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THE ROLE OF JET STREAKS IN THE TORNADIC DEVELOPMENT OF NOVEMBER 16, 1989 OVER THE NORTHEAST UNITED STATES

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1. INTRODUCTION

On November 16, 1989, severe thunderstorms produced 22 tornadoes across the northeastern United States. Eight tornadoes occurred in New York, seven in New Jersey and four in eastern Pennsylvania (U.S. Dept. of Commerce 1989). The tornadoes in New York and New Jersey were unprecedented, having been the first ever recorded in the month of November (U.S. Dept. of Commerce 1989). In general, tornadoes occur infrequently in this part of the United States with peak activity in July and August (U.S. Dept. of Commerce 1991). Viewed from both an historical and meteorological perspective, this was an extremely rare event.

This paper will focus on the role of jet streaks in this "off season" tornado outbreak, in an attempt to show that the upper- and lower-level jet streaks were the key elements in producing an environment favorable for tornado development. In Section 2, a brief review of jet-streak concepts is presented along with an overview of previous studies relating jet-streaks to the occurrence of severe weather. Section 3 details the relative strength and positions of the jet structures at the 300 and 850 mb levels, and provides and analysis of the sounding data. The jet structures are then linked to the formation of an environment favorable for severe weather. A summary is presented in Section 4, along with suggestions for future study.

2. BACKGROUND

A detailed discussion of the dynamics and kinematics of jet stream and jet-streak structure is provided by Reiter (1963); Palmen and Newton (1969); and Uccellini (1990). The principles of jet-streaks, as applied in this paper, are reviewed in this section. As defined by Palmen and Newton (1969), jet streaks are regions of isotach maxima in which the maximum wind speed is greater than the propagation of the jet. Fifty knots is the common minimum speed for defining a jet stream segment, called a jet-streak or jetlet. Regions of upper-level convergence and divergence in the vicinity of jet-streaks are generated as air is accelerated in the entrance and decelerated in the exit regions of the jet-streak. The convergence and divergence areas have been derived from analysis of the vorticity equation and have been empirically verified (Bebbe and Bates 1955; McNulty 1978; Johnson and Uccellini 1979; Uccellini 1990).

Figure 1 shows the divergence and convergence patterns for straight and curved jet streaks. Looking downstream, upper-level divergence is found in the left front and right rear quadrants of the jet-streak, while upper-level convergence is found in the left rear and right front quadrants. If the flow is cyclonically curved, the positive relative vorticity on the left side of the jet axis is increased. This enhances the divergent and convergent areas on the cold air (cyclonic) side. The opposite is true for anticyclonically curved jet-streaks.

Meteorologists have long known the role of jet stream features in the development of severe convective storms. Beebe and Bates (1955), emphasized the interaction of upper- and lower-level wind maxima in the development of severe thunderstorms. A model of intersecting jet structures was proposed whereby low-level convergence (850 mb) is surmounted by higher-level divergence (500 mb) assisting in the release of convective instability (Fig. 2).

Lee and Galway (1956 a,b; 1957), observed a relationship between speed maxima at 200 mb and tornado occurrence, emphasizing the contribution of jet maxima to higher-level divergence. In his severe storm forecasting manual, Miller (1967) set forth a summary of severe storm forecast guidelines incorporating low- and mid-level jet features as key parameters. The physical reasoning behind several of Miller's key parameters can be traced to short-wave troughs and upper-level divergence associated with jet stream wind maxima.

McNulty (1978), examined upper tropospheric wind maxima and empirically verified their associated divergence and convergence fields. From an examination of severe weather outbreaks for March 1976, four basic patterns were observed in the positioning of the 300 and 850 mb wind fields (Fig. 3). It was generally concluded that severe weather occurs between the 850 and 300 mb wind maxima under areas of upper-level divergence. McNulty noted that wind maxima alone can produce severe weather through divergence in the right rear quadrant of the jet and that no cross-over of the jets is necessary.

In general, upper- and lower-level jet maxima, as analyzed on constant pressure surfaces, are treated as separate entities. Johnson and Uccellini (1979) present a basis for the coupling of upper and lower tropospheric jet-streaks resulting in the development of conditions favorable for severe convective storms. In a case study, jet-streaks were examined on isentropic surfaces illustrating the coupling of the low-level jet within the exit region of the upper-level jet. By using a numerical model simulation, the dynamical process responsible for the coupling of jet structures was examined. The coupling was the result of mass adjustments caused by the propagation of the upper-level jet-streak. The results of the study by Johnson and Uccellini, verify the concept that conditions favorable for severe thunderstorms can be forced by propagation of upper-level jet-streaks.

Harnack and Quinlan (1989), quantified key synoptic variables associated with severe local storms in the northeast United States. The analysis of 850-300 mb jet-streaks revealed that principally the right rear, and secondarily the left front, quadrants of 300 mb jet streaks were the most favored areas for severe weather occurrence in the northeast. At 850 mb, it was concluded that severe weather tends to occur to the left of the 850 mb jet axis.

3. APPROACH AND ANALYSIS

The referenced studies tended to consider "classic" springtime severe storm conditions (i.e., intense and strongly baroclinic synoptic settings with well defined jet configurations). In such a case, the primary role of the upper tropospheric jet-streak is to provide upper-level divergence and advect cool, dry air within the middle and upper troposphere. The low-level jet transports warm, moist air toward the convective region. The combined interaction of the upper and lower tropospheric jets create a region in which severe weather ultimately develops.

The November 16, 1989 event had several conditions identifiable with a "classic" severe weather episode. In particular, a well defined jet configuration was observed. The magnitude of the tornado occurrences, suggested a highly sheared environment favorable for mesocyclone and tornado development. The scope of the analysis was then simplified by focusing on the upper- and lower-level jet configuration as the key mechanism in producing the convective instability, upper-level divergence, and shear necessary to produce tornadoes. This particular approach was supported by the research cited previously.

Standard 12-h constant pressure level plots at 300 and 850 mb, and GOES water vapor imagery, were used to identify jet features. The SHARP (Skew T/Hodograph Analysis Program) Workstation (Hart and Korotky 1991) was used to analyze the radiosonde data. Surface plots, at standard synoptic times, were also used along with a limited data set from the AFOS Data Analysis Program (ADAP). The analysis of the data focused upon: 1) the configuration of the jet features at the standard levels of 850 and 300 mb; and 2) the shear and instability as analyzed on the SHARP Workstation.

3.1 Jet Analysis

The analysis of the 300 mb data at 1200 UTC, indicated a distinct jet-streak in excess of 120 knots over western Pennsylvania (Figure 4). The core of the 120 kt winds, extended along an axis from near Elkins, WV (EKN), north to near Buffalo, NY (BUF). From the analysis, it was evident that the right rear quadrant of the 300 mb jet-streak would be in a position to impose upper-level divergence over eastern Pennsylvania, eastern New York and New Jersey.

Because of the temporal and spatial limitations of the upper air observation network, it was difficult to trace the origin and eventual destination of the 300 mb jet-streak. Inferences on the probable course were made from an examination of GOES water vapor satellite imagery. The water vapor imagery at 1200 UTC revealed a large dark band, indicative of a dry slot extending from Alabama and Georgia, north into western Pennsylvania (Fig. 5). By 2100 UTC, the dry slot had shifted east and north into eastern Pennsylvania and New York

(Fig. 6). Since the 300 mb jet-streak was initially aligned with the dry slot, it is reasonable to infer that the movement of the jet-streak followed that of the dry slot on the satellite imagery. As a result, the divergent right rear quadrant of the jet-streak at 300 mb did pass over the severe weather area, providing upper tropospheric divergence.

The analysis of the 850 mb data at 1200 UTC, showed a 60 to 80 kt jet-streak, extending from Virginia into southeast New York (Fig. 7). The core of the 80 kt winds extended along an axis from just north of Richmond, VA (RIC) to just south of Allentown, PA (ABE). The well developed 850 mb jet aided in the advection of a warm and moist air mass into the region, creating an environment favorable for convection. At the surface, a tongue of temperatures and dew points on the order of +19 to +21°C were positioned along and directly to the right of the jet axis. At the 850 mb level, along and to the right of the jet axis, corresponding values of temperature and dew point ranged from +10 to +13°C.

In general, a lower tropospheric jet is associated with fields of vertical motion, although they are not as clearly defined as they are for the upper-level jet (Doswell 1982; McNulty 1978). For example, the same divergence-convergence generalization cannot be applied to the quadrants of the 850 mb jet-streak because the inertial terms of the vorticity equation are not necessarily negligible at lower levels of the atmosphere (McNulty 1978). Based on their analysis of the vorticity equation, Beebe and Bates (1955) noted that a region of convergence does exist to the left of a southerly low-level jet (strongest in the northern end where warm advection is the greatest). This relationship to upward motion can be seen if one recalls that warm advection is related to upward vertical motion in quasigeostrophic theory. Because of the difficulty in providing a qualitative estimate of the vertical motion associated with the 850 mb jet, the emphasis here is on the destabilizing effects via advection.

A composite of the extrapolated 1700 UTC position of both wind maxima is presented in Figure 8. The relative positions are similar to those observed by McNulty (1978) (Fig. 3). A composite of all confirmed tornadoes and funnels sighted is presented in Figure 9 (Henricksen and Summer 1990). A comparison of the extrapolated jet-streak positions, to the occurrence of tornadoes in Figure 9, clearly shows that the tornadoes occurred along and to the left of the 850 mb jet core and under the right rear divergent quadrant of the 300 mb jet. In Figure 9, most of the reported tornadoes in eastern Pennsylvania and New Jersey, occurred at the time when the 850 mb jet was in close proximity. The temporal distribution is from south to north, following the northward movement of the 850 mb jet.

3.2 Sounding Analysis

The 1200 UTC sounding and hodograph for Atlantic City, NJ (ACY) were examined using the SHARP Workstation. The SHARP program allows for modification of the data based on observed or predicted conditions. The lower levels (surface to 850 mb) of the ACY sounding were modified, by using the dew point and temperature values observed upstream and incorporating the data in the modified sounding (Fig. 10). Observed surface data at 1600 UTC was also used to modify the sounding to represent the surface environment around the time the tornadoes occurred in eastern Pennsylvania and New Jersey. Values for selected parameters derived from the initial and modified soundings are summarized below in Table 1. The Lifted index, Showalter index, and the Convective Available Potential Energy (CAPE; B+), were selected to illustrate the changes in stability of the atmosphere.

By increasing the low-level moisture and temperature at ACY, a corresponding decrease in stability was observed. The initial lifted index decreased from +1 to -3, while the Showalter index decreased from -1 to -2. The modified indices are significant in the fact that they are well within the climatologically favored range associated with severe weather events in the northeast, as described by Harnack and Quinlan (1989). Harnack and Quinlan (1989) concluded that the Showalter index had an average value of +1, with minimum values of -1 to -2 for tornado occurrences in the northeast United States.

The 1600 UTC CAPE of 990 J/kg^{-1}, was also within the range of cool season values associated with tornadoes as reported in a study by Johns, Davies, and Leftwich (1991). A wide range of CAPE, 200 to 5300 J/kg^{-1}, were found to be associated with all the strong to violent tornado cases studied. For the cold season cases, 95% exhibited CAPE values less than 2500 J/kg^{-1}. While the change in the CAPE at ACY was only a modest increase, from 62 to 990 J/kg^{-1}, the latent instability was obviously sufficient for severe thunderstorms and/or tornadoes to develop.

Because of the close proximity of the 850 mb jet and the presence of a strong geostrophic wind field, a highly sheared environment was anticipated. Increased speed and veering of winds with height (vertical shear) are recognized as favorable conditions for mesocyclone and tornado development (Fawbush and Miller 1954; Miller 1967). The vertical wind shear is a crucial parameter for tornado forecasting (Davies-Jones et al. 1990). The SHARP Workstation calculates several parameters related to shear. These include the Storm Relative (SR) helicity and the Energy-Helicity Index (EHI).

Davies-Jones et al. (1990), provide a mathematical and physical explanation of helicity. For the scope of this paper it is sufficient to understand that SR helicity gives a measure of the rotational potential of a thunderstorm updraft. As such, helicity can be a valuable forecasting parameter for tornadoes. Davies-Jones et al. (1990), suggest some rough intervals for SR helicity values as related to tornado intensity (the units of m^2s^{-2} will be omitted from henceforth). The values for weak (F0, F1) tornadoes are 150-299, strong tornadoes (F2, F3) 300-449, and for violent tornadoes (F4, F5) 450 and above.

The EHI represents the potential tornadic intensity as a function of CAPE and SR helicity. The reliability as a forecasting parameter is not yet conclusive. However, based on empirical studies, values above 1 have shown the potential for strong tornadoes (F2, F3) when additional severe weather parameters are present (Hart and Korotky 1991). EHI Values around 5 are associated with violent tornadoes (F4, F5).

Figure 11 is the modified hodograph for ACY at 1600 UTC. The hodograph was modified by slightly increasing the observed 1200 UTC winds (by approximately 10 kt) near the 850 mb level. This was done to account for the northward movement of the 850 mb jet. The storm motion and resultant storm inflow was also modified using observed storm motion from local radar observations. It should be noted that the calculation of helicity is extremely sensitive to the storm motion and the storm relative winds (Davies-Jones et al. 1990). Despite the sensitivity, the modifications to the hodograph produced only a small increase in helicity, from 406 to 412 (Table 1). Note that before modifying the hodograph, the initial SR helicity value of 406 was well within the range associated with strong tornadoes. Therefore the 850 mb jet had already aided in creating an environment where updraft rotation was likely should convection occur.

The modifications to the ACY data produced a more dramatic increase in the EHI (Table 1). The EHI at ACY increased from 0.16 to 2.47. This value went from being well below the threshold for any tornadic activity, to a value associated with strong tornadoes. This is the result of the EHI being a function of SR helicity and CAPE. With a high amount of helicity already present at 1200 UTC, the increase in instability was the key in producing a high EHI index at 1600 UTC. In this case, the modified EHI would have been a very good predictor for tornadoes, lending support to the use of the EHI as a potential forecast tool.

The sounding and hodograph for Albany, NY (ALB) were also modified using the methods described previously. Note that the modified values of SR helicity and the EHI were significantly higher than those calculated for ACY (Table 1). The wind speeds were increased on the ALB hodograph to account for the northward movement of the 850 mb jet. Winds around the 850 mb level were increased to between 70 and 80 kt to reflect the proximity of the 850 mb jet at 1800 UTC (a net increase of approximately 15 kt from the 1200 UTC values). The sensitivity of helicity to the storm relative winds resulted in the large increase in SR helicity and the corresponding increase in the EHI.

In a related study, LaPenta (1991) examined helicity, buoyancy, and the EHI for several tornadic events, primarily in the northeastern United States. The November 16, 1989 event was among those sampled. The sounding for ALB was modified using observed storm motion, surface temperature, and surface dew point temperatures for 1800 UTC. The resultant CAPE was 740, the SR helicity was 591, and the EHI 2.68. Note that a more conservative approach was used in modifying both the sounding and hodograph. In particular, winds were not increased to account for the 850 mb jet resulting in the lower values of SR helicity and the EHI. However, the results are consistent in demonstrating that conditions were favorable for tornado development based on the SR helicity, CAPE, and the EHI. The results in Lapenta's study also illustrated that these indices should be used with caution, and can be misleading if not used in conjunction with other meteorological parameters.

4. SUMMARY AND CONCLUSION

The results of this study, emphasize the importance of jet stream wind maxima to the occurrence of severe weather. In this case, the 850 mb and 300 mb jet-streaks were the key elements in producing a favorable environment for tornado development. The relative position and intensity of the jet features were similar to those described by McNulty (1978) and Harnack and Quinlan (1989), and were consistent with strongly baroclinic weather systems.

The proximity of the 850 mb jet provided the necessary directional and speed shear for tornadic development. This was reflected by the high initial (1200 UTC) value of SR helicity at Atlantic City, NJ. The transport of warm and moist air by the 850 mb jet resulted in destabilization of the atmosphere as demonstrated by the 1600 UTC increases in CAPE and the EHI at Atlantic City. Movement of the 300 mb jet-streak then allowed the divergent right rear quadrant to be in a position to provide upper-level divergence, assisting in the release of convective instability.

The authors recognize that many important factors, both at the surface and aloft, attend outbreaks of tornadoes and severe thunderstorms. In this case, an intense synoptic scale system was present, characterized by well defined jet features. The dynamic forcing of positive vorticity advection (PVA) and the intense surface and mesoscale features were not addressed. In fact, severe weather can and often does occur in far more benign meteorological settings.

Several topics remain for future research. The relative positions of the upper- and lower-level jets suggest the possibility that these features were coupled as demonstrated by Johnson and Uccellini (1979). An investigation of the processes involved in the development of the low-level jet may yield further insight as to the role it played in establishing an environment favorable for severe convective storms. For example, did the lower-level jet develop as an extension of the upper-level jet? Was the strong 850 mb jet the result of synoptic forcing or did the Appalachian mountains enhance the intensity of the jet through orographic effects?

Even with well-defined and large synoptic scale features, the limitations of the 12-h upper-air observational network were evident in this study. The lack of temporal and spatial resolution made it difficult to track the origin and evolution of the 300 and 850 mb jet features. With the increased use of wind profilers, Doppler radar, and improved resolution satellites, detecting and tracking important features such as tropospheric wind maxima should become easier in the future.

ACKNOWLEDGMENTS

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Table 1. Stability and Shear parameters calculated by SHARP for November 16, 1989.

STATION AND TIME	LI (°C)	SI (°C)	B+ (CAPE) (JKg^{-1})	SR HEL $(M/S)^2$	EHI
ACY 1200 UTC	+1	-1	62	406	0.16
ACY 1600 UTC	-3	-2	990	412	2.47
ALB 1200 UTC	1	2	24	576	0.09
ALB 1800 UTC	-3	-1	993	796	4.41

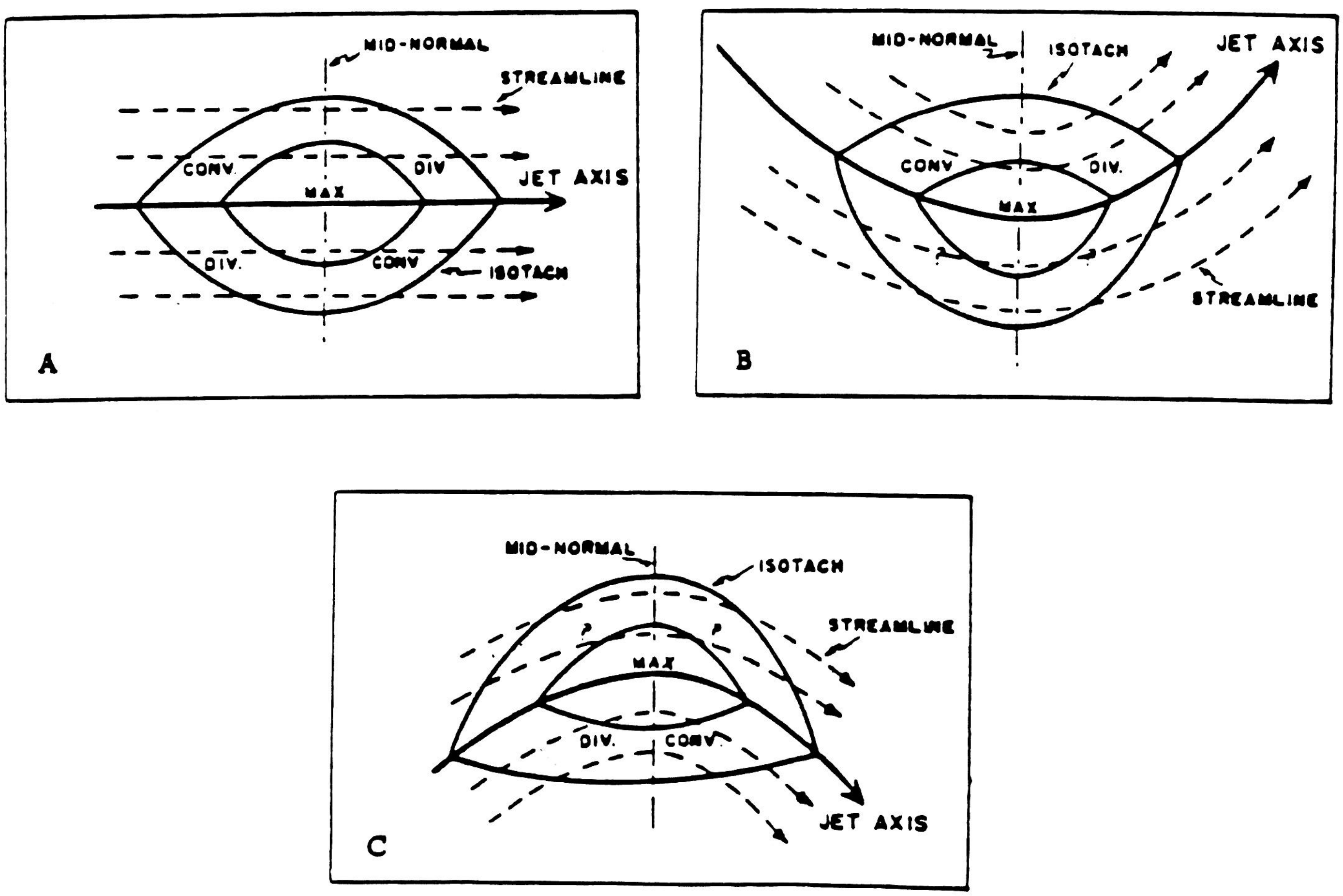


Figure 1. Horizontal divergence pattern in an upper-level jet-streak: A) without curvature; B) cyclonic curvature; and C) anticyclonic curvature. From Beebe and Bates (1955).

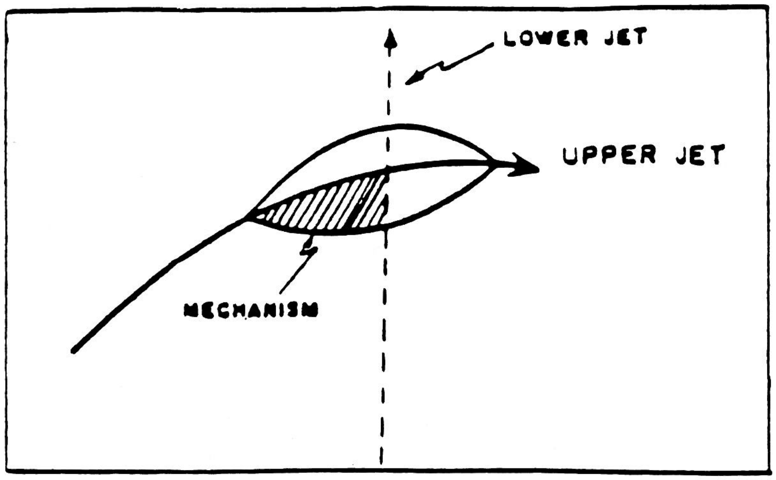


Figure 2. Crossing of upper- and lower-level jets (hatched area denotes location of severe weather. From Beebe and Bates (1955).

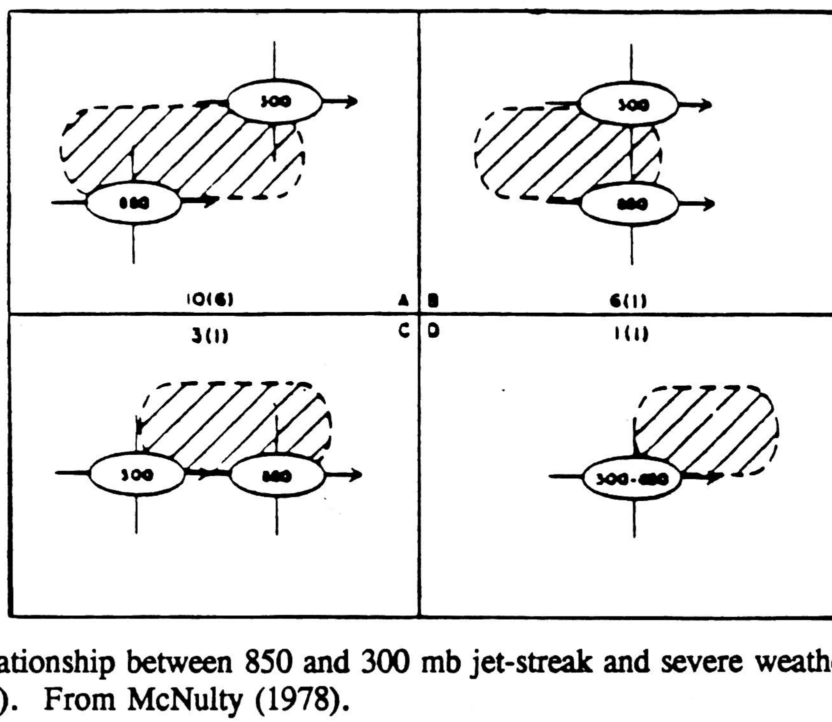


Figure 3. Relationship between 850 and 300 mb jet-streak and severe weather occurrence (hatched area). From McNulty (1978).

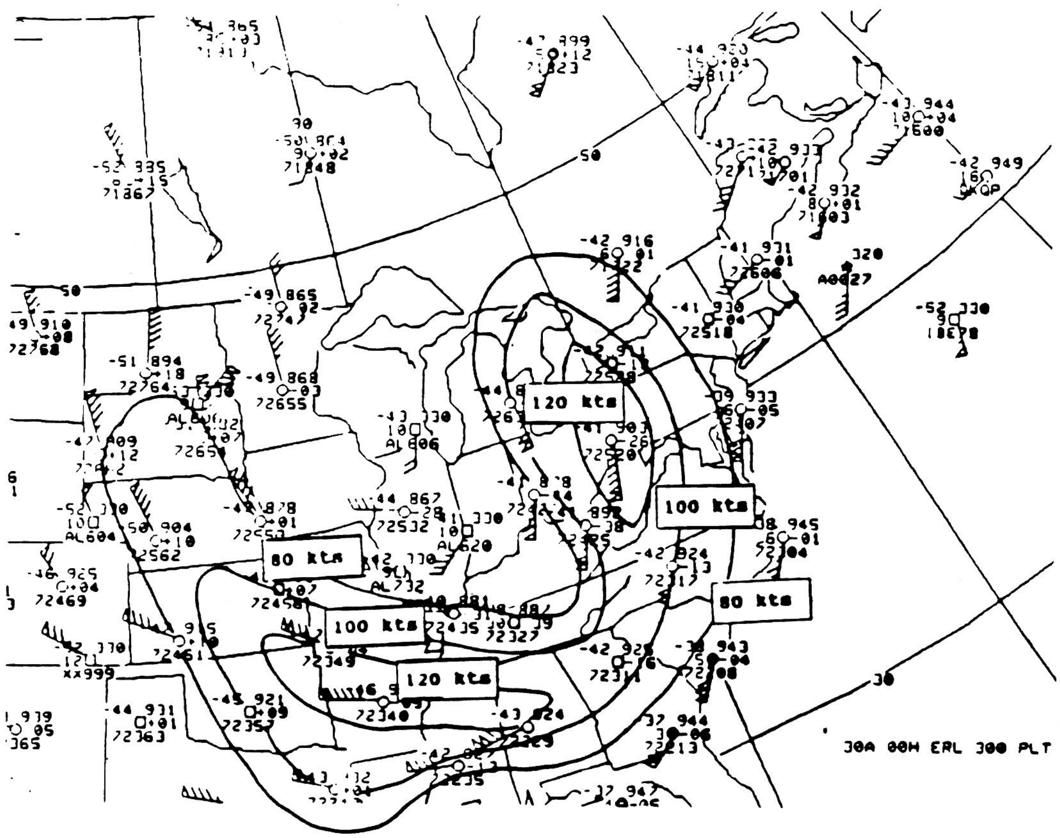


Figure 4. 1200 UTC, November 16, 1989 300 mb isotach analysis (kt).

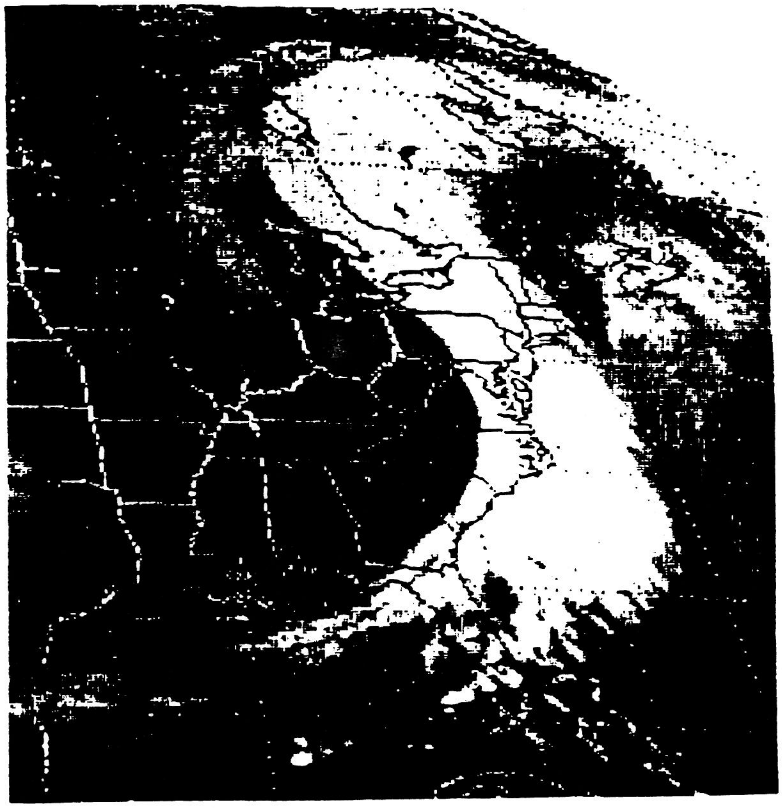


Figure 5. 1200 UTC, November 16, 1989, 6.7 μm water vapor satellite imagery.

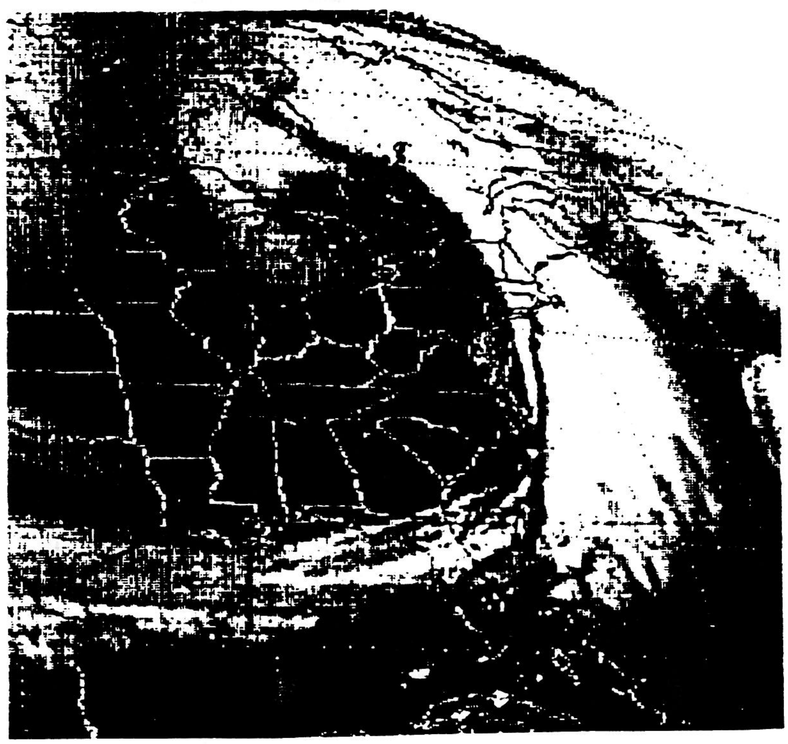


Figure 6. Same as Figure 5 except for 2100 UTC.

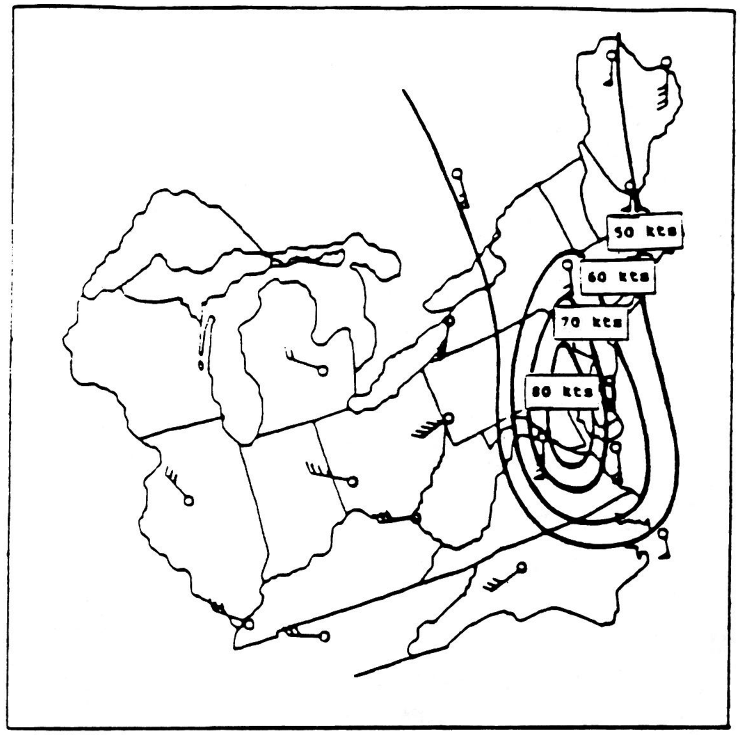


Figure 7. 1200 UTC, November 16, 1989 850 mb isotach analysis (kt).

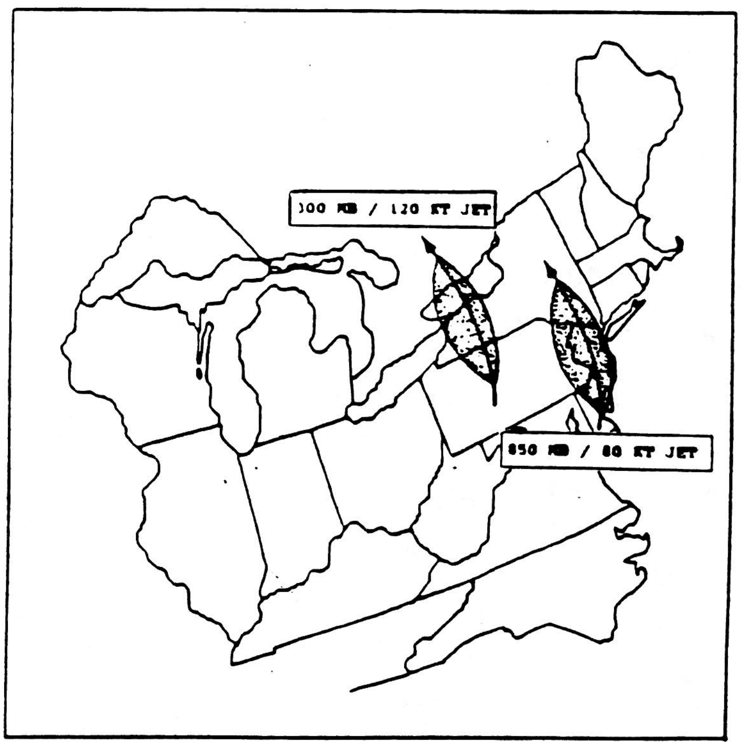
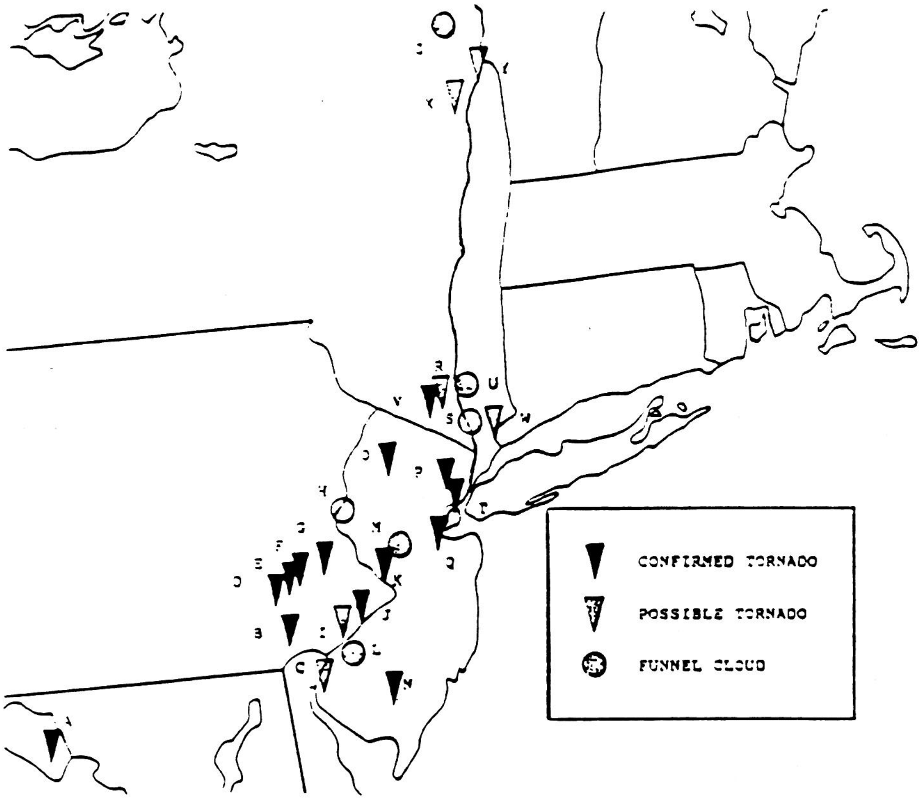


Figure 8. 1700 UTC, November 16, 1989 extrapolated 300 and 850 mb jet positions.



	EVENT	TIME		EVENT	TIME
A	TORNADO F2	5:00 AM	N	TORNADO F0	11:20 AM
B	TORNADO F0	10:24 AM	O	TORNADO F0	11:30 AM
C	POSSIBLE TORNADO	10:45 AM	P	TORNADO F0	NOON
D	TORNADO F0-1	10:45 AM	Q	TORNADO F0	12:04 PM
E	TORNADO F0	11:00 AM	R	POSSIBLE TORNADO	12:15 PM
F	TORNADO F0	11:00 AM	S	FUNNEL CLOUD	12:15 PM
G	TORNADO F0	11:00 AM	T	TORNADO F0	12:25 PM
H	FUNNEL CLOUD	11:00 AM	U	FUNNEL CLOUD	12:29 PM
I	POSSIBLE TORNADO	11:15 AM	V	TORNADO F1	12:30 PM
J	TORNADO F1	11:15 AM	W	POSSIBLE TORNADO	12:58 PM
K	TORNADO F1	11:15 AM	X	POSSIBLE TORNADO	1:50 PM
L	FUNNEL CLOUD	11:20 AM	Y	POSSIBLE TORNADO	1:55 PM
M	FUNNEL CLOUD	11:20 AM	Z	FUNNEL CLOUD	2:15 PM

Figure 9. November 16, 1989 composite plot of confirmed tornadoes and funnel clouds. From the Eastern Region Natural Disaster Survey Report (Henricksen and Summer 1990).

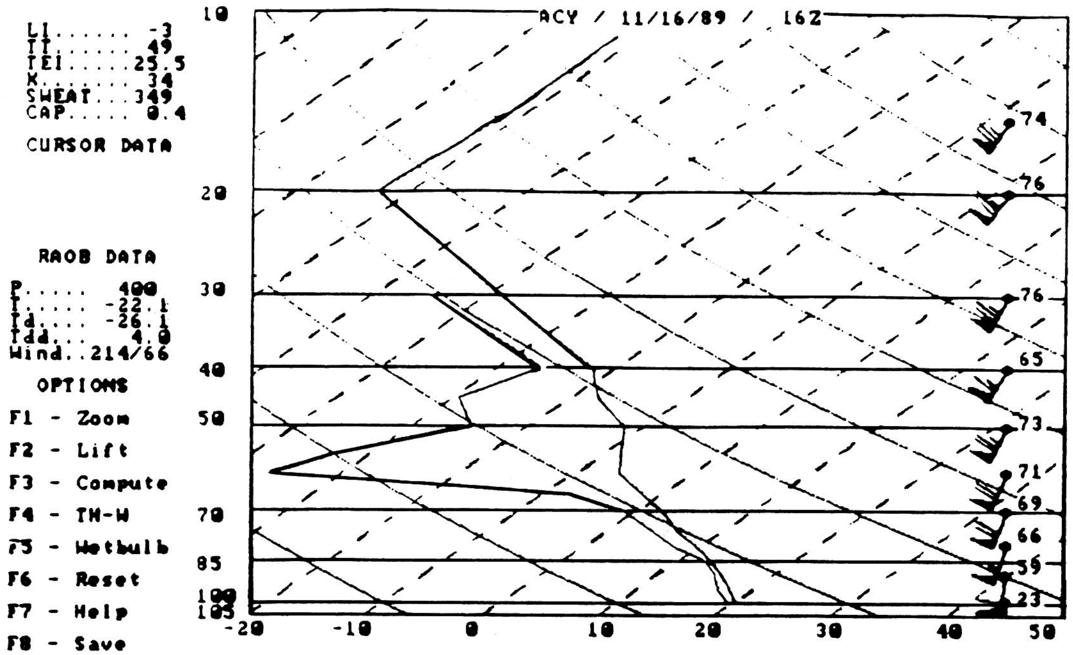


Figure 10. November 16, 1989 modified sounding for Atlantic City, NJ. From the SHARP Workstation (Hart and Korotky 1991).

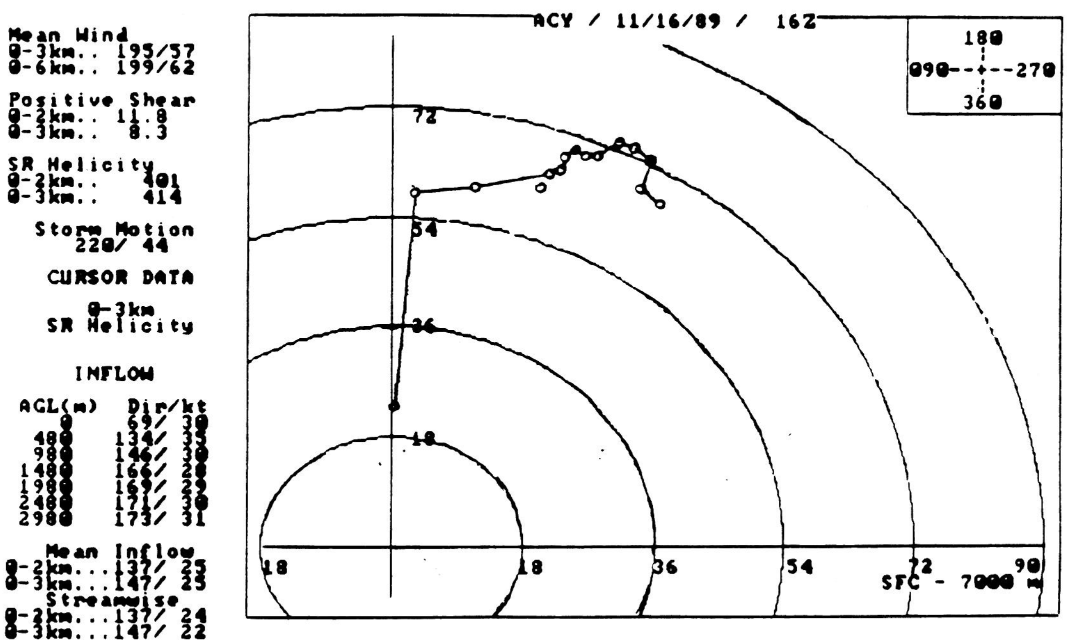


Figure 11. November 16, 1989 modified hodograph for Atlantic City, NJ. From the SHARP Workstation (Hart and Korotky 1991).