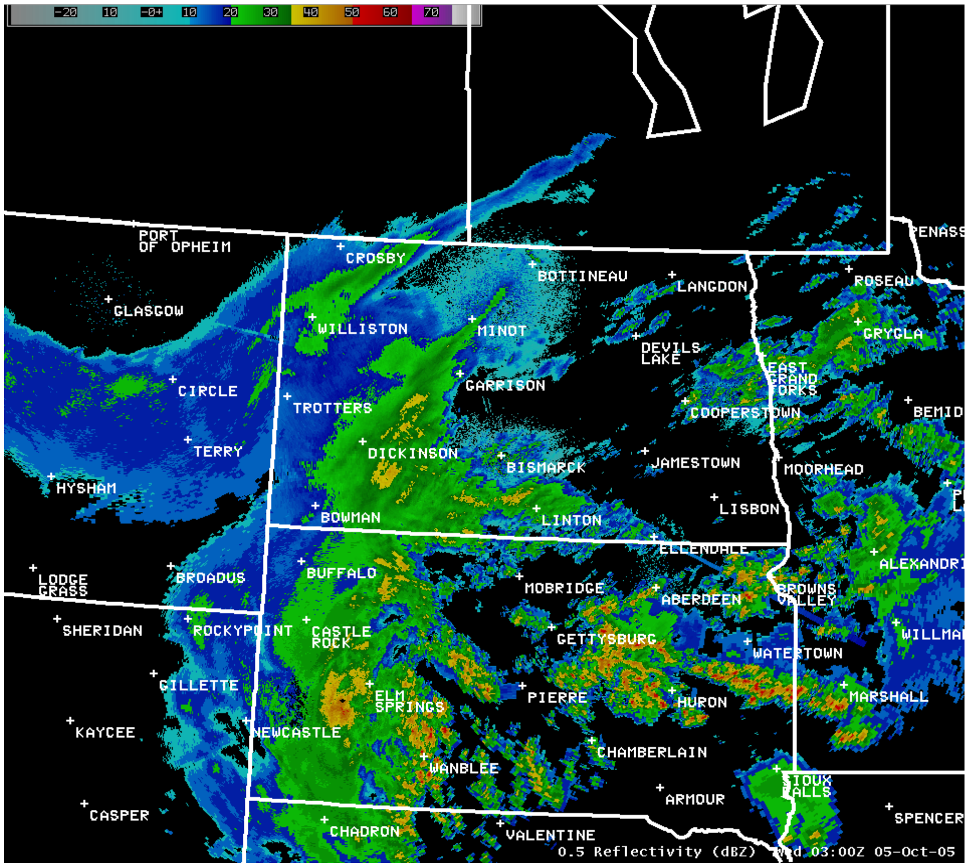

Figure 13. WSR-88D base reflectivity (0.5° elevation) mosaic valid at 0300 UTC 5 October 2005.

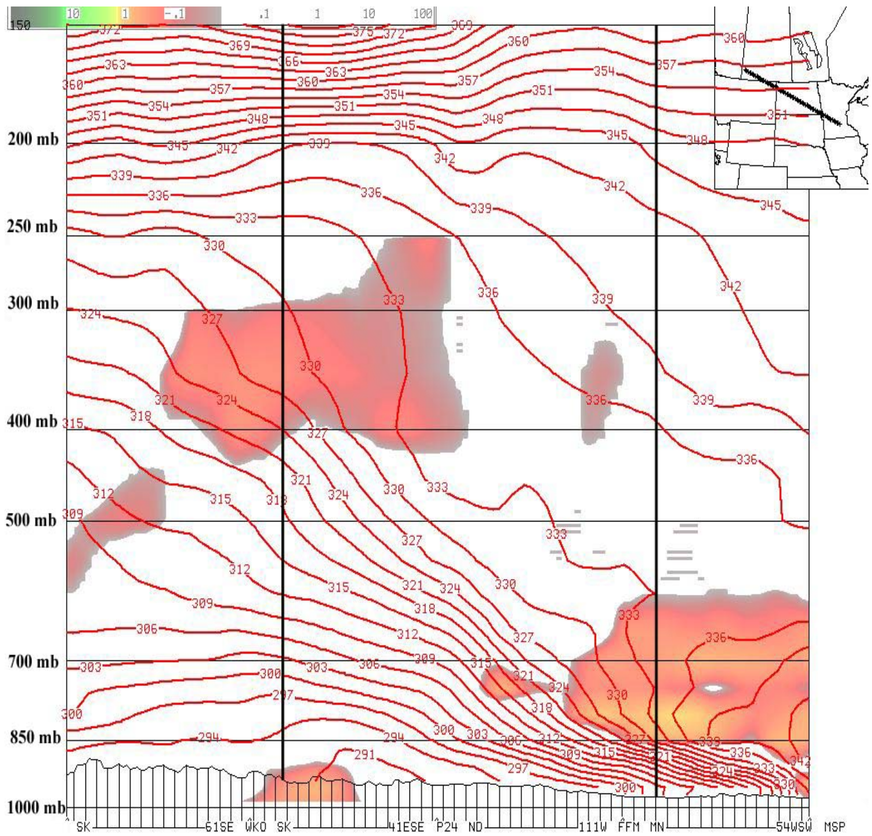

Figure 14. Same as Figure 11, except valid 06 UTC 5 October 2005.

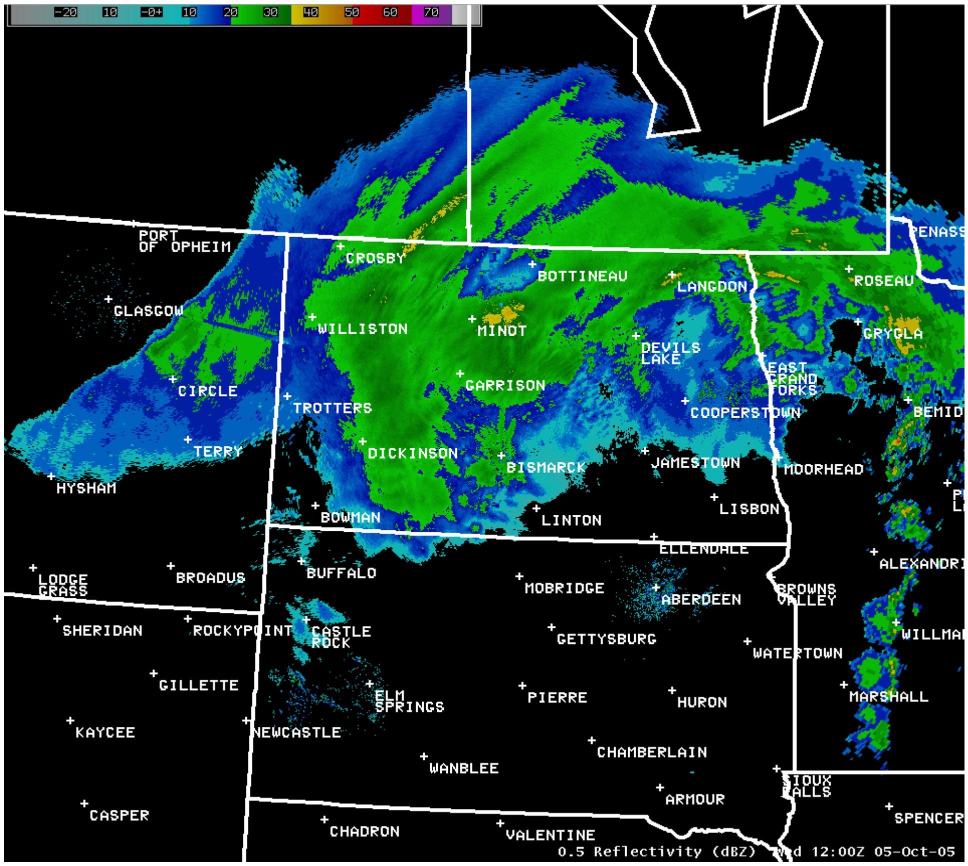

Figure 15. WSR-88D base reflectivity (0.5^{o} elevation) mosaic valid at 1200 UTC 5 October 2005.

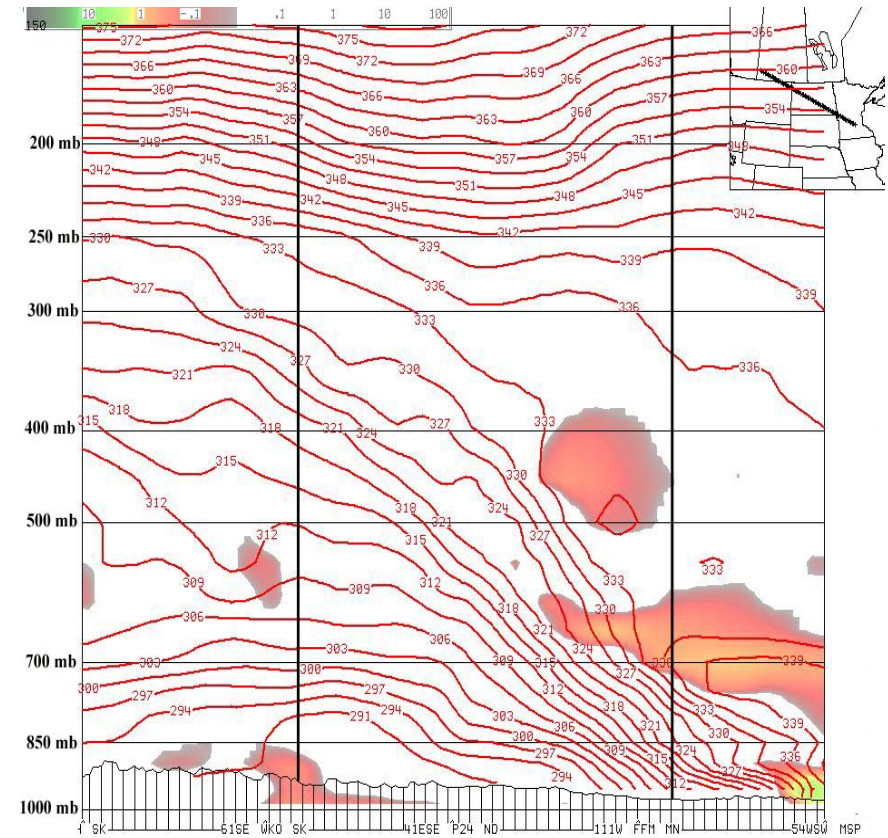

Figure 16. Same as Figure 11, except valid 1200 UTC 5 October 2005.

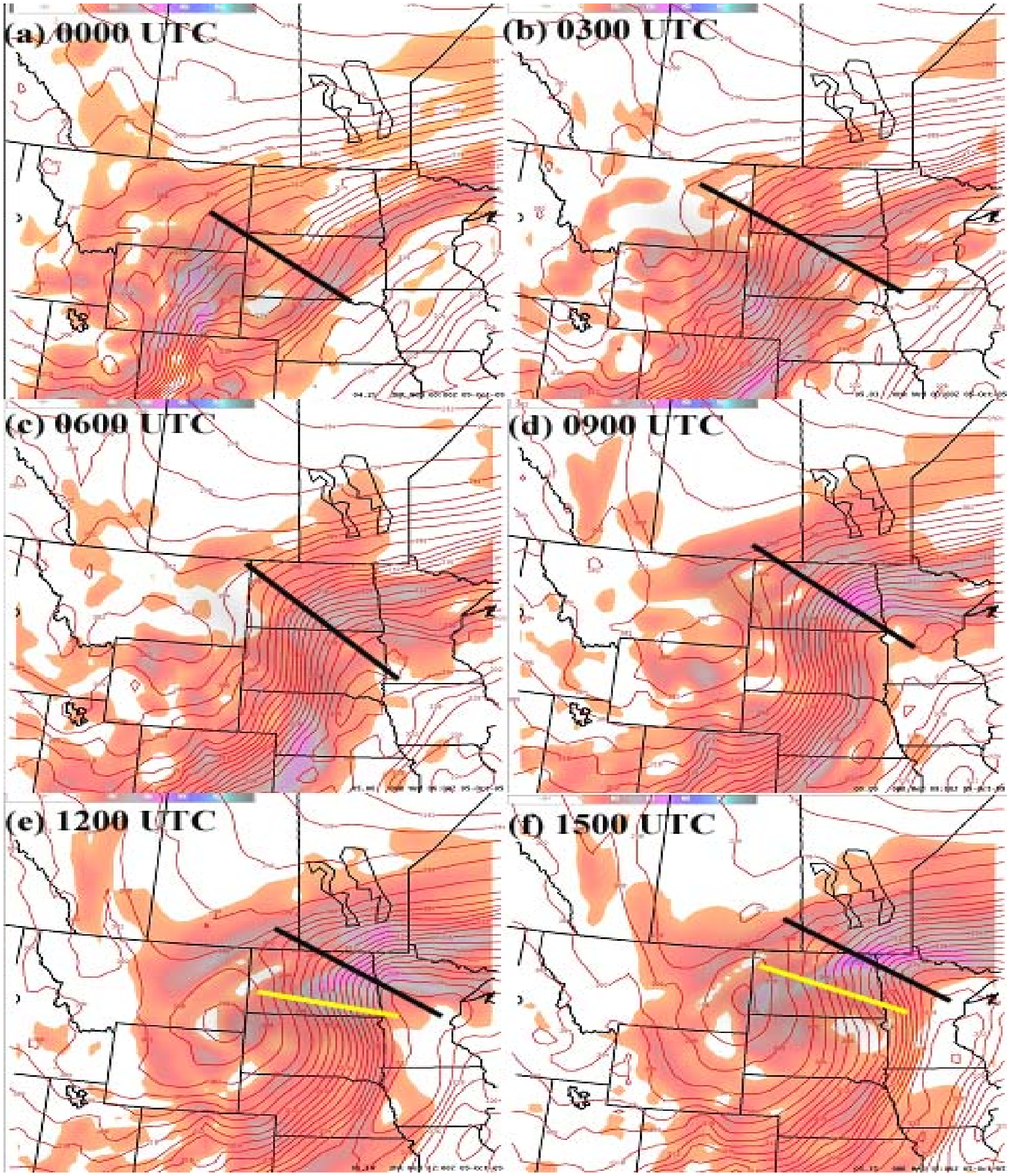

Figure 17. RUC analysis of 800-500 mb average layer θ_{es} (red lines) and deformation (shading) for (a) 0000, (b) 0300, (c) 0600, (d) 0900, (e) 1200, (f) 1500 UTC 5 October 2005.

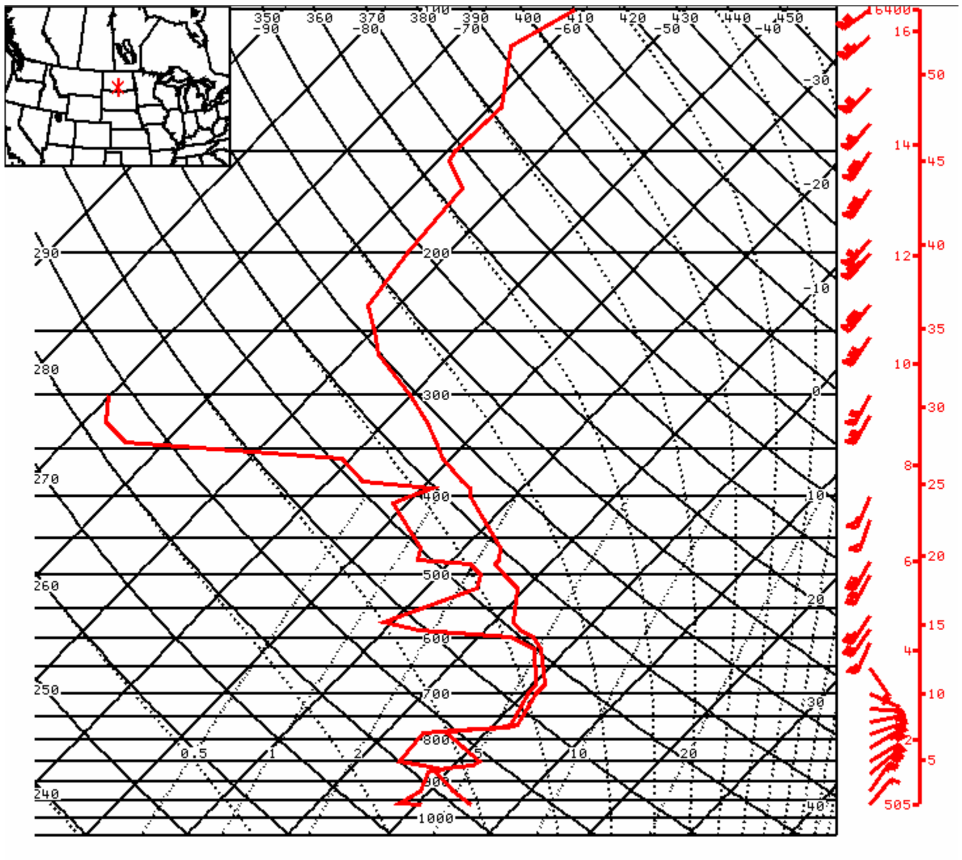

Figure 18. Observed skew *T*-log *p* sounding for Bismarck North Dakota at 0000 UTC 5 October 2005. Solid lines indicate temperature and dewpoint (^{o}C) while wind barbs are indicated on the right-hand side in knots.

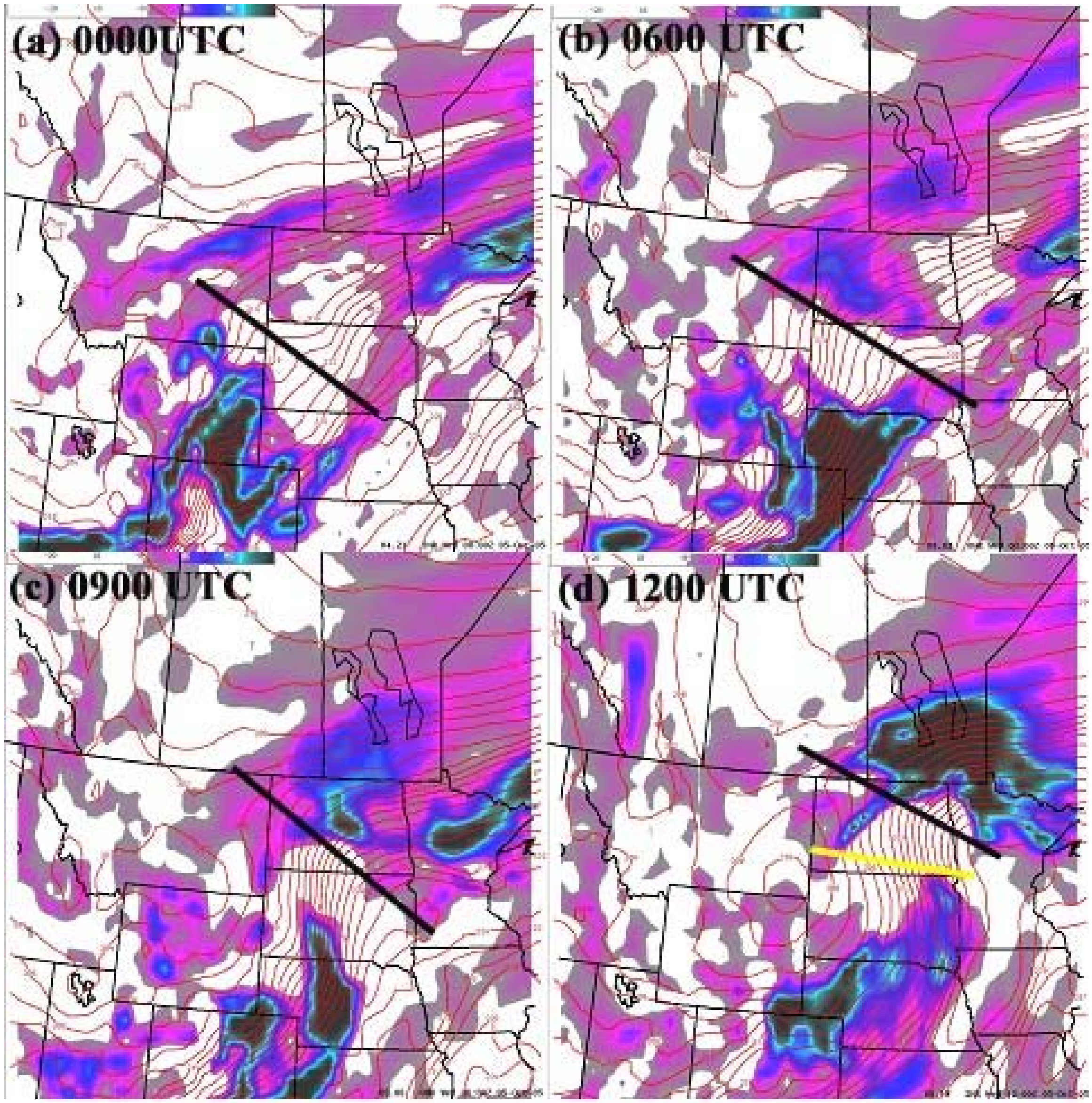

Figure 19. RUC analysis of 800-500 mb layer average frontogenesis (shaded), contours of θ_{es} (red), TROWAL axis (black line), and a secondary TROWAL axis (yellow line) for (a) 0000, (b) 0600, (c) 0900, (d) 1200 UTC 5 October 2005.

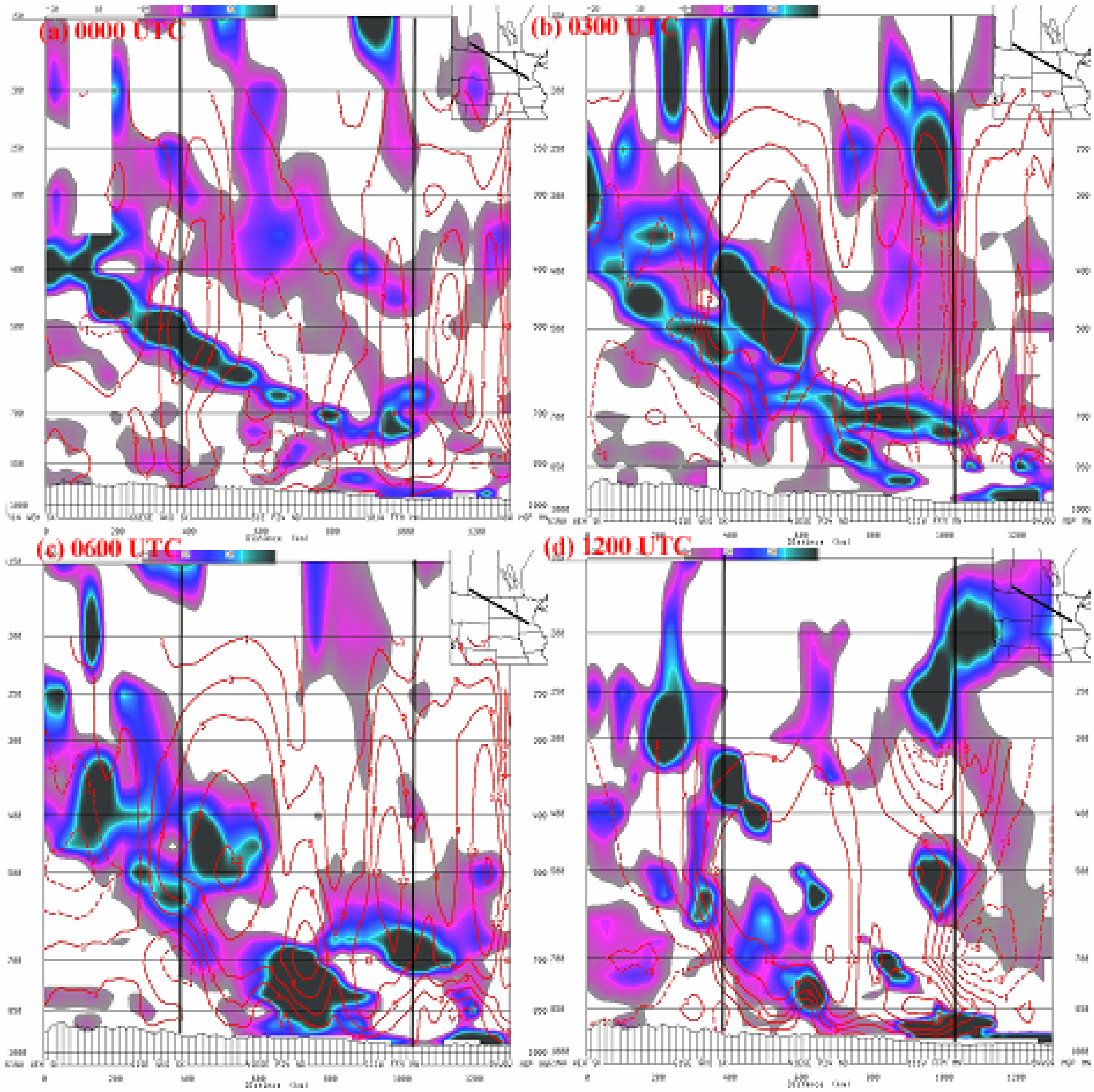

Figure 20. RUC vertical cross section of frontogenesis (shaded) and vertical motion (red contours) (a) 0000, (b), 0300, (c) 0600, (d) 1200 UTC 5 October 2005. Cross section area location is visible in upper right corner of each image.