BENSON, MINNESOTA CONVECTIVE SYSTEM ON JUNE 11, 2001

Steve Froehlich

Department of Atmospheric and Oceanic Sciences, University of Wisconsin at Madison

Abstract

On June 11, 2001 a strong convective cell formed over western Minnesota, and progressed southeastward across southern Minnesota. The cell formed in a linear structure of cells then merged into one intense cell. Shortly after this merger the storm spawned its first tornado which was an EF-2 that struck Benson, MN. The Benson tornado was the highest (EF rated) tornado to be produced from this cell. Interestingly the cell produced both cyclonic and anticyclonic mesocyclones, the cause of which is mostly unknown. The formation and trigger of the original cell was due to a narrow low level inflow of instability at low levels. Enough convective inhibition existed to trap the instability in the boundary layer until low level moisture convergence, and frontogenesis provided the quasi-geostrophic forcing for vertical motion that broke the capping inversion, and initiating convection.

INTRODUCTION

On June 11, 2001 a convective system moved through the northern plains area. The storm produced high winds, large hail, and several tornadoes. The most severe tornado started in Benson, MN, lasted for 16 minutes, and traveled 8 miles with an average velocity of 30 mph. This tornado was an EF-2 on the estimated fujita scale, and will be referred to as the Benson tornado. Due to the relatively long life span compared to other tornadoes produced by the system, the Benson tornado was the most damaging tornado the system produced. Several mobile homes were damaged, and a few people were injured as they fled the tornado on foot, from their car. The Benson tornado caused over $10 million in damages, as well as damaged over 70 different structures. The system went on to produce over 30 tornadoes from the 11th to the 13th of June, including one EF-3 tornado in Todd County on June 13th. Hail produced during the storm reached diameters up to 3 inches, and gusts up to 80 mph.

The atmosphere as a whole was very unstable to convective activity due to a variety of factors. Processes that aided in the instability were concentrated flows of low level moisture from the south influenced by a tropical cyclone which concentrated the flow in the southern plains. As well as evapotranspiration from the corn belt in the central plains. Drier air aloft was advected in from higher elevations in the west, creating large amounts of buoyant instability as it settled over the low level moisture. The main triggering mechanism can in the form of positive horizontal frontogenesis around 700mb. There is a surface low which forms downstream of the upper level trough and strengthens the cyclonic circulation at low levels which converges with the westerly flow near the northwestern edge of the surface low, this creates a deformation zone that promotes the positive frontogenesis.

Daytime differential heating, and an existing warm and cold front helped create a barrier, which pooled the moist air from the south could into a converge zone near western Minnesota.

SYNOPTIC CONDITIONS

Throughout the setup and genesis time frame there is an upper level low in southwestern Canada. There is also a tropical cyclone that makes landfall over the southern plains in eastern Louisiana, Mississippi, and Alabama. This tropical cyclone (Allison) inhibits the normal pattern of moisture advection off the Gulf of Mexico into the northern states. The tropical cyclone in the southern plains created a synoptic low at its center, and as a consequence built a ridge upstream and downstream of the tropical cyclone center. There were three surface lows in the central US. One in southwestern Wyoming, one in northeastern Colorado, and the strongest low was in eastern South Dakota at 18 UTC on the 11th. The flow at this time is such that all of the warm, moist air is being channeled in a very narrow stream up to the north (figure 1a). The low over northeastern Colorado helps create a trough that dips down through New Mexico into the western tip of Texas. This trough acts as the main mechanism to advect the high equivalent potential temperature from lower latitudes near the Gulf of Mexicxo up into the northern central plains. From figure 1(a) the trajectory of a parcel can be traced from eastern Texas near the Gulf of Mexico, up through the central plains, and converging in western Minnesota and easter South Dakota. The flow at higher levels (1b) is almost straight westerly, but has a slight southerly component to it. The overall flow regime at upper levels will advect warm, relatively dry, mixed layer air into the northern plains. The net effect of the differential moisture advection between higher and lower levels in the atmosphere is to create large instability due to a warm moist layer of air below a warm dry mixed layer aloft. The warmer mixed layer aloft creates a capping inversion at the inter phase between the different air masses which traps the moisture beneath, and keeps it from mixing with the drier air above.

The flow aloft is much faster than the flow at lower levels. This is due to a larger pressure gradient around the US/Canadian border (versus central plains), which stretches over most of Continental US. This accelerates the winds at high levels and creates large vertical wind shear. It is worth noting that there is an absence of a strong upper level jet in the region. Maximum wind velocities in the upper troposphere are on the order of 20 m/s, hence the directional change with height is more important than the magnitude.

One of the more important contributors to strong storm formation besides the buoyant instability, is the existence of large amounts of Helicity in western Minnesota.

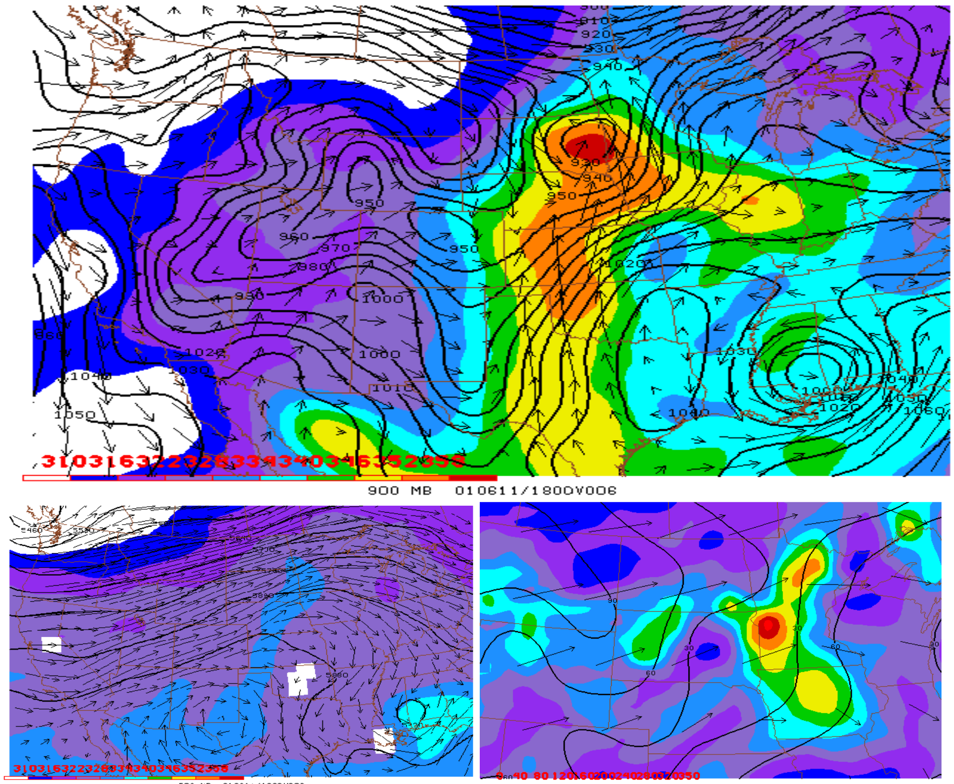

Figure 1: Upper plot (1a) is equivalent potential temperature in fill contour, contoured every 6 K. Geopotentail heights are contoured in black lines every 10 meters to emphasize mass field. Wind vectors are contoured in black arrows. Plot in lower left (1b) is 300mb geopotentail heights, contoured in black lines every 60 meters, and wind vectors contoured in black arrows. Plot in lower right (1d) is geopotential heights contoured every 30 meters, and helicity in fill contour, contoured every 40 m^2/s^2. Figure Valid 18Z

Figure 1c shows the helicity at 18 UTC which is about 3 hours before the tornado hits Benson, MN. From figure 1c the bullseye of strong helicity values are in the western region of Minnesota. It will be shown later that this is the region in which the severe weather initiates. The helicity values in this region are maximized around 360 m^2/s^2. The large helicity is important to the formation of this storm because of the lack of strong vertical wind shear in the absence of an upper level jet streak. This lack of vertical wind shear limits the production of rotating horizontal vorticity tubes, which become tilted into the vetical in severe storms. The rotation from these vorticity tubes, act to stabilize the storm by creating an inertial wall which prevents drier air at higher levels from mixing in the updraft. Helicity likewise creates stability when the velocity and vorticity vectors are aligned. In three-dimensional turbulent flows with a non zero mean helicity value, this tends to damp the inertial decay range of the storm [1]. Therefore helicity becomes important to severe storm formation, especially in an instance like that of the Benson torando, where the environment is characterized by lower vertical shear.

The Role of Surface Moisture Fluxes

The environment in the central plains is characterized by large amounts of moisture at low levels with mixing rations of 14 g/kg in western Minnesota, and maxing at around 16 g/kg in southern Iowa at 900mb. At the surface dew points over southwestern Minnesota reached 84^o Fahrenheit at 18 UTC. It has been suggested [2] that some of the large amounts of moisture came from an influx of moisture due to evaporation from the corn belt, and saturated soil from previous rainfalls.

During the spring of 2001 the rainfall totals in the Midwest were above normal with as much as 200% the mean rainfall for a period from May 1 to June 10 2001. Furthermore 3 times the mean rainfall fell in a 10 day period [2]. The largest amounts of rainfall were in southern Iowa and northern Missouri. Much of this moisture either becomes stagnant in the soil, or is used by the vegetation, and can be released in the form of transpiration.

The proximity of the corn belt to the region of instability and time of year, also influenced the strength and genesis of the severe weather via low level moisture fluxes through transpiration. From [2] in a typical year the average daily evaportranspiration is about 3.3 mm/day, or 97% of potential evapotranspiration (ET for short). During the peak ET, usually during July and August 65% of the maximum rates can be observed. This is strongly dependent on the availability of soil moisture. As mentioned earlier there was a large supply of moisture in the ground from the previous heavy rainfalls.

On June 10th and through 6 UTC on June 11th there was a region of high mixing ratio, compared to that of its surroundings (not shown) over the Iowa, Missouri, Nebraska area. This area of higher mixing ratio is coincident with the regions of high moisture flux from ET specified in [2]. Thus providing more evidence to verify the claims made in [2]. Furthermore from a Climate Diagnostics Center (CDCI) streamline reanalysis vertical transport of a parcel was less than 20 m over the five day period from June 6th to June 11th. This is significant because this indicates that there was little mixing between the moist air in the boundary layer, and the drier free atmosphere above. Therefore the moisture that evapotranspirates from the surface becomes trapped in the lower atmosphere.

From [2] this region of low level moisture becomes advected north, then northeast (wind field in 1(a)) into the western Minnesota area by 18 UTC on June 11th. Furthermore because of the orientation of the geostrophic flow and the surface low, this moisture is channeled into a relatively narrow region and brought up into western Minnesota. Here it is meet with a frontal boundary, as well a developing area of convergence in the morning hours, which strengthens as the day progresses. It will be shown that the combined effects of the channeling due to the mass driven wind field, and the boundaries in the area (which are manifested as both thermal and moisture boundaries) cause a pooling and converging of moisture into the southwestern Minnesota area. This moisture pooling becomes the main instability which allows the formation of the severe weather.

There are thermal and moisture boundaries that form in the central plains are most prevalent in eastern South Dakota, and specifically large in southwestern Minnesota. From figure 1a there is a strong gradient in equivalent potential temperature (θ_e) through the central plains. Also from figure 1(a) the flow to the west is tightening this gradient by advecting low θ_e air to the east. As mentioned earlier a parcel experiences little vertical displacement as it advected along its path. Hence the formation of these thermal and moist boundaries aid in the channeling of the moisture up toward southwestern Minnesota. In the upper northwestern portion of Minnesota the flow is west northwesterly at 18 UTC. Contrast this to the flow in the southern portion of Minnesota which is southerly (figure 2d). This creates an area of convergence along Minnesota's western boundary. This helps keep the growing pool of low level moisture from migrating too far west. There are several features of the flow which aid in this pooling effect. The most important ones are discussed above.

Demonstrated in figure 2 are the two main mechansims responsible for the creation, and or transport of low level moisture into the southwestern Minnesota area. First concentrating at lower levels, it is apparent from the plume of warmer dew point temperatures in figures 2c and 2d, that moisture is being advected from the south. At 12 UTC the effect of the decaying tropical storm Allison is not as prevalent as it is at 18 UTC near the surface. At 12 UTC the cyclonic winds associated with Allison are beginning to cut off the southerly moisture advection. This is evident in the shape of the 20^{o} C isodrosotherm in figure 2c, which bows inward due to the westerly flow associated with Allison. It should be noted that in figure 2, the flow at higher levels is more closely tied to the circulation of the tropical cyclone. As a result the flow is southerly at low levels near the surface, and is not southerly at 700mb. At 18 UTC the low level moisture begins to become squeezed by the converging synoptic and tropical flows. The flow associated with Allison, also seems to cut off most of the moisture advection, which is evident in the separation of the 20^{o} C isodrosotherm (figure 2d) leaving only a narrow path of southerly flow remaining. Figures 2c and 2d demonstrate the southerly advective component to the instability. Figures 2a and 2b demonstrate both, the narrowing effect the convergent flow has on the moisture tongue, as well as the contribution from ET. Focusing first on the narrowing effect, by comparing the 0^{o}C isotherm in figure 2a with the 0^{o}C isotherm in figure 2b, it is evident that the flow is squeezing the moisture tongue zonally. This increases the zonal moisture gradient, and hence the moisture density since the moisture now pushed over a smaller region.

Recall from [2] that the area which experienced the greatest amount of rainfall above the mean was in southern Iowa and northern Missouri. From a comparison between figures 2a and 2b it should become evident that position of the isodrosotherms do not change much (except for the loss of the low dew points over southern Texas). This is especially evident in the 0^{o}C isotherm over the Texas Panhandle which narrows, but does not really change position. Since this is in the only region of possible southerly flow, it follows that there is little moisture advection into the northern plains at this time, at 700mb from the south. Furthermore by contrasting the -4 and -8^{o}C isotherm that extends through the western plains, it does not migrate to the west a whole lot. However if one compares the dew points as the day progresses from 12 UTC to 18 UTC over southern Iowa and northern Missouri, the dewpoints increase around 4 to 8^{o}C in that time frame. Since there is little advection from the west and seemly none from the south, and the dewpoint increase is exactly over the same region that experienced the enhanced rainfall, this supports the earlier arguements that a major source of moisture is from ET. The trajectory simulation from [2] showed the origin of a parcel at 20 UTC at Benson, MN to be central, to southern Iowa, precisely where the warmest dew point temperatures exist. This can also be deduced by examining the flow regime in figure 2b, where if one follows the wind vectors as if they were continuous streamlines, it is apparent that a parcel could be pulled around a cyclonic circle, then advected up into southwestern Minnesota.

CONVECTIVE INITIATION

The synoptic conditions created the large amounts of instability, however the instability required a triggering mechanism to perturb the environment enough to break the capping inversion and release the CAPE values in excess of 3,000 J/kg in some locations. There were several factors which added to the convective initiation, however it will be shown that one of the main mechanisms is the presence of a deformation zone, along a convergent boundary, which cause positive horizontal frontogenesis at low levels. Other factors besides the frontogenetical forcing for upward vertical motion was an area of moisture convergence, a break in the cloud decks in the middle of the state which lead to enhanced solar insolation at the surface, and as mentioned earlier the presence of a helicity maxima (figure 1c) in the area of convective initiation.

From figure 1a and 1c the area of highest helicity values are coincident with the largest equivalent potential temperature values. Furthermore the frontogenesis, solar insolation, and moisture convergence are all localized within that same area. This area is almost directly over Benson, MN.

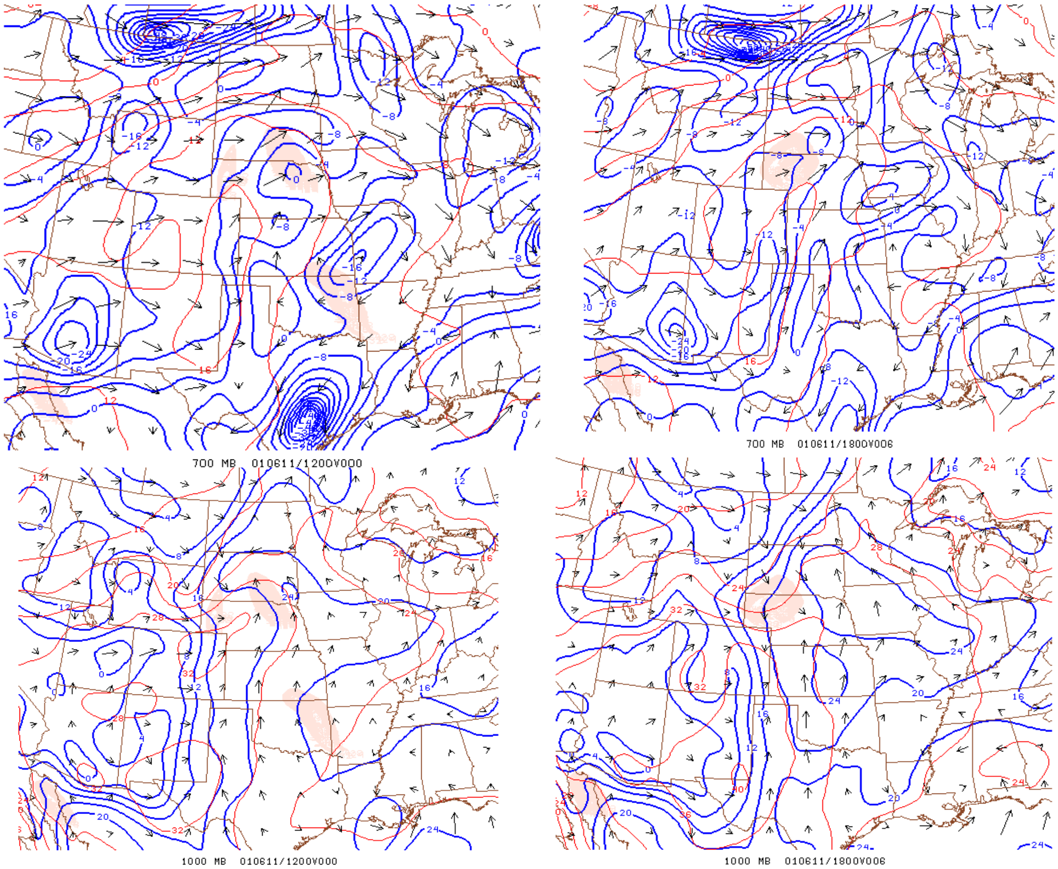

Figure 2: Plot in Upper left (2a) is 700mb dew point temperatures in blue lines, contoured every 4^{o} C. Wind vectors are plotted in black arrows. Plot valid at 12 UTC. Plot in upper right (2b) is 700mb dew point temperatures in blue lines, contoured every 4^{o} C. Wind vectors are plotted in black arrows. Plot valid at 18 UTC. Plot in lower left (2c) is 1000mb dew point temperatures in blue lines, contoured every 4^{o} C. Wind vectors are plotted in black arrows. Plot valid at 12 UTC. Plot in lower right (2d) is 1000mb dew point temperatures in blue lines, contoured every 4^{o} C. Wind vectors are plotted in black arrows. Plot valid at 18 UTC. Figure valid 6/11/2001.

The deformation zone mentioned earlier forms in the western plains at 12 UTC. By 18 UTC the strongest deformation migrates to northwestern Minnesota. This deformation zone promotes positive horizontal frontogenesis. From [3] wherever there is positive horizontal frontogenesis the tightening of the horizontal temperature gradient must induce a Sawyer-Eliasson circulation. This Sawyer-Eliasson circulation produces the upward vertical motion that forces the air through the capping inversion, breaking it and allowing convection release. The cell that produces the Benson Tornado forms to the southwest of an existing disturbance, which initiated around 03 UTC on the 11th and moved eastward. From figure 3 the cells form in a linear structure, which from [3] is indicative of a frontogenetical forcing mechanism. The cells from figure 3 merge together, and intensify in later time steps.

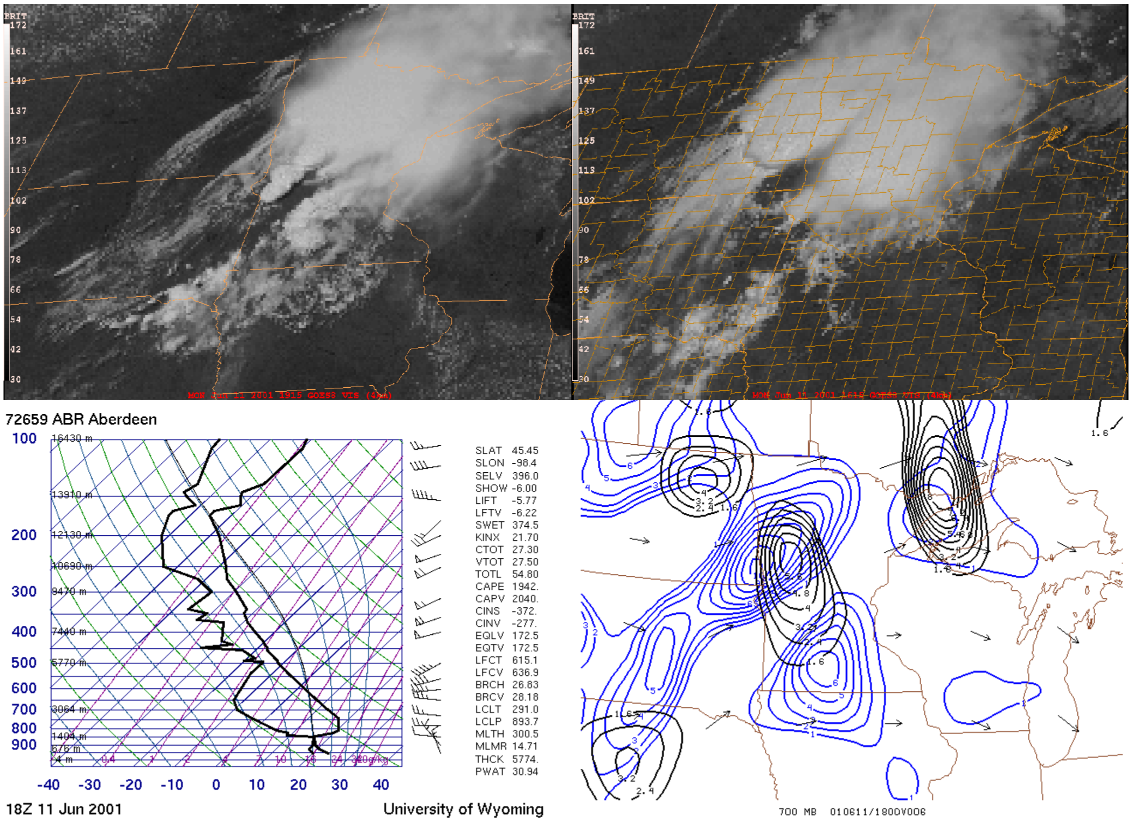

Figure 3: Plot in upper left (3a) is visible satellite imagery from the GOES East satellite. Valid 19 UTC. Plot in upper right (3b) is visible satellite imagery from the GOES East satellite. Valid 16 UTC. Plot in lower left (3c) is skew-T valid 18 UTC. Plot in lower right (3d) is positive vorticity advection by the 300-700mb thermal wind contoured in black lines. Blue lines are 700mb positive horizontal frontogenesis, contoured every 1, valid 18 UTC. Figure valid 6/11/2001.

Figure 3(a) is the first time step in the satellite data in which the convective cell that produces the Benson tornado is visible. This cell is embedded in the group of cells in 3(a), which forms in a linear structure, that jets out from the cloud deck to the northeast. From figure 3c there is a fairly strong capping inversion, which holds the lower level moisture, and hence instability in the boundary layer, and inhibits convection. Above the inversion there is an elevated mixed layer apparent in figure 3(c). The environmental lapse rate in this layer is close to dry adiabatic, and above the elevated mixed layer the lapse rate is still very steep. From figure 3(c) the level of free convection around 650mb. The capping inversion is around 850mb over Aberdeen, SD at 18 UTC. The positive vorticity advection by the thermal wind (pva) and the positive horizontal frontogenesis is maximized at 700mb over western Minnesota. The quasi-geostrophic forcing for vertical motion in this area provides strong enough lift to break the capping inversion, and because it is maximized between the capping inversion and level of free convection this will allow a parcel that has broken through the initial inversion to accelerate to the level of free convection. Not shown but worth noting is that the frontogentical forcing for vertical motion is near the same magnitude and location throughout the depth of the boundary layer, and lower troposphere. This frontogenetical forcing is more important at lower levels than the pva. From [3] precipitation forced by pva tends to form more circular, splotchy, structures. The cells in figure 3(a) form a in a linear structrue, which indicates pva is not as important to convective initiation as other forcings. Because the frontogenetical forcing for vertical motion is localized in the lower troposphere, and given the banded nature of the cell formation, it is thought that the frontogenesis contributes largely to the initial break of the capping inversion, and hence is considered the main triggering mechanism. The reader should keep in mind that figure 3(a) is an hour ahead of figures 3(c) and 3(d). This is why the small growing convective cells appear to initiate slightly to the southeast of the max of frontogenitical forcing.

Once a parcel reaches the level of free convection it will be positively buoyant until it hits the equilibrium level around 190mb. It is worth noting that the equilibrium level is at the same level as the tropopause. This implies more than 6500m of positive buoyancy (firgure 3(c)). This results in the near 2000 J/kg of convective available potential energy (CAPE). Recall from figures 3(d) and 1(c) the region of strong quasi-geostrophic forcings for vertical motion is coincident with a region of very high helicity. Hence once the updrafts form the large helicity values helps them become a stable disturbance, which is important in the generation of the mesocylone that spawns the Benson tornado.

The flow at 18 UTC is southerly, and as mentioned earlier, warmer temperatures are being advected from south at low levels. This helps lesson the temperature difference between the boundary layer and the free atmosphere above. Also the Benson, MN region is exposed to surface insolation which helps to erode the capping inversion. From figure 3(c) the boundary layer is saturated near the capping inversion. This would cause a cloud layer to form around the lifting condensation level which is slightly above the 900mb isobaric surface. Between 14 UTC and 18 UTC there is a break in both the upper level and lower level decks in far eastern South Dakota and far western Minnesota (figure 3(b)). This break allows the earth's surface exposure to solar insolation. Even though the ground is only exposed to the insolation for four hours, the insolation is intense due to the time of year. At this time of year the angle the incoming solar radiation makes with the surface, is closer to a 90 degree angle, which maximizes the surface heating. A comparison between figures 3(a) and 3(b) shows that the convective cells formed in the same location of the cloud break. This implies that while the frontogenetical forcing probably was most important in the convective initiation, surface heating in the cloud break further weakened the capping inversion. In this sense the surface heating was more of a steerer of where the convective initiation was, because the cells could have been initiated anywhere along the line of intense frontogenesis. Wherever the solar insolation weakened the capping inversion, this weakening determined where along the frontogenetical forcing the convective cells formed.

Figure 4 shows the moisture convergence at 850mb valid at 18 UTC. It should be noted that the region of maximum moisture convergence is located over the same region of convective initiation. The moisture convergence is maximized at 850mb which is directly under the capping inversion. This will act to put a large strain on the capping inversion from both a dynamic and buoyant perspective. From [4] moisture convergence is important for locating regions where convective initiation takes place. There are several reasons for this claim. First the most obvious advantage of knowing regions of moisture convergence is to locate where there is a large amount of moisture is being condensed into a smaller region. It should be noted that moisture convergence at higher levels in the atmosphere has been thought to inhibit convection, while the moisture convergence at low levels produces conditions favorable for convection. Intuitively if moisture is being converged at low levels of the atmosphere this will generate instability due to simple density or buoyancy arguments. However a scale analysis performed in [4] shows that the moisture convergence is directly proportional to the horizontal mass convergence field. This means that moisture convergence has dynamical implications as well as buoyancy proxies. Since the moisture convergence is proportional to the mass field this implies that moisture convergence can be used to identify mesoscale boundaries. Furthermore this proportionality implies that the moisture convergence can also dynamically lift a parcel due to the horizontal convergence component inherent in its definition from [4]. It is this combination of lower level mass convergence convergence, and the piling of moisture, that creates (climatologically) the close correlation between moisture convergence and convective inititaion, and this case is no exception.

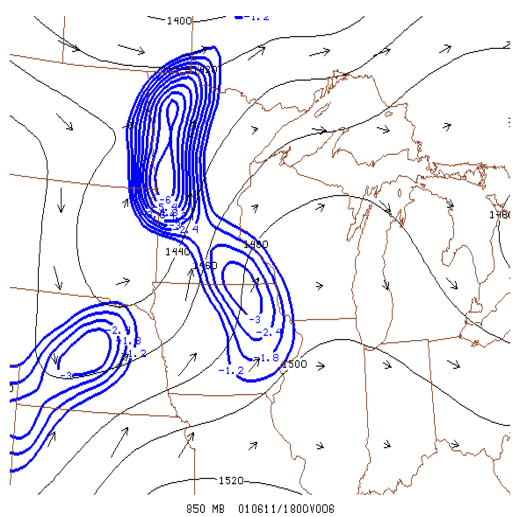

Figure 4: Plot of geopotential heights contoured every 20 m in black lines, and moisture convergence contoured every .6 1/s in blue lines.

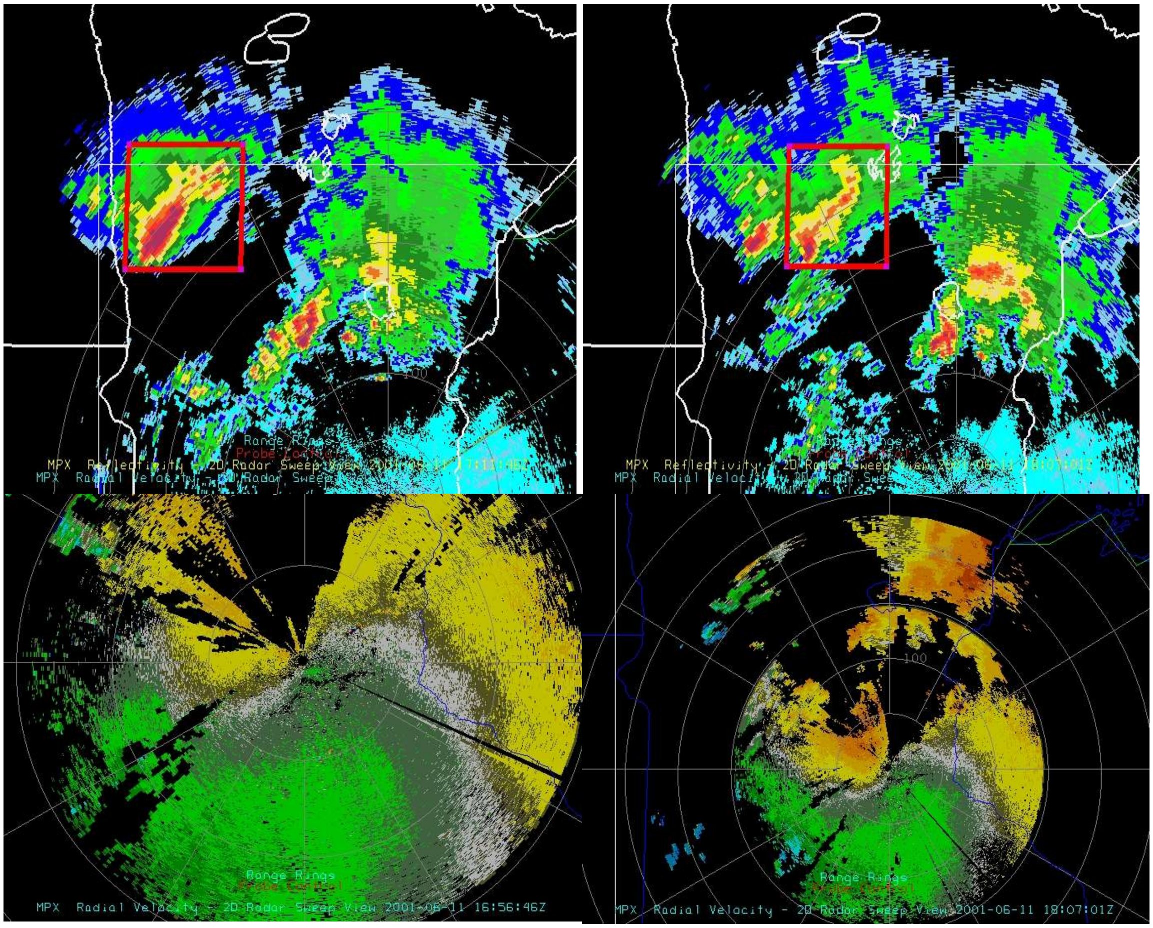

Figure 5: Plot in upper left (5a) is base reflectivity, valid 17:11 UTC. Plot in upper right (5b) is base reflectivity, valid 18:07 UTC. Plot in lower left (5c) is base radial velocity, green values represent velocities radially advancing toward the radar, while orange/red values represent velocities retreating radially away from the radar site, valid 16:56 UTC. Plot in lower right is base radial velocity, same color scheme as 5c, valid 18:07 UTC. Figure is at MPX, valid 6/11/2001.

MESOSCALE ANAYLSIS

As mentioned earlier the convective storm that forms the Benson tornado forms to the southwest of the convective system shown in figure 5. It should be noted that the time frame in figure 5 is before the convective cell that produces the Benson tornado forms. This figure is used to verify the arguments thus far.

From figure 5(a) note the high reflectivties in the cell marked with the red box. Max reflectivities at this time in that particular cell is around 62 dBZ. Note the position of the storm in figure 5(a) compared to the position of the quasi-geostrophic forcing for vertical motion in figure 3(d), and the low level moisture convergence in figure 4. Since the moisture convergence is related to horizontal mass convergence, and the mass convergence in this particular case is creating a deformation zone that is promoting the positive horizontal frontogenesis, these forcings are fundamentally related in that sense. Hence from here forward both of these forcing will be referred to as the convective initiative forcings (CIF). At 17:11 UTC the highlighted cell is over the region of maxima in CIF. At 18:07 UTC the cell migrates to the west away from the CIF maxima. Note the change in dBZ values between figures 5(a) and 5(b). As mentioned earlier at 17:11 UTC reflectivity values are around 62 dBZ, however at 18:07 UTC reflectivity values drop to around 42 dBZ. This is a substantial drop in refectivity (20 dBZ). Also note the orientation of the convective cell in relation to the radar site, by using the range rings which are spaced in 50 km increments, the cell is approximately 50 km closer to the radar station, from 5(a) to 5(b). Due to the parabolic surface the radar projects on the two dimensional plot, the closer the cell is to the radar the lower in the atmosphere the radar sees. Hence the image in figure 5(a) are taken at a higher altitude than that of figure 5(b). This means the drop in reflectivities is even more significant than 20 dBZ. The difference in reflectivities is attributable to the position of the cell in relation to the CIF. When the cell is over the maxima CIF, the dBZ values are the highest, however once the storm migrates to the east of this forcing, the dBZ values decrease. It is also worth noting that a smaller updraft forms to the west of the highlighted cell in figure 5(b), which also emphasizes the importance of the CIF in forming strong convection.

Figures 5(c) and 5(d) show the radial velocities. The radial velocities can be used as a measure of the vertical wind shear. From figure 1(c) the area is characterized by high helicity values in the region to the northwest of MPX. Figures 5 (c) and 5(d) give a sense of the vertical structure of the wind shear. The main point this figure is emphasizing in the radial velocity plots, is the constant large directional wind shear with height in the MPX area. Note the time frame between 5 (c) and 5(d) is about an hour. This shows the steadiness of the shear instability in the environment prior to the initiation of the cell that produced the Benson tornado.

At 18:58 UTC the first echo of the cell that produces the Benson tornado becomes identifiable. From figure 6(a), at first several convective cells form in the region of max CIF. However the cells begin to merge with each other. The cells that move with a slight northerly component lose strength in their updraft, and fade. Conversely the cells that form further south, and move with a southerly component strengthen. This strengthening is evident in the greater reflectivities from figure 6(a) to figure 6(b). The reason for the dissipation of the north cells and the strengthening of the cells to the south has to do with the way they move relative to the mean wind. The cells in the north move to the left of the mean wind, while cells to the south move to the right of the mean wind. This causes the right movers to benefit from continued helicity, while the left movers weaken.

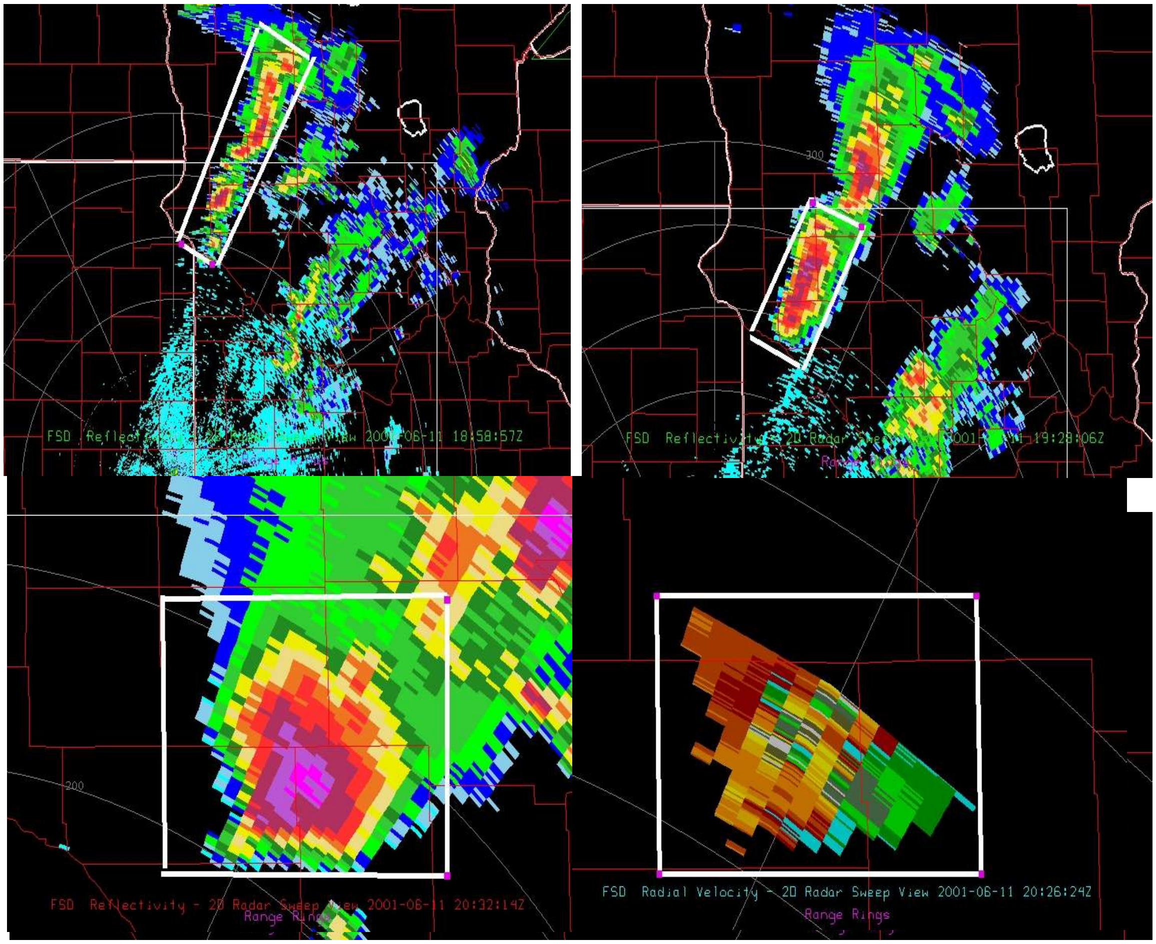

Figure 6: Plot in upper left (6a) is base reflectivity, valid 18:58 UTC. Plot in upper right (5b) is base reflectivity, valid 19:28 UTC. Plot in lower left (6c) is base reflectivity, valid 20:32 UTC. Plot in lower right is base radial velocity, green values represent velocities radially advancing toward the radar, while orange/red values represent velocities retreating radially away from the radar site, valid 20:26 UTC. Figure is at FSD, valid 6/11/2001.

Figures 6(a) and 6(c) highlight the merger of the smaller cells from 18:58 UTC to 19:28 UTC on the 11th. The cell furthest south in figure 5(a) is the cell which produces the Benson tornado. It is also the cell that exhibits the strongest southerly component in it's mean path. The intensity of the storm after the merger of the cells suggests some kind of constructive interference between the individual cells that helps to organize the storm into one stronger cell. From 19:28 UTC to 20:16 UTC one can see the storm gaining strength as the individual updrafts either die out or enhance the original updraft. From figure 6(c) the a hook echo is present at 20:32 UTC. This indicates the presence of the mesocyclone. From the storm reports, the tornado is on the ground at this time. Reflectivities throughout the time period when the tornado is on the ground, are near 70 dBZ. Not shown but evident in the satellite data are the overshooting tops of the updrafts. The overshooting tops are consistent with the high reflectivities, sustained by strong updrafts that persist throughout the life cycle of the tornado.

The presence of the mesocyclone is also seen in 6(d) in the radial velocity couplet. In figure 6(d) the tight couplet of green and orange colors indicate the large cyclonic shear within the mesocyclone. In the satellite data the radial velocity couplet is a better proxy for the mesocyclone, since the hook echo in the reflectivity is not as identifiable throughout the tornado's life cycle. The lack of a prominent hook echo is due to the fact that the cell is over 200km from the radar site, hence the reflectivities are from relatively high up in the storm. This is because the hook echo is produced by the low level inflow, and the rear flanking downdraft, and this is most prominent at lower levels (figure 7(a)).

As mentioned earlier the Benson tornado lasts for almost 20 minutes. The longevity of this storm could be attributed to the fact that the cell moves to the southeast, and benefits from the tongue of high θ_e air at lower levels. This gives the storm the energy it needs to sustain the tornado for a relatively long time (compared to the other tornadoes that broke out the second longest tornado, after the Benson tornado was only a few minutes).

The longevity of this cell could also be attributable to the synoptic induced shifts in low level vorticity regions, and persistent synoptic inflow of moisture at low levels. From figure 7(b) the surface low pressure maxima has migrated to the east, which bring it right in the path of the Benson cell. Not only does this create vorticity associated with the low pressure center, but because the low is still deepening at this time, the upper level divergence responsible for deepening the low could also aid in enhancing the Benson cell.

There is also a horizontal shear line in figure 7(b) across southern Minnesota. This particular shear line causes cyclonic shear in the area where the Benson storm is migrating. This increases the sources of low level vorticity that the Benson cell can feed off of.

Along the shear line is the persistent southerly flow, that will continue to advect higher θ_e air from the south. Notice that the high θ_e air comes from both the corn belt, as well as near the Gulf of Mexico. From figure 7(d) the convective region is identifiable northwest of Winona, Minnesota. The boundary layer is still unstable due to the persistent low level moisture advection. This is evident from figure 7(d) because of the increasing θ_e values that decrease with height, and the unstable layer above it.

Before the death of the storm the Benson cell produced an anticyclonic tornado. From figure 7(c) the hook echo has an anticyclonic curvature to it, indicating the anticyclonic mesocyclone. This could have formed due to the backing of wind with height. From figure 3(c) the backing winds form vorticity tubes that when tilted in the vertical would produce and anticyclonic rotation. The dynamical

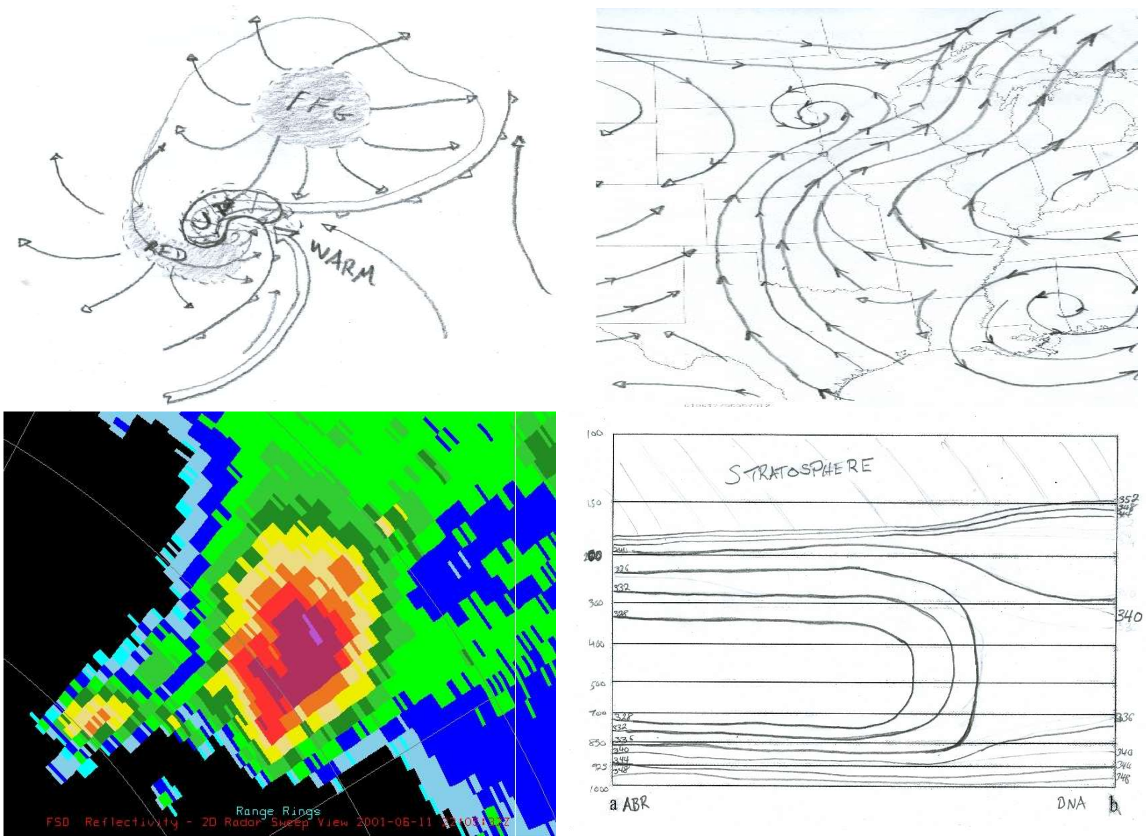

Figure 7: Plot in upper left (7a) is conceptual model of airflow in a supercell. Plot in upper right (7b) is 900mb streamlines, valid 00 UTC 6/12/2001. Plot in lower left (7c) is base reflectivity valid 22:03 UTC 6/11/2001. Plot in lower right is hand drawn cross section of equivalent potential temperature contoured every 4 K between ABR and ONA, valid 00 UTC 6/12/2001.

reasons for the switch from producing a cyclonic hook echo and an anticyclonic one are unknown. The possible causes of this change could be that from figure 1(a) the winds in southern Minnesota are southerly at low levels, however the winds are northerly to the northwest. Since at upper levels the flow is mainly westerly (figure 1 (b)) this means that the winds are backing with height in the northwestern portion of Minnesota and veering with height in the southwestern portion of Minnesota. This could cause both cyclonic and anticyclonic horizontal vorticity tubes to form. When the updrafts tilt cyclonic tubes into the vertical, a cyclonic mesocyclone forms. When updrafts tilt and stretch anticyclonic vorticity tubes into the vertical, a anticyclonic mesocyclone forms. From figure 7(a) this anticyclonic voticity would cause the rear flanking cold front to switch to the left side, and the weak warm inflow jet would form to the left as well. This would explain the shift in hook echoes from figure 6(c) to figure 7(c). Other possibilities could be associated with mesoscale alterations in the small scale, shear and baroclinicity induced by the storm, and which could or could not be dependent on synoptic conditions.

CONCLUDING REMARKS

The synoptic conditions in this instance produce large amounts of instability that are concentrated over a small area. Furthermore all of the triggering mechansims are located in the same area. From figure 1 the largest helicity values are in the same region as the high θ_e. From figure 3 the cloud break, pva, frontogenesis, and moisture convergence in figure 4 are all located in the same area, however the pva is not thought to contribute substantially to the convective trigger, and is shown just to exmplify the concentration of quasi-geostrophic forcings for upward vertical motion, which led to the precise and timely NWS forecasts.

Tropical storm Allison allowed for the flow to advect instability into a narrow band. The evapotranspiration added to the instability at low levels. The wind field advected this instability into the western Minnesota area. The convergent flow in the western Minnesota area was critical in allowing for the lower level moisture convergence, and frontogenesis that aided in convective initiation. The net result of the of concentration of both, the instability, and triggering mechanisms in the same region allowed for a narrowing in the threat for severe weather. This allowed the National Weather Service to precisely identify the region that was in the most danger of convective activity. This allowed for a minimization in the amount of people hurt or injured during the convective outbreak. Hence when the Benson tornado struck, the people in that area were well aware that severe weather was headed in their direction and took appropriate action.

REFERENCES

[1]-American Meteorological Society Website, Glossary of Meteorology. http://amsglossary.allenpress.com/glossary/help

[2]–Cheresnick, D,. R., Jeffrey, B., B. The Impact of Land-Atmosphere Interactions on the Benson, Minnesota, Tornado of 11 June 2001. BAMS, 637-624. *Weather and Forecasting*. 20., 351-366.

[3]–Martin, J. Mid-Latitude Atmosphere: A First Course. West Sussex, England. Wiley. 2006.

[4]–Schultz, D., M., Banacos, P., C. The Use fo Moisture Flux Convergence in Forecasting Convective Initiation: Historical and Operational Perspectives. *Weather and Forecasting*. 20., 351-366.