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A CASE STUDY OF A SEVERE THUNDERSTORM OUTBREAK IN SOUTHERN VIRGINIA

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1. INTRODUCTION

The Blue Ridge in western Virginia often limits the potential for approaching thunderstorms to produce severe weather in the west central and southern counties of Virginia. Uncharacteristically, on June 4, 1993, a cluster of thunderstorms reached intensity levels in this region similar to those associated with the Great Plains. As the storms moved over Franklin, Bedford, Campbell, and Pittsylvania counties, and the cities of Bedford and Lynchburg, hail from 1 1/2" to 4 1/2" in diameter and wind gusts in excess of 58 mph were observed. The greatest property damage was a result of intense straight-line winds in the Lynchburg area, with an estimated $21 million damage in the city alone. Additionally, $430,000 of damage occurred in the area immediately outside of Lynchburg. Downed trees caused power outages which affected 95% of the residents. In addition, baseball- to softball-sized hail fell across a two-county area and caused extensive crop and property damage.

The purpose of this paper is to investigate the atmospheric conditions that led to the development of the thunderstorm complex on June 4, 1993. An examination of the general thunderstorm outflow, and how this feature acted in concert with a rear inflow, jet-induced downdraft to enhance the damage in the Lynchburg area, also will be presented.

2. SYNOPTIC SCALE ANALYSIS

At 1200 UTC on June 4, 1993, a stationary front extended from a surface low over northern Arkansas, through northern Kentucky and southern West Virginia, and into southern Virginia (Fig. 1). The airmass along and south of the front was characterized by a layer of moisture (>50% mean RH) from the surface to nearly 700 mb. A ridge axis of +12°C to +14°C 850 mb dewpoint air extended along the front across Kentucky and into southwest West Virginia. A considerable surface temperature gradient also was evident (not shown).

Johns and Hirt (1987) found that quasi-stationary thermal and moisture boundaries play a key role in the initiation and development of derechos, which are widespread convectively induced windstorms that may produce damaging winds. Johns and Hirt (1987) also found that derechos take one of two forms. The *progressive* derecho usually originates along and north of the thermal/moisture boundary. The resulting convective line moves in a direction nearly parallel to the boundary at a slight angle toward the warm sector, usually to the right of the mean wind. The *serial* derecho differs in that it moves in a direction nearly perpendicular to the mean flow and usually at a slower speed.

By 1450 UTC, damaging winds were occurring across northeastern Kentucky in the region of the first convective development north of the surface frontal zone. During the early afternoon, the axis of damaging winds propagated into southern West Virginia and finally moved into west central and southern Virginia by late afternoon.

An analysis of the mid- and upper-level wind field for this case was performed by using the PC-GRIDDS software (National Weather Service 1993) and the gridded data from the Nested Grid Model (NGM). A cross section along 36° N, near the surface front, revealed that nearly uniform southwesterly flow existed at 1200 UTC from central Kentucky into Southern Virginia (Fig. 2). The analysis indicates winds of 40 to 45 kt at 500 mb above the surface boundary, with a 50 kt maximum located over southeast Kentucky. The flow at 700 mb was equally strong and approached the 34 kt average Johns and Hirt (1987) found to be associated with derecho events.

At 1200 UTC, a broad, and somewhat flat, anticyclonic flow dominated the 500 mb wind field across the eastern third of the United States, while a short wave trough was lifting out of the central plains (Fig. 3). A weak vorticity maximum (10×10^{-5} sec^{-1}) was embedded in the flow downstream from the trough over central Kentucky, near the development of the initial convection.

In order to examine the role synoptic scale forcing had on the thunderstorm development, a Q-vector analysis was performed. According to quasi-geostrophic theory, large scale vertical motion is dependent on differential vorticity advection with height plus the laplacian of the low-level thickness advection. This relationship is shown mathematically by the omega equation (Holton 1979):

$$\left(\nabla^2 + \frac{f_0^2}{\sigma}\frac{\partial^2}{\partial p^2}\right)\omega = \frac{f_0}{\sigma}\frac{\partial}{\partial p}\left[\boldsymbol{V}_g \cdot \nabla\left(\frac{1}{f_0}\nabla^2\phi + f\right)\right] + \frac{1}{\sigma}\nabla^2\left[\boldsymbol{V}_g \cdot \nabla\left(-\frac{\partial\phi}{\partial p}\right)\right]$$

where, the first term on the right side of the equation represents differential vorticity advection, and the second term on the right side is proportional to thickness advection. Since in the atmosphere the two terms on the right hand side often act to cancel each other out, it is best to combine the two terms to get a quantitative estimate of omega. This can be achieved graphically through the use of Q-vectors. From Hoskins et al. (1978), we learned that upward vertical motion can be inferred from a convergence of the Q-vector field, while divergence of the Q-vectors represents downward vertical motion.

An analysis of the Q-vector fields between the 850-700 mb layer and the 500-300 mb layer was performed by using PC-GRIDDS and the 1200 UTC, June 4 NGM gridded data.

In the 850-700 mb layer, an area of convergence (upward vertical motion) extended across eastern Kentucky into extreme southwest Virginia (Fig. 4a). In the 500-300 mb layer, the convergence of the Q-vector field was near zero over the same area (Fig. 4b).

As previously mentioned, by late morning on June 14, severe thunderstorms were occurring across eastern Kentucky. Therefore, it can be inferred from the Q-vector analyses, that the large scale upward vertical motion over eastern Kentucky likely played a role in the development of the initial, and later severe, convection.

3. MESOSCALE ENVIRONMENTAL ANALYSIS

An analysis of the pre-storm mesoscale environment was performed by using the Skew-T Hodograph Analysis and Research Program workstation (SHARP; Hart and Korotky 1991). The most representative upper-air sounding for the airmass south of the frontal boundary was from Greensboro, NC (GSO). The lowest layer of the 1200 UTC June 4 sounding was modified by using the maximum temperature from a thermograph trace at the Weather Service Office in Lynchburg (WSO LYH) prior to the development of the convective storms. The upper-level environment was not modified, as little change occurred. [A time series cross section of the temperature and dew point field based on PC-GRIDDS and the 1200 UTC NGM gridded data centered over WSO LYH (not shown) revealed little change in the temperature profile aloft through the 36-h forecast].

The modified sounding illustrated that high instability was present immediately south of the surface front, with a surface to 500 mb lifted index of -7°C, and Convective Available Potential Energy (CAPE) of 2268 J/kg.

As mentioned earlier, very little vertical directional wind shear was present. However, the SHARP hodograph for GSO illustrated considerable speed shear in the 0-5 km level (Fig. 5). It has been shown in 3-dimensional simulations, that speed shear magnitudes near 15 m s^{-1} in the lowest 5 km of the atmosphere were found to be sufficient to tilt a thunderstorm updraft and contribute to the development of an elevated rear inflow (Weisman 1992). An estimate of the speed shear for the 1000-700 mb layer in this case, (which most closely resembled the Weisman (1992) dataset), was derived by using PC-GRIDDS and the 1200 UTC NGM gridded data. Speed shear magnitudes of 14-16 m s^{-1} were found in the area of study extending across the frontal zone from central Kentucky into the tidewater of Virginia (Fig. 6). Weisman (1992) also noted that the structure and strength of these convectively generated rear inflow jets depended on not only large shear values, but also on the amount of buoyant energy ahead of the storms (represented by CAPE).

A conclusion is drawn from these analyses that the highly sheared and marginally unstable environment present across southern Virginia, could have aided in the development of a convectively generated rear inflow jet. Further indication of the presence of a rear inflow jet was found with a close examination of radar data discussed in the next section.

The downdraft strength was approximated from the sounding by using the Downdraft Available Potential Energy (DAPE) equation (Foster 1958; Caracena and Maier 1987; National Weather Service 1993). The equation, which integrates DAPE from the surface to the level of free sink (LFS) is calculated by:

$$DAPE = 1/2\left[g\int_{LFS}^{sfc}\frac{(T_{pd}-T_e)}{T_e}dz\right]$$

where, g = acceleration due to gravity ($9.8\ ms^{-2}$);
T_{pd} = temperature of the downdraft parcel at the surface; and,
T_e = env. surface temperature.

To obtain the maximum downdraft wind speed (w_{max}), the following equation may be used:

$$w_{max} = -\sqrt{2DAPE}$$

Using the 1200 UTC June 4 modified sounding for WSO GSO (Fig. 7), this equation yielded a downdraft speed approximation of 36 m s^{-1} or 70 kt.

4. RADAR METEOROLOGY

Radar film obtained from the WSR-74S at the National Weather Service Meteorological Observatory in Volens, VA (WSMO VQN) was used for the analysis of the structure and movement of the severe storms through central and southern Virginia. By 1800 UTC, severe cells were detected over southern West Virginia, approximately 140 n mi west of WSMO VQN (not shown). The first cluster of cells passed through Roanoke, VA around 1945 UTC, producing softball sized hail over Franklin county. A bounded weak echo region (BWER) that represented the storm updraft, was detected within one storm, as it moved across southern Bedford and Northern Pittsylvania county, producing baseball sized hail as well as isolated wind damage. Lemon (1980) and others have determined that it is during the BWER existence and collapse stages that the largest hail develops. A second cell exhibiting a BWER was observed just south of the previous one shortly before 2030 UTC. As these storms moved generally east at around 40 kt, golfball sized hail was observed at VQN as the main cores of the two storms passed north and south of the station around 2100 UTC.

By 2130 UTC, the two cells had begun to merge over Mecklenburg county, southeast of VQN and a rapid increase in intensity occurred. At 2138 UTC, shortly after the merger, a pendant or hook echo was observed over central Mecklenburg county. The hook echo was only visible for a few minutes before another BWER was evident. By 2146 UTC, the storm exhibited a single intense updraft with a decreased forward speed and a sharp turn to the right. As the cell moved southeast at 30 kt, it possessed a Digital Video Integrator and Processor (DVIP) level 6 core up to 25,000 feet and a maximum echo top of 63,000 feet. The scope of this paper does not allow a full explanation of the observed radar features However this same progression of events, from BWER formation and collapse to low-level pendant formation, has been outlined as a severe multicell thunderstorm complex evolving into a supercell (Lemon 1980).

Reflectivity data obtained from the Weather Surveillance Radar 1988-Doppler (WSR-88D) in Sterling, VA (KLWX) was also analyzed for this event. However, the evolving supercell was beyond the quantitative velocity range of 124 n mi. The reflectivity data at 2140 UTC (not shown) indicated the merger of the two cells over Mecklenburg county when the supercell was at the strongest stage. A 65 dBZ reflectivity core was detected at a range of 140 n mi from the KLWX WSR-88D.

At the same time of the apparent supercell evolution in the southern counties of Virginia, another area of storms was entering the western portion of the state. A line of strong storms quickly developed from Roanoke and Lynchburg, VA between 2030 and 2100 UTC. By 2145 UTC, a strong reflectivity gradient developed along the leading edge of these storms, with the strongest cells in the central portion of the line over Bedford county. Damaging winds were reported in the city of Bedford with widespread power outages. As the line advanced eastward, the central portion quickly accelerated, forming a bow echo. Close inspection of the WSR-74S radar film showed a distinct weak echo region developed on the rear flank at the center of the line as it reached Bedford. The weak echo notch continued its penetration into the back of the storm at 2155 UTC in the vicinity of Lynchburg (Fig. 8). As the line accelerated eastward, this spearhead broadened and became the vertex of the bow echo. Schmidt and Cotton (1985) found a considerable correlation between a rear inflow jet and this indentation in the reflectivity field. Other weak echo channels, presumably from similar jetlets, developed further south along the line by 2201 UTC (Fig. 9).

These results suggest that a rear inflow jet played a key role in generating the intense winds experienced at Lynchburg and the subsequent bow echo. The environmental conditions that produce rear inflow jets were present as previously mentioned. The WSMO VQN WSR-74S radar imagery confirmed its probable existence.

The bow echo propagated eastward, increasing in forward speed to over 45 kt as observed by both the WSMO VQN radar and the KLWX WSR-88D. Over a dozen reports of damaging winds were logged between 2200 and 2300 UTC, spreading out along the length of the convective line.

An aerial survey over Lynchburg revealed several small bursts of wind damage, which fanned out from the apparent path of the storm. Fujita and Wakimoto (1981) identified this type of pattern as a downburst cluster. One microburst produced a peak wind of 64 kt (68 mph) at WSO LYH, as the center of the storm passed north of the station. The anemometer trace shows this microburst superimposed on the main thunderstorm outflow at 2154 UTC (Fig. 10). Two other weak perturbations in the wind flow (labelled A and B) can also be distinguished in the following 20 minutes. In addition, the barograph trace showed a significant pressure jump at the onset of the peak wind. Fujita (1985) has observed that, on average, pressure increases of only several tenths of a millibar may be associated with microbursts. The barograph trace recorded at WSO LYH depicts a 7.8 mb jump in pressure (Fig. 11).

Likewise the thermograph trace recorded a sharp departure from pre-storm readings with a temperature drop of 7.5°C (13.5°F) at the time of the peak wind (Fig. 12). A barograph trace from Brookneal, VA, located 30 miles south of Lynchburg, showed a similar pressure jump of nearly 7.0 mb shortly after 2200 UTC as the outflow passed over that location (Fig. 13).

5. SUMMARY AND CONCLUSION

This study has illustrated that the environmental conditions present on June 4, 1993 favored convective development. Even though a widespread severe weather outbreak did not occur, the isolated reports of large hail and damaging winds were significant. The pre-storm synoptic and mesoscale environments across western and southern Virginia were unstable enough to generate numerous severe convective storms over the area. These conditions were evident not only by the development of a bow echo initiated by strong inflow into the rear of the storm, but also by supercell development, as outlined by the radar depiction of these storms. A detailed analysis of the atmosphere was performed to illustrate the role of synoptic scale forcing in the generation of the severe convection. Spatial and temporal cross sections were used to reveal the character of the pre-storm environment. Also, layer-averaged wind fields, and 850-700 mb and 500-300 mb layer Q-vector analyses, were examined based on NGM gridded data and the application program PC-GRIDDS. These results suggest that use of the NMC gridded data and PC-GRIDDS in the future, should help to improve operational forecasts during these types of weather events.

ACKNOWLEDGMENTS

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REFERENCES

Caracena, F. and M. W. Maier 1987: Analysis of a microburst in the FACE meteorological mesonetwork in southern Florida. *Mon. Wea. Rev.*, **115**, 969-985.

Foster, D. S., 1958: Thunderstorm gusts compared with computed downdraft speeds. *Mon. Wea. Rev.*, **86**, 91-94.

Fujita, T. T., 1985: *The Downburst.* Satellite and Mesometeorology Research Project (SMRP). Univ. of Chicago, 122 pp.

____, and R. M. Wakimoto, 1981: Five scales of airflow associated with a series of downbursts on 16 July 1980. *Mon. Wea. Rev.*, **109**, 1438-1456.

Hart, J. A., and J. Korotky, 1991: The SHARP Workstation V1.50. A Skew-T Hodograph Analysis and Research Program for the IBM and Compatible PC. NOAA/NWS, Charleston, WV, 30 pp.

Holton, J. R., 1979: *An Introduction to Dynamic Meteorology*. Academic Press, 391 pp.

Hoskins, B. J., I. Draghici, and H. C. Davies, 1978: A new look at the ω-equation. *Quart. J. Roy. Meteor. Soc.*, **106**, 707-719.

Johns, R. H., and W. D. Hirt, 1987: Derechos: Widespread convectively induced windstorms. *Wea. and Forecasting*, **2**, 32-49.

Lemon, L. R., 1980: Severe thunderstorm radar identification techniques and warning criteria. *NOAA Technical Memorandum NWS NSSFC-3*, National Oceanic and Atmospheric Administration, U.S. Department of Commerce, 60 pp.

Schmidt, J. M., and W. R. Cotton, 1985: Bow echo structure and evolution within a high plains meso-β squall line. *Preprints, 14th Conference on Severe Local Storms*, Indianapolis, Amer. Meteor. Soc., 77-80.

National Weather Service, 1993: OTB Training Module 4: Basic Convection-II. United States Department of Commerce, National Oceanic and Atmospheric Administration, 43 pp.

Weisman, M. L., 1992: The role of convectively generated rear-inflow jets in the evolution of long-lived meso-convective systems. *J. Atmos. Sci.*, **49**, 1826-1847.

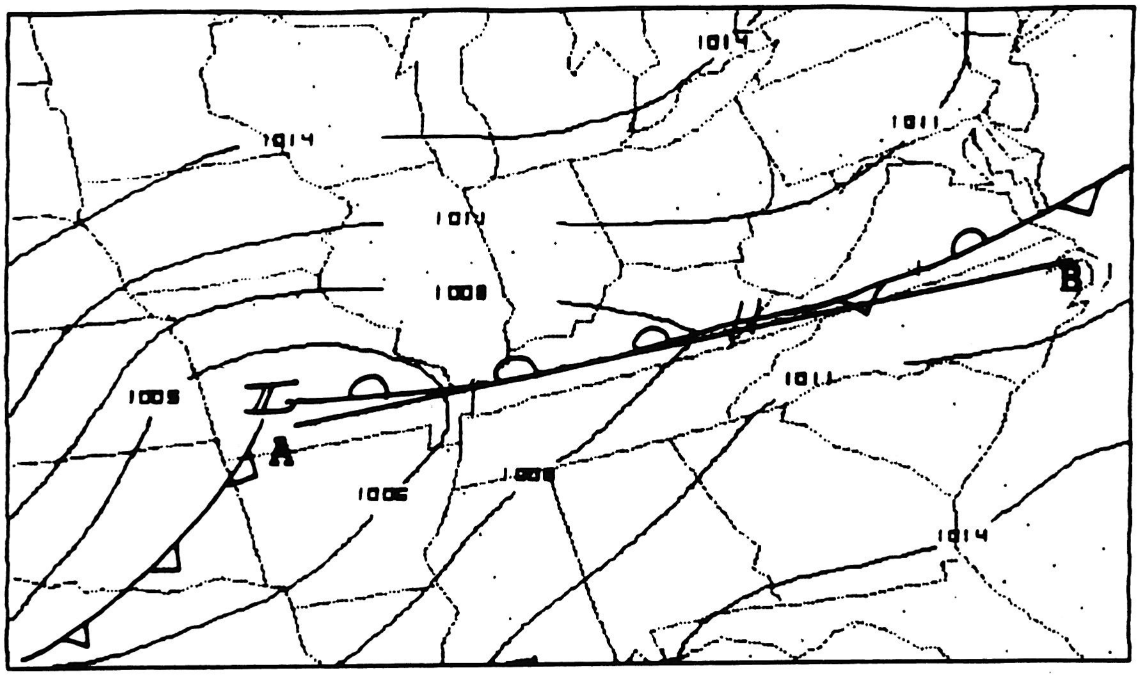

Figure 1. 1200 UTC, June 4, 1993 surface pressure and frontal analysis. Note that the cross section in Figure 2 is defined by the points A - B above.

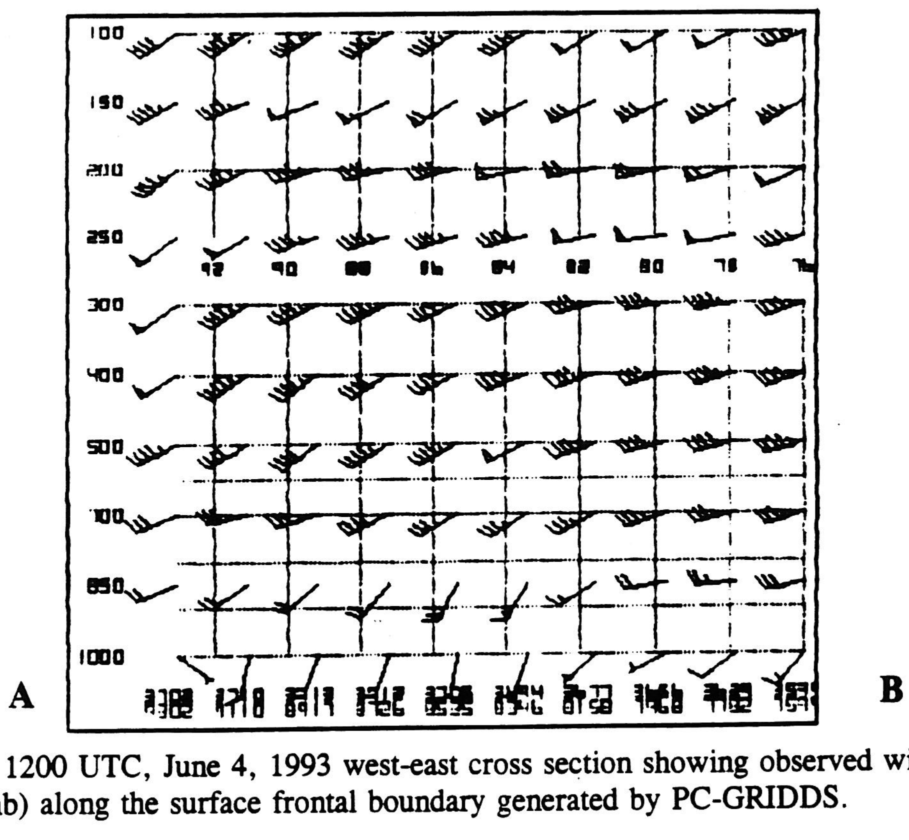

Figure 2. 1200 UTC, June 4, 1993 west-east cross section showing observed winds (kt) with height (mb) along the surface frontal boundary generated by PC-GRIDDS.

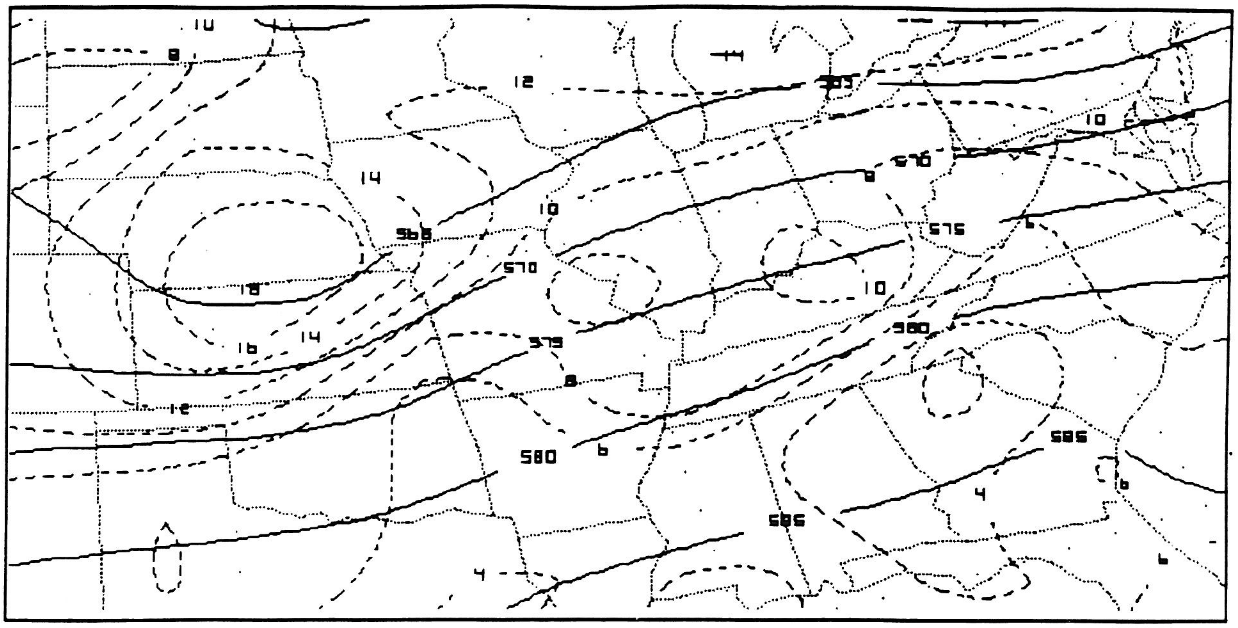

Figure 3. 1200 UTC, June 4, 1993 500 mb height contours (solid) and absolute vorticity (dashed) generated by PC-GRIDDS.

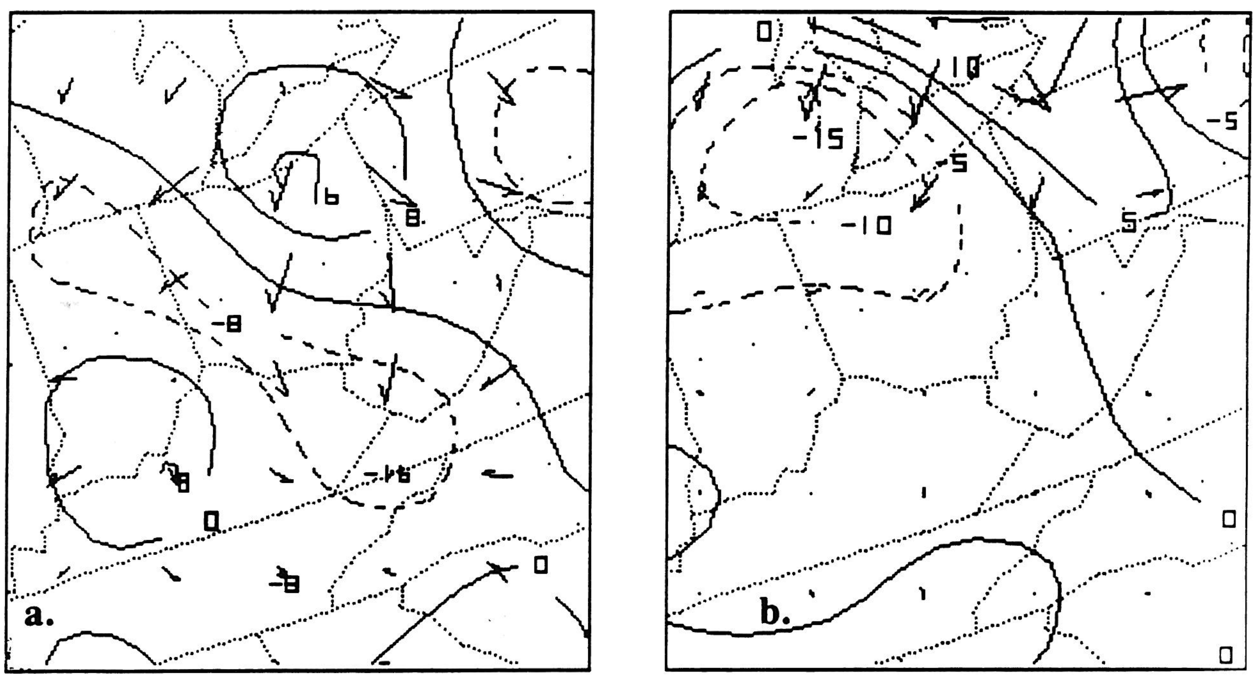

Figure 4. 1200 UTC, June 4, 1993, a) 850-700 mb layer Q-vectors (arrows) and divergence of Q-vectors, and b) 500-300 mb layer Q-vectors and divergence of Q-vectors generated by PCGRIDDS. Note that negative (positive) values represent upward (downward) vertical motion.

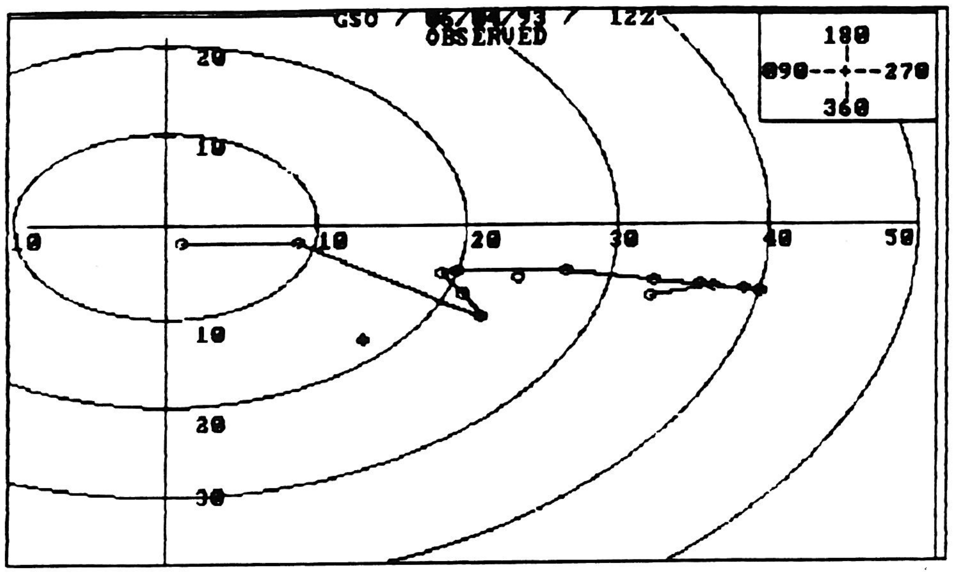

Figure 5. 1200 UTC, June 4, 1993 hodograph for Greensboro, NC. Each successive point along the hodograph represents an increase in elevation of 1,000 ft. From the SHARP Workstation (Hart and Korotky 1991).

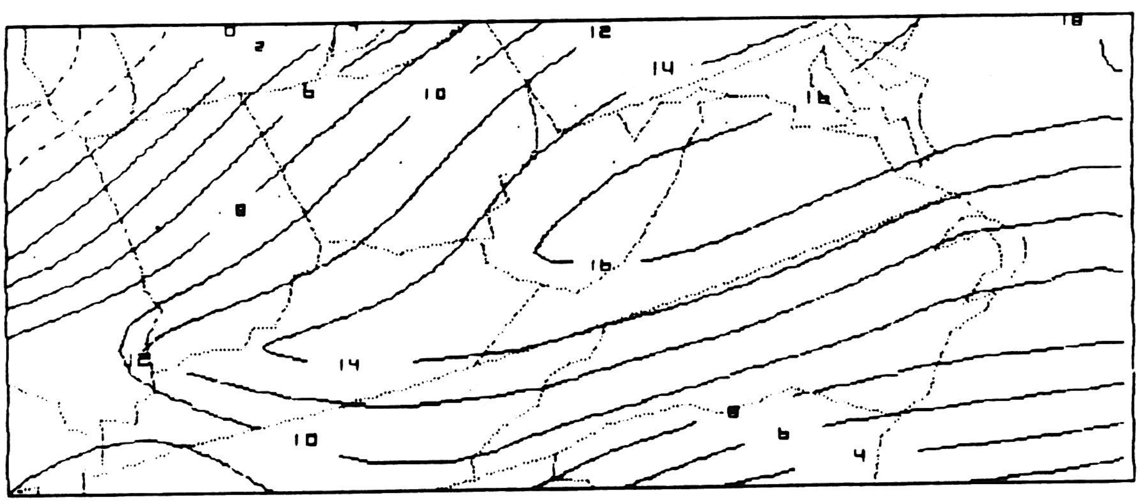

Figure 6. 1200 UTC, June 4, 1993 00-h NGM forecast of the 1000-700 mb layer wind speed shear (m/s) generated by PC-GRIDDS. The wind speed increasing with height is denoted by positive values.

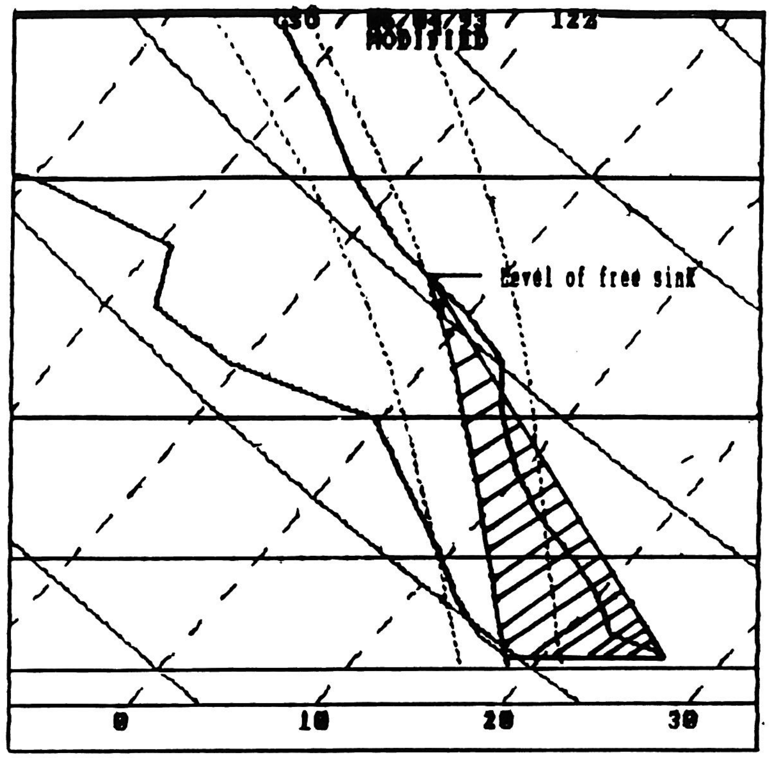

Figure 7. 1200 UTC, June 4, 1993 modified sounding for Greensboro, NC. The shaded area represents DAPE. From the SHARP Workstation (Hart and Korotky 1991).

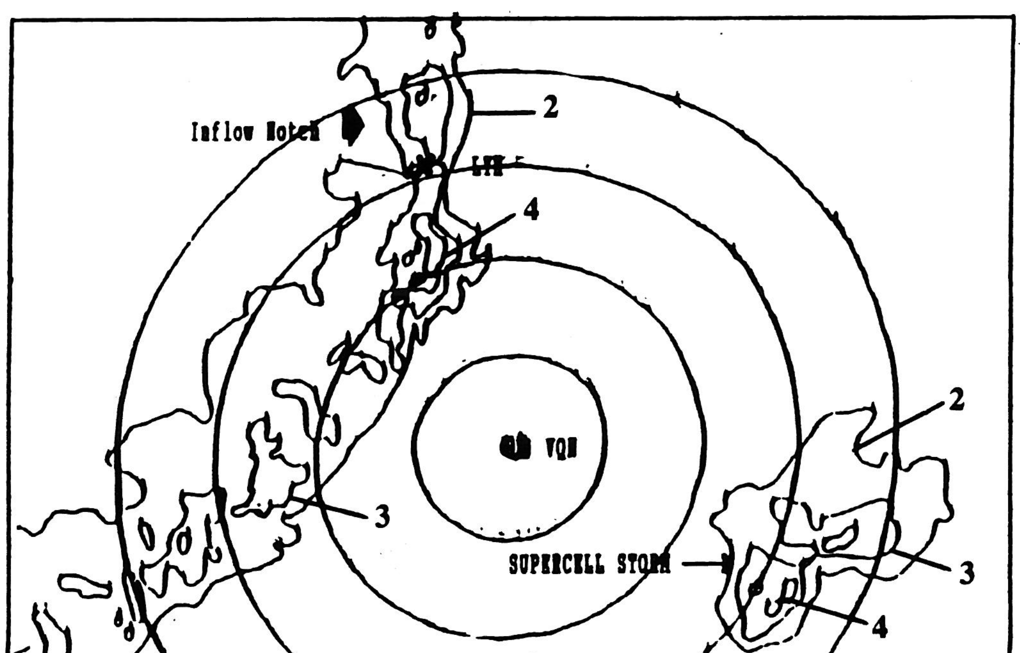

Figure 8. 2155 UTC, June 4, 1993 WSMO Volens WSR-74S radar imagery. The outermost contour depicts DVIP level 2 reflectivity returns (labled 2), the next inner contour depicts DVIP Level 3 reflectivity returns (labled 3), and the innermost contour depicts DVIP Level 4 reflectivity returns (labled 4). The range rings represent a radial distance of 25 n mi.

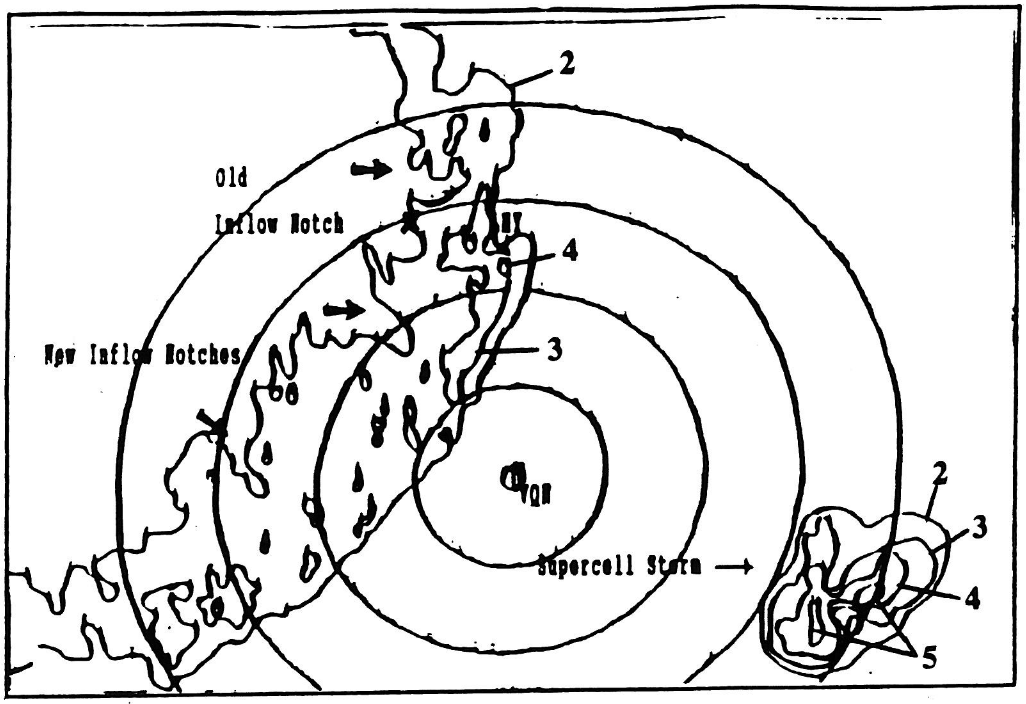

Figure 9. As is Figure 8, except for 2201 UTC, June 4, 1993. Note that the innermost contours of the supercell storm in the lower right of the figure are DVIP Level 5 reflectivity returns (labled 5).

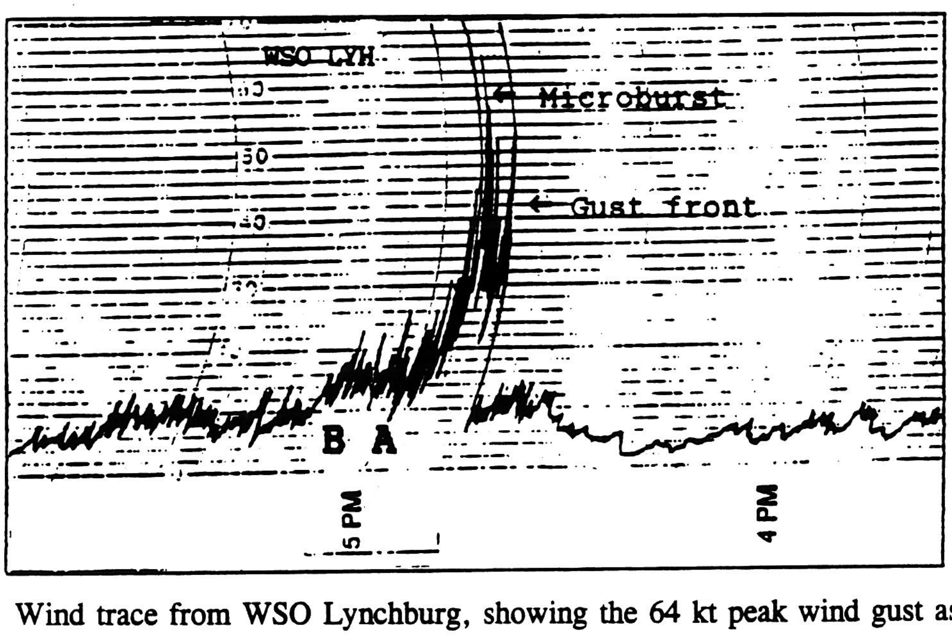

Figure 10. Wind trace from WSO Lynchburg, showing the 64 kt peak wind gust associated with a microburst. Note that the microburst occurred 4 minutes after the passage of the gust front. (Time is in EDT).

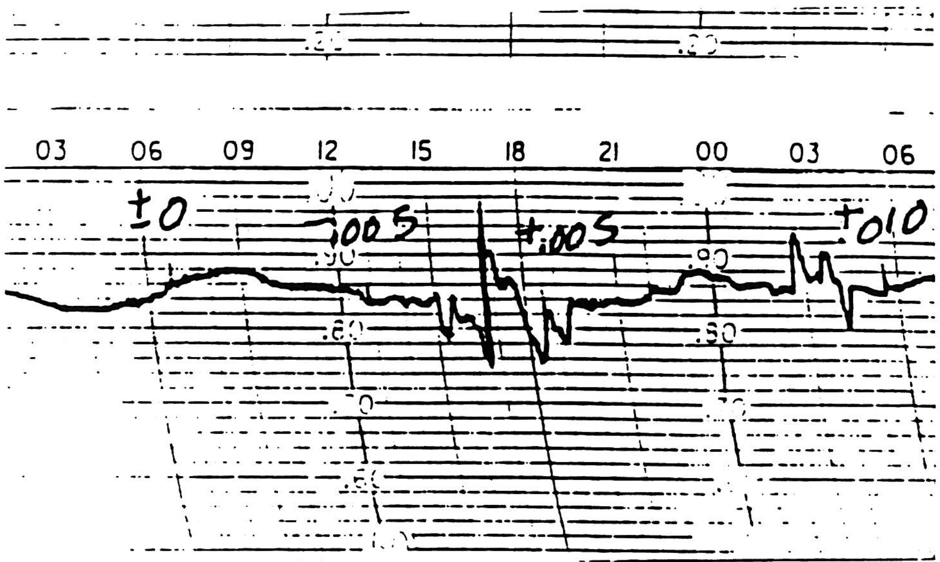

Figure 11. June 4, 1993 barograph trace from WSO Lynchburg. Note the 0.23 inch (7.8 mb) pressure jump at 4:55 p.m. (Time is in EDT).

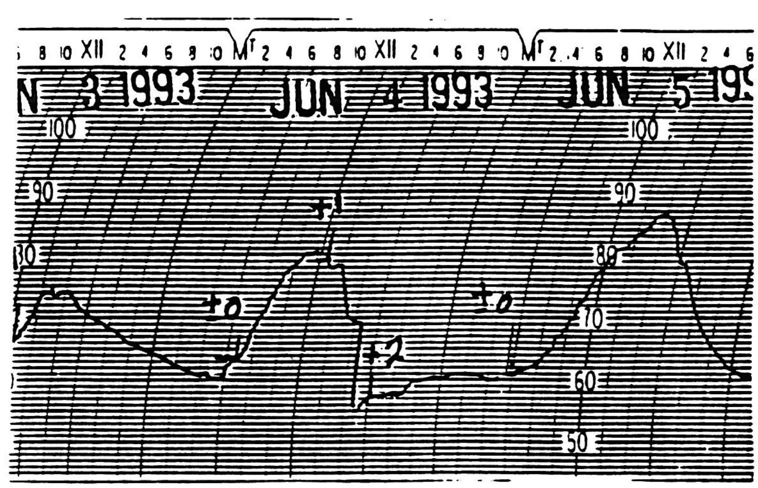

Figure 12. June 4, 1993 thermograph trace from WSO Lynchburg. Note the 13.5°F (7.5°C) temperature drop at 4:55 p.m. (Time is in EDT).

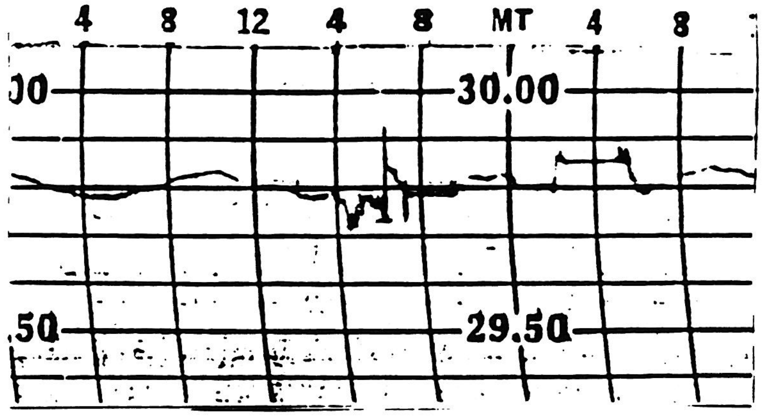

Figure 13. June 4, 1993 barograph trace for Brookneal, VA. Note the 0.21 inch (7.0 mb) pressure jump at approximately 6:00 p.m. (Time is in EDT).