ANALYSIS OF THE 3 – 4 JUNE 2002 EXTREME RAINFALL EVENT OVER IOWA AND ILLINOIS

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Abstract

During the period from 1800 UTC 3 June to 2100 UTC 4 June 2002, heavy convective rainfall of over eight inches in east-central Iowa and northwestern Illinois resulted in extreme flash flooding and river flooding. The elevated convection was episodic in nature and formed north of a quasi-stationary frontal boundary. Weather Surveillance Radar - 1988 Doppler (WSR-88D) data revealed at least four mesoscale convective systems (MCSs) that contributed to the heavy rainfall. Diagnostic storm motion and propagation vectors were computed and compared to estimates using both the upwind and downwind Corfidi vector methods. The first two MCSs' storm motion vectors were well predicted by the downwind Corfidi vector method. However, the motion of the latter two MCSs, which contributed the most to the total rainfall, were not predicted well by either of the Corfidi vector methods. It is shown that in the latter two MCSs, cold pool outflows and the development of a mesoscale pressure ridge enhanced the frontal boundary and organized deep moisture convergence for a sustained period of time, thus altering both the propagation of new cells and the motion of the existing convection. A contrast is also noted between low-centroid echoes within MCS #3 that created heavy rainfall and a high-centroid storm within MCS #4 which dropped one-inch hail. Finally, operational considerations associated with the real-time forecasting of this event are discussed to fully appreciate the difficulty of predicting this type of heavy rain event.

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1. Introduction

During the period from approximately 1800 UTC 3 June to 2100 UTC 4 June 2002, convective rainfall resulted in extreme flash flooding and river flooding over portions of east-central Iowa and northwest Illinois. As seen in Fig. 1, rainfall amounts over four inches were common in these

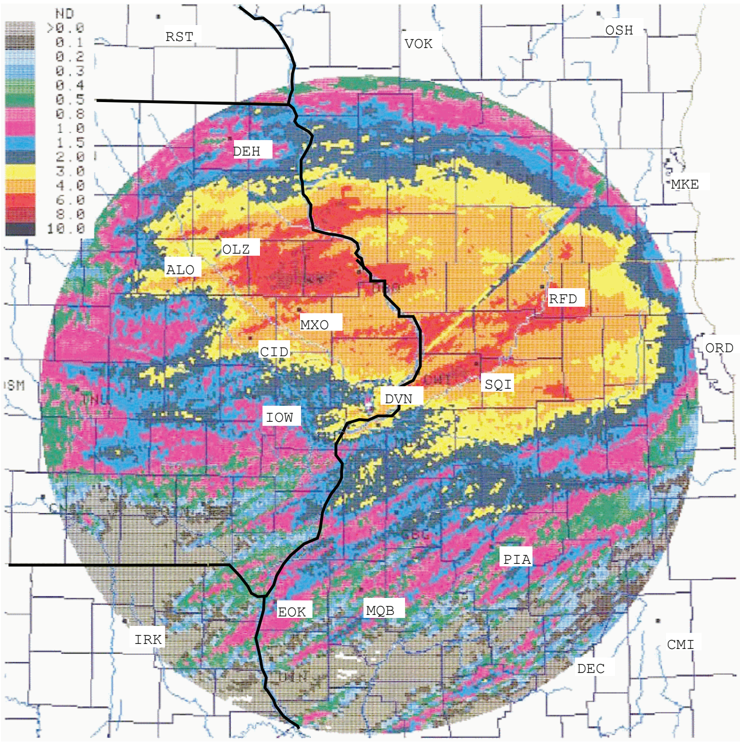


Fig. 1. WSR-88D storm total precipitation (STP) from Davenport, Iowa for the 3-4 June 2002 heavy rainfall event. The storm totals in inches are from 0002 UTC 2 June 2002 through 0213 UTC 5 June 2002. Labeled on the figure are surface observing sites using their the 3-letter identifier.

areas with extreme amounts, as high as eight to eleven inches, reported in Delaware and Dubuque counties in Iowa (see Fig. 2 for pertinent geographic information). The maximum rainfall for this event exceeded the 1 in 100 year event for both 48-h and 72-h periods (7.83 in and 8.42 in respectively; Huff and Angel 1992). According to *Storm Data* [NOAA/National Climatic Data Center (NCDC) 2002] the Rock River reached flood level near Joslin, Illinois while the Maquoketa and Wapsipinicon Rivers rose well above flood stage. The Maquoketa River rose high enough to force closing of the water treatment plant in Monticello (MXO), in northeast Jones County (see storm rainfall totals for MXO area in Fig. 1), which did not occur even during the historic floods of 1993. President G. W. Bush declared 17 counties in eastern Iowa disaster areas as over 7.2 million dollars of property damage occurred. In northwest and west-central Illinois, rainfall of six to ten inches also resulted in significant property damage (around 3 million dollars), rivaling that of the summer of 1993 and the spring snowmelt of April 2001 (Zogg et al. 2002). During the height of the storm, rainfall rates of over two inches in an hour were recorded.

The goal of this paper is to document those meteorological processes that forced the successive, episodic mesoscale convective systems (MCS) that repeatedly moved over the region during this 27-h period. Secondly, processes which affected MCS motion and increased the precipitation efficiency (PE) of the convective cells are described. Towards these ends, section two reviews the major parameters that contribute to MCS initiation and motion. Section three describes the synoptic-scale regime within which deep convection was initiated and nurtured in this particular event. Section four examines the mesoscale characteristics of the event through an analysis of surface, WSR-88D radar, and RUC-II data. In this latter section the focus is upon the episodic nature of the convection and triggering mechanisms. Section five discusses the operational considerations that forecasters faced during this heavy rainfall event. Finally, section six presents an overall discussion of our results and concluding remarks.

2. Meteorological Factors Contributing to MCS Initiation and Sustenance

According to Fritsch and Forbes (2001), MCSs can be categorized as being one of two types. Type 1 events result when a narrow region of conditionally unstable low-level air is forced to ascend in a frontal zone such as a stationary front. Type 2 events occur in a more barotropic environment and depend upon the production of convectively-generated cold pools interacting with the ambient vertical wind shear to produce a region of mesoscale ascent. In this sense type 1 events are due to externally imposed forcing, while type 2 events depend

Fig. 2. Geographic information (cities, rivers, etc.) pertinent to this case as discussed in text.

strongly on "features and processes imposed by the convection itself" (Fritsch and Forbes 2001). However, it would seem that in the case to be described herein there is evidence of both types of forcing as MCSs were produced north of a stationary boundary, yet showed distinct evidence of being affected by outflow boundaries produced from the convection.

The "frontal" type heavy rain scenario described by Maddox et al. (1979) involves the presence of a quasi-stationary boundary with deep convection forming on the north side of the boundary, often in the presence of a diurnally-varying low-level jet (LLJ) which transports conditionally unstable air northward and upward along the boundary. Frontal type heavy rain events are climatologically favored to occur during the warm season (Maddox et al. 1979), often forming overnight as the LLJ increases in strength from the south-southwest. Moore et al. (2003) noted that many frontal type events can also be characterized as elevated MCSs since they initiate not only north of a quasi-stationary boundary but also above a cool, stable boundary layer and thus are not connected to diabatic processes within the planetary boundary layer (PBL; see Colman 1990). In their composite study of warm season elevated thunderstorms, Moore et al. (2003) found that convection was focused:

- on the cool side of a strong north-south surface equivalent potential temperature (θ_e) gradient,
- within a southwest-northeast elongated moisture convergence axis in the 925-850 hPa layer,
- within a maximum of positive 850-hPa θ_e advection,
- downstream from a southwesterly LLJ,
- in the right-entrance region of an upper-level jet (ULJ) streak,
- within an axis of high convective available potential energy (CAPE) and low convective inhibition (CIN) as computed from the maximum unstable parcel, and
- on the warm side of a region of low-middle tropospheric frontogenesis.

Recently, Banacos and Schultz (2004) have noted that surface horizontal moisture flux convergence may not always be representative of a deeper layer and suggest that an integrated value of moisture convergence may be more useful for estimating convective initiation. They also note that observations from the National Weather Service's Storm Prediction Center (SPC) indicate that as many as 50% of thunderstorms have an updraft source level above the surface. Thus, for elevated thunderstorms it is important to estimate both lifting mechanisms as well as the source of the moist, unstable air that feeds the thunderstorms.

Doswell et al. (1996) stated that most long duration rainfall events are associated with MCSs which exhibit slow motion and high PE. Junker et al. (1999) found that most of the heavy rain events associated with the flooding of the summer of 1993 in the Midwest had cloud-layer winds which were nearly parallel to the low-level boundary with a small component toward the cold air, thereby increasing the likelihood of "training" (i.e., cells that repeatedly move over the same geographic location). Chappell (1986) described storm motion as the sum of the advection and propagation vectors. Corfidi et al. (1996) used the 850-300 hPa mean winds to estimate the advective component of a MCS's heaviest rainfall cells (i.e., meso-β scale elements) and a vector equal in magnitude but opposite in direction to the LLJ as an approximation of the propagation component to arrive at the storm motion vector (illustrated in Fig. 3a). Mean absolute errors for this simple technique for MCS speed and direction were 2.0 m s^{-1} and 17°, respectively (Corfidi et al. 1996). Although the original "Corfidi vector" technique (called the "upwind-propagating" technique) can be useful for MCSs associated with training (e.g., quasi-stationary or back-building convective storms), there were many cases in which it failed to describe the rapid downstream motion of MCS's such as that which occurs during fast-moving bow echo events. In these cases, MCSs propagate downstream as is typical during fast-moving bow echo events. For the later cases, when outflow boundaries contribute to new cell development *downstream* from the convection, Corfidi (2003) described a new approach for estimating storm motion. In this case the cloud-layer winds (850-300 hPa mean winds) are an approximation of the cold-pool motion since momentum transfer from the mean cloud layer contributes most to the gust-front/cold pool velocity. The storm motion obtained from the original upwind-propagating vector technique represents the "negative of the cold pool-relative flow" and is used as the propagation vector. This propagation vector has a large component parallel to the cold pool motion vector, resulting in a MCS motion vector that invariably "points" downwind (illustrated in Fig. 3b). This "downwind-propagation" technique is most applicable to those situations when mid-level dry air (relative humidity less than 50%) is present in the storm environment, increasing the likelihood of cold pool formation due to evaporatively-cooled downdrafts from convection. The original "upwind propagating" technique appears to be most useful in near-saturated storm environments. It should be noted that in both cases a uni-directional wind shear profile is typically found, although north of a stationary boundary the winds in the first 100 hPa or so usually display moderate shear and veer. In the present case both the upwind-and downwind-propagation techniques will be tested to assess their applicability in the elevated thunderstorm environment.

As noted earlier, MCSs associated with intense rainfall usually consist of high-precipitation efficient cells, where PE is defined as the ratio of storm rainfall to the moisture ingested by the storm over its lifetime (Doswell et al. 1996). Assessing whether a storm-environment is capable of supporting convective elements which have high PE is not a trivial task as PE is a function of both storm-environment variables and cloud microphysical processes, the latter not being easy to quantify in an operational environment. However, there are some parameters such as the mean environmental relative humidity and wind shear, and storm-relative moisture influx, which can be used to estimate the ability of the environment to support convective cells with high PE. Also, several studies (e.g., Pontrelli et al. 1999; Kelsch 2004) have shown that in real time high PE cells can usually be identified by their low-centroid (i.e., high reflectivities greater than 50 dBZ concentrated below 5 km) echoes. Kelsch (2004) has noted that storms displaying this trait have rainfall rates that are more accurately described by a tropical Z-R relationship. Some aspects of this PE problem will be addressed later in this paper.

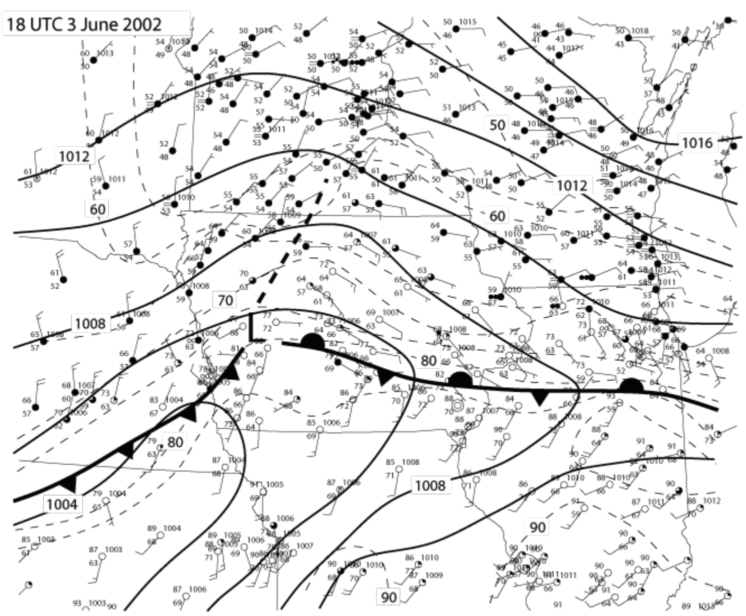


Fig. 4. Surface mesoanalysis for 1800 UTC 3 June 2002. Station data are plotted according to standard station model and fronts are analyzed using standard symbols. Solid lines are isobars in hPa and dashed lines are isotherms every 5° F. Thick dashed lines denoted troughs and wind shift lines.

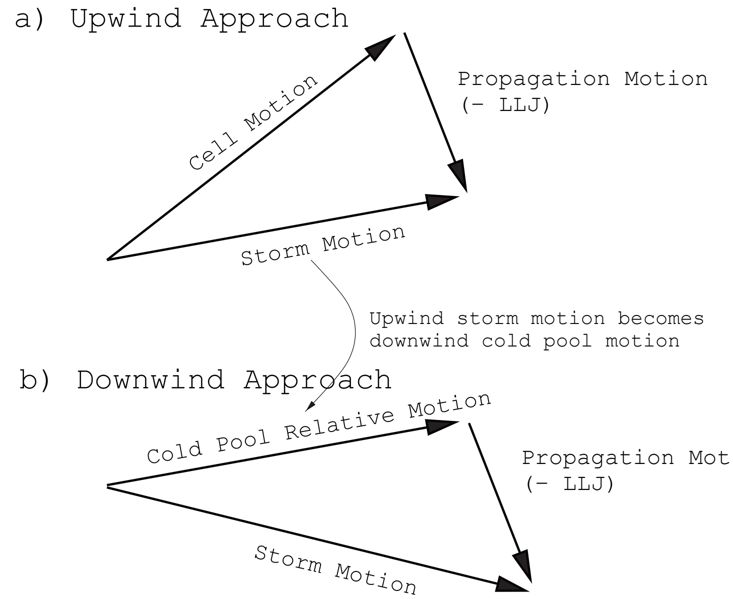


Fig. 3. An illustration of the Corfidi vector technique for both a) upwind and b) downwind approximations (adapted from Corfidi 2003).

3. Synoptic-Scale Environment

At 1800 UTC 3 June the surface analysis (Fig. 4) revealed a cold front trailing southwestward from a low in southwestern Iowa and a quasi-stationary front extending eastward from the low into northern Indiana. The thermal gradient across the stationary front was quite strong, especially across Iowa where temperatures ranged from the low 60s (°F) north of the front to the middle 80s (°F) south of the front. The weak cyclonic circulation still remained anchored in southwest Iowa at 2100 UTC (Fig. 5) along with two inverted troughs to the north of the low in central Iowa; one in west-central Iowa and another in eastern Iowa. At this time, two outflow boundaries were identified by radar, one in northern Illinois associated with MCS #1 (see discussion in section 4a) and another in east-central Iowa associated with the new convection of MCS #2 (section 4b).

Based upon radar analysis, most of the heavy convective rainfall appeared to fall after 0000 UTC 4 June 2002, therefore the rest of the discussion of the synoptic environment attending this case will focus upon this time. Surface conditions at 0000 UTC 4 June (Fig. 6) were similar to 3 h earlier, except for the movement of the two previous outflow boundaries. The westernmost outflow boundary in Fig. 5 in Iowa moved east-southeast into northwest Illinois, while the east-west outflow boundary in Illinois moved slightly to the south over the last 3 h. In addition, the second inverted trough in east-central Iowa had dissipated by 0000 UTC 4 June. The stationary boundary to the east of the weak low pressure

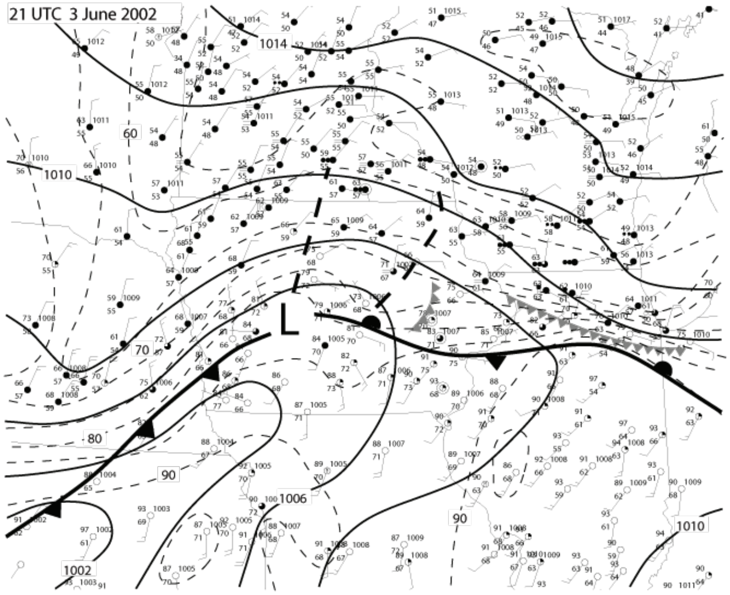


Fig. 5. Same as Fig. 4 except for 2100 UTC 3 June 2002 and with outflow boundaries depicted using small frontal pips.

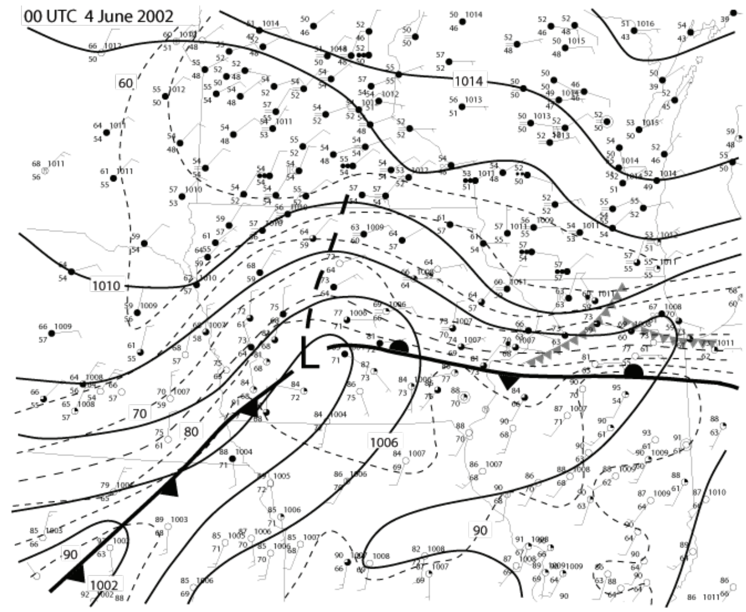


Fig. 6. Same as Fig. 5, except for 0000 UTC 4 June 2002.

in southwestern Iowa was a focus for strong surface moisture convergence (Fig. 7) where values of greater than 3.0 g (kg-h)$^{-1}$ were diagnosed in northwestern Illinois. During the previous two hours (2200 and 2300 UTC), a surface moisture convergence maximum of over 3.5 g (kg-h)$^{-1}$ was located in south-central Iowa, thereby confirming the spatial and temporal continuity of the maximum seen in Fig. 7. Dew points in the low 70s (°F) were found just southwest of the weak cyclone, nearly coincident with the warmest surface air. The surface equivalent potential temperature (θ_e) field for 0000 UTC 4 June (Fig. 8) reveals values over 350 K over a broad area to the south of the surface front, with a maximum of 355 K in southern Iowa. Later, between 0000 and 0400 UTC 4 June, surface air cooled by Lake Michigan was carried inland over northeast Illinois on northeasterly winds, strengthening the baroclinic zone over northeast Illinois.

Analyses of the 850, 500, and 250 hPa surfaces (Fig. 9) reveal important clues related to the environment supportive of convective development. At 850 hPa (Fig. 9a) a weak inverted trough was found from southeast Colorado northeastward into Minnesota. Warm (temperatures greater than 20°C), moist (dew points greater than 15°C) air was being advected into southwest Iowa by southwesterly flow of about 10 m s^{-1} (~20 knots). An 850 hPa frontal boundary can be identified in central Iowa extending into northern portions of Illinois and Indiana. The 500 hPa flow (Fig. 9b) reveals a broad ridge of west-southwesterly flow dominating the north-central Plains states with a weak trough well upstream from the incipient convection. As noted by many authors (e.g., Maddox et al. 1979; Moore et al. 2003) heavy convective rainfall often occurs near the inflection point in the mid-tropospheric flow, in a region of weak positive to neutral absolute vorticity advection. At 250 hPa, the objective analysis (Fig. 9c) diagnosed a similar flow as at 500 hPa, with a weakly anticyclonically-curved ULJ streak over the Dakotas extending into Minnesota. This would place the Iowa heavy convective rainfall event on the anticyclonic side of the ULJ, south of the maximum mean 300-200 hPa divergence (Fig. 10). Junker et al. (1999) and Moore et al. (2003) have noted that heavy convective rainfall events are frequently located near the southern extent of the maximum upper-level divergence in their composite studies, confirming the reliability of this signature.

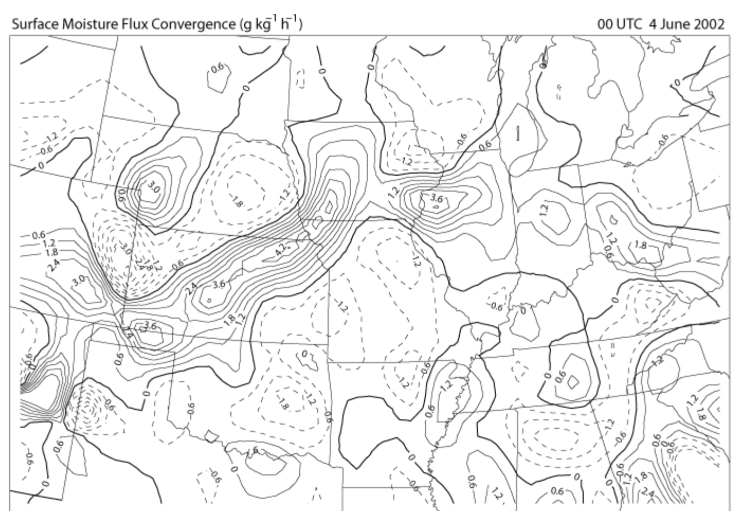


Fig. 7. Surface moisture flux convergence for 0000 UTC 4 June 2002 in units of g (kg-h)$^{-1}$. Dashed lines indicate moisture flux divergence while solid lines indicate moisture flux convergence.

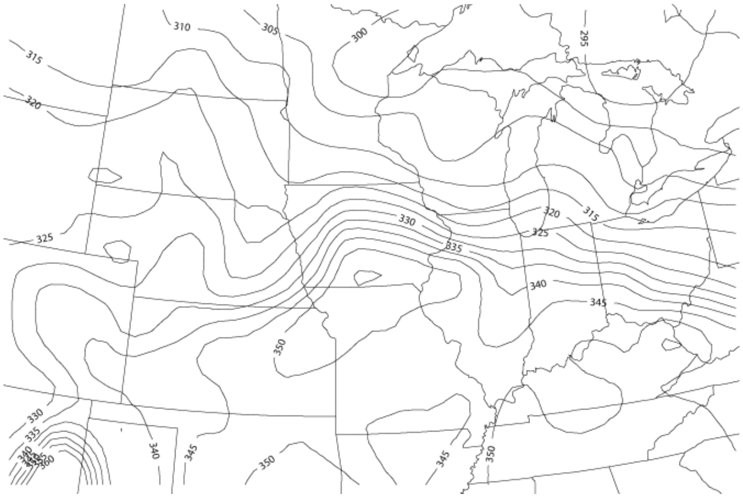
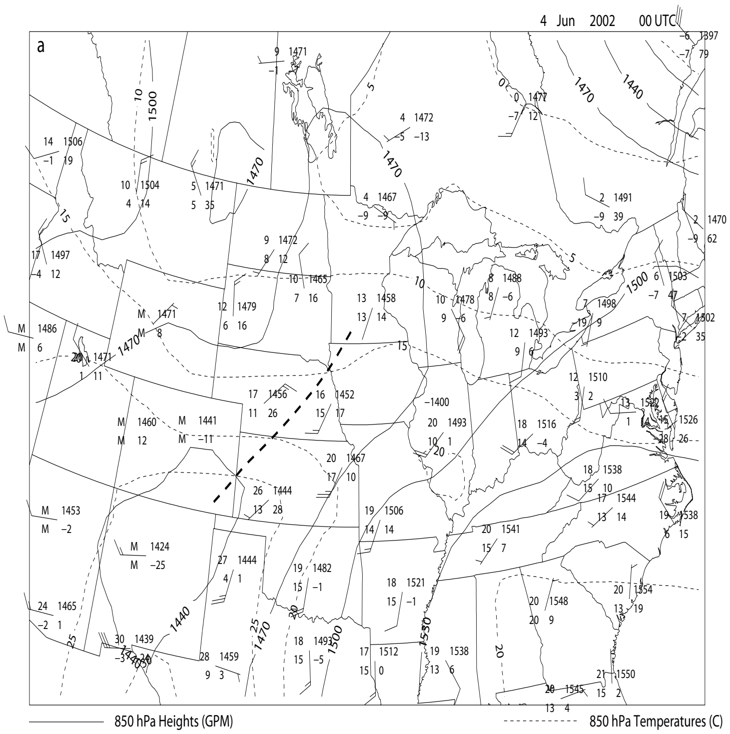


Fig. 8. Surface equivalent potential temperature (θ_e) in K for 0000 UTC 4 June 2002.

The 900-700 hPa average two-dimensional frontogenesis field computed following Petterssen (1956) and derived from rawinsonde data reveals two maxima; one centered in southeast South Dakota with an axis extending into central Iowa, and another in northwest Kansas (Fig. 11). The axis of the former maximum is located approximately 200 km north of the surface-based boundary and is associated with a direct thermal circulation, shown in Fig. 12a, by the cross section of tangential ageostrophic winds and kinematic vertical motions taken along the 91° longitude meridian cutting through eastern Iowa (as indicated in Fig. 11). Figure 12b displays an analysis of θ_e and relative humidity along the same cross section. It reveals an axis of high θ_e and relative humidity sloping upward to the north above the frontal zone, seen as a layer of strong convective stability.

Unfortunately, the rawinsonde for 0000 UTC 4 June from the NOAA/National Weather Service (NWS)



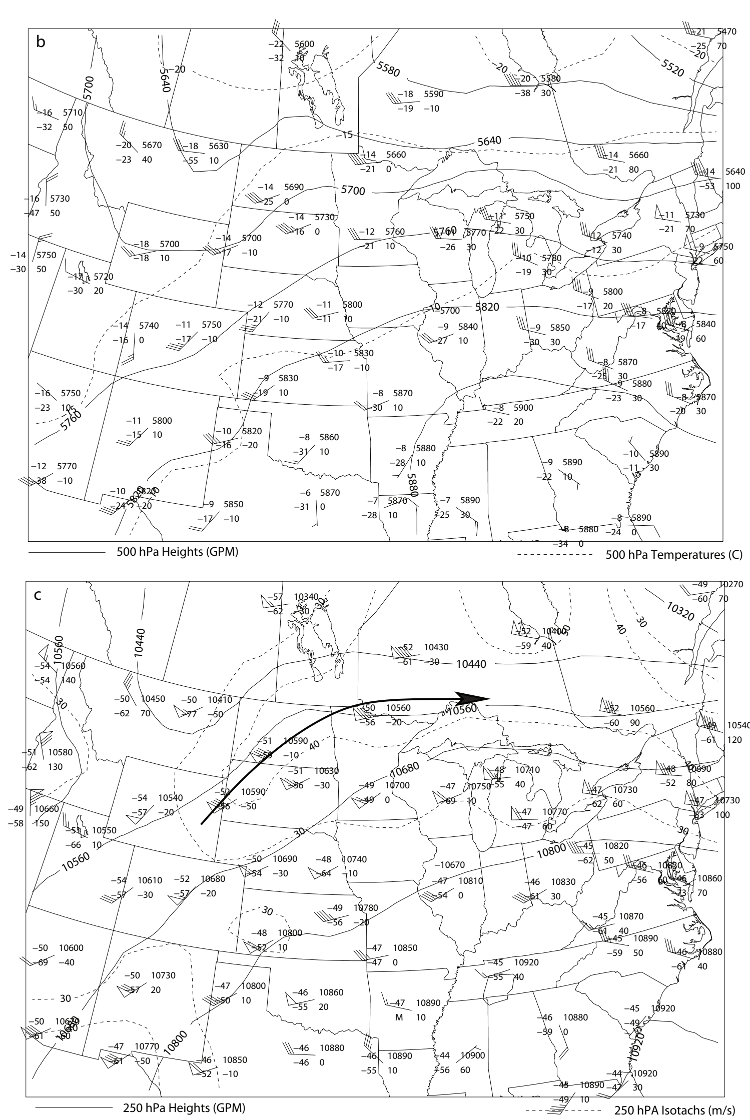


Fig. 9(a-c). Objectively-analyzed upper-level charts for 0000 UTC 4 June 2002.

(a) 850 hPa, where solid lines are isohyets (every 30 gpm) and dashed lines are isotherms (every 5° C); thick dashed line indicates a trough axis,
(b) 500 hPa, where solid lines are isohyets (every 60 gpm) and dashed lines are isotherms (every 5° C), and
(c) 250 hPa, where solid lines are isohyets (every 120 gpm), dashed lines are isotachs (every 10 m s^{-1}), and the bold arrow is the jet axis.

Weather Forecast Office (WFO) in Davenport, Iowa (DVN) prematurely ended at 733 hPa, probably because it encountered strong convection during ascent. However, it is instructive to look at the inflow air approximated by the WFO Topeka, Kansas (TOP) sounding (Fig. 13) as it represents the air mass being advected into east-central Iowa. The sounding had high precipitable water (PW; 1.73 inches) and a deep layer of instability (maximum θ_e CAPE of 4401 J kg^{-1}). The inflowing PW values in northeastern Kansas were at least 130% of normal for this time of year (early June). However, two things likely prevented convection in northeast Kansas; the absence of a boundary along which to focus convection, and a substantial maximum θ_e CIN of 175 J kg^{-1} (note the presence of two small inversions at ~830-800 hPa and ~765-755 hPa). The plan view of maximum θ_e CAPE (Fig. 14) depicts a maximum over 4000 J kg^{-1} in northeast Kansas with decreasing values to the northeast. Interestingly, the WFO Lincoln, Illinois sounding (not shown) was much drier than the TOP sounding resulting in a maximum θ_e CAPE of only 1603 J kg^{-1}. Thus, the objectively analyzed plan view of maximum θ_e CAPE is not representative of the conditions in eastern Iowa as the DVN sounding was not included in the objective analysis.

From the preceding discussion it can be seen that the precursor conditions for heavy convective rainfall were present in eastern Iowa for this event – moisture, lift, and instability. High PW and low-level θ_e values were part of the inflow into the area. Lift was present in the form of a direct thermal circulation associated first with moderate low-level frontogenesis and later, as will be shown, by outflow from subsequent thunderstorm activity. Lastly, high values of instability (maximum θ_e CAPE greater than 4000 J kg^{-1}) were streaming northeastward along and over the frontal zone.

4. Mesoscale Environment

Archive level II data captured from the WSR-88D radar at DVN and the WSR-88D Algorithm Testing and Display System (WATADS) developed at the NOAA/ National Severe Storms Laboratory (NSSL) were utilized to analyze the MCSs from this event. Analysis revealed that five significant MCSs affected the DVN County Warning Area (CWA), whose characteristics are described in Table 1. Note that the movements listed in Table 1 represent an average over the lifetime of the MCS.

a. MCS # 1

The first MCS formed approximately 100 km north of the stationary front over northwest Illinois around

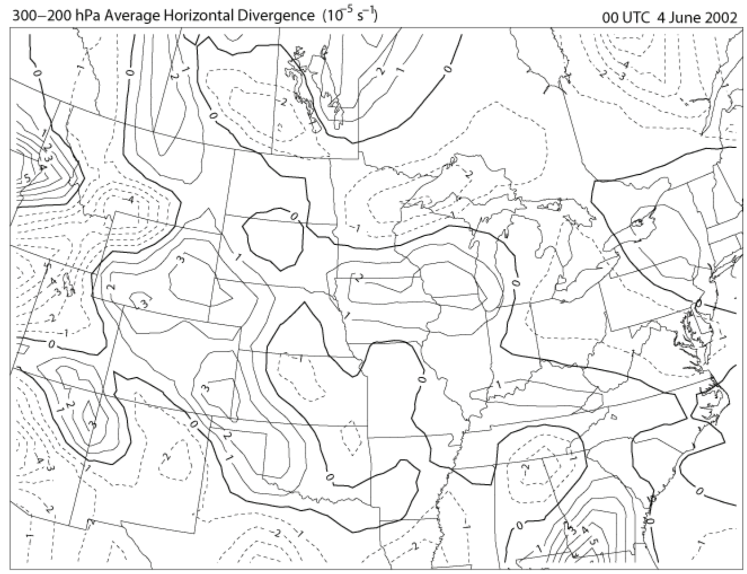


Fig. 10. 200-300 hPa average divergence x 10^{-5} s^{-1} (where solid lines indicate divergence and dashed lines indicate convergence) for 0000 UTC 4 June 2002.

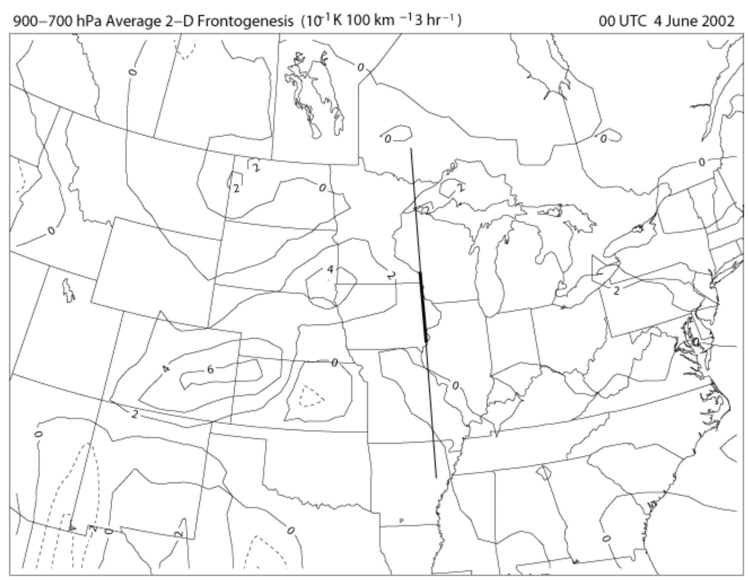
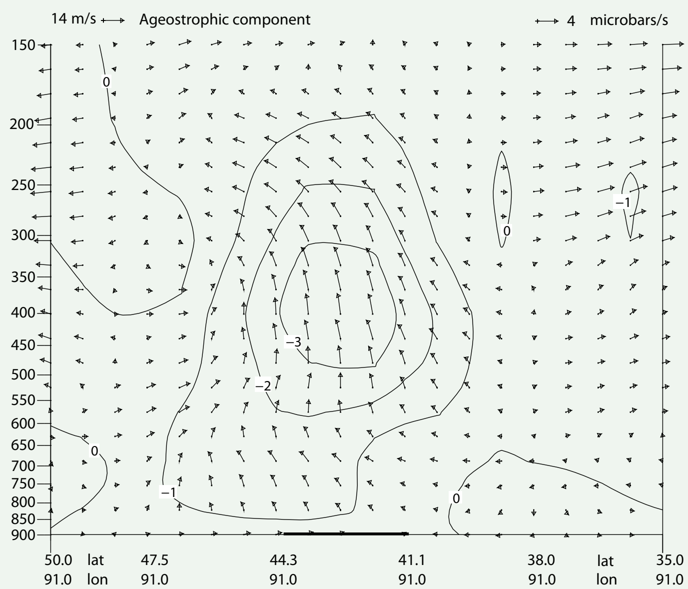


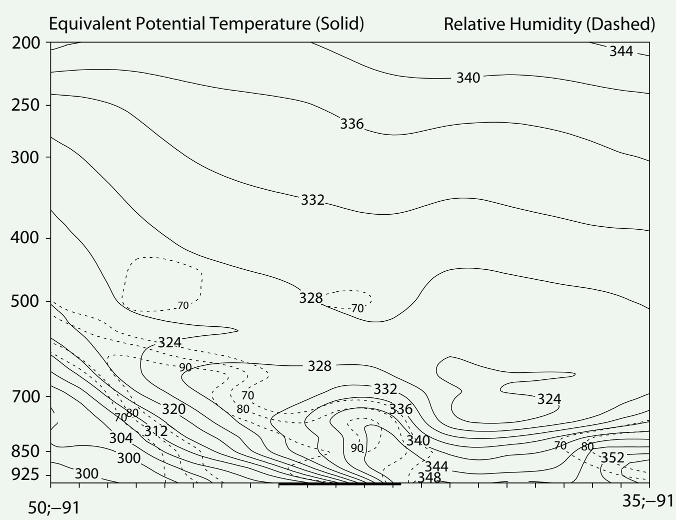
Fig. 11. 900-700 hPa average two-dimensional frontogenesis [10^{-1} K (100 km-3 h)$^{-1}$] for 0000 UTC 4 June 2002. Long solid line indicates location of cross section for Fig. 12. Thick portion of line denotes the region of upward vertical motion depicted in Fig. 12a.

1800 UTC 3 June 2002. It traveled eastward at about 13 m s^{-1}, just south of the Illinois-Wisconsin border. This MCS was relatively small and displayed little organization throughout its lifetime (Fig. 15), therefore its contribution to the rainfall totals for the day was relatively minor. However, close inspection of the animated radar reflectivity field and surface observations revealed that by the end of its lifetime MCS #1 produced a weak but discernable outflow boundary in Illinois, oriented west-east just to the southwest of Lake Michigan (see Fig. 6, surface analysis for 00 UTC 4 June). The second outflow boundary to the west was associated with MCS #2, which is described in more detail in the next subsection. MCS #1 developed north of the axis of maximum surface moisture convergence within a strong gradient of surface θ_e (see Figs. 7-8, respectively), which is typical for elevated thunderstorms (Moore et al. 2003; Banacos and Schultz 2004).

Fig. 12(a-b). Vertical Cross Sections 4 June 2002 00 UTC.



(a) Cross section of transverse ageostrophic circulation taken along the line indicated on Fig. 11 for 0000 UTC 4 June 2002. Solid lines are isopleths of vertical motion in :bars s-1.



(b) Cross section of θe (K; solid lines) and relative humidity (%; 70% or greater, dashed lines) taken along the line indicated in Fig. 11 for 0000 UTC 4 June 2002.

Corfidi vectors were computed using the 850-300 hPa mean wind to estimate the cell motion and a wind vector equal and opposite of the 850 hPa wind to approximate the propagation vector. Although the mean wind vector was computed over the MCS location, the representative

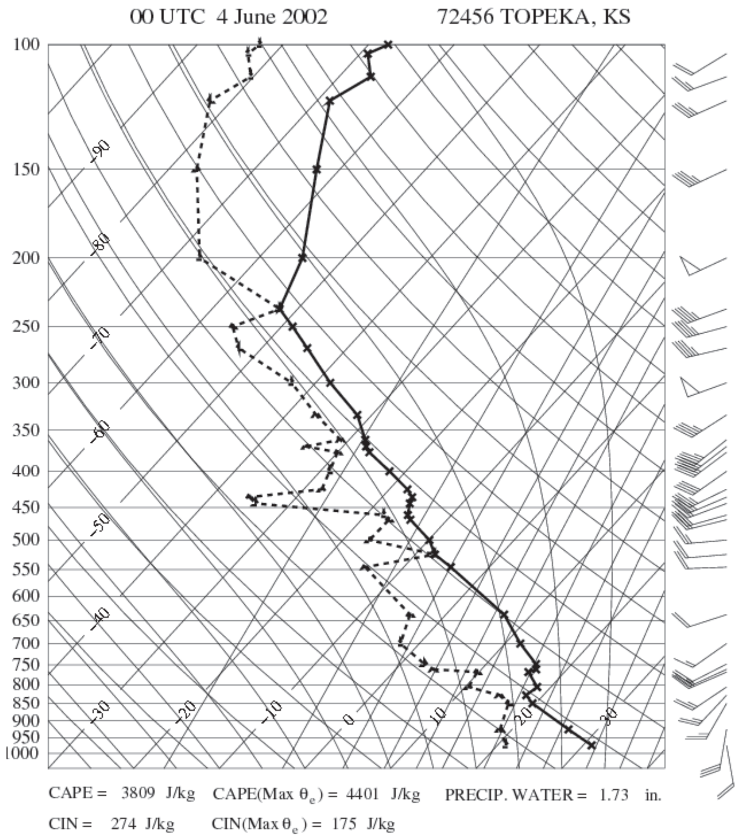


Fig. 13. Skew-t Log-P sounding for Topeka, Kansas for 0000 UTC 4 June 2002, along with parameters computed from this sounding.

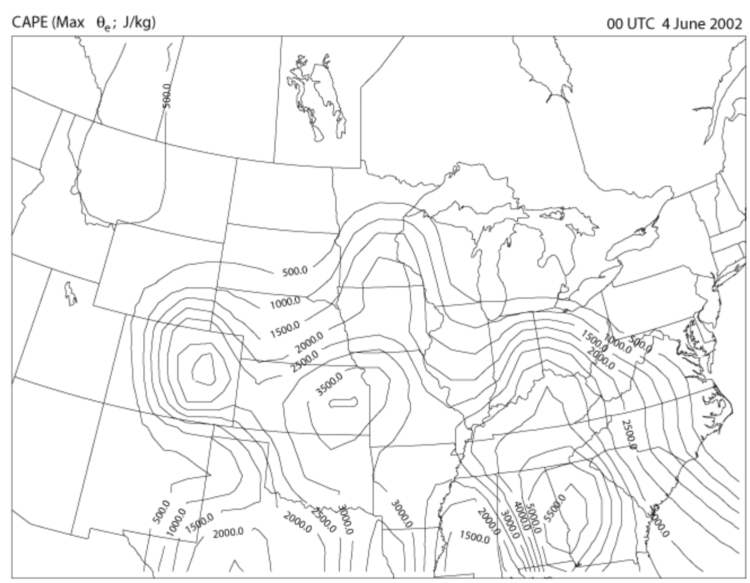


Fig. 14: Maximum-θe CAPE (J kg-1) for 0000 UTC 4 June 2002.

location for the propagation vector was chosen from a grid point approximately 100 km southwest (upstream) from the MCS. The eastward movement of this MCS was well predicted by the downwind Corfidi vector method (Fig. 16a, 2100 UTC Corfidi downwind estimate) at 2100 UTC which estimated a storm motion of 272° at 17.5 m s^{-1}. A diagnostic propagation vector, computed by estimating the cell motion from the WATADS display over a one hour period and using infrared (IR) satellite imagery to compute storm motion, is shown in Fig. 16b. Note that the *actual* storm motion was 281° at 9.8 m s^{-1}, revealing that the predicted storm motion was too fast and more westerly than observed. This is due to the fact that the *actual* propagation vector was directed to the southwest (upstream) as opposed to the propagation vector estimate by the Corfidi method, which was decidedly downstream. It seems that the presence of high values of surface moisture convergence, θ_e, and CAPE located to the southwest played a significant role as to where new cells would develop, thereby reorienting the propagation vector.

MCS#	Formation Region/Time	Dissipation Region/ Time	Movement
1	Northwest IL, 18 - 19 UTC, 3 June	Northeast IL, 23-00 UTC, 3-4 June	East at 13.3 m s^{-1}
2	East-Central IA, 20 - 21 UTC, 3 June	Northeast IL, 03-04 UTC, 4 June	East at 17.8 m s^{-1}
3	Northeast IA, 01-02 UTC, 4 June	Northeast IL, 09-10 UTC, 4 June	Southeast at 13.3 m s^{-1}
4a	Northeast IA, 06-07 UTC, 4 June	Northeast IL, 17-18 UTC, 4 June	Southeast at 13.3 m s^{-1}
4b	East-Central IA, 14-15 UTC, 4 June	Northeast IL, 19-21 UTC, 4 June	East at 17.8 m s^{-1}

Table 1. List of MCSs observed by WSR-88D during the heavy rainfall event near Davenport, Iowa.

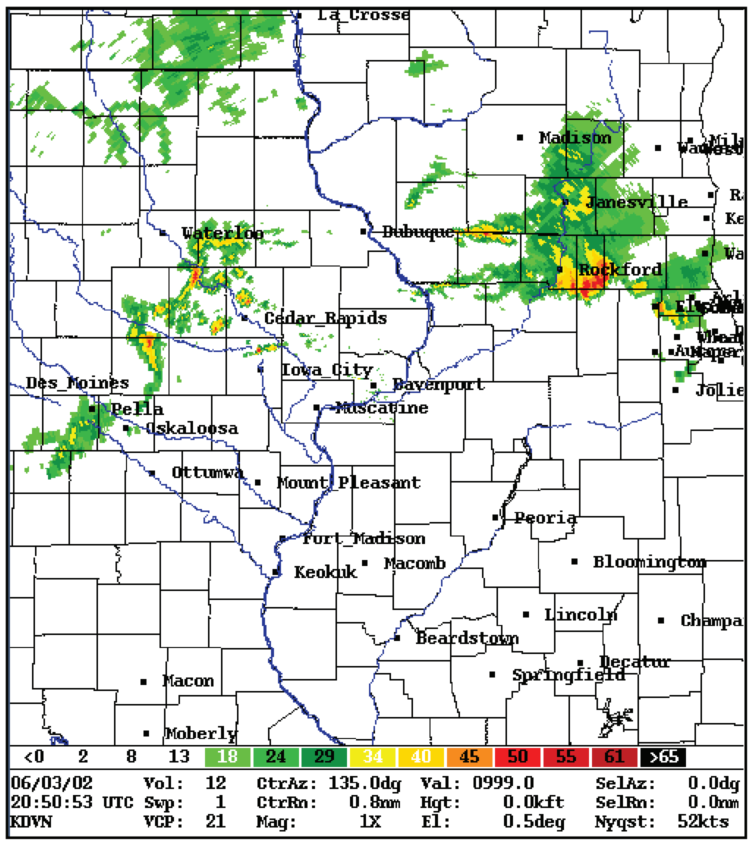


Fig. 15. WSR-88D 0.5° elevation reflectivity from Davenport, Iowa for 2051 UTC 3 June 2002.

b. MCS # 2

MCS #2 also initiated approximately 100 km north of the quasi-stationary front around 2000 UTC 3 June, along a weak wind shift line that extended from the stationary front to northeast Iowa (see Fig. 3). This wind shift line separated dominantly easterly flow in northern Illinois, Wisconsin, and eastern Iowa from weak northeasterly flow in north-central Iowa. As with MCS # 1, MCS #2 formed on the cool side of the surface θ_e gradient northeast of the maximum surface θ_e value of 355 K located in southern Iowa. The MCS grew spatially and in intensity as it moved eastward nearer the cold pool/outflow boundary created by MCS #1. The mesoscale support for MCS #2 was stronger and deeper than with MCS #1. Surface moisture convergence values increased to over 4.0 g (kg-h)$^{-1}$ by 2200 UTC 3 June in south-central Iowa (Fig. 17a). In addition, moisture convergence averaged over the 950-850 hPa layer (Fig. 17b; RUC-II analysis) also increased throughout south-central Iowa from 0.8 to 1.2 g (kg-h)$^{-1}$ during the same time period, indicating a deep layer of moisture convergence to support convection. Average θ_e advection over the 950-850 hPa layer reveals that the MCS activity was focused along an axis of maximum positive θ_e advection (Fig. 18; RUC-II analysis).

MCS #2 moved through east-central Iowa into northern Illinois, evolving into an asymmetric squall line (Hilgendorf and Johnson 1998, Parker and Johnson 2000) with an elongated line of intense convection with reflectivities exceeding 60 dBz in northwestern Illinois

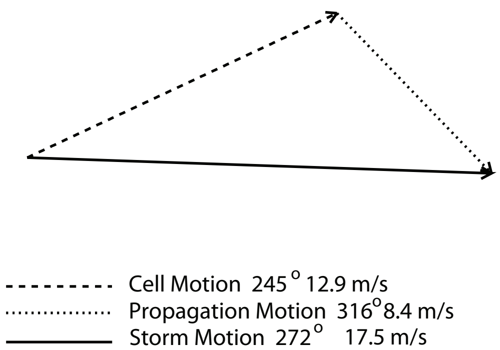


Fig. 16(a). Downwind Corfidi vector method for 2100 UTC 3 June 2002.

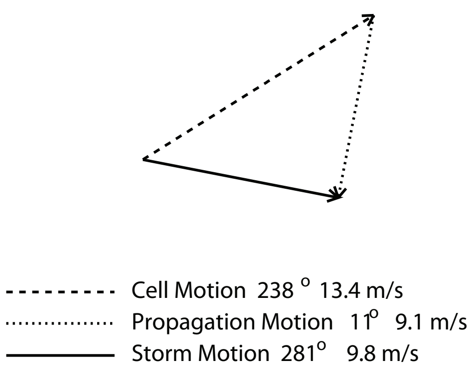


Fig. 16(b). Observed cell, propagation, and storm motion vectors for the same time.

and a large stratiform region of precipitation in southern Wisconsin (Fig. 19). This MCS eventually dissipated in northeastern Illinois around 0400 UTC 4 June. Over its 8 h lifespan MCS #2 moved to the east at an average speed of 17.8 m s^{-1}, similar to MCS #1, but was larger and more organized than MCS #1.

The downwind Corfidi vector method predicted an MCS motion of 270° at 16.5 m s^{-1} at 2300 UTC 3 June (Fig. 20a). This compares favorably to the actual MCS motion of 274° at 19.5 m s^{-1} (Fig. 20b) as analyzed from the IR satellite imagery. Although MCS #2 was stronger and more widespread than MCS #1, its contribution to the total precipitation was likely limited due to its fast forward motion.

c. *MCS # 3*

MCS #3 formed between 0100-0200 UTC 4 June to the northwest of the outflow boundary laid down by MCS #2, very close to the initiation region of MCS #2. This MCS developed into an extensive west-east band of convection from north-central Iowa into extreme northwest Illinois and southwest Wisconsin (Fig. 21). MCS #3 was oriented approximately parallel to the large-scale frontal boundary, in contrast to MCSs #1 and #2, which were more normal to the boundary. Apparently, the outflow boundaries of MCSs #1 and #2 collectively acted to reinforce the cool boundary-layer air north of the frontal boundary in eastern Iowa and northern Illinois, moving it about 50 km south, especially in Illinois, as seen in the surface mesoanalysis for 0200 UTC 4 June (Fig. 22). It is likely that this enhanced boundary layer thermal gradient provided the additional lift that air parcels needed to reach their lifting condensation level (LCL) and eventual level of free convection (LFC) for convection to initiate. Also, the orientation of the outflow boundary helped to organize the convection along a broader west-east axis than found in the previous two MCSs. In this way the first two MCSs enhanced and restructured the mesoscale environment, favoring a larger outbreak of convection oriented in a west-east fashion, thereby illustrating the non-linearity of convective processes that are so difficult to capture in a numerical model. The west-east orientation of the storm complex coupled with the motion of the individual cells to the east (following the mean cloud-layer winds), resulted in locally excessive rainfall accumulation (Doswell et al. 1996). In addition, the consequence of the first two MCSs was likely to moisten the atmosphere, thereby increasing the PE of the latter or third MCS.

A deep layer of moisture convergence over eastern Iowa helped to sustain and strengthen MCS #3. Evidence of this can be seen in Fig. 23a-b, which displays the surface and RUC-II 950-850 hPa average moisture convergence, respectively, at 0500 UTC. Strong positive values of moisture convergence greater than 2.0 g (kg-h)$^{-1}$ can be seen over eastern Iowa. In fact, the 950-850 hPa average moisture convergence of 2.0 g (kg-h)$^{-1}$ axis is oriented west-east and covers most of central and eastern Iowa. Moore et al. (2003) and Junker et al. (1999) have noted that training of convection is favored under conditions when the moisture convergence axis is parallel to a quasi-stationary boundary. Junker et al. (1999) has also noted that training is more likely when the mean cloud-layer flow (approximately 850-300 hPa) is nearly parallel to the frontal boundary with a small component of motion toward the cold side of the front. The mean cloud-layer flow at this time (Fig. 24; RUC-II analysis) show that the flow was decidedly from the southwest at about 25-30 knots, directed across the frontal boundary. Having the mean flow cross the frontal boundary at an angle acts to slow down the advance of outflow boundaries, thereby "locking in" the zone of moisture convergence associated with the MCSs.

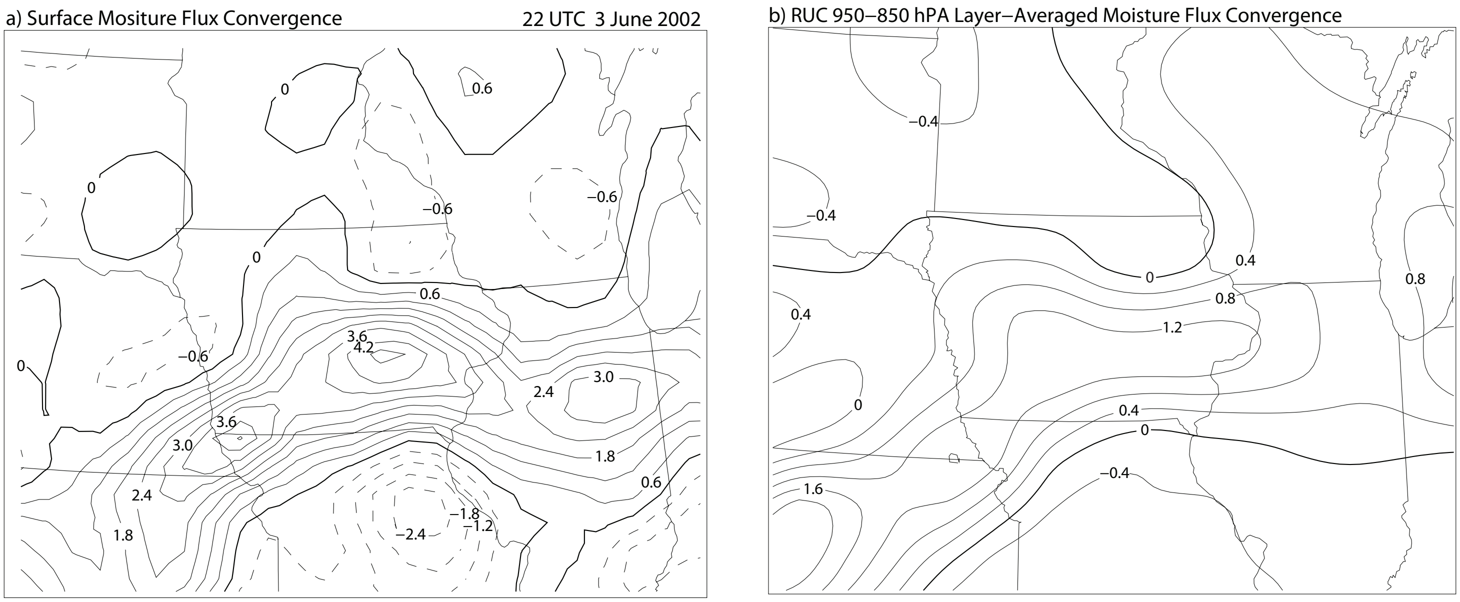


Fig. 17(a). Same as Fig. 7, except for 2200 UTC 3 June 2002. (b) 950-850 hPa layer-averaged moisture flux convergence for 2200 UTC 3 June 2002 from RUC-II analysis; units are 10^{-1} g $(kg\text{-}h)^{-1}$.

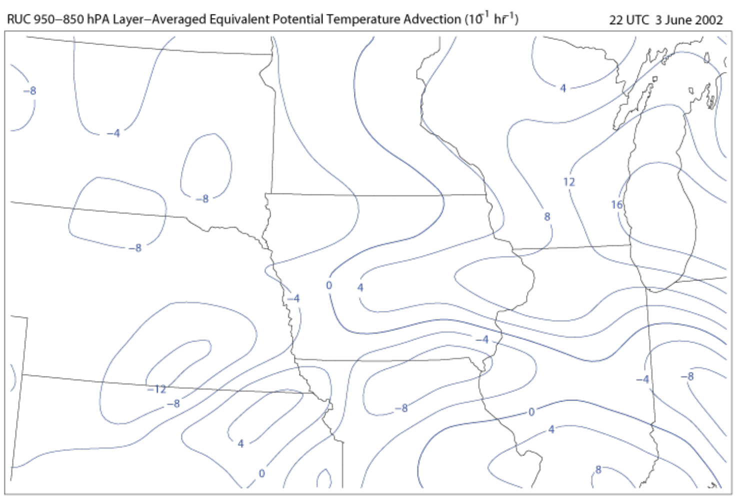


Fig. 18. Layer-averaged 950-850 hPa θe advection for 2200 UTC 3 June 2002 from RUC-II analysis in units of 10-1 K h-1.

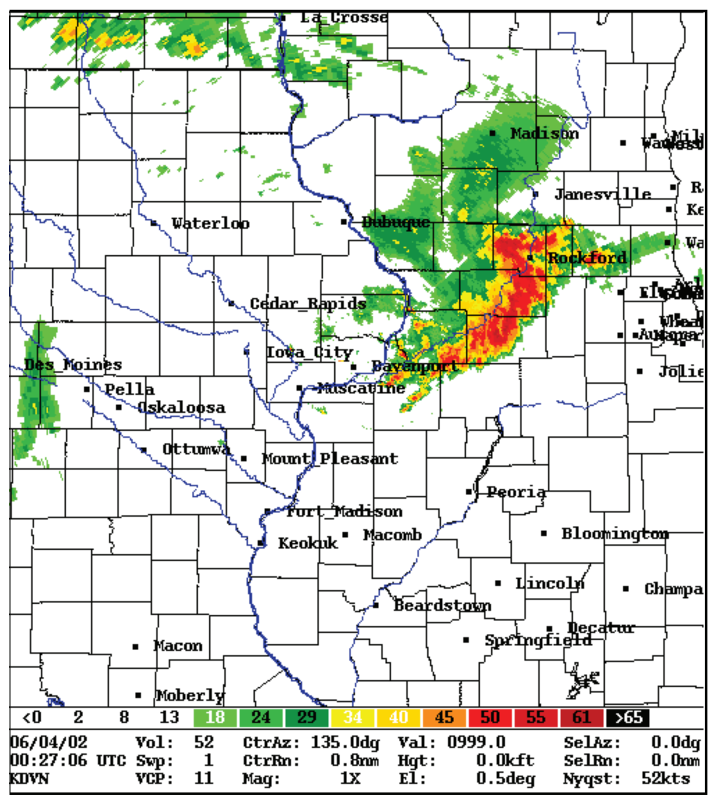
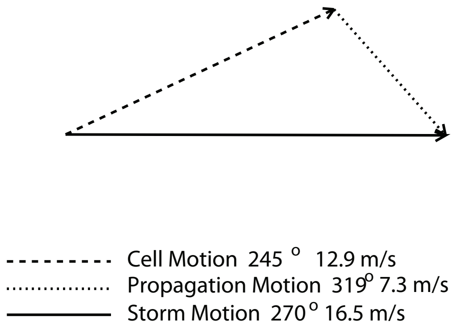


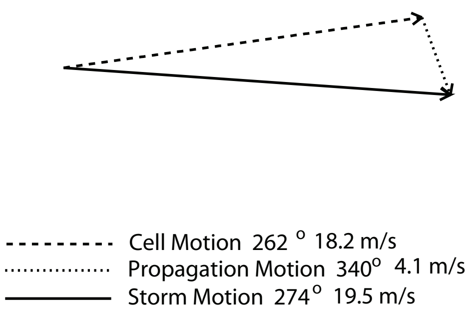
Fig. 19. Same as Fig. 15, except for 0027 UTC 4 June 2002.

Another influence on the MCS activity at this time was the changing character of the LLJ. Many researchers have noted the importance of the LLJ in transporting high-θ_e air into the storm environment and creating low-level moisture convergence downstream from the wind maximum. Figure 25 shows the LLJ at 850 hPa at 0300, 0600, 0900, and 1200 UTC 4 June, as determined from RUC-II analysis data. One can see a steady increase in wind speed from 0300-0900 UTC from about 30 knots to 40 knots with a slight decrease in wind speed from 0900 to 1200 UTC to about 35 knots. Further, over this 12 h period inertial veering of the LLJ is apparent as the wind direction over eastern Kansas (the upstream flow for Davenport, Iowa) changes from 213° to 254°. This narrowly focused and strengthening LLJ was instrumental in sustaining the convective cells associated with MCS #3 through its 9 h lifetime, contributing to the

Fig. 20(a-b). Same as Fig. 16, except for 2300 UTC 3 June 2002.



a) Downwind approximation



b) Observed cell, propagation, and storm motion vectors for the same time.

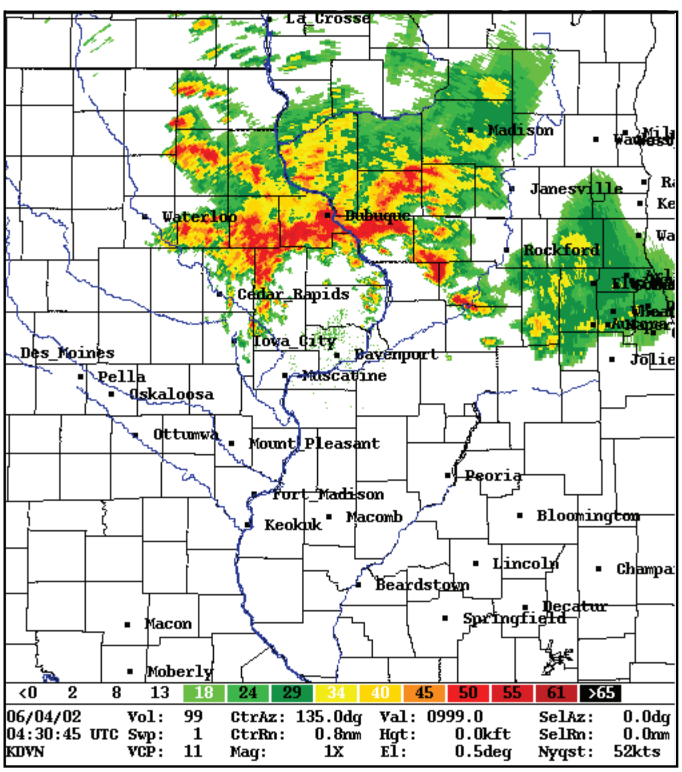


Fig. 21. Same as Fig. 15, except for 0431 UTC 4 June 2002.

extensive area of heavy rainfall. As MCS #3 moved toward the Davenport, Iowa area, several cells exhibited a low-centroid of high reflectivity (Fig. 26). Storms with high reflectivity (greater than 50 dBz) in the lower levels of the atmosphere (below 15,000 ft) tend to produce intense, heavy rainfall representative of high precipitation efficient thunderstorms (Pontrelli et al. 1999).

This intense, long-lasting MCS was observed to move from 297° at approximately 14.2 m s^{-1} (Fig. 27a). For this case, the downwind Corfidi vector method estimated the storm motion to be 278° at 12.2 m s^{-1} (Fig. 27b). Thus, the speed was reasonably close but the "predicted" direction was shifted 19° counterclockwise from the observed direction. The upwind Corfidi vector method estimated the storm motion to be about 353° at 8.3 m s^{-1} (Fig. 27c). Thus, it had the right idea concerning the southward motion, but it underestimated the speed by about 6 m s^{-1}. An inspection of the surface mesoanalysis at 0500 UTC (Fig. 28) provides some clues as to why neither Corfidi vector method was correct for this time period. Note the presence of a significant mesoscale ridge extending from extreme southwest Wisconsin into east-central Iowa. This mesoscale ridge developed as pressure rises of ~2 hPa occurred in the vicinity of Dubuque, Iowa. To the west and east of the mesoscale ridge, inverted troughs of low pressure can be found. Thus, an isallobarically-driven outflow was directed from the pressure ridge to the southwest and southeast thereby enhancing the mesoscale convergence along a region west and east of Davenport, Iowa. This mesoscale ridge began to build around 0400 UTC and was a recognizable feature through

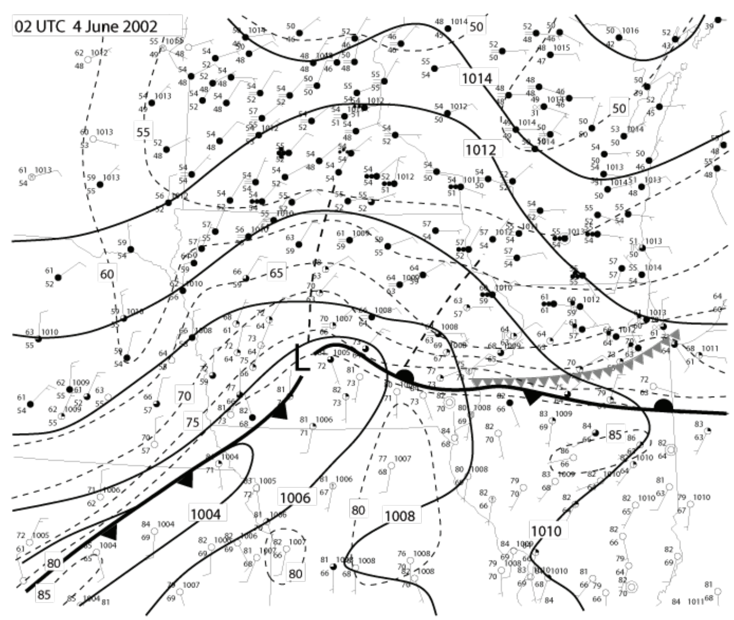


Fig. 22. Same as Fig. 5, except for 0200 UTC 4 June 2002.

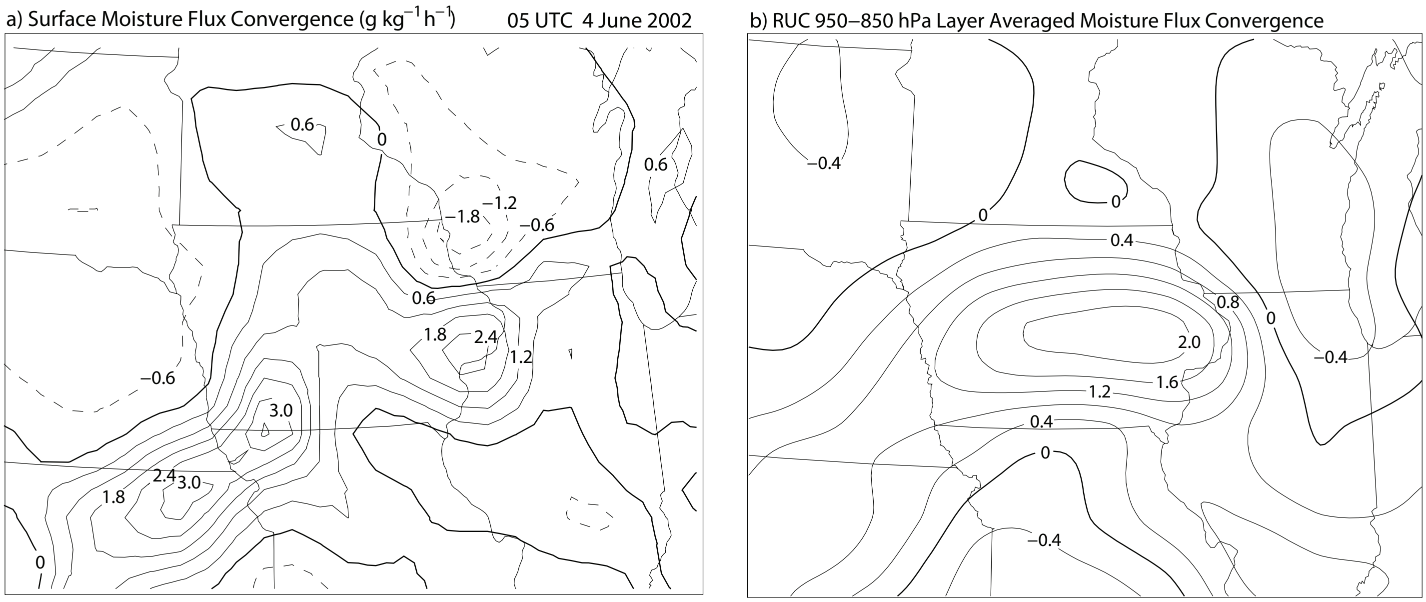


Fig. 23. Same as Fig. 17, except for 0500 UTC 4 June 2002.

0700 UTC. Thus, new convective cell development was to a large degree dictated by the evolution of this mesoscale convergence zone and to a lesser degree by the LLJ which was gradually veering during this time period. So, the actual propagation vector seen in Fig. 27a is directed to the southwest. Due to its shallow nature the low-centroid storm did not move with the average 850-300 hPa winds, a key assumption in the Corfidi vector method. Secondly, one must also use the Corfidi vector method with caution when convective outflow boundaries and mesoscale surface pressure perturbations alter the low-level wind to enhance convergence in a region that may be different from what the LLJ may infer.

d. MCS # 4a-b

MCS #4 was comprised of two components, one initiating around 0600 UTC and the other around 1400 UTC 4 June. Since these two MCSs were in close proximity to each other they were designated MCS #4a and MCS #4b. MCS #4a formed in northeast Iowa around 0600 UTC 4 June, while MCS #4b formed around 1400 UTC 4 June in east-central Iowa. An interesting aspect of these two MCSs is how long they lasted. Normally, nocturnal MCSs dissipate by late morning. However, both of these MCSs continued into the early afternoon hours of 4 June.

MCS #4a initiated as a cluster of cells near the east-central Minnesota-Iowa border, a considerable distance to the northwest of MCS #3. It expanded as it moved south, with new cells forming along the western periphery, thereby exhibiting backward propagation eventually forming a west-east line of strong convection in eastern Iowa (Fig. 29). Backbuilding of convection was undoubtedly influenced by the LLJ which was basically from the west-southwest at 35 knots at 1200 UTC (see Fig.

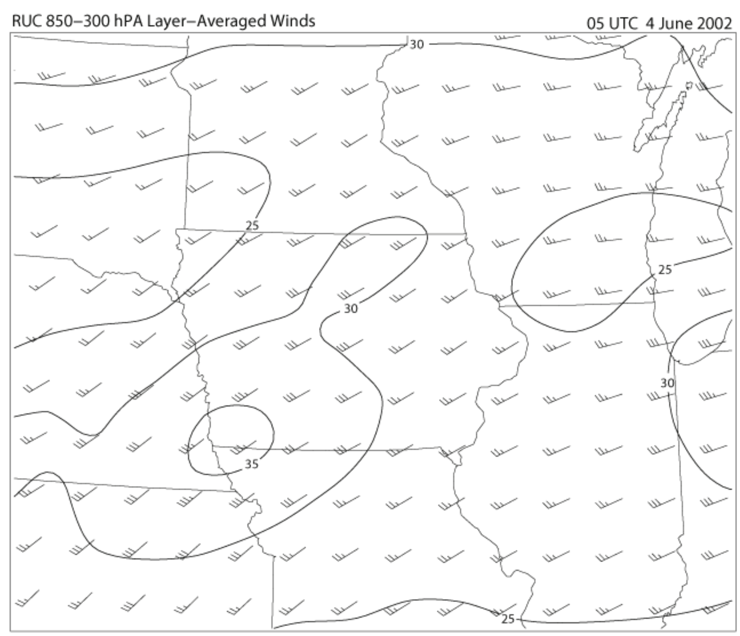


Fig. 24. RUC-II analysis 850-300 hPa layer-averaged winds at 0500 UTC 4 June 2002. Solid lines are isotachs for every 5 knots.

25d). The 950-850 hPa average moisture convergence field, early in the development of MCS #4a was oriented primarily west-east (see Fig. 23b). However, with time the axis of strong low-level moisture convergence evolved into a predominately north-south axis (Fig. 30; RUC-II analysis). The surface moisture convergence field reflected this change in orientation as well (not shown). Thus, as MCS #4a moved through the DVN CWA, the convective line orientation became more northeast-southwest. The 950-850 hPa average θ_e advection for 1100 UTC (Fig. 31: RUC-II analysis) also reveals a distinct north-south axis with the first signs of drier air being transported into western Iowa consistent with negative θ_e advection. Eventually MCS #4a dissipated in northeast Illinois around 1800 UTC.

Through WSR-88D radar imagery, MCS #4a was observed to move to the southeast at approximately 13 m s^{-1}. Early in the evolution of MCS #4a (from 0700-0900 UTC), the downwind Corfidi method generally predicted a storm motion from 283° at 13.5 m s^{-1} (Fig. 32a). Later (from 1000-1700 UTC), the downwind Corfidi method predicted storm motion toward the northeast at increasingly unrealistic speeds approaching 33 m s^{-1} (not shown). However, the upwind Corfidi method for the same period (Fig. 32b) predicted backbuilding convection noted by the strong propagation vector pointing upstream, although there are significant directional errors in later time periods. It is difficult to say which Corfidi method was "correct" for MCS #4a, since there was evidence in the radar imagery of both forward and backward propagation. Animation of the DVN WSR-88D radar imagery revealed that while MCS #4a moved southeast into Illinois, new convection erupted in eastern Iowa in its wake.

MCS #4b formed around 1400 UTC to the northwest of Davenport, contributing not only to the flash flood threat, but also severe weather in the form of one inch hail, primarily over Delaware County, Iowa (NCDC 2002).

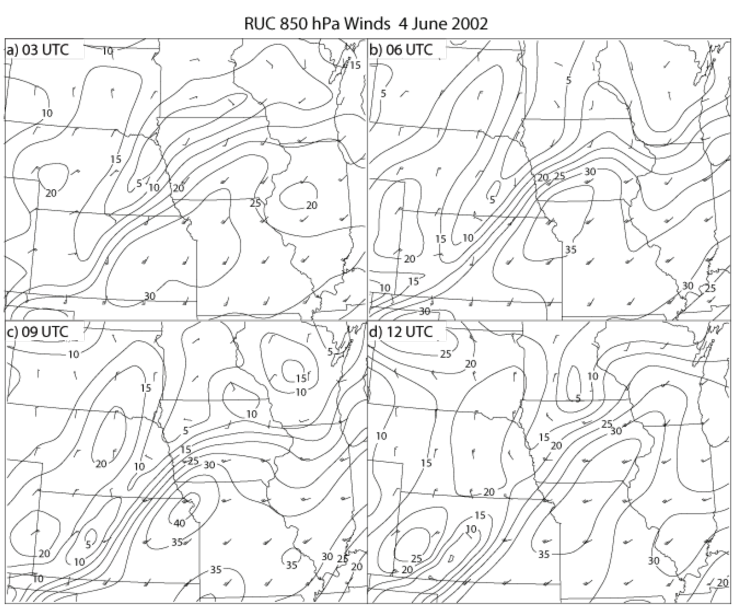


Fig. 25(a-d). RUC-II analysis 850 hPa winds for (a) 0300 UTC, (b) 0600 UTC, (c) 0900 UTC, and (d) 1200 UTC 4 June 2002. Solid lines are isotachs for every 5 knots.

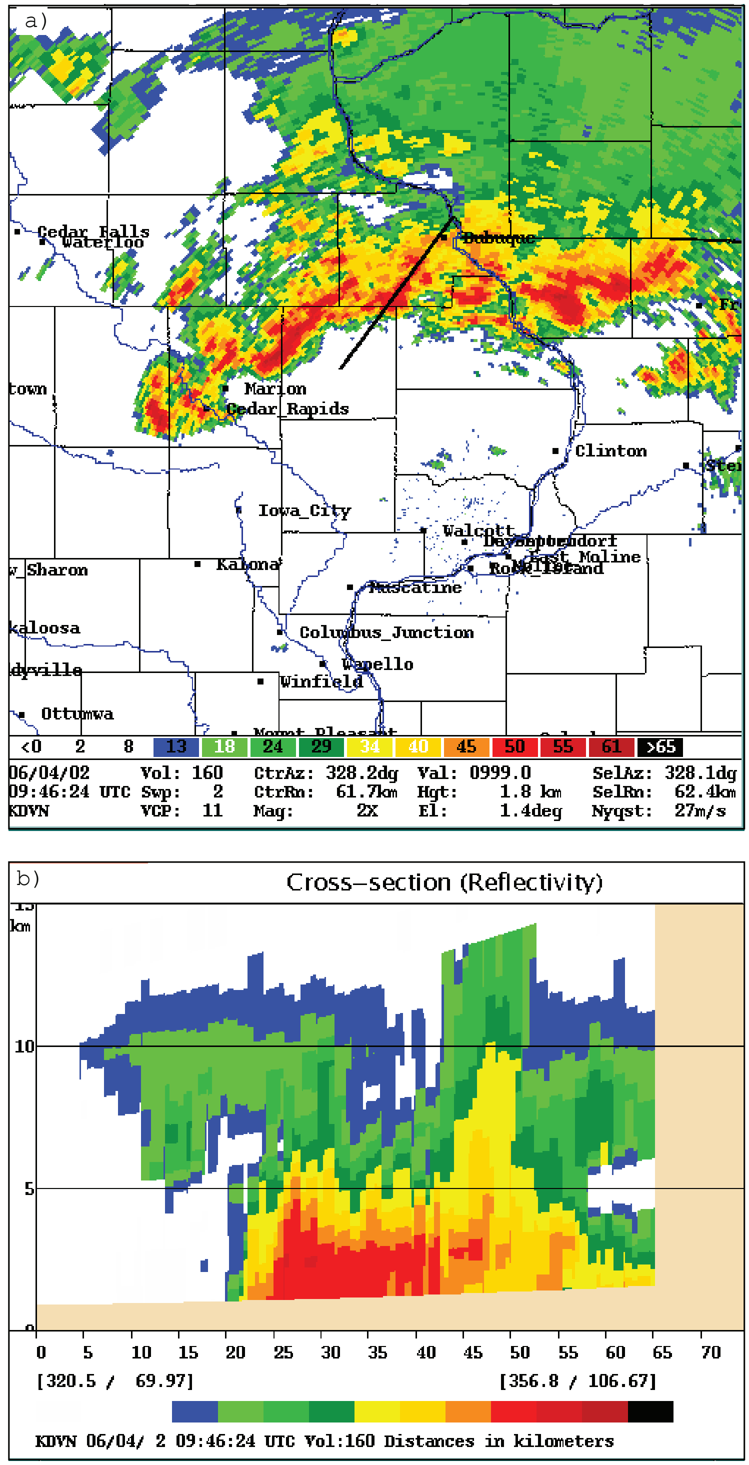
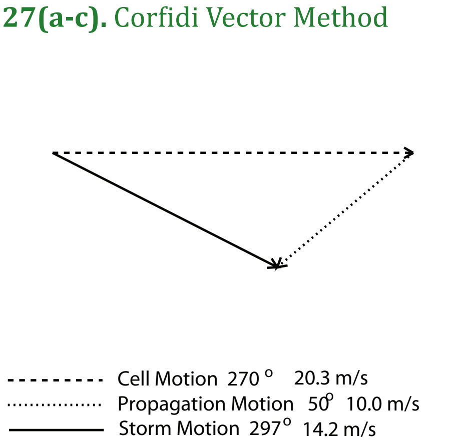
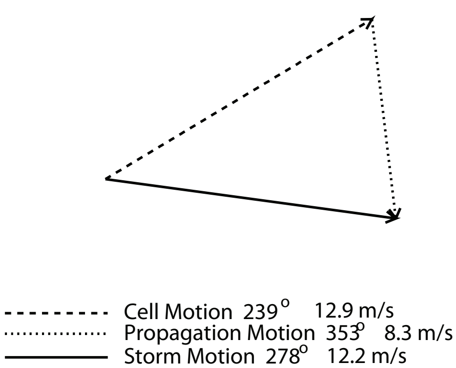


Fig. 26(a). Same as Fig. 15, except for 0946 UTC 4 June 2002. Solid line indicates location of cross section shown in (b). (b) Cross section of reflectivity along solid line shown in (a) indicating low-centroid cell echo.

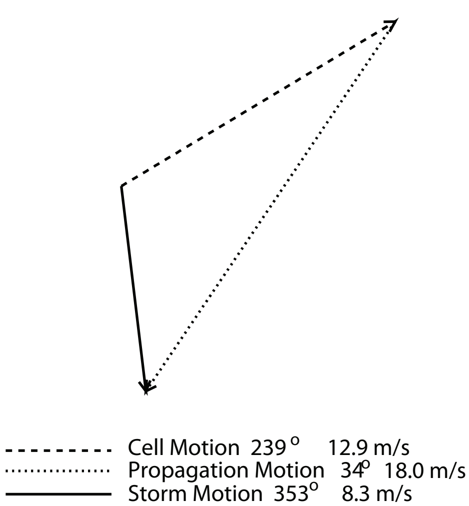
Fig. 27(a-c). Corfidi Vector Method



(a) Observed cell, propagation, and storm motion vectors for 0500 UTC 4 June 2002.



(b) Downwind Corfidi vector method for 0500 UTC 4 June 2002.



(c) Upwind Corfidi vector method for 0500 UTC 4 June 2002.

The WSR-88D radar reflectivity at 1400 UTC (Fig. 33a) indicates values in excess of 70 dBz. The reflectivity cross section taken across one of the severe convective cells reveals a classic structure of a hail-producing storm with reflectivities greater than 50 dBz extending to at least 35 000 feet with a core of 70 dBz values indicative of a hail shaft (Fig. 33b). This image is in stark contrast to the low-centroid, high precipitation efficient convective cell seen in Fig. 26b. It is likely that the dry air advection upstream from this MCS (see Fig. 31) invigorated the convective instability and lowered the height of the wet-bulb zero, thereby increasing the likelihood of hail production in several cells within this convective cluster.

As MCS #4b evolved into a primarily north-south oriented quasi-linear convective system (Fig. 34), new cells developed along the southwestern periphery (upwind) of the MCS. MCS #4b translated to the east at about 18 m s^{-1} and dissipated in northern Illinois. For this last MCS, the upwind Corfidi method would have been useful since it predicted upwind propagation; however, it also showed storm motion to the northeast at approximately 15 m s^{-1} (not shown); the downwind Corfidi vector method resulted in a storm motion that was twice as large (not shown). Thus, even in the last MCS neither method would have resulted in a good estimate of storm motion. Again, this is likely due to the fact that the LLJ, although an important parameter to consider for storm propagation, was not the only factor influencing new cell growth. For one thing, the background synoptic scale frontal system and associated cyclone finally began moving eastward after about 0700 UTC. Also, the cold pool in the wake of MCS #4a created a northeast-southwest mesoscale ridge of high pressure over northern Illinois (Fig. 35). This mesoridge acted to: 1) force the frontal boundary further south into north-central Illinois

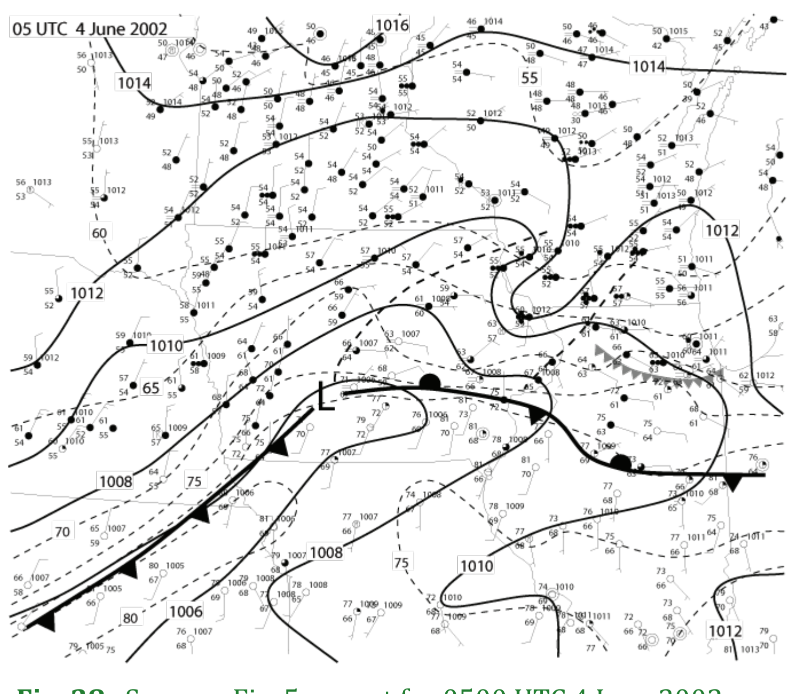


Fig. 28. Same as Fig. 5, except for 0500 UTC 4 June 2002.

and 2) create an easterly cross-isobaric flow of air across northern Illinois vectored into the incipient convection in east-central Iowa. Thus, new convection formed both downstream from MCS #4b as well as to the southwest of the original convection.

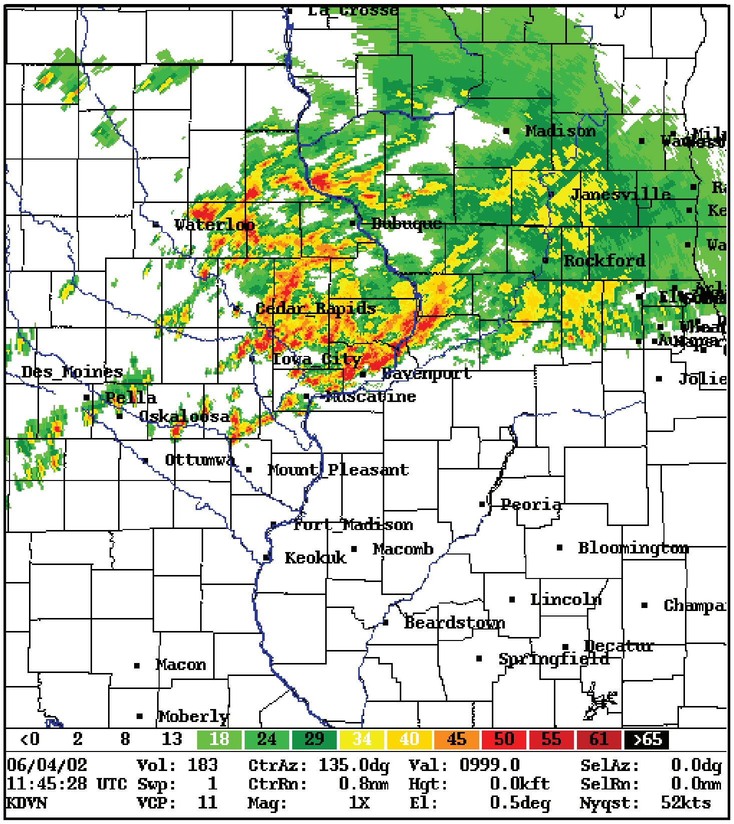


Fig. 29. Same as Fig. 15, except for 1145 UTC 4 June 2002.

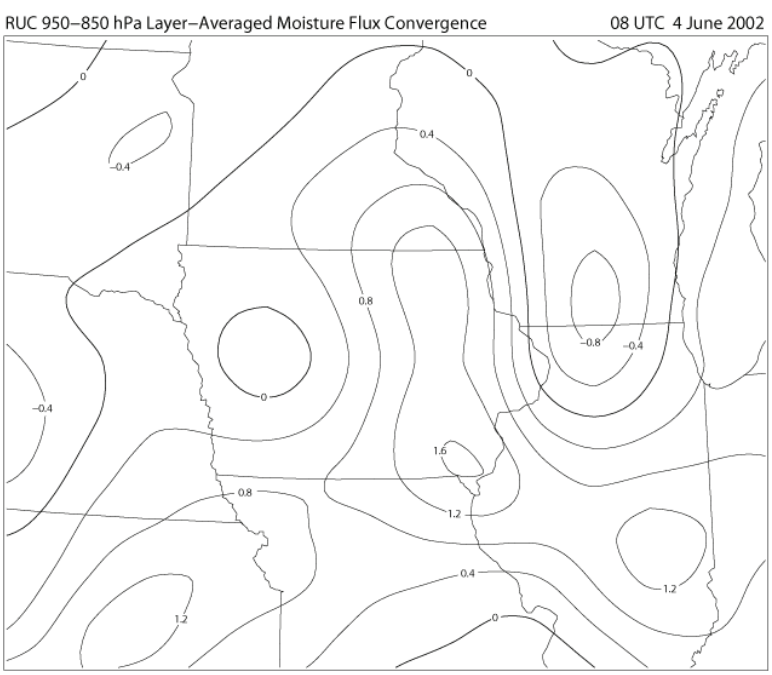


Fig. 30. 950-850 hPa layer-averaged moisture flux onvergence for 0800 UTC 4 June 2002 from RUC-II analysis; units are g (kg-h)$^{-1}$.

5. Operational Considerations during the Event

Operational challenges during this event focused on two areas: (1) evaluating operational numerical model output to determine the relative risk of severe thunderstorms vs. flooding, and (2) determining the location and magnitude of heavy rain. These are typical issues associated with the weather pattern described in section 2.

On the synoptic-scale, the NOAA/NWS National Center for Environmental Prediction (NCEP) models generally forecast the timing and location of the forcing mechanisms (upper and low-level jets and weak short-wave troughs) satisfactorily; however, the models diverged on some key details in the surface fields, specifically the location of the stationary surface front. Additional information provided by the European Centre for Medium-Range Weather Forecasts (ECMWF) and United Kingdom Meteorological Office (UKMET) added to the diversity of possible outcomes, thus the forecasters used an "ensemble" average, i.e., tossed out two outlying model solutions and took the consensus of the rest.

Several reasons led to the assessment of heavy rain and the likelihood of flash flooding as a primary threat and severe weather in the form of hail as a secondary and minor threat. These included:

- model soundings which indicated a strongly capped warm sector and an unstable and very moist air mass along and north of the stationary front, thereby suggesting elevated rather than surface-based convection, with limited mid-level dry air to support hail production;
- forecasts which kept most of the area in question north of the stationary front, due in part to reinforcing cold northeast flow off Lake Michigan;
- a climatologically favored pattern of nocturnal thunderstorms north of a boundary posing a heavy rain threat rather than the threat of tornadic or straight-line winds, and
- nearly-saturated soils which increased the likelihood of flash flooding.

Understanding and anticipating threat types is critical when planning staffing needs for significant events, especially when they occur during the overnight hours. This task is particularly challenging when the dual threats of severe weather and flooding co-exist.

6. Summary and Conclusions

The heavy rainfall which caused flash flooding and river flooding across eastern Iowa and northwestern Illinois was caused by episodic MCSs which formed in this bi-state region and moved east-southeast over a 27-h period. Most of the convection was elevated, as it formed to the north of a quasi-stationary boundary and above the surface. However, WSR-88D radar data and surface station observations revealed the presence of outflow boundaries and cold pools, which altered subsequent convection and its motion. Apparently, the convective downdrafts from this strong convection were able to reach close to the surface, penetrating the relatively stable boundary layer. Thus, although the basic synoptic-scale characteristics of this warm season, elevated convection agreed with the composite analysis described by Moore et al. (2003), mesoscale features created by the convection affected the orientation of the secondary and tertiary convection, as well as the propagation characteristics.

In general, the downwind Corfidi vector method was useful for estimating the propagation and storm motion vectors of the first two MCSs. However, the characteristics of the propagation and therefore storm motion of MCSs #3 and #4a-b were quite complex, owing to the veering of the LLJ, evolution of the mesoscale surface pressure pattern, and changing position of the large-scale frontal boundary. MCS #3 displayed strong evidence of high-precipitation convective cells that contributed to the heavy rainfall in and near Davenport, Iowa. The orientation of the cold pool and the mesoscale ridging, produced by earlier convection, enhanced the frontal boundary and organized deep-layer moisture convergence for a sustained period of time for MCS #3.

MCS #4a was characterized by backbuilding convection in which new cells actually formed upstream from the original convection. It, too, was basically west-east in orientation, reflecting the fact that the deep-layer moisture convergence was strong and organized in like fashion. Neither the downwind or upwind Corfidi vector methods completely captured the observed MCS motion in this case as there was evidence of downstream and upstream propagation at different times in its life cycle. Finally, MCS #4b which actually developed during the morning hours of 4 June had a dominantly north-south orientation which was due to the fact that the surface cyclone and associated frontal features moved eastward, while the surface and low-level moisture convergence changed into a north-south orientation. Also, as dry air entrained into the MCS from upstream, some cells became severe and dropped one-inch hail in parts of DVN's CWA.

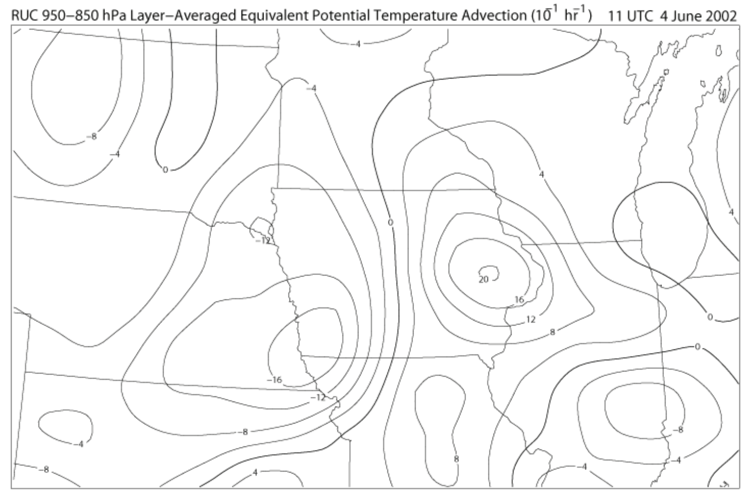
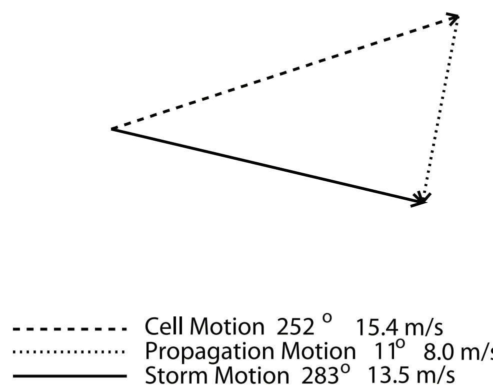
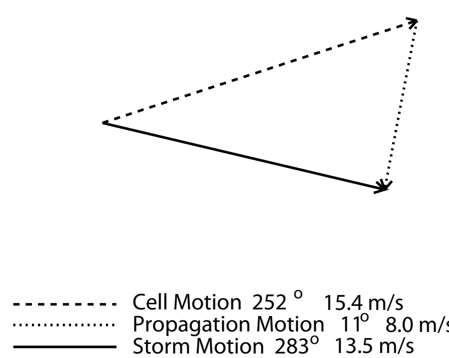


Fig. 31. Same as Fig. 18, except for 1100 UTC 4 June 2002.

Fig. 32(a-b). Corfidi Vector Method



(a) Downwind Corfidi vector method for 0800 UTC 4 June 2002.



(b) Upwind Corfidi vector method for 0800 UTC 4 June 2002.

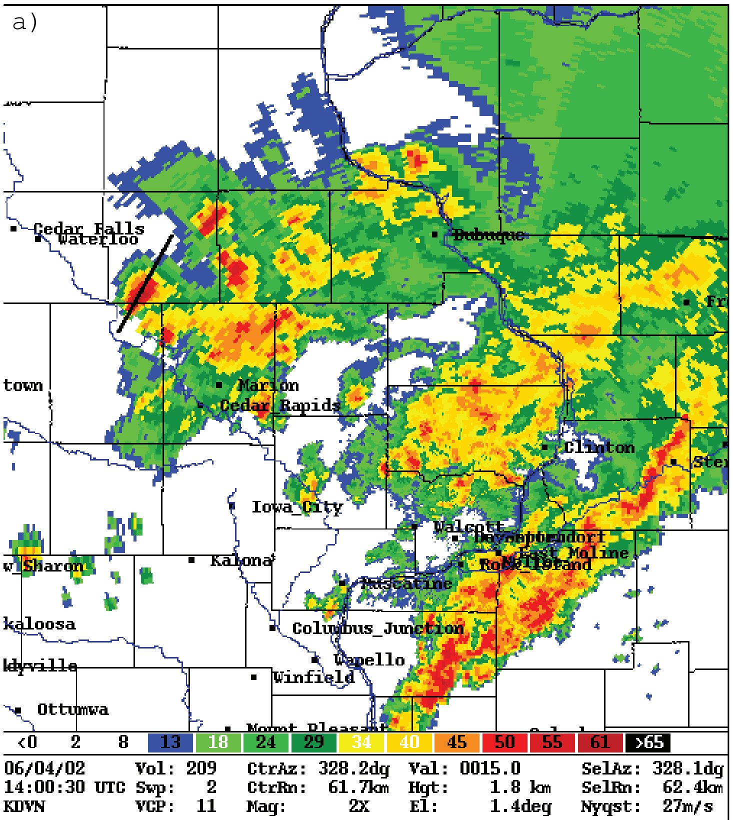


Fig. 33(a). Same as Fig. 15, except for 1400 UTC 4 June 2002. Solid line indicates line of cross section shown in (b).

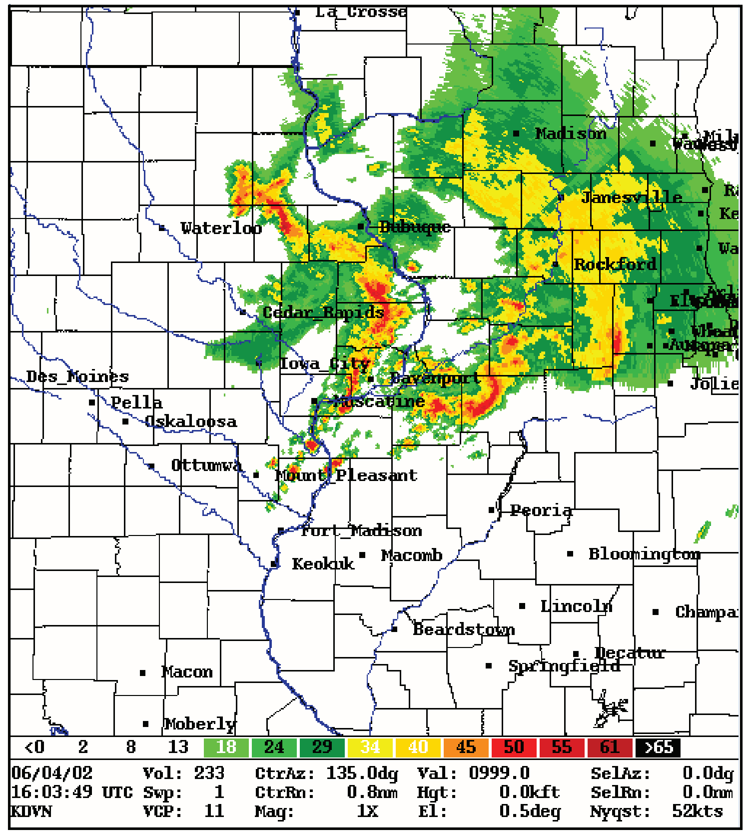


Fig. 34. Same as Fig. 15, except for 1604 UTC 4 June 2002.

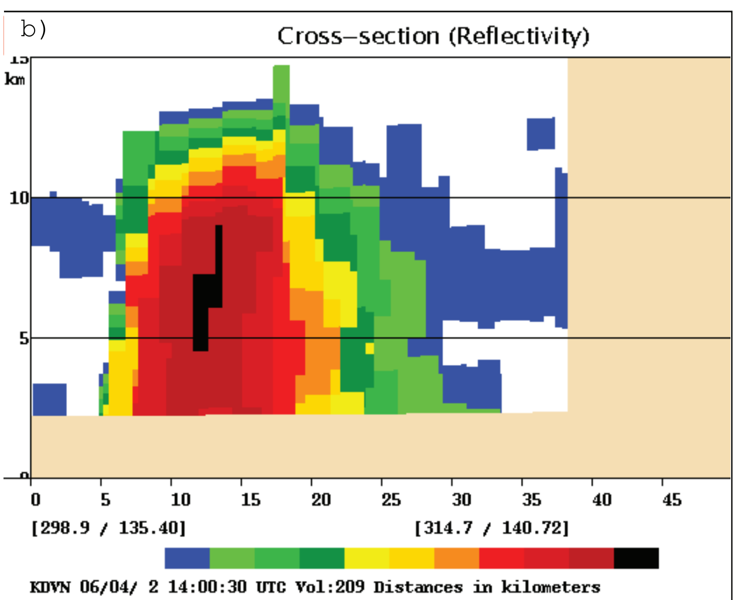


Fig. 33(b). Cross section of reflectivity along line shown in (a) indicating high-centroid cell echo which produced one-inch diameter hail.

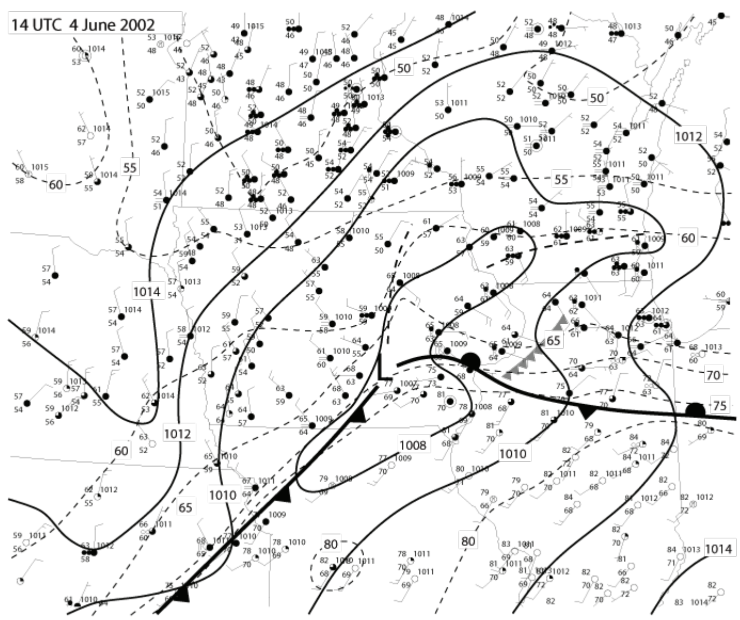


Fig. 35. Same as Fig. 5, except for 1400 UTC 4 June 2002.

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