The Glass Ceiling: Analysis of 19 July 2006 Western Iowa Null Severe Weather Event

Daniel R. Chavas, *University of Wisconsin-Madison, Madison, WI*
AOS 453 Final Case Study
13 May 2007

ABSTRACT

On July 19th, 2006 strong near-surface warm moisture convergence over western Iowa to the southwest of an intensifying warm front boosted temperatures above 90F with dewpoints as high as 85F in some locations. Meanwhile, a very dry, generally well-mixed air mass was in place aloft. The combination of these two ingredients resulted in extreme levels of instability that, coupled with a nearly idealized wind profile and strong frontogenetical forcing, produced conditions ripe for a severe weather outbreak. However, strong warm air advection associated with the southwesterly low-level jet just above the moist boundary layer acted to intensify and deepen the capping inversion, inhibiting convective initiation. Analysis of the effects of evapotranspiration from crops is also performed.

1. Introduction

Too much instability can sometimes be a self-limiting condition for convective initiation. Such was the case during the 19 July 2006 null severe weather event over northwestern Iowa. A dominant upper ridge permitted the development of a deep, hot, dry well-mixed layer at the surface over the Great Plains, while near-surface winds advected moisture northwestward from the Gulf of Mexico as a weak surface cyclone developed over the north-central Great Plains. As warm frontogenesis intensified, staggering indices of convective instability and supercell potential were observed by late afternoon as temperatures soared into the mid-90s F and dewpoints rose into the mid-80s F.

However, as observed directly by the author and his companion, not a single significant cumulus cloud was observed throughout the afternoon and evening, as a shallow boundary layer "dome" of what was effectively fog (visibilities were reduced to 1-2 miles) due to the extreme moisture content trapped beneath a very strong inversion created oppressively humid near-surface conditions.

The author seeks to assess the precise mechanisms that trapped this unstable air mass so close to the surface and how, despite the presence of anomalously strong forcing for vertical motion associated with intense warm frontogenesis combined with the arrival of a robust mid-level trough, not even a single convective cloud was able to form. Furthermore, a simple formulation is employed to assess the effect of evapotranspiration on boundary layer equivalent potential temperature values and the associated potential for convection.

2. Data

Analyses and 6-hour forecasts are taken from 6-hourly Eta model runs and 3-hourly RUC model runs from 0000 UTC 19 July to 0000 UTC 20 July 2006. Radar is taken from the archived NEXRAD local radar reflectivity data. Satellite images are taken from the GOES-8 satellite image archive.

3. Synoptic Overview

The synoptic set-up for the 19 July 2006 null severe weather event includes many of the key ingredients that typically contribute to a truly explosive outbreak. At 1200 UTC (6a local time) on the 19th, a strong westerly jet stretches across the United States-Canada border, as depicted in the Eta model analysis 250 hPa plot (fig. 1), including a 90-knot maximum wind speed located along the northwestern border of North Dakota, northern Minnesota, and reaching the western edge of Lake Superior. In the vicinity of the jet exit region, the jet streak curves clockwise across central and southeastern Wisconsin. Curiously, this places the null event location interest just to the south of a region strong upper-level ageostrophic convergence and associated subsidence in the right exit region of the jet, which would be expected to suppress convection. However, the shear associated with the jet maximum would also be expected to enhance the initiation and maintenance of supercellular activity.

Figure 2 depicts the 700 hPa large-scale flow, which includes a dominant

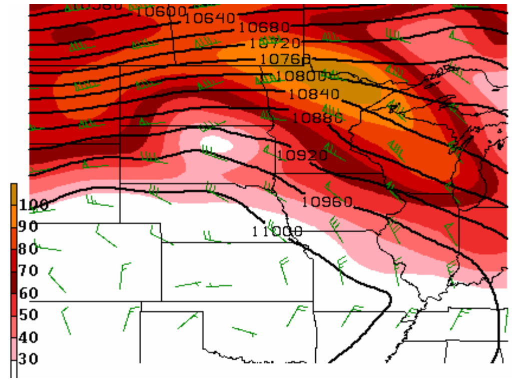


Fig 1. 250 hPa heights, wind speeds, wind barbs from 1200 UTC 19 July 2006 Eta model analysis.

anticyclone over southeastern Kansas. The circulation associated with this feature provides moderate northwesterly winds, with speeds at approximately 20 knots. Thus, there exists moderate speed and directional shear between middle and upper levels, but, more importantly, the flow throughout this layer is westerly and of significant magnitude to efficiently tilt an updraft and advect convective outflow away from the updraft core, creating the classic cumulonimbus anvil that is characteristic to a powerful supercell. A strong shortwave is visible in the 3160m height contour over the west-central Dakotas, and a second high-amplitude shortwave trough is visible in the 3200m height contour over eastern Colorado. These two shortwave features combine to create one of the two strong dynamical forcings for vertical motions during this null case. Finally, two moist pockets are located to the northwest and to the south of Iowa, with a channel of very dry air in between. Given the flow at 700 hPa based upon the height contours, one would expect the moist pocket to the northwest to be advected just to the north of Iowa, while the tongue of very dry air would be advected overtop of the region. This dry channel clearly indicates that the free troposphere will be dominated by

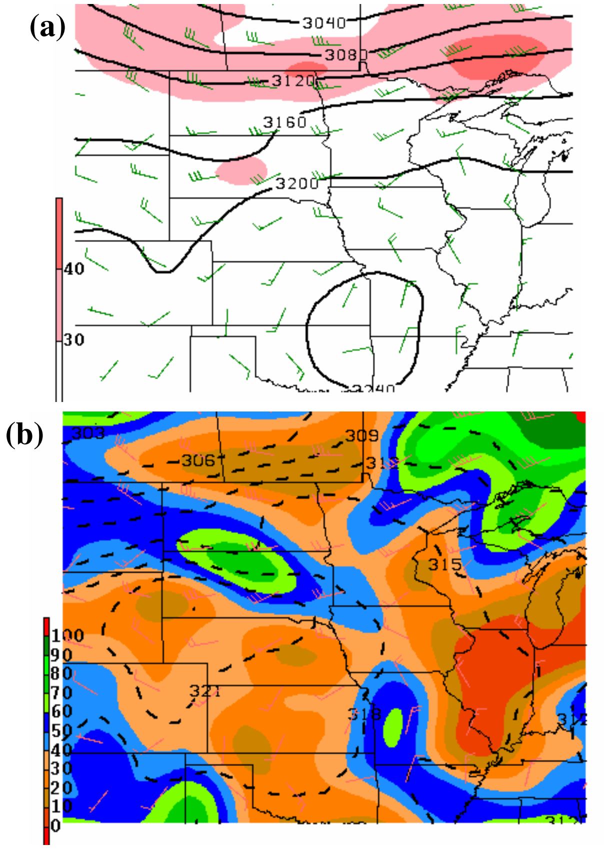


Fig 2. 700 hPa 1200 UTC 19 July Eta model run. (a) wind speeds (fill, kts), heights (m), wind barbs (kts); (b) relative humidity (fill, %) and potential temperature (K).

very low θ_e air, suggesting high convectively instability of moist low-level air.

The 850 hPa flow is depicted in figure 3, where the flow has shifted towards a more southwesterly pattern. Furthermore, a small 40-knot low-level jet maximum is visible across the eastern border between South Dakota and Nebraska and stretching into northwestern Iowa. The existence of this feature indicates the availability of strong speed shear at low-to-mid levels, which is vital to the development of strong convective updrafts and downdrafts, as well as the associated shear that enables a mesocyclone to form. The two short waves noted at 700 hPa are partially merged at 850 hPa, visible in the 2040m height contour, creating a two-pronged, high-vorticity trough oriented southwest-northeast across Nebraska and South Dakota. In addition, the initial stages of warm frontal development are clear in southern South Dakota, as the potential temperature (and relative humidity) gradient begins to be compacted by the strongly confluent flow across the upper Midwest. The existence of the low-level jet further enhances the frontogenetic process due to the region of convergence downstream of the jet speed maximum across northern Iowa. Finally, it is crucial to recognize the circular region with a potential temperature maximum (318K) collocated with a relative humidity minimum (10-20%) given the nature of the flow at this level. It is clear that one might expect this very warm, very dry pocket to be advected into northwestern Iowa by mid-late afternoon and into the evening, impinging on the existing moist pocket that persists over southwestern Minnesota. Dry air above the boundary layer implies a high probability of a very prominent mid-level θ_e minimum, which is conducive to highly conditionally unstable conditions. However, the warmth of this layer is certainly disconcerting to the storm chaser, as mid-level warm advection will reduce the buoyancy of a parcel lifted from below, potentially beyond a threshold that can be overcome by calculated atmospheric forcings for vertical motion. This warm bubble will be discussed in much greater detail in a later section.

Finally, figure 4 exhibits the 950 hPa boundary layer synoptic pattern. Stark contrasts between this level and the aforementioned level only 100 hPa above are evident. First, a weak but developing surface cyclone is located in western South Dakota and Nebraska, which has

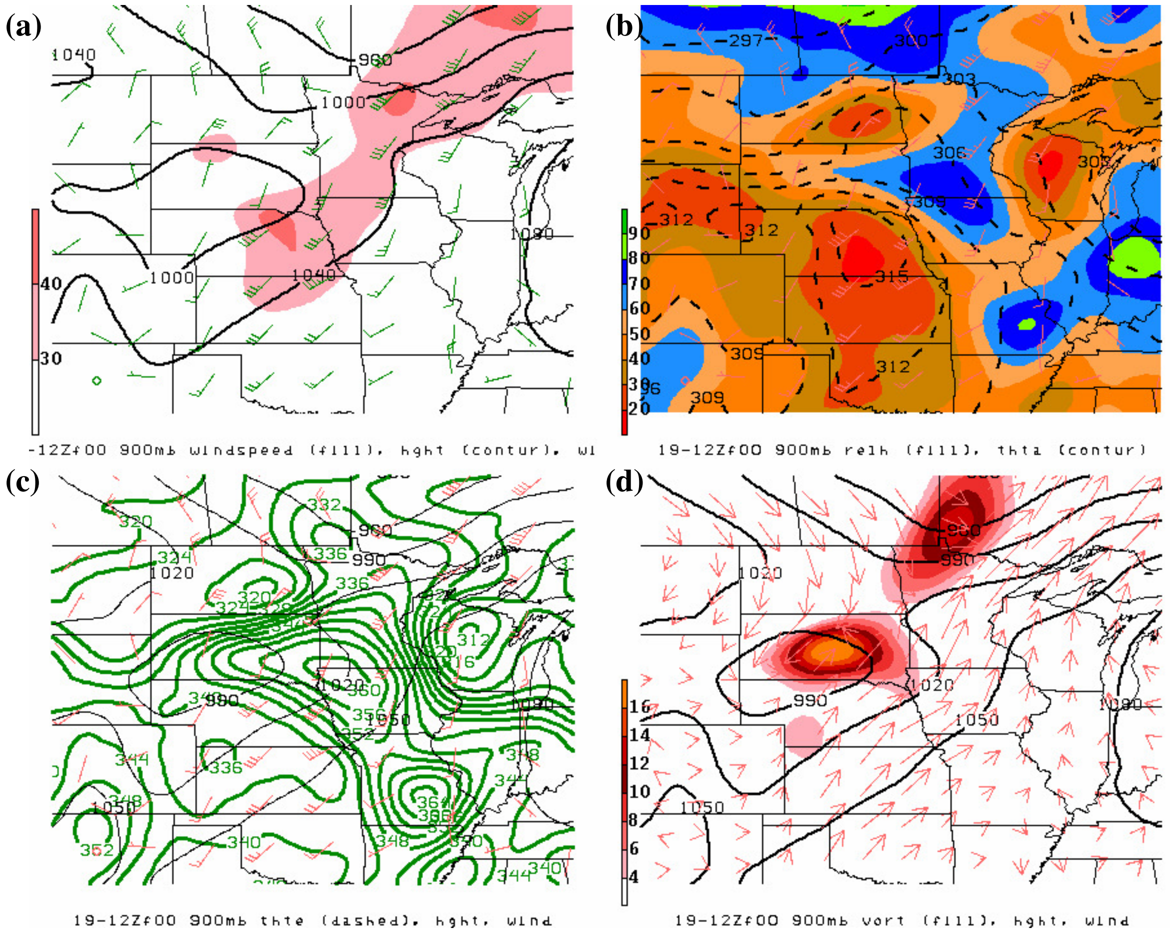


Fig 3. 850 hPa 1200 UTC 19 July Eta model run. (a) wind speeds (fill, kts), heights (m), wind barbs (kts); (b) relative humidity (fill, %) and potential temperature (K); (c) θ_e (K, green), heights (m), wind barbs (kts); (d) vorticity (fill, s^{-1}), heights (m), wind arrows (kts).

associated with it cyclonic flow, including southerly and southeasterly flow across Iowa. The directional shear within the boundary layer is further enhanced by the effects of friction, which causes the flow to turn and cross the height contours towards lower heights (pressure)—thus, a southerly flow becomes southeasterly. It is clear, then, that potent directional shear is maintained by this pattern within the bottom 100-150 hPa of the troposphere. This shear is further enhanced by the developing surface cyclone, which exhibits a tightening pressure (height) gradient and thus an accelerating flow that can act as a greater reservoir for momentum to drive a supercellular circulation. Wind speeds at this level reach 30 knots, which, when coupled with a frictional lower boundary condition, implies tremendous shear and intense mixing. A second important contrast with the 850 hPa level is the prominent moist region stretching across Iowa and maximizing over northern Illinois. Thus, there is a significant drop-off in moisture content evident, likely implicating a strong shift between boundary layer and upper level air that is separated by a significant inversion. This inversion is enhanced by the large-scale flow, with south-southeasterly flow advecting in (relatively) cool but humid air originating from the Gulf of Mexico

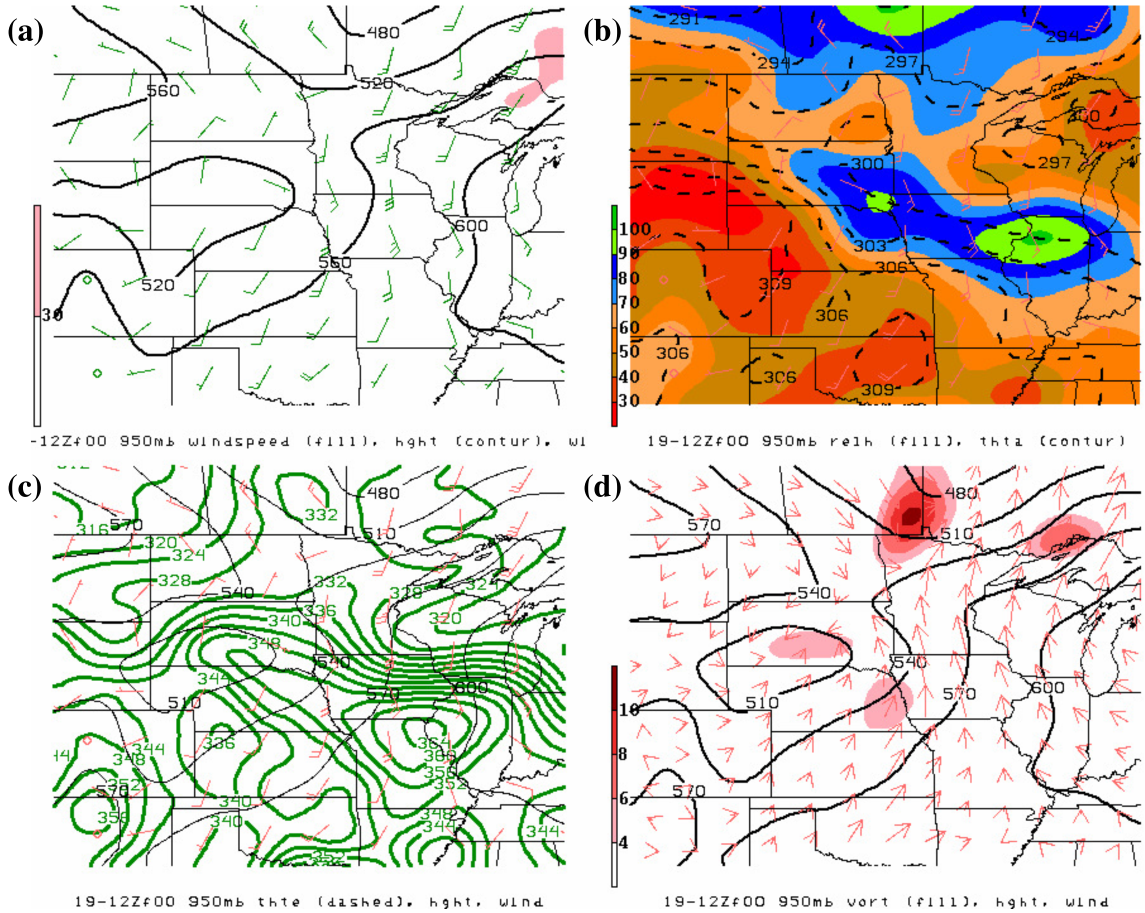


Fig 4. As in fig. 3, but at 950 hPa.

(although air remains relatively dry near Arkansas) near the surface and southwesterly flow advecting in hot, dry Great Plains air overhead. Ultimately, comparison of differential advection of equivalent potential temperature between the 850 hPa level (low θ_e air from Nebraska/Kansas) and the 950 hPa level (high θ_e air from northwestern Missouri) encapsulates the onset of a large contrast between boundary layer and free-tropospheric air that suggests the capping inversion may approach a threshold of convective inhibition.

Despite concerns of a growing capping inversion, two very prominent forcing features would appear set to break through even a very strong cap in the region of interest. First, the 700 hPa shortwaves (discussed earlier) embedded within the confluent flow at that level are converging overhead of the developing cyclone. This serves to enhance forcing for vertical motion through differential positive vorticity advection, as well as to add to the boundary layer circulation that is already becoming highly frontogenetic. Figure 5 depicts the moderate 850 hPa warm frontogenesis at 1800 UTC 19 July that can reasonably be expected to intensify given the approaching shortwaves aloft and the nature of the low-level flow, which is confluent and parallel to the baroclinic zone at and above 850 hPa and then convergent at 950 hPa, with southerly flow south of the front and suddenly southeasterly flow just to the

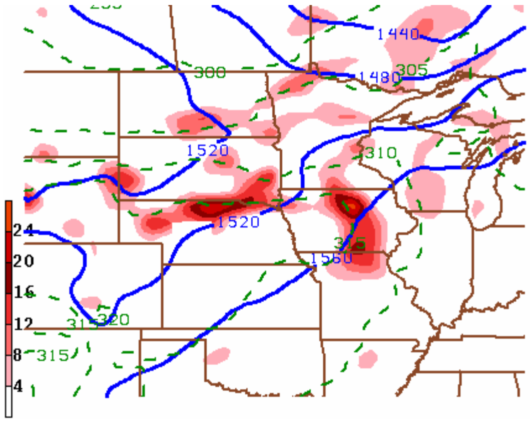


Fig 5. 850 hPa frontogenesis (fill, K*$(100\text{km})^{-1}(3\text{ hr})^{-1}$), heights (m, blue), potential temperature (K) from the 1800 UTC RUC model analysis.

(a)

(b)

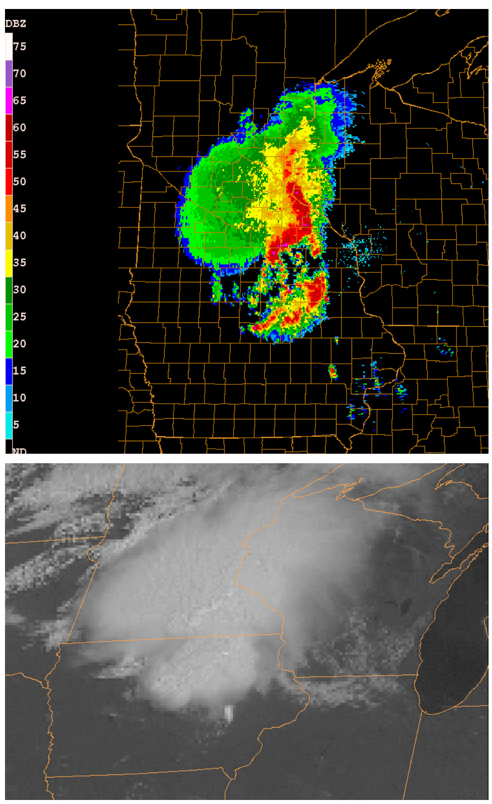


Fig 6. (a) Radar reflectivity at 1525 UTC 19 July; (b) Satellite image at 1515 UTC July 19.

north. Not only are both of these forcings significant and concurrent, but also the timing of their arrival and maximization in the mid-late afternoon is impeccable for the promotion of severe weather. Moreover, figure 6(a) shows the radar reflectivity from 1525 UTC (approximately 9a), where vigorous convection is evident over east-central Iowa (and Minnesota). This convection developed as elevated convection just north of the developing warm front—and to the east of the bust storm chase region—a few hours prior, resulting in significant precipitation and reduced solar insolation (satellite image in figure 6(b)) during much of the morning to the east and north of the warm front. Meanwhile, to the south and west, skies remained relatively clear throughout the day and little precipitation was recorded. This horizontal contrast in both solar insolation and evaporative cooling at the surface provides an additional diurnally-forced mechanism for warm frontal intensification. Thus, the overall effect of these forcings is to further enhance frontogenesis at a time when the general synoptic pattern at all levels is already conducive for a severe convective outbreak.

4. Mesoscale Discussion

a. *The warm front*

The warm front acts as the dominant forcing mechanism for vertical motion during this case as well as modulator of the boundary layer processes that contribute to the incredible instabilities that are observed. Figure 7(a) depicts the 10m streamlines across the Midwest at 0000 UTC 20 July. Convergence along the warm front is evident stretching from the low pressure center in northeastern Nebraska to the northwest corner of Iowa and diagonally

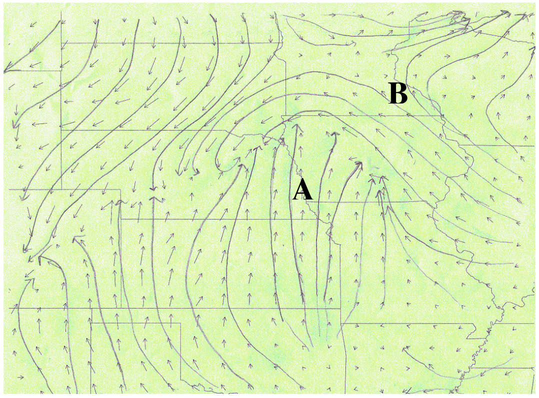


Fig 7. 10m streamlines at 0000 UTC 20 July, from Eta model analysis. 'A' and 'B' denote endpoints of the cross-section in fig. 8.

across the state towards the southeast corner. Strong southerly flow extends from the south-central U.S., undoubtedly advecting warm, moist boundary layer air northward that originates from near the Gulf of Mexico. This flow pattern is typical of a classic severe weather set-up, as high θ_e values near the surface increasingly destabilize the lower troposphere. Meanwhile, the sharp wind shift associated with the surface warm front is clear, as the southerly flow suddenly backs approximately 75 degrees over a distance of at most 200 miles. This convergence region is precisely where severe weather would have the highest probability of firing, as the boundary layer is pumped full of "fuel" for convection that builds up throughout the day as the airmass is partially blocked from advancing northward by the warm frontal boundary. Furthermore, the cyclonic shear along the front would only further enhance any convection, especially supercells whose mesocyclones rely on significant values of low-level 3-dimensional speed and directional shear.

Figure 8 depicts a southwest-northeast cross-section of frontogenesis, potential temperature, mixing ratio, and vertical motion taken across Iowa from Omaha, Nebraska to Minneapolis, Minnesota (see figure 7 for locations) from the 0000 UTC 20 July RUC model analysis. First, the warm front is indicated by the strong surface potential temperature gradient at the center of the cross-section. The front extends nearly vertically up 75 hPa-100 hPa from the surface and then becomes almost instantaneously horizontal, extending northwards over southern Minnesota. Such a structure implies a very sturdy northern boundary that inhibits advancement of the steadily warming and moistening airmass in the lowest 50-75 hPa. This phenomenon is obvious in the bubble of incredibly high values of mixing ratios, including a maximum of 26 g/kg just to the left of the intersection of the 308K isentrope with the surface. The spatial structure of this bubble suggests non-advective processes at work, given that the cross-section is taken approximately parallel to the observed flow at that level. Closer examination of such processes is discussed in a later section.

The cross-section also indicates intense frontogenesis of up to 1 $K*(100km)^{-1}*(1hr)^{-1}$ at 950 hPa. This region of frontogenesis is located along the warm edge of the warm front. Thus, the strongest vertical motions would be expected just on the warm side of this frontogenesis maximum, since quasi-geostrophic secondary circulations induce a thermally-direct circulation in order to reduce to the increasing temperature gradient. Instead, though, ascent is curiously located directly above the region of frontogenesis, with maximum values of 8 mb/s. Both vertical motion and frontogenesis slant northeastward with height—hugging the warm front—suggesting that convection is in fact elevated over the warm front rather than

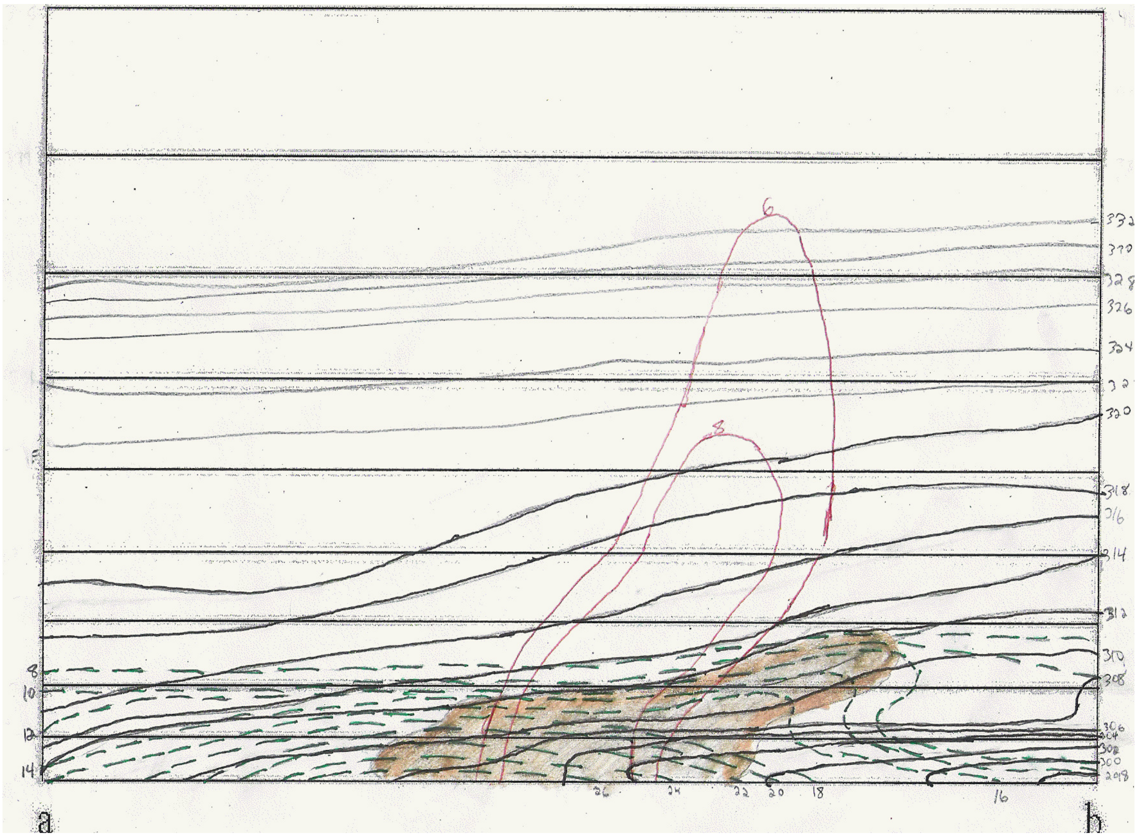
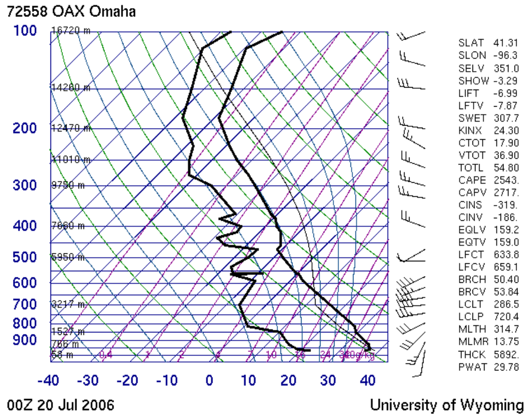


Fig 8. Cross-section from OAX to MPX at 0000 UTC 19 July of frontogenesis (fill), vertical motion (red, negative mb/s), mixing ratio (green dashed, g/kg), and potential temperature (K).

surface-based on the warm side of the front. This is indicative of a strong capping inversion in place, as the rising air is effectively forced to the north, where a weaker cap is in place, before it can convect. The regions of forcing and of observed vertical motions are also collocated with the moist bubble at the surface. Thus, more than sufficient boundary layer moisture, heating, and forcing to release conditional instability are focused into a single region in north central Iowa. Even a strong capping inversion could be reasonably expected to be broken under such conditions. Unfortunately, the radar reflectivity (not shown) displays only weak elevated convection to the north of the front, which corresponds well with the vertical motion contours in the cross-section.

Soundings on the warm side (Omaha) and cold side (Chanhassen) of the warm front at 0000 UTC 20 July are depicted in figures 9(a) and 9(b), respectively. At Omaha, a nearly perfectly mixed layer stretches across an incredibly deep layer from 910 hPa up to 475 hPa, with only a very weak "notch"—likely the remnants of an old boundary layer inversion top—just below 800 hPa. This deep layer thus appears composed of a deep elevated mixed layer that may have developed over the Rockies and the intense surface heating due to solar insolation that boosted surface temperatures to around 36C (97F), which is nearly sufficient to make the entire layer absolutely unstable. However, Omaha is at the boundary of a region where there is a dearth of moisture to the west and south, as the flow is south-southwesterly rather than south-

(a)


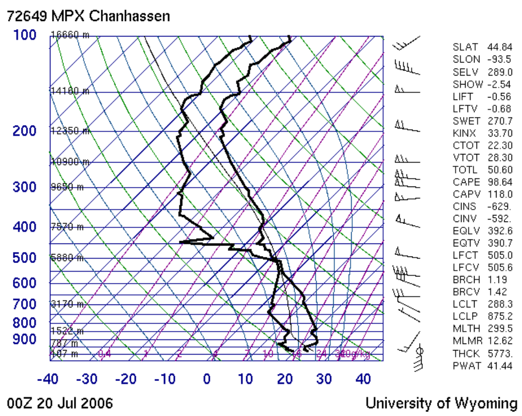
(b)


Fig 9. 0000 UTC 20 July 2006 soundings from (a) Omaha and (b) Chanhassen.

southeasterly as it is to the east, and originates from the Mexican plateau and the dry Great Plains rather than from the Gulf of Mexico. Surface dewpoints near 20C (i.e. 15C dewpoint depressions) rapidly decline with height, resulting in relative humidity values of only 10-30% even as low as 950 hPa to the southwest of Omaha, which implies that surface parcels must be lifted to considerable heights in order to condense. This surge of dry air manifests itself at Omaha in the 700 hPa lifting condensation level (LCL) depicted on the sounding. Even with such low boundary layer relative humidities, the Omaha sounding *still* has a calculated convective available potential energy (CAPE) value of 2543 J/kg. However, if moisture were increased and the LCL (and LFC) were lowered to around 850 hPa, for example, CAPE values could easily reach 4-5 times the observed value. Furthermore, the vertical wind profile places a 40 knot westerly jet streak between 600 and 700 hPa that greatly enhances both speed and directional shear within the lower half of the troposphere to support supercell formation. Overall, in the vicinity of Omaha, the atmosphere is highly unstable and has virtually no capping inversion, implying that very little forcing would be required to initiate convection, but the boundary layer water vapor content is too low to realize this convective potential.

Meanwhile, to the northwest of the warm front, the Chanhassen, MN sounding (fig 9(b)) illustrates quite the opposite situation. At the surface, temperatures are much cooler (22C) but dewpoints are comparable to Omaha (20C), as a large low-level moist pocket prevails over southern Minnesota. The warm frontal boundary is evident in the very large temperature inversion from 950 hPa-925 hPa, creating a strong low-level convective inhibition that blocks any surface-based convection. However, if parcels lifted northward over the front were to saturate through expansional cooling, elevated convection would likely take place given that moderate CAPE values would be available with little convective inhibition. This is indeed observed closer to the front in the cross-section (fig 8), but by the time northward-moving parcels reach Chanhassen, enough moisture may have been rained out to prevent further elevated convection from occurring. Ultimately, it is important to note that the vertical temperature profiles from approximately 650 hPa to the tropopause are nearly identical in both the Omaha and Chanhassen soundings despite the drastic differences in the lower half of the troposphere; Chanhassen is embedded

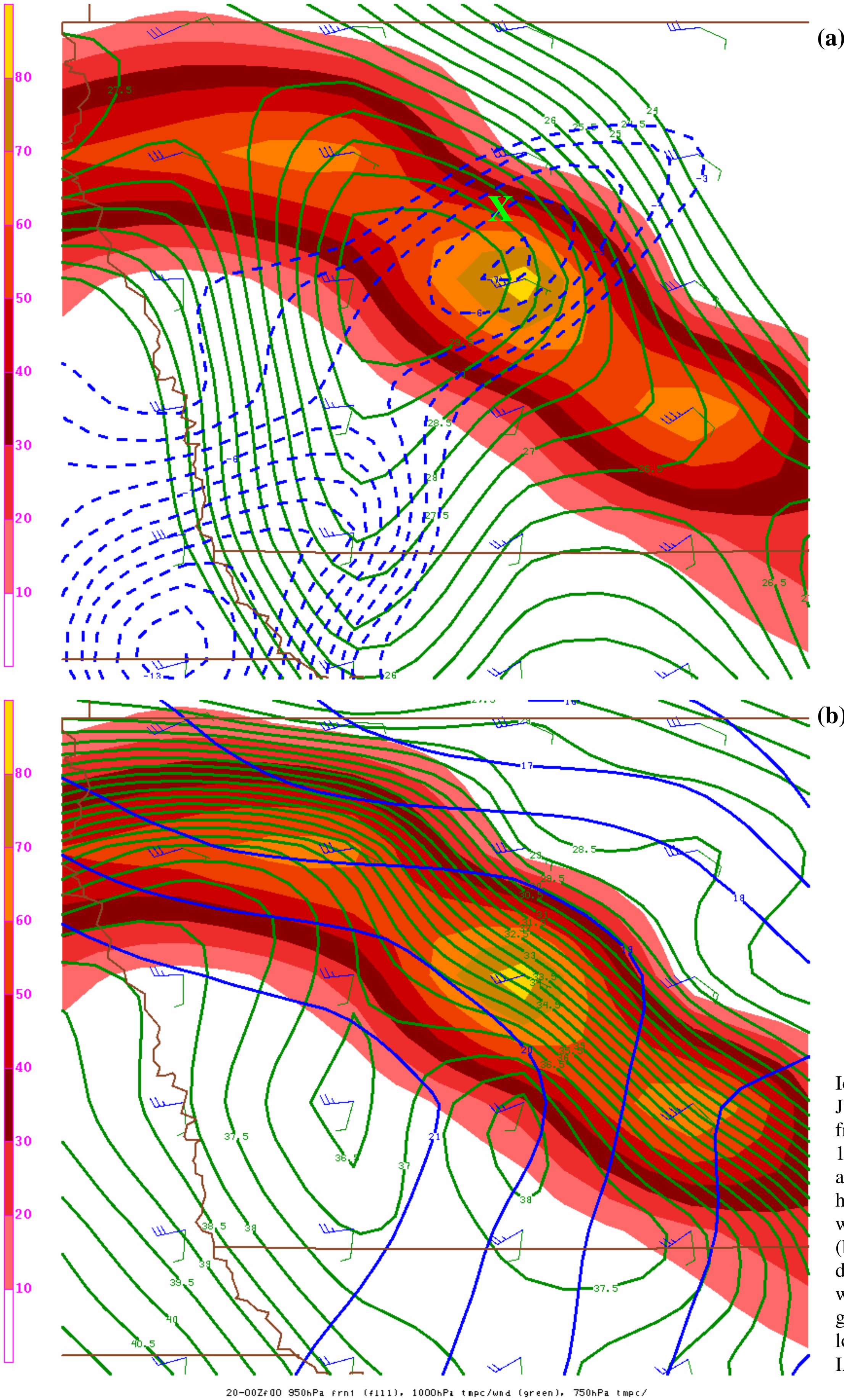


Fig 10. Western Iowa at 0000 UTC 20 July. (a) 950 hPa frontogenesis (fill), 1000 hPa dewpoint and winds (green), 750 hPa dewpoint and winds (blue dashed); (b) as in (a), but with dewpoint replaced with temperature. The green 'X' is the location of Humboldt, IA.

within a much stronger deep-layer 50-65 knot westerly jet. Both upper-level profiles are characterized by a moist-adiabatic lapse rate above 400 hPa, a layer of relatively stable air between 400 hPa and 450 hPa, and then a nearly dry-adiabatic lapse rate from 450 hPa to 650 hPa. This implies that the disparity in CAPE values between the two cities is entirely determined by the boundary layer dynamics and thermodynamics. In the case of Omaha, the low surface moisture content inhibits convection, while in Chanhassen, the elevated warm frontal boundary acts as an impenetrable capping inversion. Thus, the mesoscale set-up for this event confines the region of possible severe convective outbreak to a small region in northwestern Iowa: necessarily on the warm side of the front, but still far enough east that the low-level flow acts as a moisture and CAPE source rather than a sink.

b. *Humboldt, IA*

A close-up of this "wedge" region of potential convective initiation is shown in figures 10(a) and 10(b). Figure 10(a) depicts frontogenesis, 1000 hPa dewpoints and winds, and 750 hPa dewpoints and winds according to the 0000 UTC 20 July RUC model analysis. At and to the west of the location of maximum frontogenesis, the 1000 hPa dewpoints attain a local maximum value of 29.5C, which could be categorized as oppressively humid. This indicates very high moisture content in this region and thus low LCLs, even given such hot surface temperatures. Meanwhile, in the 750 hPa dewpoint field, this very same location attains a local *minimum* value of –7C. Thus, there exists a very small, focused region where a low-level moist conveyor from the southeast and a mid-level dry conveyor from the southwest converge, implying extreme conditional instabilities for surface based ascent.

On the other hand, though, figure 10(b), which is equivalent to figure 10(a) but with dewpoint replaced by temperature, tells a crucially different story. A local maximum, in temperature rather than dewpoint, is once again located very near to the aforementioned region of maximum frontogenesis, with temperatures approaching 38C in south-central Iowa. Furthermore, weak warm air advection is indicated by the 1000 hPa winds, which would suggest a weakening of the capping inversion and perhaps increased potential for a convective outbreak. However, at 750 hPa, in direct contrast to the local minimum in dewpoint observed at the same level in figure 10(a), there again exists a local maximum in temperature. More importantly, the 750 hPa winds indicate much stronger warm air advection than at the surface due to the presence of the low-level jet maximized at approximately 800 hPa, suggesting that the capping inversion would be expected to intensify as the day progresses. At 0000 UTC, in the region of maximum surface heating (~38C), the 750 hPa temperature is approximately 20C. Given the height of the 750 hPa surface at approximately 2600m, the local lapse rate is only ~6.9 C/km—i.e. barely moist unstable at the most unstable location in the region. In other words, throughout most of this region, surface air parcels are *absolutely stable* to vertical displacement throughout the lowest 250 hPa of the troposphere. Thus, for a parcel to convect, it would require lifting to at least the 750 hPa level, which, even given the very strong frontogenetical forcing

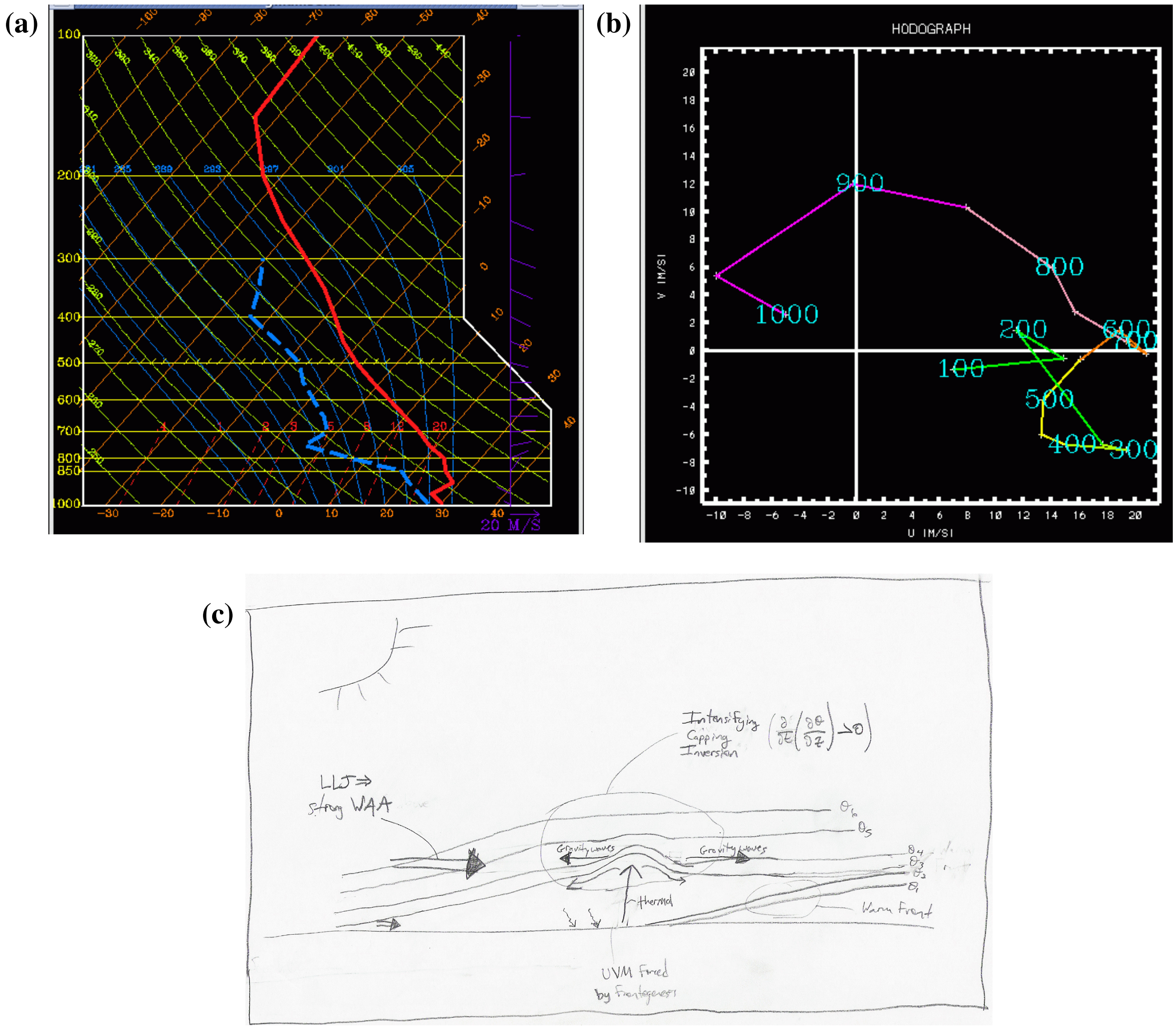


Fig 11. 0000 UTC 20 July RUC model generated (a) vertical sounding, and (b) hodograph, for Humboldt, IA. (c) conceptual model of convective inhibition as seen at Humboldt.

present, would be very difficult to achieve.

With this knowledge unavailable at the time, however, the author and his companion identified Humboldt, IA (see fig 10(a) for location) as the ideal location for *in situ* observation on the day of the event. Figure 11(a) illustrates the model-generated vertical sounding at Humboldt using the 0000 UTC 20 July RUC-model analysis. Humboldt is located just to the north of the location of greatest low-level instability discussed earlier. While very warm and very humid air is recorded near the surface, the capping inversion is dominant, extending over a 200 hPa layer form 750 hPa to 950 hPa. Calculated CAPE values of 8208 J/kg denote the staggering instability of the atmosphere, but examination of the sounding reveals that nearly all of this instability exists between 750 hPa and 150 hPa; the anomalously high tropopause (150 hPa) contributes to this extreme CAPE value by elevating the equilibrium level. However, the convective inhibition is calculated at 78 J/kg; above 50 J/kg is generally considered to be too strong for convective initiation. Furthermore, close analysis of the sounding profile reveals that the capping inversion reaches so close to the surface that, even with very high dewpoints, if one were to lift a parcel dry-adiabatically from the surface, it is likely that this parcel would reach the base of the stable inversion *before* it would reach the LCL, and thus even initial cumulus cloud formation would be inhibited. This corresponds well with observations: no cumulus cloud ever formed in the vicinity of Humboldt; instead, there appeared to be a foggy moisture "dome" with an abrupt upper boundary—a boundary layer glass ceiling—that persisted throughout the afternoon and evening. This discontinuity in moisture was very likely the lower edge of the capping inversion that separated the very moist boundary layer air and the very dry air located just above.

For the storm chaser, this situation was indeed unfortunate. The calculated helicity value at Humboldt was 325 m/s^2, including a nearly ideal, semi-circular vertical hodograph (figure 11(b)) with strong low-level shear and moderate westerly winds aloft; the lifted index (surface to 500 hPa) was approximately 13C; the difference between θ_e maximum and minimum was 41.6K. Overall, nearly every index of instability surpassed the highest typical scale readings. But so too was the 3.2K capping inversion, suggesting not only a deep inversion but a very strongly stable one as well. Ultimately, the small region of tremendous convective potential was doomed by a flow pattern that featured stronger warm advection at just above the boundary layer than within it; this differential advection pattern was sufficiently potent to suppress the strong forcing for ascent associated with the warm frontogenesis. A conceptual model of this process of convective inhibition, such as that seen in the Humboldt sounding, is displayed in figure 11(c).

c. Evapotranspiration

The role of evapotranspiration has been a long-debated issue with regards to convection and convective potential, especially over the midwestern Corn Belt. The landscape of the state of Iowa is dominated by fields of corn and soybeans, whose plants transpire water onto their surfaces that is evaporated into the boundary layer air. The quantitative importance of this water vapor source is not clear, but how this mechanism may contribute to convective activity is: increasing boundary layer water vapor increases low-level θ_e and lowers the LCL, resulting in higher CAPE values and reducing convective inhibition associated with a capping inversion. Thus, evapotranspiration may act to intensify convective activity and, in some cases, even permit convective initiation when it would have been inhibited by the inversion without its added effects. The impact of this process would likely be enhanced under warm, sunny conditions with strong low-level winds for sufficient mixing to maintain efficient evaporation rates.

Repeated examination of weather maps seems to indicate that this indeed is an important source of moisture in the summertime boundary layer. A bullseye local maximum of θ_e values is often observed in Iowa that cannot possibly be advected in from a nearby location. This case in particular is no different, although the proximity to the warm front and associated strong gradient in θ_e makes

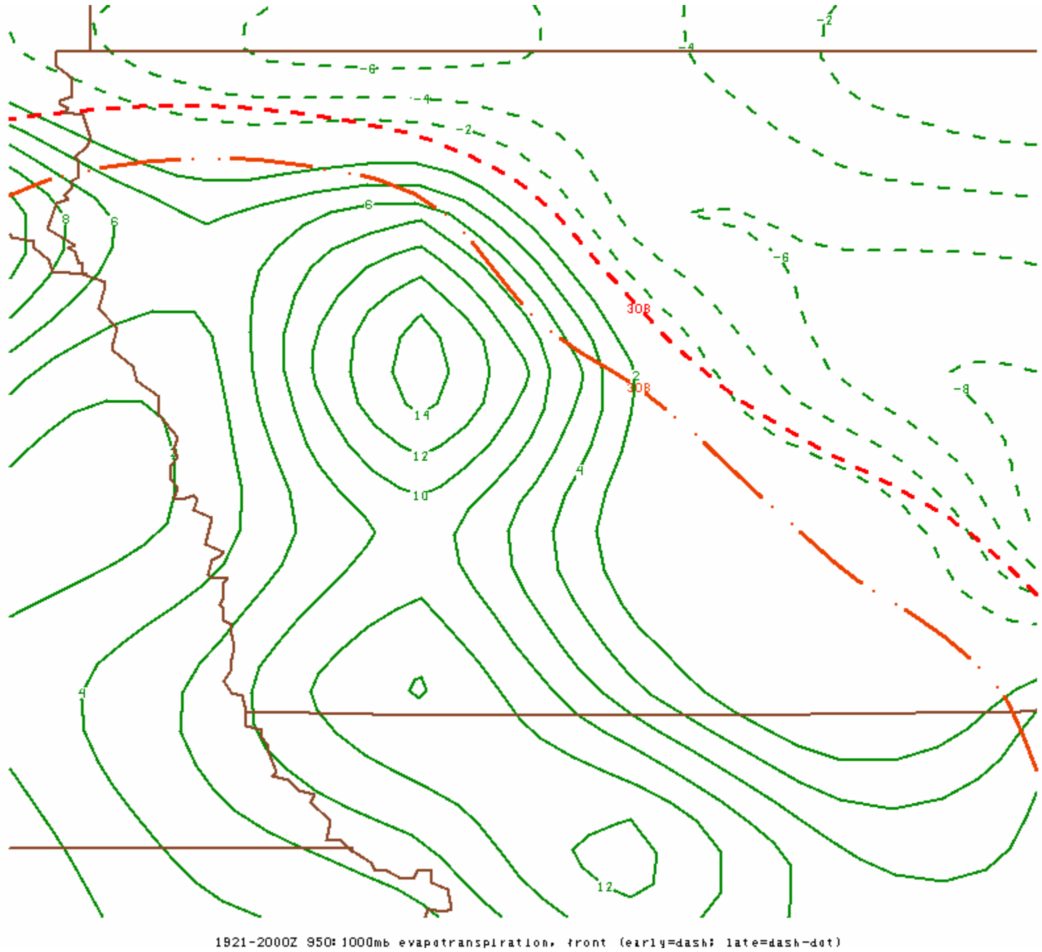


Fig 12. Calculated net water vapor moist entropy flux (K) into the 950:1000 hPa layer by processes other than horizontal advection from 2100 UTC 19 July to 0000 UTC 20 July using RUC model data. See text for formulation. Warm edge of front at 2100 UTC (red) and 0000 UTC (orange), as marked by 308 K isentrope.

data analysis of this subject more prone to error.

Nonetheless, estimation for the net contribution to low-level θ_e by non-advective processes is calculated in GEMPAK. First, the component of θ_e related to moisture can be calculated simply by subtracting the θ from the θ_e:

$$\theta_{wv} = \theta_e - \theta \, . \qquad (1)$$

Then, expansion of the Lagrangian derivative into its Eulerian and advective terms:

$$\frac{d}{dt} = \frac{\partial}{\partial t} + \vec{V} \bullet \nabla \qquad (2)$$

allows for estimation of a surface moisture source, given that water vapor can be viewed as a passive tracer assuming no phase changes. Employing a finite differencing approach within the 950 hPa-1000 hPa layer using RUC model analysis data, the net water vapor moist entropy flux moving with the flow over a 3-hour interval is approximated by:

$$\Delta\theta_{wv} = (\theta_{wv})_{t2} - (\theta_{wv})_{t1}$$
$$-10800 * \frac{1}{2}\left[adv(\theta_{wv})_{t1} + adv(\theta_{wv})_{t2}\right]$$

where advection over this period is assumed to be the mean of the horizontal advection at the beginning and at the end of the period and 10800 is the time interval length in seconds. This crude approach assumes that: 1) the majority of the water vapor flux from evapotranspiration is retained within the lowest 50 hPa of the boundary layer, and 2) net vertical advections across the 950 hPa surface are negligible. Both assumptions seem reasonable given the penetration of the capping inversion down to within the 900-950 hPa, which would limit the majority of mixing to below the 950 hPa level.

When this formulation is applied from 1800 UTC to 2100 UTC 19 July, a weak *negative* water vapor moist entropy flux is calculated over the region of interest in the vicinity of Humboldt, IA (not shown). However, when applied for the period 2100 UTC 19 July to 0000 UTC 20 July, shown in figure 12, significant values for net Lagrangian water vapor moist entropy flux are noted, including a bullseye with a maximum value of +14K in west-central Iowa. Figure 12 also depicts the location of the 308K contour at both 2100 UTC 19 July and 0000 UTC 20 July as a proxy for the warm edge of the warm front. Clearly, the front does not shift significantly during this period, suggesting that frontal interactions and associated vertical motions are at most a minor cause for this notable moisture flux based on the particular formulation used. Instead, the existence of such a focused bullseye may be related to the horizontal gradients in temperature and solar insolation between western and eastern Iowa, as the western Iowa crop belt sat beneath a very warm air mass and clear skies throughout the day while eastern Iowa received little solar insolation.

This calculation is certainly not capable of attributing the entire non-advective θ_e increase to evapotranspiration alone, nor can the author claim that the values given for changes in θ_e are precise given the crudeness of the method employed. However, it is noteworthy to recognize that this method output a *negative* water vapor moist entropy flux into the layer due to non-advective (horizontal) processes over the prior 3-hour time period in this region. This implies that the method used is likely not grossly overestimating the quantities calculated over the latter period when a marked shift occurred. Furthermore, because no precipitation was recorded in this region over the prior 24 hours, evaporation of earlier rainfall can be excluded as a moisture source. Finally, it is important to note that the RUC model analyzed considerably higher surface θ_e values at 0000 UTC 20 July than did the Eta model valid at the same time. However, Aligo et. al. (2003) assessed evapotranspiration in the context of forecast errors in precipitation in a case study of the 22 July 2002 Midwest precipitation event using both the RUC and Eta models and demonstrated that the RUC model more accurately predicted precipitation totals despite in fact forecasting *less* evapotranspiration than the Eta model. Thus, there is no clear indication that the RUC model parameterization for this process is inherently overpredictive.

A 14K 3-hour water vapor moist entropy flux maximum is a very significant contribution to the high instability of this region. The observed θ_e

at Humboldt at 0000 UTC was 375.7K; assuming a reduction of 14K (as an applied example), the θ_e decreases to 361.7K while still maintaining a 1000 hPa temperature of 30.4C. Thus, the dewpoint would be reduced from 27.7C to 24.3C, a decrease of 3.4C. This implies that a parcel must be cooled expansionally by an additional 3.4C beyond the original LCL temperature to reach the new LCL. Since a parcel below the LCL would be lifted dry adiabatically (10 C/km), a parcel must be lifted an additional 340m, or approximately 25 hPa in the lower troposphere, before it can condense and begin to rise moist adiabatically. Given the 4 C/km difference between the dry and moist adiabatic lapse rates, this implies a 1.36C increase in the strength of the capping inversion. Furthermore, given the well-mixed lapse rate observed in the layer directly above the inversion, a 25 hPa increase in the height of the LCL also implies a 25 hPa increase in the height of the level of free convection (LFC), which indicates that the inversion has deepened by 25 hPa in addition to becoming more stably stratified. Finally, the CAPE value in this modified sounding (taken as the original sounding with a 3.4C reduction in 1000 hPa and 950 hPa dewpoints) is only 4984 J/kg—a reduction of 3224 J/kg; the CIN value is 277.9 J/kg—an astonishing 199.6 J/kg increase! These values should certainly be viewed with guarded skepticism, but the mere fact is that even if the calculated effects were halved, this would still have an enormous impact, most notably on the convective inhibition. Overall, then, it is hypothesized that addition of non-advective water vapor moist entropy fluxes into the lowest 50 hPa, much of which is likely attributable to evapotranspiration from crops, is capable of dramatically reducing the strength of the capping inversion while enhancing CAPE and thus increasing the probability of convective initiation.

5. Conclusion

The 19 July 2006 null severe weather event over northwestern Iowa was characterized by very high values of nearly every major index of severe weather potential—as well as convective inhibition. Unfortunately for the severe weather aficionado, the capping inversion was too strong even for the intense forcing for vertical motion associated with warm frontogenesis proximal to the developing surface low pressure center. While intense daytime heating and anomalous low-level moisture transport in western Iowa boosted surface equivalent potential temperature values above 380K—numbers generally seen only at the core of tropical cyclones or at very high elevations—enhanced warm and dry advection just above the very moist boundary layer associated with a southwesterly low-level jet at 800-900 hPa was sufficiently more intense than near the surface to gradually intensify the capping inversion throughout the day. Surface heating was so intense over the Great Plains beneath the dominant upper ridge and relative humidities were so low that a very deep, hot, dry mixed layer developed that was advected over the top of the relatively cool, moist air mass in the lowest 50 hPa with origins from the Gulf of Mexico. This large warm pool was built over the prior couple of days, during which a large anticyclone dominated over the central Great Plains. The result was a capping inversion that was simply too warm and too deep, even despite the incredible surface warming and moistening on the warm side of the warm front. Ultimately, then, the convective outbreak was inhibited by one of the very ingredients typically required for a strong severe weather event: the veering winds within the lowest 200 hPa, which shifted from southeasterly (from the Gulf of Mexico) to southwesterly (from the plains), and the strong speed shear between the surface and the low-level jet at the top of the boundary layer acted as a self-capping mechanism that only strengthened as the afternoon proceeded.

The nature of the low-level moisture convergence was examined to attempt to quantify the much-debated effects of evapotranspiration on convective instability in Iowa. Using a simple formulation to determine the net increase in moist entropy within the lowest 50 hPa due solely to water vapor fluxes, the author calculated a maximum 3-hour net non-advective (horizontal) increase in equivalent potential temperature of 14K over west-central Iowa. Whether or not this seemingly extreme value for such a short time period is an overestimate due to the crudeness of the method employed, the fact that the same method used over the prior 3-hour period (still during strong daytime heating) output a *negative* water vapor moist entropy flux suggests that the method does not necessarily show a tendency for exaggerated miscalculation. Furthermore, given the preponderance of similar θ_e "bullseye" maxima over Iowa throughout the summer, including values typically associated with the core of a tropical cyclone, a 14K increase due to evapotranspiration may in fact not be unreasonable. The Humboldt, IA sounding was then modified according to this 14K increase to determine the effects on stability if this external moisture source is turned off: CAPE values were decreased by **** and, more importantly, convective inhibition was increased by ****. These numbers constitute significant shifts in the parameters that govern convective initiation and thus, based on this case study example, the author concludes that it is likely that evapotranspiration can dramatically increase the probability of a particular forcing to overcome convective inhibition and initiate convection on any given day in high crop density regions within the corn belt. Evapotranspiration would be expected to have a greater impact on a hot, sunny day with strong turbulent boundary layer mixing to maintain efficient evaporation throughout the day.

Additional analysis of the effects of evapotranspiration is certainly warranted, perhaps with a more rigorous method for evaluation of net Lagrangian water vapor moist entropy flux. An ideal formulation would take into account the varying height of the boundary layer (i.e. the lower boundary of the capping inversion) across space and time. However, the case study examined in this work provides nearly ideal conditions to examine these fluxes, given that the capping inversion was so strong and so low that no cumulus clouds (a water vapor sink) ever formed and, given the incredible contrast in relative humidity across the inversion base, very little entrainment mixing between moist and dry air would be expected across this interface. Furthermore, the warm front was nearly stationary during this period, suggesting that data problems associated with the nearby tight moisture gradient along the front can likely be neglected. Analysis of parameterizations of evapotranspiration across models and comparison with *in situ* observations would be useful given the large contrast between RUC model and Eta model surface equivalent potential temperature values during this null event.

ACKNOWLEDGEMENTS

I would like to thank Ross Lazear for leading the storm chase that ultimately led to the topic of this paper. We saw no severe weather, but we were fortunate to learn about the culture of west-central Iowa and the terrific diners along the road home.

REFERENCES

Aligo, E. A., W.A. Gallus Jr.and T.-C.Chen, 2003: Using Iowa environmental mesonet (IEM) data to assess the effects of small-scale variations in soil moisture and sources of errors in precipitation forecasts. Tenth Conference on Mesoscale Processes, Portland, Oregon, Jun 23-27 2003.