WSR-88D OBSERVATIONS OF MESOSCALE PRECIPITATION BANDS OVER PENNSYLVANIA

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Abstract

Weather Surveillance Radar-1988 Doppler (WSR-88D) located in central Pennsylvania detected a large area of snow, mixed precipitation, and rain over the central part of the state on 30-31 October 1993. It also detected an area of locally heavy rain over southeastern Pennsylvania on 4-5 December 1993. In both cases, distinct bands of enhanced precipitation were embedded within broader areas of precipitation. The bands were oriented along the 1000- to 500-mb thermal wind vector and were nearly parallel to the 850-mb isotherms. For both cases, the WSR-88D velocity azimuth display (VAD) wind profile (VWP) indicated a low-level northeasterly flow, which veered with height and became southwesterly aloft. Base velocity data from the 0.5 degree elevation scan revealed the classic warm advection "S"-shaped pattern over central Pennsylvania. The base reflectivity data showed several bands of precipitation, which were later determined to have formed in a conditionally stable to conditionally neutral atmosphere. For the October case, the rain / snow line remained near or just east of the Radar Data Acquisition (RDA) site during the event, which resulted in bright band contamination of the base reflectivity data in the band near the RDA. In both cases, the precipitation bands formed in a jet entrance circulation with low-level northerly ageostrophic winds that reinforced the cold air. For the October case, the surface low and warm front remained well south of the area. Soundings from Pittsburgh, Pennsylvania, and Dulles, Virginia, showed near neutral static stability and strong vertical wind shear. These two conditions are necessary for the development of Conditional Symmetric Instability (CSI) precipitation bands. Cross sections of equivalent potential temperature and geostrophic angular momentum (M) surfaces from both rawinsonde and model gridded data revealed that areas susceptible to CSI were present over central Pennsylvania for both cases. Low-level frontogenesis was also present for both cases and contributed to the formation of the banded precipitation over Pennsylvania in conjunction with the CSI present.

1. Introduction

The banded nature of precipitation associated with extratropical cyclones has been well documented (see Browning and Reynolds 1994; Browning 1986; Bennetts and Hoskins 1979; Hobbs 1978; Houze et al. 1976). Reuter and Yau (1990) showed that banded precipitation events were not a rare phenomenon during the Canadian Atlantic Storms Project (CASP). Hobbs (1978) noted that regions of locally heavy precipitation associated with oceanic cyclones occurred from mesoscale precipitation bands. The bands were typically 1000 km in length but contained within them smaller features on the order of 100 km. Sanders and Bosart (1985) showed that the width of the mesoscale band in the "megalopolitan snowstorm" of 11-12 February 1983, was also on the order of 100 km.

Conditional Symmetric Instability (CSI; Emanuel 1983) is one mechanism which can create banded precipitation. CSI is often referred to as **"slantwise"** convection because the instability causes **tilted** or **slanted** mesoscale circulations, similar to the way conditional instability causes upright convection. It has been shown by Wolfsberg et al. (1986), Moore and Blakely (1988), and Sanders (1986) that areas conducive to CSI are often located north of an approaching warm front or developing surface cyclone. The heavier precipitation appears as distinct mesoscale precipitation bands embedded in a larger area of synoptic scale ascent and precipitation. Due to the enhanced ascent within these mesoscale bands, areas affected by CSI receive more precipitation than surrounding areas.

Frontogenesis is another mechanism which can create banded precipitation (see Sanders and Bosart 1985; Sanders 1986; Wolfsberg et al. 1986; Gyakum 1987; and Moore and Blakely 1988). Petterssen (1936) defines frontogenesis in terms of an increase in the horizontal potential temperature gradient. Geostrophic confluence acting on a horizontal potential temperature gradient, in addition to increasing the temperature gradient, leads to a disruption of thermal wind balance by making the temperature gradient too large for the associated geostrophic vertical wind shear. A thermally direct transverse ageostrophic circulation is induced to restore thermal wind balance and is a manifestation of frontogenesis (Sanders and Bosart 1985). Mesoscale banded precipitation is thought to form in frontogenetic regions, since thermally direct circulations favor ascent in the warm air when the potential vorticity in the warm air is small (Emanuel 1985; Sanders and Bosart 1985).

The roles of CSI and frontogenesis in forming banded precipitation appear to be interrelated. Wolfsberg et al. (1986) note that symmetric instability and frontogenesis may not act independently, since the condition for ellipticity in the Sawyer-Eliassen equation for transverse frontal circulations is symmetric stability. Sanders and Bosart (1985) concluded that the banded precipitation associated with the "megalopolitan snowstorm" of 11-12 February 1983, was a result of frontogenetic forcing within or adjacent to a region of small symmetric stability. Wolfsberg et al. (1986) and Sanders (1986) both concluded that the band formation in the New England snowstorm of 11 December 1982 was a product of both frontogenesis and CSI.

On 30 October 1993, a developing surface cyclone along the East Coast produced a combination of rain and snow over Pennsylvania and New York. This storm brought unusually early heavy snows to the Northeast, with over 30 cm falling in many portions of northern and western Pennsylvania and upstate New York. On 5 December 1993, a slow moving surface low along the East Coast produced heavy rains across the Mid-Atlantic region. Between 7.5 and 12.5 cm of rain fell over portions of southeastern Pennsylvania. In both cases, the heaviest precipitation over Pennsylvania fell from organized linear bands which were detected by the Central Pennsylvania (KCCX) Weather Surveillance Radar-1988 Doppler (WSR-88D). Similar to Sanders and Bosart (1985), Sanders (1986), Wolfsberg et al. (1986), and Moore and Blakely (1988), the precipitation bands, for both cases, were aligned nearly parallel to the thermal wind.

The purpose of this paper is to demonstrate that the WSR-88D was able to identify mesoscale precipitation bands for both cases mentioned above. Observed and model data were used to diagnose the environment in which the bands formed during each event. In both cases, short-term model forecasts indicated that conditions were favorable for the development of mesoscale precipitation bands prior to their detection on radar. We will emphasize the operational use of model and observed data for the prediction of banded precipitation areas. In addition, the use of the WSR-88D to detect mesoscale precipitation bands will also be discussed.

2. Methodology

The WSR-88D data used in this case is from the central Pennsylvania Doppler radar (KCCX radar). The Radar Data Acquisition (RDA) site is located in central Pennsylvania approximately 17 km to the northwest of State College and University Park airport (KUNV), Pennsylvania. During both events, the KCCX radar had no digital archive level II capabilities. Archive level II data from the Sterling, Virginia (KLWX) radar was available for 5 December 1993, and was examined in this study. All other WSR-88D data used in this study are from the archive level III database from the KCCX radar. The reader is referred to Klazura and Imy (1993) for a description of WSR-88D products.

Conventional meteorological surface and upper-air data were obtained from The Pennsylvania State University Research, Operational Meteorological, and Training System (PROMETS; Cahir et al. 1981). Gridded numerical weather prediction model data were obtained from the NOAA/NWS National Centers for Environmental Prediction (NCEP) in Camp Springs, Maryland. Gridded data from the NCEP operational Regional Analysis and Forecast System (RAFS; Hoke et al. 1989) Nested Grid Model (NGM) were also used in this study. For the 30 October case, NGM data were limited to 6-h forecasts out to 48-h. For the 5 December case, forecasts were available at 3-h intervals.

NGM gridded data were displayed using the GEneral Meteorological PAcKage, version 5.1 (GEMPAK 5.1; desJardines et al. 1991). All derived quantities were obtained from built-in GEMPAK functions. These functions were used to compute geostrophic angular momentum surfaces (MSFC) and frontogenetic forcing (FRNT). The geostrophic wind was used to compute momentum surfaces, thus the geostrophic angular momentum was examined. Model-gridded winds were used unless otherwise noted.

3. Halloween Snowstorm, 30–31 October 1993

a. Synoptic overview

On 30 October 1993, a large area of precipitation moved across Pennsylvania from the south and west. The northern and western edge of the precipitation was primarily in the form of snow. The area of snow extended from northern West Virginia across western Pennsylvania into southern New York State. Heavy snow fell across much of northern and western Pennsylvania and south central New York State, while most of the East Coast of the United States received significant rainfall from this system.

The large-scale flow over the United States on 30 October 1993, is depicted in Fig. 1. Over the northeastern United States and Mid-Atlantic states region, conditions were setting up for a large-scale precipitation event. At 0000 UTC 30 October, a broad 250-mb jet extended from northeastern Mexico across the Mississippi Valley and into southern New England (Fig. 1a). A jet streak was embedded within this broad jet core over the Ohio Valley (Fig. 1a). A second surge of short wave energy, evident over Montana and Wyoming, was moving down the west side of the 250-mb trough. At lower levels, an 850-mb low was located over northern Louisiana with a broad baroclinic zone extending northeastward from this low into northern New York and southern Canada (Fig. 1c). Twelve hours later the 250-mb jet streak moved northward over northern New England and intensified (Fig. 1b). A second jet streak became apparent over the Ohio Valley and Pennsylvania by 1200 UTC. The 850-mb low was positioned in the entrance region of the 250-mb jet streak over the Ohio Valley (Fig. 1b and Fig. 1d).

Detailed surface analyses for 1200 and 1800 UTC 30 October are shown in Figs. 2a and 2b. At 1200 UTC, a surface low was located over western North Carolina with an inverted trough extending northeastward from the low into eastern Pennsylvania. The 850-mb thermal pattern at this time (Fig. 1d) suggests that this trough was the leading edge of the baroclinic zone at the surface. In addition to this trough, there was a hint of low-level cold air damming along the East Coast as shown by the pressure ridge extending from New Jersey southward into Virginia (Fig. 2a).

By 1800 UTC, the surface low over the Appalachian Mountains began to weaken as it moved into southwestern Virginia (Fig. 2b). A secondary surface cyclone developed over coastal Virginia. The isobaric pattern still suggested the presence of the inverted trough extending northward from the secondary low along the coast.

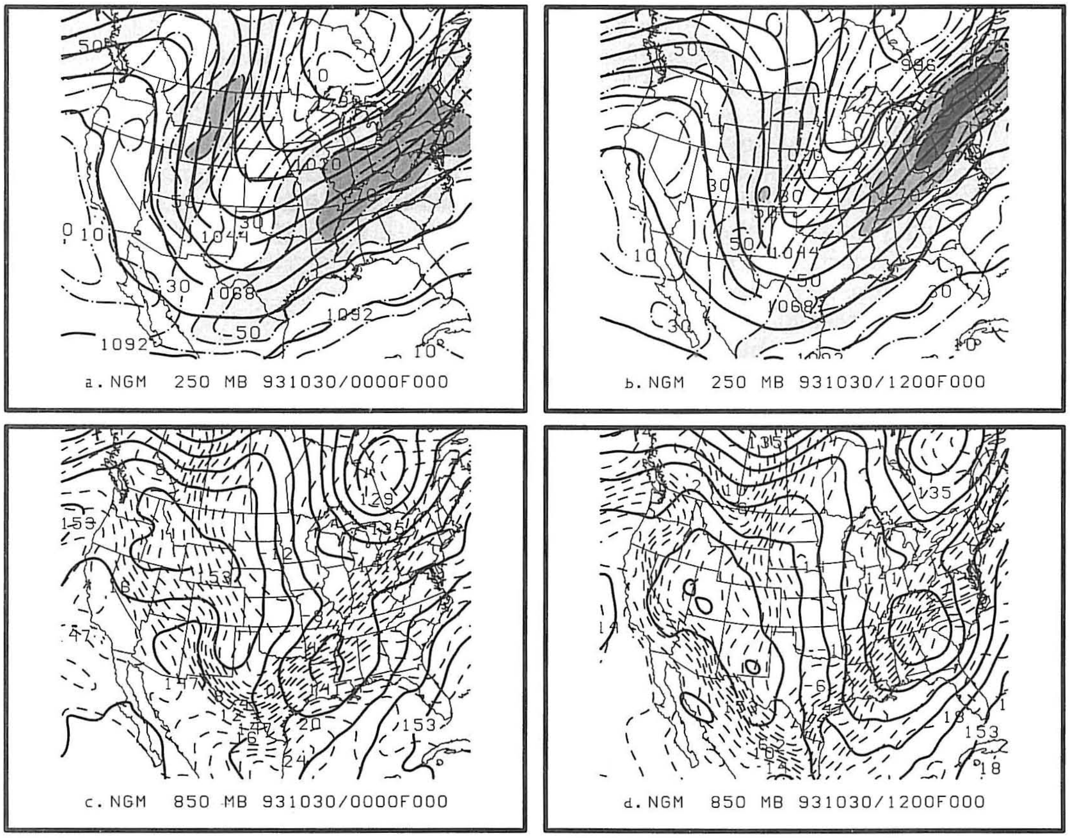

Fig. 1. NGM 00-h forecast of a) 250-mb heights and isotachs valid at 0000 UTC 30 October 1993, b) 250-mb heights and isotachs valid at 1200 UTC 30 October, c) 850-mb heights and isotherms valid at 0000 UTC 30 October, and d) 850-mb heights and isotherms valid 1200 UTC 30 October. Isotachs are contoured every 10 m s^{-1}, isotherms every 2 °C, 250-mb height contours every 12 dm, and 850-mb height contours every 30 dm.

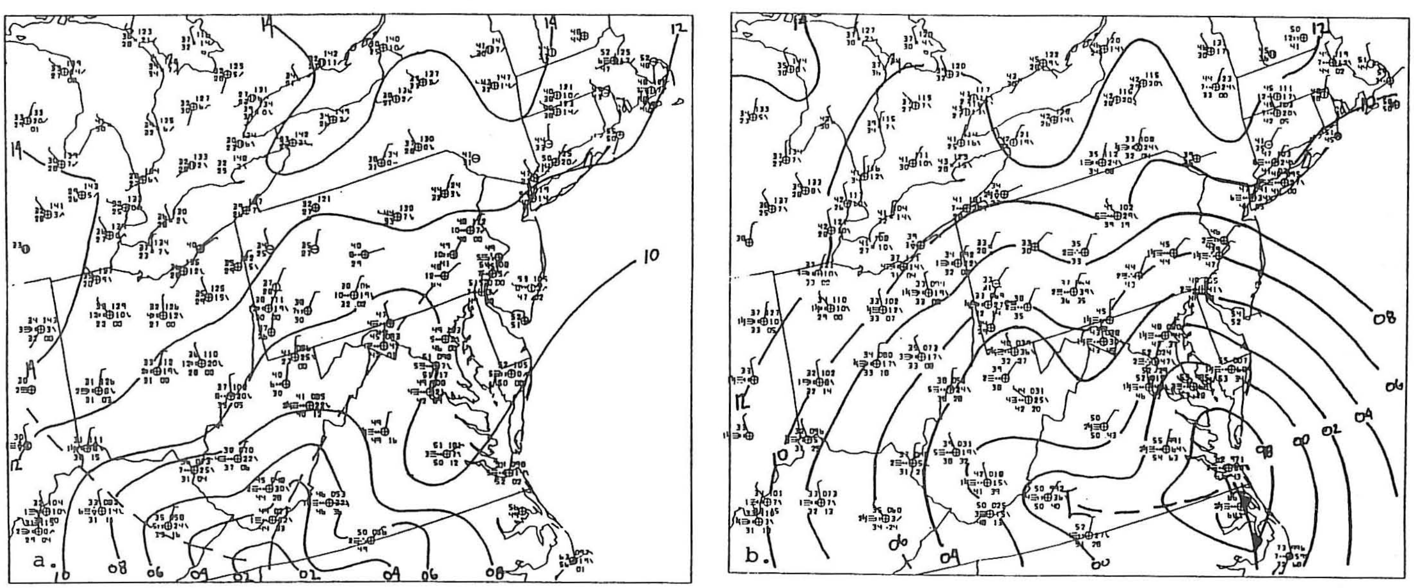

Fig. 2. Surface analyses for a) 1200 UTC and b) 1800 UTC 30 October 1993. The surface pressure is contoured every 2 mb.

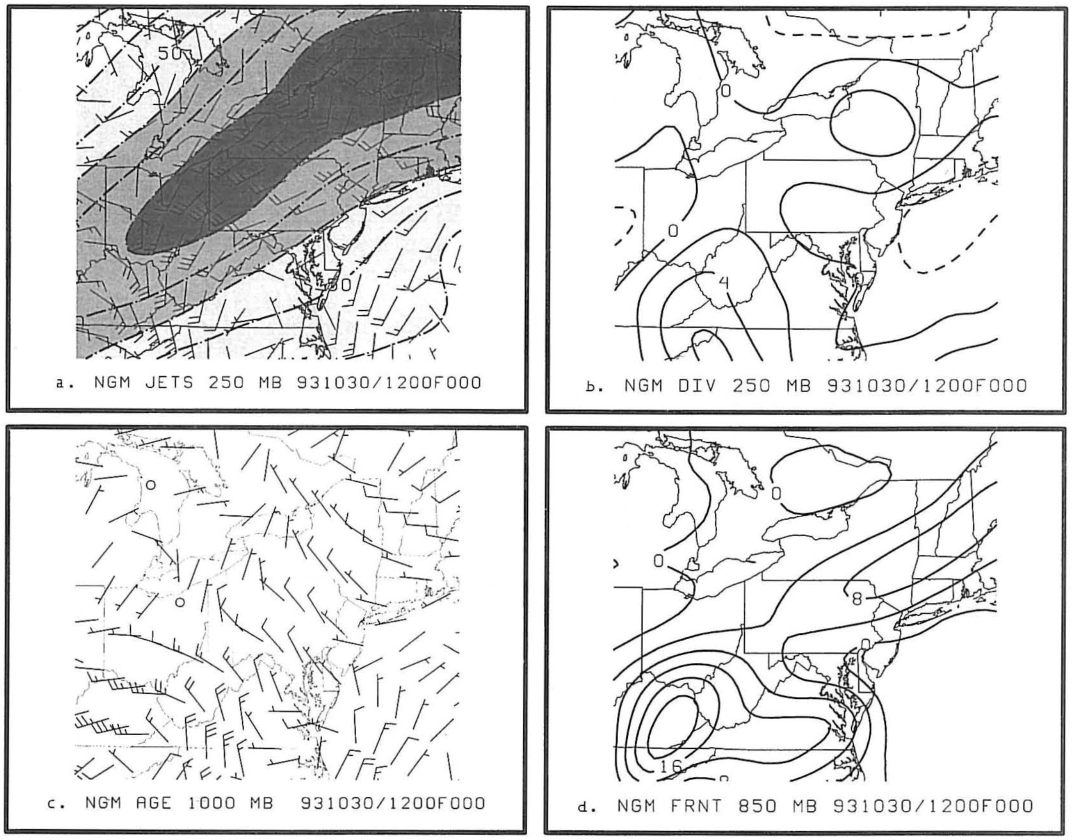

Fig. 3. NGM 00-h forecast valid at 1200 UTC 30 October 1993 of a) 250-mb isotachs and ageostrophic winds, b) 250-mb divergence, c) 1000-mb ageostrophic winds and d) 850-mb frontogenesis. Isotachs are contoured every 10 m s^{-1} with shading beginning at 40 m s^{-1}. Wind barbs are in m s^{-1}; divergence is 2 x 10-5 s^{-1}; and frontogenesis is every 4 x 10-9 K s^{-1}.

The NGM 00- and 06-h forecasts of 250-mb isotachs, ageostrophic winds and divergence; the 1000-mb ageostrophic winds; and 850-mb frontogenetic forcing are shown in Figs. 3 and 4, respectively. At 1200 UTC 30 October, a large area of upper-level 250-mb divergence (Fig. 3b) was located over the central Appalachian Mountains near the entrance region of the Ohio Valley jet streak (Fig. 3a). The upper-level ageostrophic winds were from the south in the entrance region from eastern North Carolina into West Virginia and became more southwesterly over Pennsylvania. At 1000 mb (Fig. 3c), northerly ageostrophic winds dominated the central Appalachians. The associated secondary ageostrophic circulation with this jet entrance region contributed to the area of strong frontogenesis that extended eastward from eastern Kentucky to coastal Virginia (Fig. 3d).

Figs. 3a and 3b reveal the second, more subtle jet streak over central New York and western New England. The signature is evident in both the ageostrophic circulation (Fig. 3a) and the 250-mb divergence field (Fig. 3b). The northwesterly lower-level ageostrophic winds across New York, Pennsylvania and New England supported the secondary frontogenetic maxima which extended from central Pennsylvania northeastward into central New England (Fig. 3d). The precipitation bands began to form around 1200 UTC, in the vicinity of this smaller secondary frontogenetic region over central and northeast Pennsylvania.

By 1800 UTC, both jet streaks had moved northeastward and were located over New York State and New England (Fig. 4a). A large area of 250-mb divergence (Fig. 4b) and 850-mb frontogenetic forcing (Fig. 4d) shifted to the Mid-Atlantic region. As a result of strong frontogenesis, the 850-mb baroclinic zone (not shown) shifted to the coast. The weaker secondary area of frontogenesis shifted into northern New England and, subsequently, the precipitation bands over Pennsylvania became less organized.

In order to assess the potential for CSI with this case, a cross-section of geostrophic angular momentum (M) and theta-e surfaces, valid at 1200 UTC 30 October 1993 from the 00-h NGM, was taken normal to the thermal wind across central Pennsylvania (Fig. 5). Figure 5 clearly reveals that the atmosphere over central Pennsylvania was favorable for the formation of CSI bands close to the times at which the bands were observed on radar. It should be noted that areas supporting CSI occur when the theta-e contours slope more steeply than the M-surfaces on a cross section taken orthogonal to the thermal wind. This is evident between 800 mb and 700 mb over central Pennsylvania at 1200 UTC (Fig. 5).

b. WSR-88D observations of banded precipitation

A few select WSR-88D radar products from 30 October are shown in this study. However, this section was written after examining time loops of the data (not shown). Around 1200 UTC, a broad area of light precipitation was falling over central Pennsylvania. However, at 1204 UTC, two distinct southwest to northeast areas of banded precipitation were observed over the region (Fig. 6a; for clarity, all returns less than 20 dBZ were filtered out).

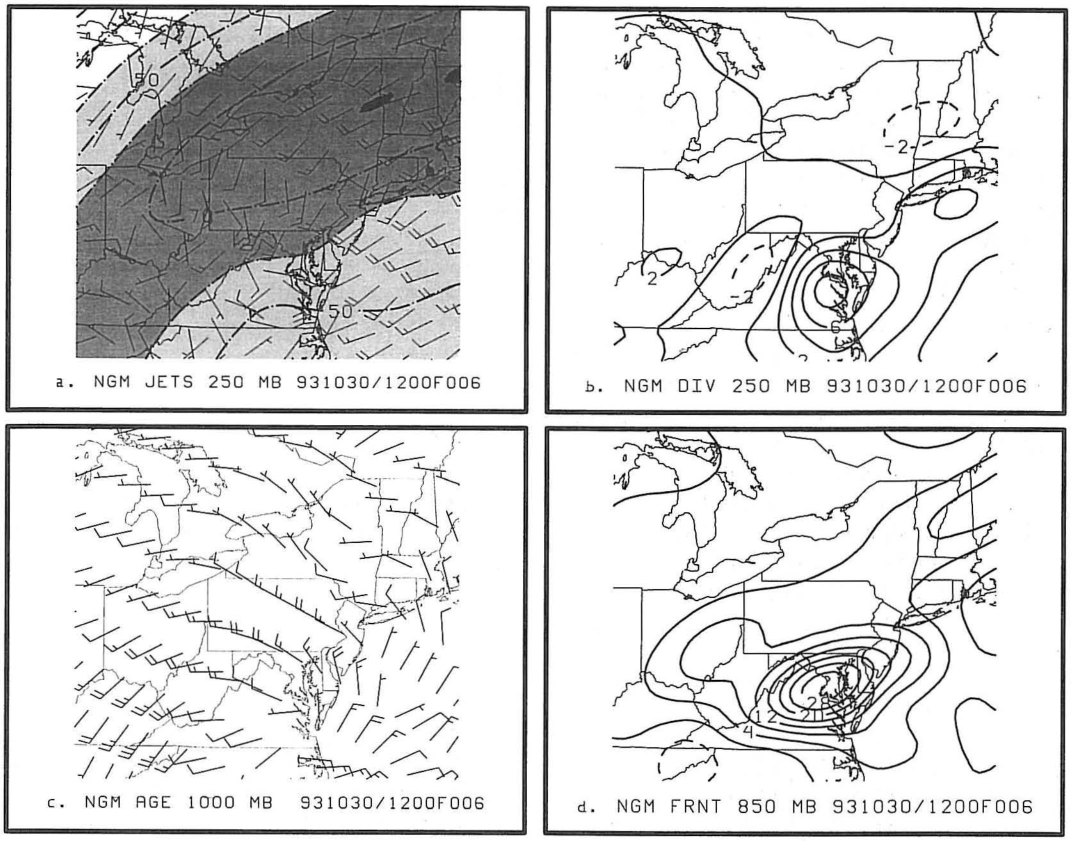

Fig. 4. As in Fig. 3, except for the NGM 06-h forecast valid at 1800 UTC 30 October.

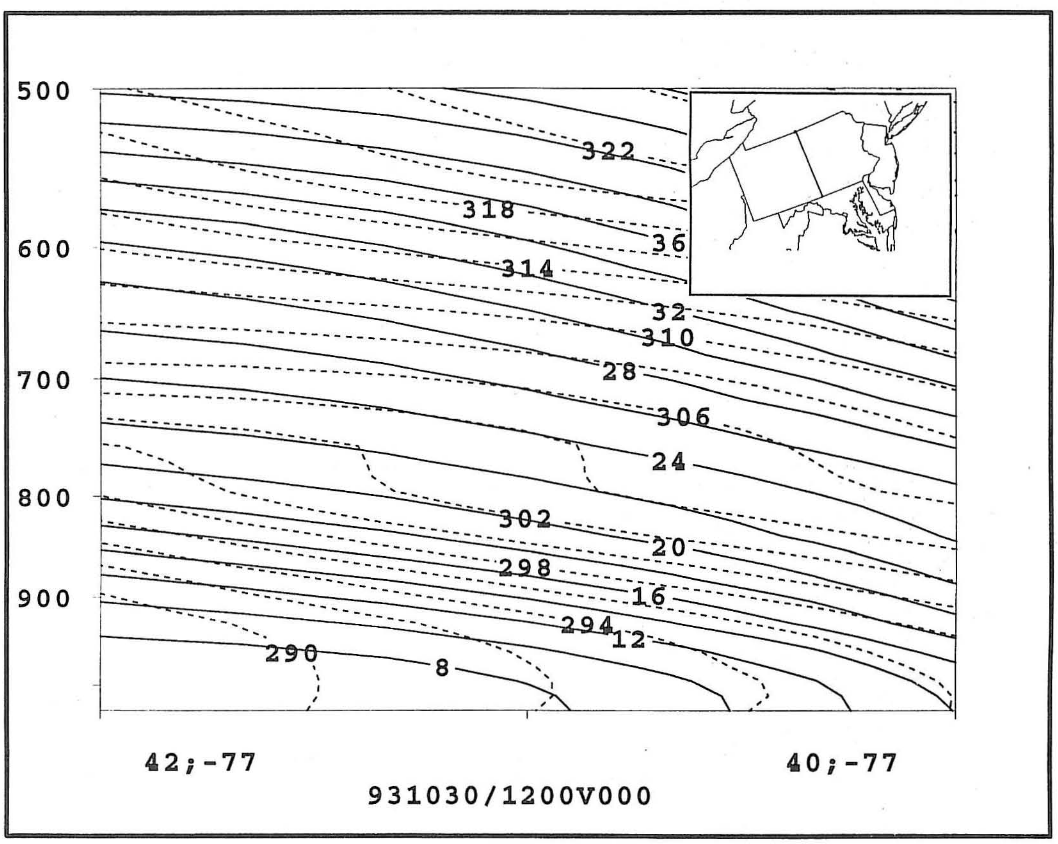

Fig. 5. Vertical cross section of geostrophic angular momentum (M) surfaces (solid) and equivalent potential temperature (dash) valid at 1200 UTC 30 October 1993. The contour interval is 2K for equivalent potential temperature and 2 s^{-1} kg^{-1} for M. Cross section latitudes and longitudes are labeled along the ordinate (cross section extends across Pennsylvania). Vertical axis is pressure (mb).

[Ed. Please see color figures 6, 7, 8 and 14 on page 18 and 19.] These bands were nearly parallel to the low-level shear vector indicated in the velocity azimuth display (VAD) wind profile (Fig. 7; hereafter, VWP). By 1417 UTC (Fig. 6b), the northern band extended from north of Pittsburgh northeastward to north central Pennsylvania. This band was the strongest, most organized and persistent band of the event. The band reached its maximum extent and intensity at 1423 UTC (Fig. 6c) with a maximum reflectivity 48 dBZ. After this time, the northern band began to weaken and fracture. By 1510 UTC (Fig. 6d), the band had moved northeastward and split into several smaller, weaker bands. Shortly after this time, it was difficult to discern any organization in the precipitation bands over northern Pennsylvania.

South of the strong linear band, several weaker, less organized bands formed. One persistent band can clearly be seen in Figs. 6b and 6c, approximately 80 km south of the northern band. This band also remained nearly parallel to the low-level wind shear vector. However, the southern band never attained as high a degree of linearity as the northern band.

The 0.5 degree elevation angle base velocity data valid at 1406 and 1510 UTC are shown in Fig. 8. During the time the precipitation bands attained their maximum extent and linearity, weak low-level northeasterly flow was present near the RDA (Fig. 8a), as indicated by the inbound (green) and outbound (red) couplet near the RDA. A broad zero isodop (gray shade in Fig. 8) separated the inbound and outbound velocities, indicating warm advection as characterized by the "S" shape in the isodop. The 1406 UTC base velocity product revealed that the upper-level winds were from the southwest, in close agreement with the upper-level wind patterns shown in Fig. 3a. By 1510 UTC (Fig. 8b), the low level northeasterly winds had increased in intensity, as illustrated by the increasing areal extent of the maximum inbound and outbound velocities located near the RDA. This increase in low-level northeasterly flow occurred near the time when the mesoscale precipitation bands became less organized.

4. December Heavy Rain Case

a. Synoptic overview

On 4-5 December 1993, a surface cyclone developed along the East Coast and moved northeastward over the western Atlantic Ocean. The surface pressure was relatively low over the northeastern United States, and there was no source of low-level cold air, resulting in rain. In contrast to the 30-31 October case, the mesoscale precipitation bands that formed in this case were not stationary, but slowly propagated to the north.

The large scale flow over the United States at 0000 and 1200 UTC 5 December 1993 is shown in Fig. 9. Similar to the 30-31 October event, a broad 250-mb jet extended from the Gulf Coast northeastward into New England (Fig. 9a). An 850-mb low was located over central Kentucky with a baroclinic zone extending northeastward from this low (Fig. 9c). Over the next 12 hours, the upper-level trough and 850-mb low deepened and

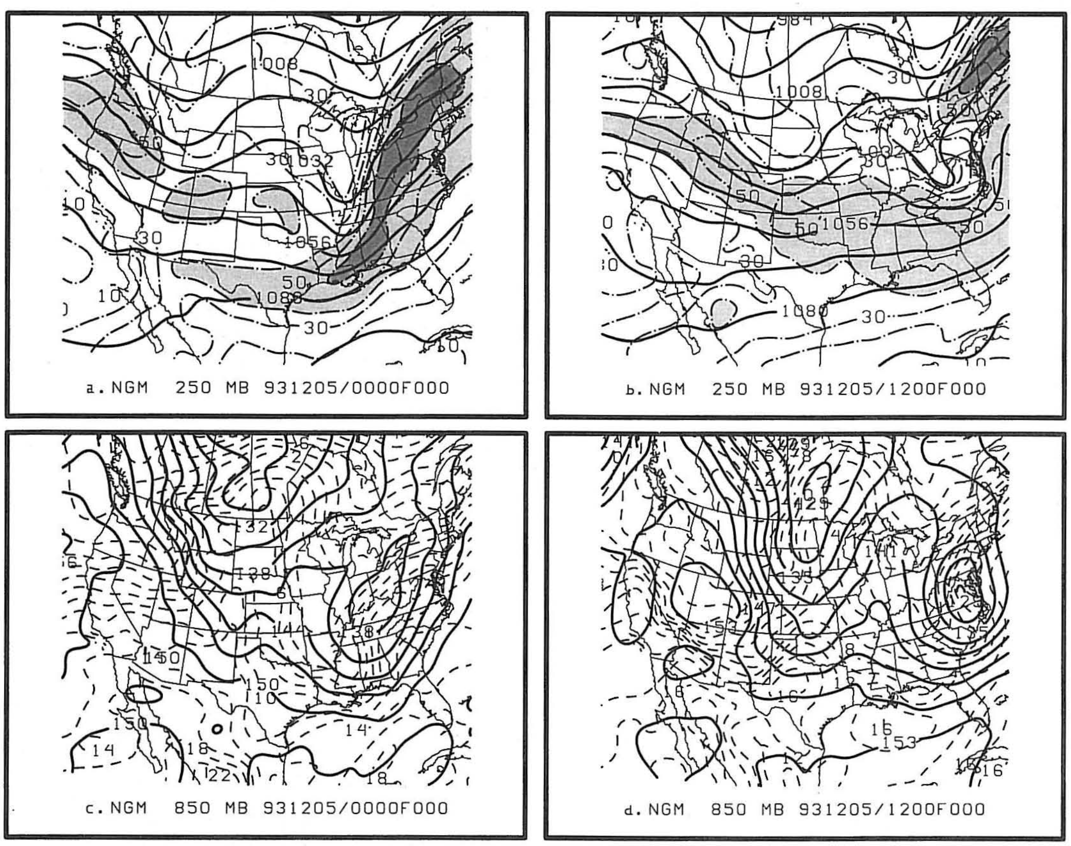

Fig. 9. As in Fig. 1, except for 5 December 1993.

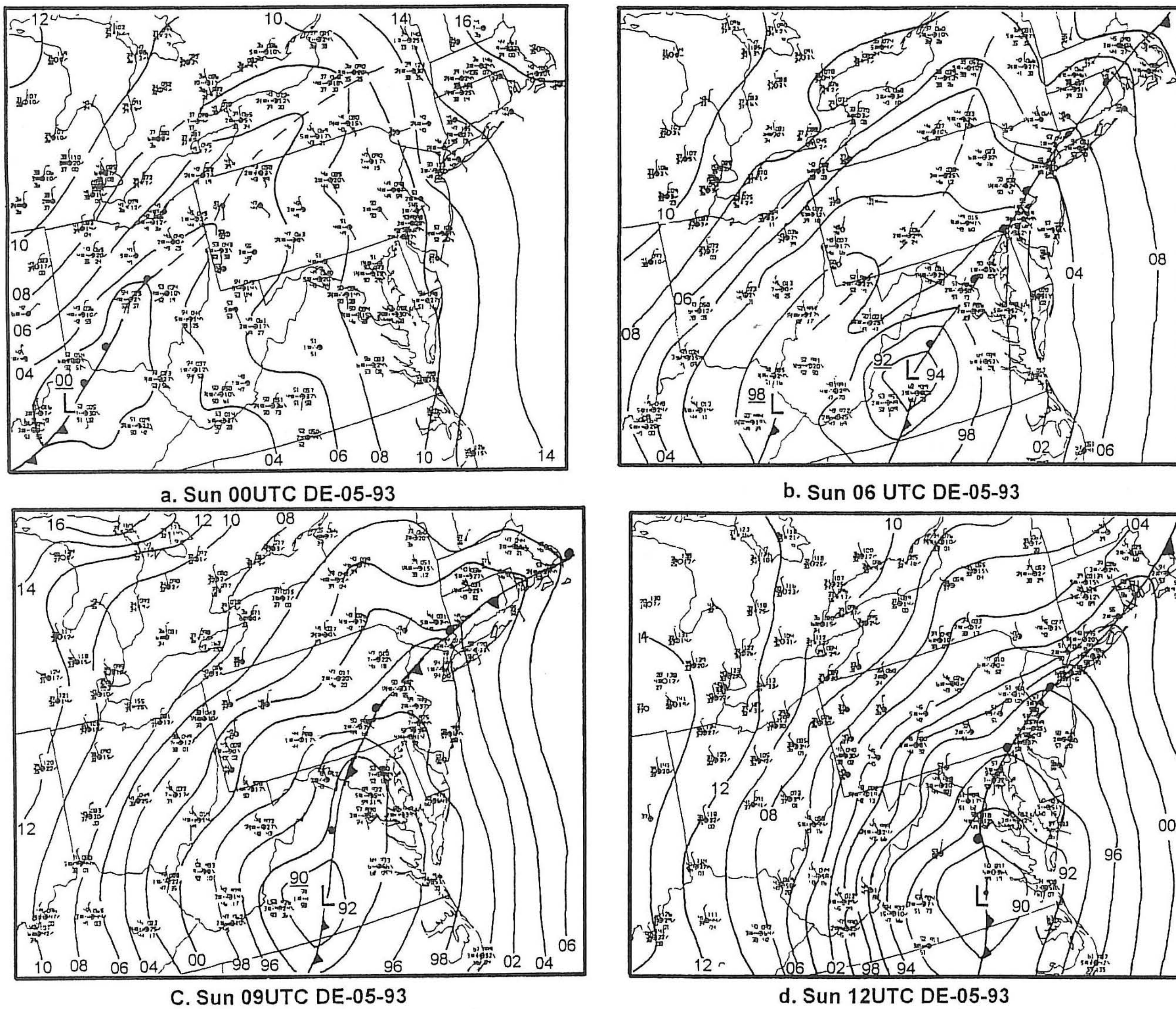

Fig. 10. Manual surface pressure and frontal analyses valid at a) 0000 UTC, b) 0600 UTC, c) 0900 UTC and d) 1200 UTC 5 December 1993.

moved to Virginia. From Fig. 9b, it is evident that the upper-level trough was becoming more negatively tilted.

Manual surface pressure and frontal analyses from 0000 UTC through 1200 UTC 5 December 1993, are shown in Figs. 10 a-d. A surface low was located over northern Kentucky at 0000 UTC (Fig. 10a) with a weak frontal boundary extending northeastward from the low. By 0300 UTC (not shown), a secondary surface low began to develop to the lee of the Appalachian Mountains, in Virginia. By 0600 UTC, the secondary low circulation in Virginia (Fig. 10b) became the primary surface circulation and the low in eastern Kentucky began to fill. The low-level baroclinic zone moved southeastward to the New England coast. By 0900 UTC (Fig. 10c), the low in the Appalachian Mountains was no longer identifiable, and the quasi-stationary front that extended northward from the surface low had buckled over the Mid-Atlantic states region suggesting that weak waves were moving along the front. By 1200 UTC, the low and associated fronts had shown little movement (Fig. 10d). Strong cross-isobaric flow was evident from New York to northern Virginia on the cold side of the front, suggesting a strong low-level ageostrophic circulation. This ageostrophic flow was tied to the thermally direct jet entrance circulation of the upper-level jet shown in Figs. 9a and 9b.

The NGM 03- and 06-h forecasts of 250-mb isotachs, ageostrophic wind, and divergence; 1000-mb ageostrophic wind; and 850-mb frontogenetic forcing are shown in Figs. 11 and 12, respectively. For the 03-h forecast, valid at 0300 UTC 5 December, a large area of 250-mb divergence was present over the central Appalachians, with a secondary area of divergence over Pennsylvania (Fig. 11b). The NGM depicted two distinct jet maxima within the broad upper-level jet (Fig. 11a), which produced the sinusoidal divergence/convergence couplets evident in Fig. 11b. A broad area of 1000-mb northerly ageostrophic winds (Fig. 11c) was present across much of the eastern United States, indicative of cold air damming (Forbes 1987) and the presence of a thermally direct jet entrance circulation (Uccellini and Kocin 1987). The strongest 850-mb frontogenesis was located over West Virginia (Fig. 11d), but frontogenetic forcing extended northward into New England.

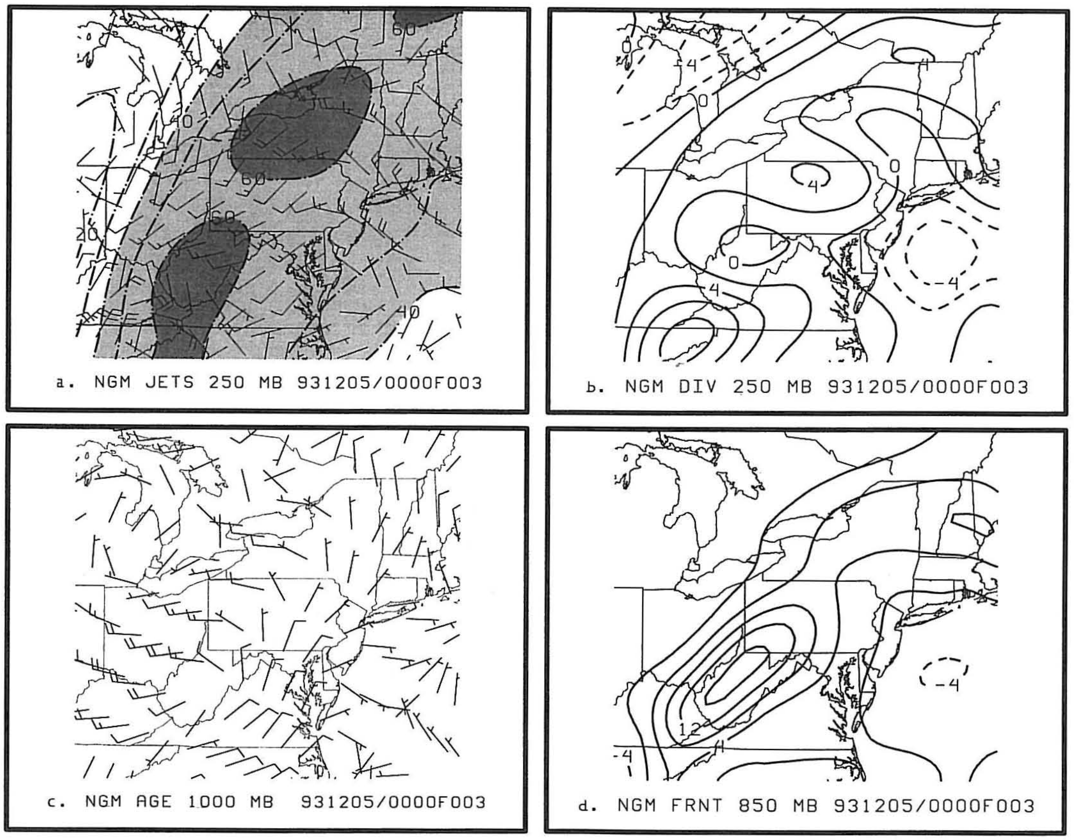

Fig. 11. As in Fig. 3, except for the 3-h forecasts from the 0000 UTC 5 December NGM run valid at 0300 UTC 5 December 1993.

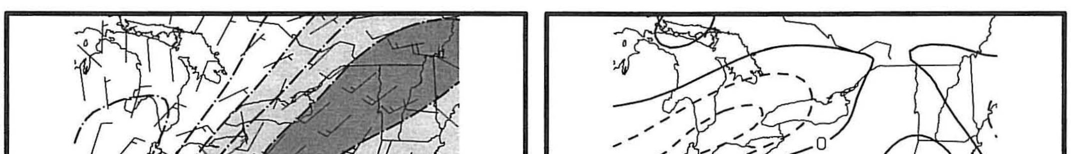

Fig. 12. As in Fig. 4, except for the 6-h forecast valid at 0600 UTC 5 December 1993.

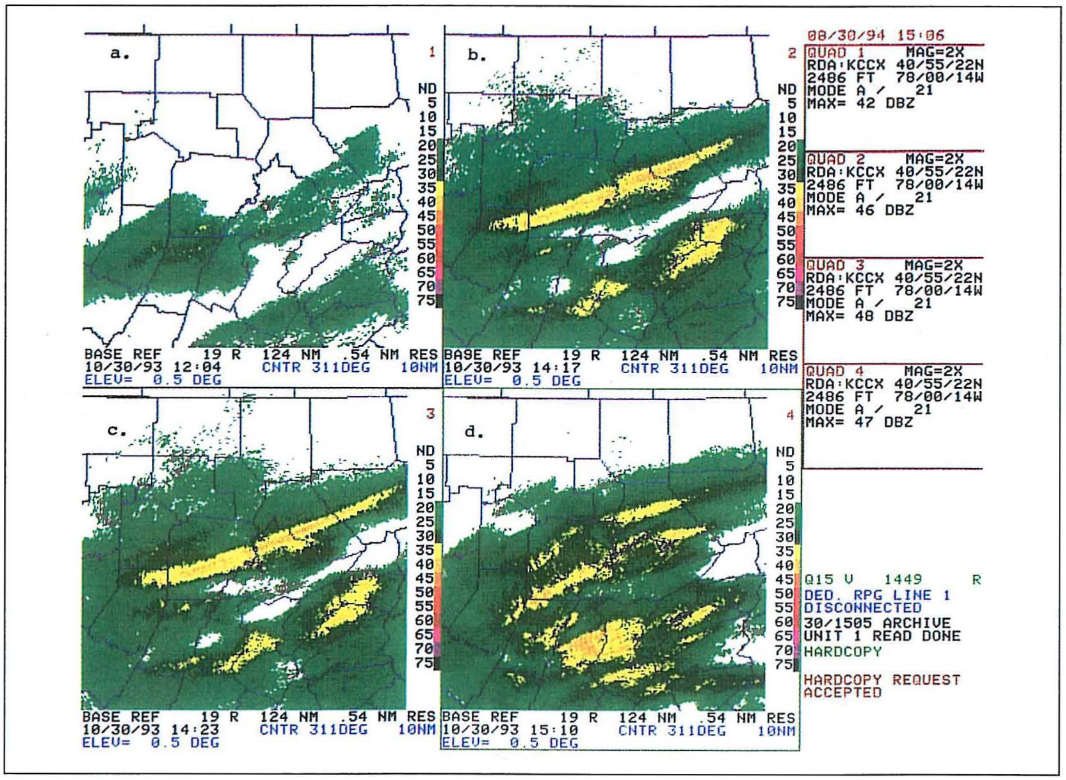

Fig. 6. Central Pennsylvania WSR-88D base reflectivity products valid at a) 1204 UTC, b) 1417 UTC, c) 1423 UTC, and d) 1510 UTC 30 October 1993. The color tables on the right depict the intervals of the reflectivity in units of dBZ. Reflectivity values less than 20 dBZ are not shown. The data resolution is 1° by 1 km and the elevation angle is 0.5°.

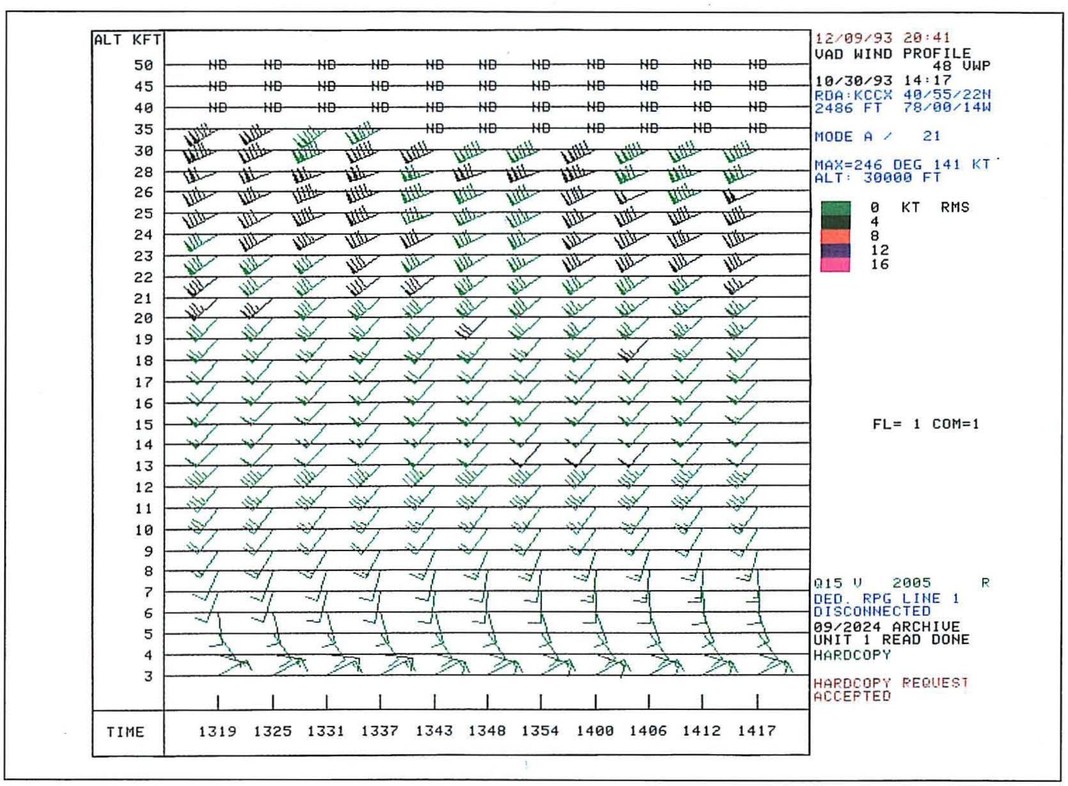

Fig. 7. Central Pennsylvania WSR-88D velocity azimuth display wind profile product on 30 October 1993 for 1319 (lower left) to 1417 (lower right) UTC. Vertical axis is altitude in thousands of feet. Here, ND indicates no data. Wind barbs are in knots. Color table on the right depicts the root-mean-square error (in knots) for each wind estimate.

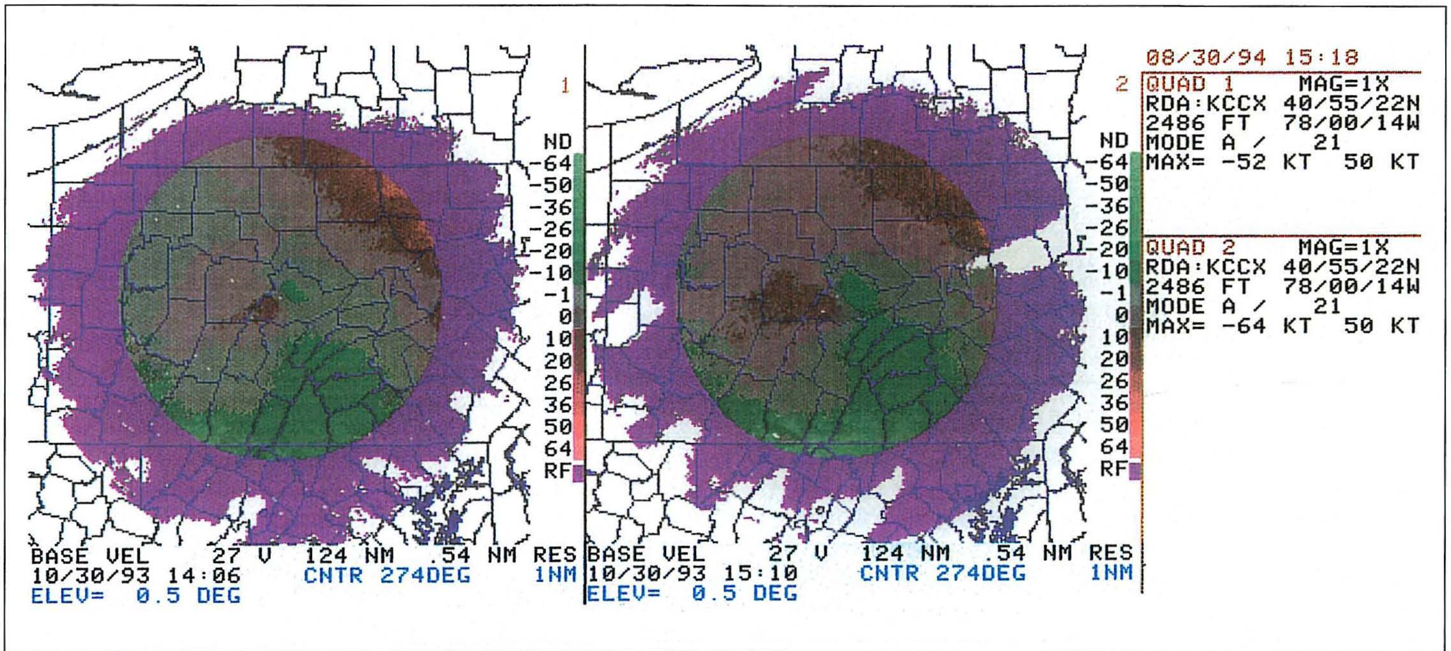

Fig. 8. Central Pennsylvania WSR-88D 0.5 degree base velocity products valid at a) 1406 UTC and b) 1510 UTC 30 October 1993. Resolution is 1° x 0.54 nm (1 km). The color tables on the right of each panel depict the intervals of velocity in units of knots and the "RF" denotes range folding. Green colors (negative values) indicate radial velocities toward the radar and red colors (positive values) indicate outbound velocities.

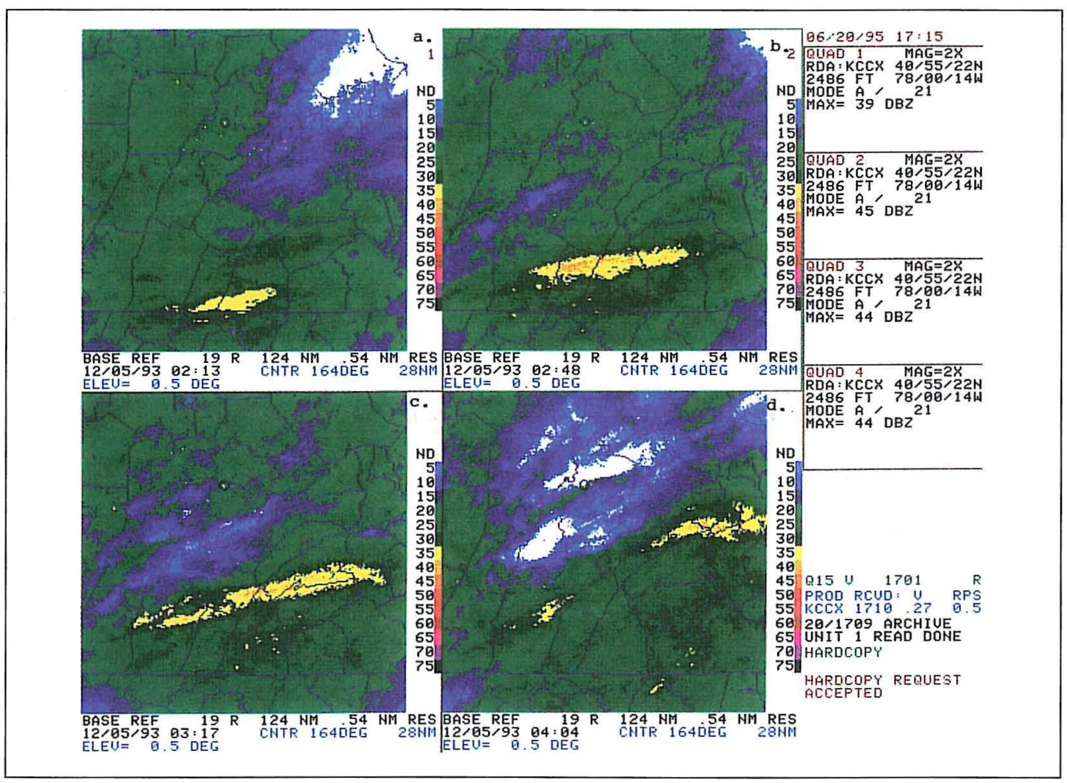

Fig. 14. As in Fig. 6, except valid at a) 0213 UTC, b) 0248 UTC, c) 0317 UTC, and d) 0404 UTC 5 December 1993.

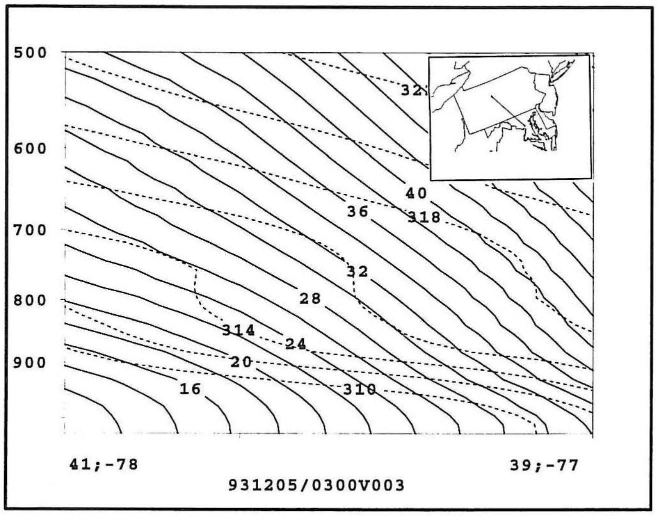

Fig. 13. As in Fig. 5, except for 0300 UTC 5 December 1993.

By 0600 UTC, the jet streak indicated by the NGM became more organized over New York and New England (Fig. 12a) and the area of 250-mb divergence shifted eastward (Fig. 12b). The 850-mb frontogenetic region had also shifted eastward and intensified (Fig. 12d). A broad area of frontogenesis was present over southeastern Pennsylvania close to where the precipitation bands had formed.

A cross-section of geostrophic angular momentum (M) and theta-e surfaces, valid at 0300 UTC 5 December 1993 from the 00-h NGM, was taken normal to the thermal wind across central Pennsylvania (Fig. 13). Figure 13 reveals a region between 900 and 700 mb where the theta-e surfaces slope more steeply than the M-surfaces. Thus, it can be concluded that there was a layer of CSI over central Pennsylvania close to the times at which the bands were observed on radar.

b. WSR-88D observations of banded precipitation

The first indications of banded precipitation occurred shortly after 0100 UTC on 5 December (not shown). By 0213 UTC, the KCCX WSR-88D detected a pronounced precipitation band over southeastern Pennsylvania (Fig. 14a; *page 19*). The band obtained a linear appearance by 0230 UTC (not shown) and increased in horizontal extent as it moved northward. By 0248 UTC, the band extended across southeastern Pennsylvania with a maximum reflectivity of 45 dBZ (Fig. 14b). The band began to rotate counterclockwise as it continued to move northward, reaching its largest horizontal extent at 0317 UTC (Fig. 14c). After 0317 UTC, the band began to weaken and fracture as it rotated counterclockwise as shown in Fig. 14d. The band moved approximately 100 km to the north in the 2-h period between 0200 and 0400 UTC.

5. Discussion

Similar to Sanders and Bosart (1985), Wolfsberg et al. (1986), Sanders (1986), Gyakum (1987), and Moore and Blakely (1988), the precipitation bands for each case developed within regions of large scale frontogenesis and CSI. The mesoscale precipitation bands for each case had time scales on the order of 3- to 6-h, which was similar to the findings of Reuter and Yau (1990), Wolfsberg et al. (1986), and Kreitzberg and Brown (1970).

The precipitation bands observed over central Pennsylvania on 30 October 1993, were oriented parallel to the shear vector in the 1.5 to 5.5 km layer. As shown in Fig. 6, the largest band remained quasi-stationery during the event and lasted for several hours. Wolfsberg et al. (1986) had a similar finding with the larger bands persisting for several hours and smaller bands lasting for 30 minutes to about an hour. Throughout the event of 30 October, the bands remained oriented from about 240° to 060° (west-southwest to east-northeast). As the event progressed the bands became less organized and smaller in size. However, they had a significant impact on the distribution of precipitation across Pennsylvania between 1200 and 1800 UTC 30 October (Fig. 15). A narrow band of 19.5 mm and greater precipitation was observed from north-central Pennsylvania east-northeastward into the southern tier of New York, in close proximity to the location of the largest precipitation band (Fig. 6). Most of southeastern Pennsylvania received greater than 19.75 mm of precipitation . Much of this precipitation was the result of a prolonged period of rain associated with the synoptic scale weather system. However, the local maxima in excess of 25 mm of precipitation was associated with CSI bands. The western portion of one of these bands can be seen in Fig. 6d. These bands were also observed on the KLWX radar (not shown).

Unlike the multi-banded, quasi-stationary 30 October case, the 5 December case was dominated by a large, single transient band, which moved northward with time (Fig. 14). Sanders (1986) observed a counterclockwise rotation of precipitation bands in a New England snowstorm. Similarly, the band on 5 December 1993, rotated counterclockwise as it moved northward (Fig. 14), remaining nearly parallel to the thermal wind vector.

Like the October case, the mesoscale precipitation bands on 5 December had a significant impact on the total precipitation in southeastern Pennsylvania (Fig. 16). Some areas under the northward moving band received between

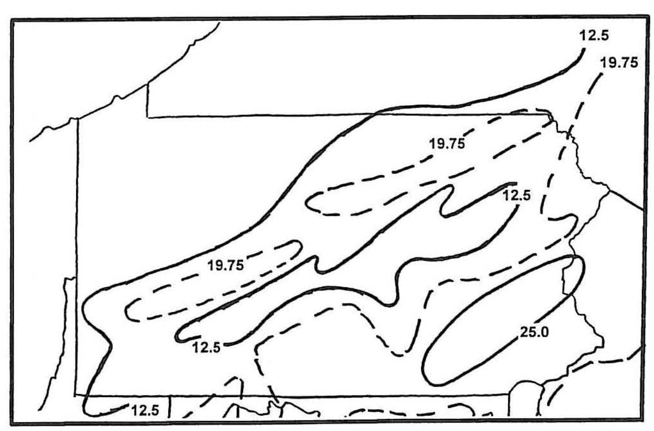

Fig. 15. Total observed precipitation (mm) for the 24-h period from 1200 UTC 30 October through 1200 UTC 31 October 1993. The contour interval is 12.5 mm. Dashed line shows the intermediate 19.75 mm contour.

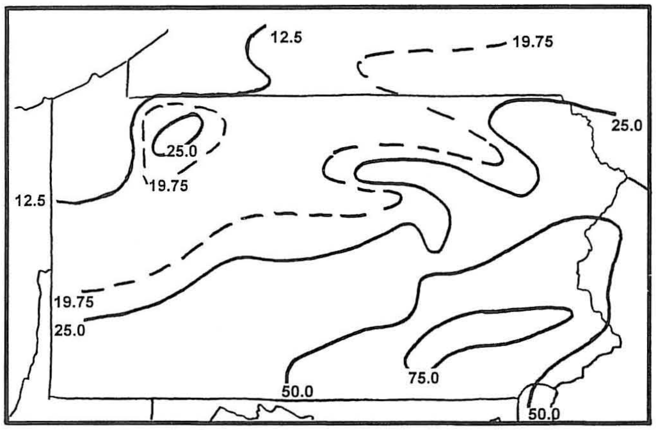

Fig. 16. As in Fig. 15 except for the 24-h period ending 1200 UTC 5 December 1993. Contour interval is 25 mm above 25 mm.

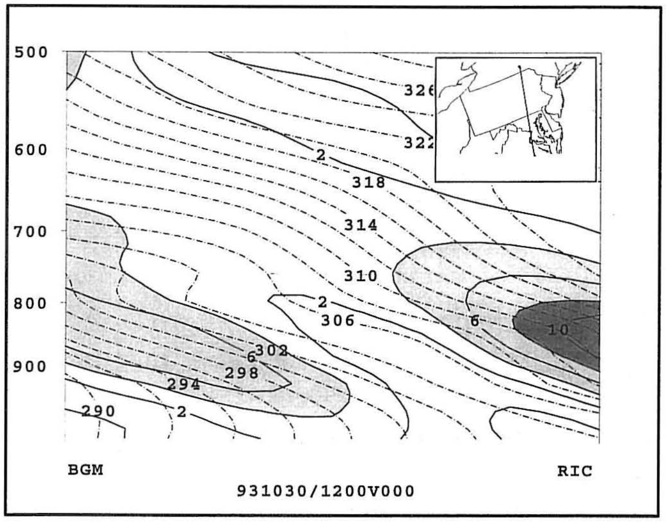

Fig. 17. Vertical cross section of equivalent potential temperature (dash) and frontogenesis (solid) valid at 1200 UTC 30 October 1993. The contour interval is 2K for equivalent potential temperature and every 2 x 10-9 K s^{-1} for frontogenesis. Cross section extends from Binghamton, New York to Richmond, Virginia across Pennsylvania (see inset in upper right). Vertical axis is pressure (mb). Shading denotes frontogenesis greater than 4 x 10-9 K s^{-1}.

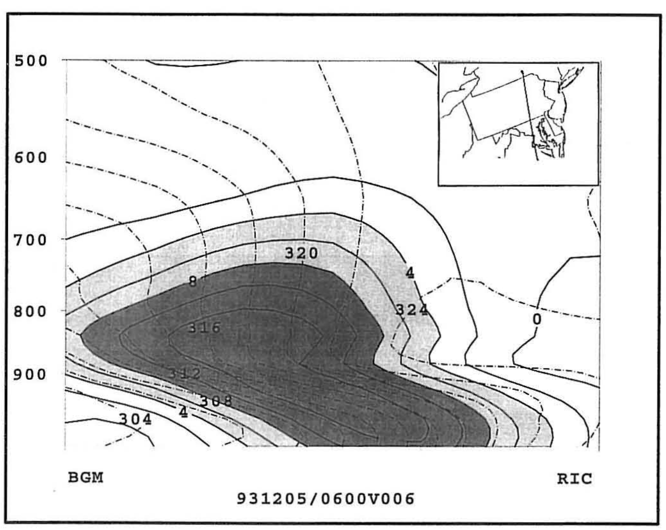

Fig. 18. Same as Fig. 17 except valid for 0600 UTC 5 December 1993.

50 and 75 mm of rainfall. The majority of this precipitation fell between 0000 and 1200 UTC.

Radar images from both the 30-31 October and 5 December 1993 events clearly indicate distinct precipitation bands embedded within a larger scale precipitation event. However, the nature of the bands differed between the two events. This distinction may be attributed to the differences in the low-level frontogenesis and the associated steepness of the equivalent potential temperature surfaces during the two events. Figures 17 and 18 show cross-sections of equivalent potential temperature and frontogenesis for the 30-31 October and 5 December 1993 cases respectively. The cross sections were taken normal to the 1000 to 500 mb thickness. The left edge of the cross-sections were located near Binghamton, New York and the right edge was located at Richmond, Virginia. The slope of the equivalent potential temperature surfaces were much steeper for the 5 December 1993 case where one single band dominated (Fig. 18) than in the 30-31 October 1993 case (Fig. 17). Similarly, the axis of maximum frontogenesis for the 5 December 1993 case was much steeper than the October case. This implies that the more stable case (30-31 October 1993) led to classic quasi-stationary CSI bands while the less stable (5 December 1993) case led to a single transient band. Detailed mesoscale model data could be used to test this hypothesis in future events.

Operationally, the development of mesoscale banded precipitation can be inferred from synoptic-scale signatures. Typically, band formation is favored in a thermally direct jet entrance circulation with low-level frontogenesis present. Strong vertical wind shear, and low inertial instability, which are typically found on the anticyclonic shear side of an upper-level jet, are two other synoptic features that favor band formation. With access to gridded model data, forecasters can better diagnose the potential for banded precipitation. Cross-sections of geostrophic angular momentum and equivalent potential temperature can be constructed normal to the thermal wind vector to diagnose regions susceptible to CSI. Gridded model data can be used to compute frontogenesis at various levels in the atmosphere. Forecasters can also examine absolute vorticity at several different layers to assess areas of low inertial instability. Finally, isotach fields and the associated ageostrophic winds can be diagnosed for the upper-level jet streak circulations.

When the synoptic-scale conditions favor the development of meso-scale banded precipitation, Doppler radar data should be examined to identify the development of the bands. National Weather Service short term forecasts can be modified to address the enhanced potential for heavy rain in locations near these bands. Furthermore, forecasters can anticipate changes in the orientation of the thickness patterns with time. If little change is expected, then band orientation and location is less likely to change, increasing the potential for heavy precipitation in areas near the bands. However, if model guidance suggests a change in the orientation of the thickness patterns, it is likely that new bands will develop reflecting these changes in the vertical wind shear. If it is determined that bands will be nearly stationary, then forecasters can anticipate that heavier precipitation will last on the order of 3 hours, which is a typical time scale for banded precipitation (Reuter and Yau 1990; Wolfsberg et al. 1986; and Kreitzberg and Brown 1970).

At the current time, mesoscale bands are not readily resolvable in numerical models, and accurate forecasting falls into the realm of the 0- to 6-h temporal frame. However, extensive use of 6-h model gridded data to identify short term changes in the thickness pattern, thermal fields, and vertical wind shear, is of great value. Although not shown, the analysis of absolute vorticity on theta-e surfaces (Hoskins et al. 1985) may further assist in identifying regions conducive for the formation of CSI bands.

6. Conclusions

The results of this study show that the KCCX WSR-88D, located in central Pennsylvania, did an excellent job detecting the mesoscale precipitation bands which formed on 30 October and 5 December 1993. Gridded numerical model and observed data, indicated that the atmosphere was nearly conditionally stable and that the necessary conditions were met to produce mesoscale banded precipitation. Clearly, the available data presented in this study showed the potential for mesoscale banded precipitation. The orientation of the bands could be anticipated prior to their appearance on radar.

In both cases the bands lasted only a few hours. Still, they had a significant impact on the storm total precipitation. From a short term forecasting perspective, it is important for the forecaster to be aware of the impact these mesoscale bands can have on the storm total precipitation, and to recognize the environment in which these bands are likely to form. Once the forecaster has diagnosed the conditions favorable for the formation of mesoscale precipitation bands, their appearance on radar can be anticipated.

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