A Winter Mesocyclone over the Midwestern United States

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ABSTRACT

A diagnostic analysis is made of a midwinter mesoscale vortex that developed over the Mississippi Valley and produced moderate to heavy snow with gale force winds (>18 m s⁻¹), lightning, and thunder along a narrow track approximately 1500 km in length. The mesoscale vortex resembled the so-called "polar lows" that form over the subpolar seas. The similarities include development on the cyclonic shear side of a long-wave trough, strong positive vorticity advection associated with a 500 mb short-wave trough, upstream tilt of the geopotential heights, conditional instability in the lower troposphere, a southeastward track, small (~200 km) diameter, and a 9 mb deepening of the surface mesoslow in 12 h. The most intriguing features of the present analysis are the extremely large potential vorticities and horizontal temperature gradients in the midtroposphere, indicating an extrusion of stratospheric air down to levels below 700 mb.

1. Introduction

During the period 4–6 February 1984, a mesoscale vortex that developed over the upper Mississippi Valley produced moderate to heavy snow and strong winds along a narrow track approximately 1500 km in length. The mesoslow reached its greatest intensity over eastern Missouri and southwestern Illinois, where the storm disrupted airline and highway traffic. The storm's development in a data-rich area permits a relatively precise description of its evolution, which is the subject of this paper.

The mesoscale vortex resembled the small-scale cyclones that form over the subpolar seas. These so-called "polar lows" typically form on the cyclonic side of the midtropospheric jet when cold air is destabilized during a southward trajectory over warmer waters (Rasmussen 1985; Mansfield 1974; Harrold and Browning 1969). Like the system analyzed here, polar lows are often accompanied by gale-force winds (>18 m s⁻¹), low visibilities, and sometimes thunder. Detailed studies of mesoscale vortices over the subarctic seas are difficult because of the storms' small scales and the sparsity of conventional observations from their remote location, although special programs [e.g., the Norwegian Polar Lows Project (see Rasmussen and Lystad 1987)] have focused on these maritime systems by utilizing non-conventional sources of data, especially satellite imagery. Shapiro et al. (1987a) describe a unique set of research aircraft measurements obtained in a polar low over the Norwegian Sea.

Several examples of mesoscale vortices over the wintertime continents have also recently been reported (e.g., Mullen 1982; Smart and Carr 1986). These continental systems are similar to the one described here in the sense that the mesoscale vortex developed in the absence of an obvious surface source of heat and moisture. The vortex in the present study appears to have been stronger than those of Mullen (1986) and Smart and Carr (1986). However, as discussed in sections 3 and 4, the distinguishing feature of our analysis is the apparent association with a midtropospheric front and an apparent extrusion of stratospheric air to elevations as low as 700 mb (~3 km). Similar stratospheric extrusions have been detected by Uccellini et al. (1985) and Shapiro et al. (1987b) in association with synoptic-scale events, but wintertime continental mesoslows have yet to be studied in a framework that includes both surface mesoslow development and troposphere-stratosphere interaction.

The goals of the analysis that follows are 1) to document a wintertime mesoslow that produced severe winter weather in the heart of the conventional observing network of the United States, and 2) to suggest that the upper-tropospheric dynamics of corresponding maritime systems be investigated for similarities to the present system and for possible roles in mesoslow development.

2. Data and analysis methods

The primary database for this study consists of all conventional hourly surface and 12-h upper air reports from stations in the United States and Canada for the period 0000 UTC 4 February–1200 UTC 6 February...
1984. National Meteorological Center (NMC) analyses and forecasts were provided by the National Climate Data Center, Asheville, North Carolina. A series of high-resolution Defense Meteorological Satellite Program (DMSP) images in both visible and infrared were obtained from the National Snow and Ice Data Center, Boulder, Colorado.

The NMC products were supplemented with hand-drawn analyses of standard meteorological variables (sea level pressure, upper-air heights, temperatures, winds) as well as vertical velocity, divergence, and vorticity. In addition, objective analyses were made of derived kinematic quantities such as temperature and vorticity advection, potential vorticity, frontogenesis, and various contributions to frontogenesis.

3. Analysis

As background for the analysis of the storm’s evolution, Fig. 1 shows the time series of hourly surface synoptic reports from various stations near the storm’s track through the Midwest. The compact size of the storm is apparent. Near-blizzard conditions occurred for several hours at major reporting stations such as Des Moines, Iowa, and St. Louis, Missouri, while stations only ~200 km distant (e.g., Champaign, Illinois, and Columbia, Missouri) experienced few effects of the storm. The mesoscale nature of the system will also be apparent in the satellite imagery discussed later.

At 0000 UTC 5 February (Fig. 2a), the first indications of a developing wave appeared on a cold front
which had moved from the United States–Canada border to central Minnesota during the past 12 h. There was no evidence of a closed circulation, but pressure falls of 2–3 mb over the past 3 h were common. The NMC radar summary for 0035 UTC showed cloud tops to 3400 m (11 000 ft) over eastern South Dakota and western Minnesota. Watertown, South Dakota reported a cumulonimbus cloud before the frontal passage.

The 500-mb height/vorticity analysis (Fig. 3) for 0000 UTC showed a ridge over western North America and a deep long-wave trough over the central United States. On the eastern side of the ridge, a short-wave trough over southern Manitoba contained a vorticity maximum of $22 \times 10^{-5}$ s$^{-1}$. The speed of the polar jet at 500 mb had increased by 40–50 knots during the past 12 h. Diffluent flow was indicated over northern Minnesota, where kinematically computed divergences were as large as $4.0 \times 10^{-5}$ s$^{-1}$. Height falls of 30–80 m and a maximum of vorticity advection ($5.6 \times 10^{-9}$ s$^{-2}$) were also found in this region.

By 0600 UTC 5 February, the frontal wave had moved approximately 450 km southeast to central Iowa (Fig. 2b). The surface winds indicated the presence of a cyclonic vortex with a diameter less than 200 km. The central pressure was approximately 1003 mb. Stations northwest of the storm center were reporting sustained winds of 13–18 m s$^{-1}$ (25–35 knots), moderate-to-heavy snow, near-zero visibility, and rapidly rising pressures. Wind gusts in excess of 26 m s$^{-1}$ (50 kt) were reported during the next several hours at Des Moines.

Diagnostic computations with the data for 0600 UTC showed a band of strong frontogenesis extending from east central Minnesota to central Nebraska. Local
maxima of the rate of change in potential temperature gradient were 1.76 K (100 km)$^{-1}$ (3 h)$^{-1}$ over east-central Iowa. Closer examination of the frontogenesis showed that the changes were dominated by the geostrophic wind in the initial stages (≈0000 UTC). However, once the mesocyclone formed, the ageostrophic contribution to the adiabatic frontogenesis became dominant.

At 1200 UTC 5 February 1984, the center of the mesolow was approximately 50 km northeast of St. Louis. In the past 6 h, the central pressure had decreased by 3 mb to 1000 mb. Several different fronts were identified in the hand-drawn analysis (Fig. 2c). The cold front southwest of the low center, and the stationary front to the northeast are consistent with the NMC analysis for 1200 UTC. Our analysis of a cold front extending northward from the low in Fig. 2c is supported by the surface winds, which are easterly at 5 m s$^{-1}$ (10 kt) at Peoria and Springfield, Illinois, but northwesterly at 13 m s$^{-1}$ (25 kt) at Quincy, Illinois and 18 m s$^{-1}$ (35 kt) at Burlington, Iowa. Peoria and Springfield are 6–8°C colder than the stations west of the analyzed front, while the pressure tendency is negative ahead of the front and positive behind it. Stations behind this front were reporting gale-force wind gusts (>18 m s$^{-1}$), moderate to heavy snow, and rapidly falling temperatures (see Fig. 1). Heavy snow was also reported immediately ahead of the front in southern Illinois, as shown by Carbondale’s report for 1350 UTC (Fig. 1).

The mesocyclone is strikingly evident in the DMSP imagery for 1128 UTC (Fig. 4). While the “head” of the cyclone’s comma cloud is only ≈200 km in diameter, the satellite image shows cumulonimbus-like cloud features near and south of the vortex center. Indeed, lightning and thunder accompanied the snow squalls as the system passed over St. Louis shortly after 1200 UTC. The NMC radar summary for 1135 UTC
indicated maximum cloud tops of 3660 m (12 000 ft) immediately west of St. Louis.

The surface diagnostic computations for 1200 UTC showed that the strongest frontogenesis (see Appendix) had increased over 70% from 0600 UTC to a maximum of 3.02 K (100 km)⁻¹ (3 h)⁻¹ northwest of the cyclone’s center. The band of strongest frontogenetic activity extended southwestward from the Iowa–Illinois border, in close agreement with the satellite depiction of cloud activity. An objective analysis of the surface wind and air temperature fields (Fig. 5) illustrates the relatively sharp changes of wind speed and direction at the frontal boundary, together with the strong advection of cold air behind the front.

Objective analysis of the 500 mb data for 1200 UTC showed a long-wave pattern similar to that in Fig. 3, but with a vorticity maximum of 27.2 × 10⁻⁵ s⁻¹ in southeastern Iowa (Fig. 6). The strong vorticity near the maximum was due primarily to the wind shear. The maximum 500 mb vorticity advection, 8.4 × 10⁻⁹ s⁻¹, was found over western Illinois, while the maximum computed divergence was 35.0 × 10⁻⁶ s⁻¹ near the storm center. The objective analysis procedure on which these values (and Fig. 6) are based is a Barnes analysis scheme utilizing rawinsonde data from ten stations as input at each grid point. The grid point spacing of approximately 80 km is comparable to the 90-km resolution of NMC’s now-operational Nested Grid Model. By contrast, the analysis produced by NMC's operational Limited Fine-mesh Model (LFM) shows a vorticity maximum of only 23 × 10⁻⁵ s⁻¹ in eastern Iowa (Fig. 7). Since the response functions are different in the two analysis procedures, these differences are not surprising.

The area of strong 500 mb divergence in western Illinois is found directly above a layer of strong convergence (see Figs. 2c and 5). Convergence was indeed the major contributor to surface frontogenesis in this region. The shallow layer of intense convergence existed in a region of weak static stability (discussed later), thereby favoring the rapid spinup of the disturbance.

Although the LFM correctly forecasted the position
of the 500 mb vorticity maximum, the forecast maximum value was only $21 \times 10^{-3}$ s$^{-1}$. Moreover, the 24- and 12-h LFM forecasts failed completely to indicate the presence of a surface mesocyclone (e.g., Fig. 8). Both forecasts moved the cold front farther south than observed as indicated by the forecast isolines of 1000-500 mb thickness (Fig. 8). The 12-h LFM forecast of sea level pressure was 6-10 mb too high over western Illinois and southeastern Missouri. The LFM forecast did predict a trailing surface trough (and hence surface vorticity) over Arkansas, although the forecast was too fast.

It should be emphasized that the forecast of a 200 km mesoscale vortex is clearly not within the theoretical capability of the LFM, since 200 km is less than twice the grid spacing of the LFM. Even the innermost C-grid of the currently operational Nested Grid Model is likely to be too coarse for the accurate simulation of a feature having a characteristic horizontal scale on the order of only two grid points. Moreover, if frontoge- netic forcing was important to the development, as it appears to have been, then a key component of the forcing is also beyond the LFM’s theoretical capability.

Vertical velocities computed by both the kinematic and adiabatic methods (see Appendix) indicated strong ascent ahead of the mesolow and cold front at the 850 and 700 mb levels. The kinematically derived maxima at these levels were $-2.3 \times 10^{-3}$ and $-5.2 \times 10^{-3}$ mb s$^{-1}$, respectively, near the storm’s center. The corresponding adiabatically derived values computed from
the sounding for Salem, Illinois (~150 km southeast of the storm center) were $-2.8 \times 10^{-3}$ and $-2.1 \times 10^{-3}$ mb s$^{-1}$ (see Fig. 9). While the adiabatically derived vertical velocities at Salem (SLO) were moderately strong and upward at all levels from 850 to 500 mb, the vertical motion was quite weak at all levels 250 km to the north at Peoria (PIA), again indicating the relatively small scale of the system. The adiabatic computations showed strong descent behind the front at Topeka (TOP), Kansas, especially below 700 mb (Fig. 9). The upper troposphere over southwestern Missouri (UMN), eastern Kansas (TOP), and Nebraska was characterized by moderately strong descent ($\sim 4 \times 10^{-3}$ mb s$^{-1}$) in both the adiabatically and kinematically derived fields of vertical velocity.

Figure 10 shows the 1200 UTC sounding for Salem, Illinois (SLO), which was near the leading edge of the mesocyclone comma cloud at this time. A conditionally unstable layer is found between 800 and 850 mb. While the surface convective temperature is approximately 5°C, the lifting index required for a positive "energy area" is +1.6, a value generally conducive to shower and thunderstorm activity.

By 1800 UTC 5 February, the mesolow had moved into west-central Kentucky and was still producing

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**Fig. 4.** A DMSP infrared photograph of mesocyclone at 1128 UTC 5 February 1984.

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**Fig. 5.** Objective analysis of surface air temperatures and wind vectors at 1200 UTC 5 February 1984. Dashed lines are isotherms at intervals of 4°C.
moderate to heavy snow. Even at 0000 UTC 6 February, a well-defined cyclonic circulation was indicated over central Tennessee in the GOES infrared imagery (Fig. 11). The storm then weakened as it moved over the Appalachians between 0000 and 1200 UTC, although it still produced substantial amounts of unforecast snow at several locations in the Appalachians (Brody 1985). By 1200 UTC 6 February, the pressure was 1007 mb in the storm’s center over eastern North Carolina. The cyclone moved offshore of North Carolina late 6 February. The narrowness of the band of heavy snow deposited by the storm is apparent in the DMSP visible image of the middle Mississippi and Ohio River valleys for 1129 UTC 6 February (Fig. 12). New snow amounts were 2–8 cm (1–3 inches) over eastern Missouri, southern Illinois, and southwestern Indiana, although amounts of 15–18 cm (6–7 inches) were recorded in the mountains of western North Carolina and southwestern Virginia (Brody 1985). None of these

Fig. 6. Objective analysis of 500 mb height (m, solid lines) and vorticity \((\times 10^{-5} \text{ s}^{-1})\), dashed lines) for 1200 UTC 5 February 1984. Analysis procedure is described in text.

Fig. 7. The LFM analysis of 500 mb height (m, solid lines) and vorticity \((\times 10^{-5} \text{ s}^{-1})\), dashed lines) for 1200 UTC 5 February 1984.
areas reported snow on the ground prior to the passage of the mesocyclone.

After moving offshore over the warm water, the storm reintensified and moved rapidly northeastward. By 1200 UTC 7 February, the NMC analyses show a low pressure center of 995 mb, approximately 450 km east of Halifax. The storm's trajectory was sufficiently far offshore that measurable precipitation was not reported along the mid-Atlantic or New England coasts. The system then tracked across Newfoundland into the Labrador Sea, where its central pressure fell to 991 mb at 1200 UTC 8 February. At all times after its passage over the Appalachians, the storm was depicted as a synopticscale cyclone on the surface analyses of NMC and the U.S. Navy.

A final portion of the analysis was motivated by the apparent presence of stratospheric air at relatively low (~3 km) levels during the storm's mature phase. Figure 13 shows a vertical north-south cross section of potential temperature, $\theta$, between Amarillo, Texas, and Sault Saint Marie, Michigan at 1200 UTC 5 February. The $\theta$-plot suggests the presence of a midtropospheric front between 500 and 700 mb in the Peoria, Illinois-Topeka, Kansas portion of the domain. The $\theta$-values (305-310 K), characteristic of the lower stratosphere over Sault Saint Marie are found at midtropospheric depths over Topeka and Dodge City, Kansas. A corresponding analysis of the 12-h change, $\Delta \theta$, of potential temperature between 0000 and 1200 UTC 5 February indicated warming of 5-10 K in a band descending from 300 mb over Green Bay, Wisconsin, and Peoria, to 700 mb over Topeka (where a 12.4 K warming was found at 500 mb).

Finally, cross-sectional plots were constructed of potential vorticity computed as $(\xi + f)(\partial \theta/\partial p)$, where $\xi$ is the relative vorticity on an isentropic surface and $f$ is the Coriolis parameter. Potential vorticity is a quasi-conservative quantity following a parcel's motion, and values exceeding $10^{-5}$ K mb$^{-1}$ s$^{-1}$ generally indicate stratospheric air (Uccellini et al. 1985; Shapiro et al. 1987b). Figure 14 shows that potential vorticities exceeding $2.0 \times 10^{-5}$ extend down to 700-800 mb over eastern Kansas and Nebraska at 1200 UTC 5 February, the time of the mesocyclone's lowest central pressure.

High values of potential vorticity often indicate extrusions of stratospheric air and are usually associated with midtropospheric frontogenesis (Danielsen 1968; Bosart 1970). Danielsen showed that the frontogenesis was actually a folding initiated by geostrophic motion, while Bosart showed that the frontogenesis occurs when the tropopause is high and cold to the southwest of the jet axis and low and warm to the northeast. Folding of the tropopause and subsequent frontogenesis can occur when there is a large gradient of potential temperature along the inclined tropopause. If a baroclinic zone exists in the vicinity of the tropopause and the isotherms are parallel to the flow, an upward increase in anticyclonic vorticity advection exists. According to Bosart's (1970) schematic representation, warming of the northeast section (the stronger side) of the jet axis, relative to the weaker winds southwest of the jet axis, causes a geostrophic imbalance due to the weakening of the pressure gradient force. Soon afterwards, the tropopause descends southwestward towards the surface. A large increase in kinetic energy can occur as the tropopause crosses a constant-pressure surface during its descent (Danielsen 1968). Bosart notes that a thermally indirect
Fig. 10. Sounding for Salem, Illinois, at 1200 UTC 5 February 1984. Temperature is shown by heavy solid line, dewpoint by heavy broken line, and pseudoadiabats by thin dashed lines.

Fig. 11. A GOES infrared photograph of mesocyclone at 0000 UTC 6 February 1984.
In the present study, a short-wave trough was indeed moving south-southeast around a long-wave trough in the vicinity of a baroclinic zone. Tropopause descent and/or an upper-level front were also apparent between Topeka and Green Bay. As the stratospheric air descended into the troposphere, vortex tubes were stretched. Conservation of potential vorticity then requires a decrease of static stability. It is quite likely that the combination of vertical stretching and baroclinicity increased the absolute vorticity sufficiently in the lower troposphere to lead to cyclogenesis.

4. Discussion

The primary purpose of this case study was to illustrate that rapid development of a surface mesolow can occur even without moisture in very cold environments over land. This development can occur beneath short-wave vorticity maxima that appear to be relatively minor perturbations on the backside of a long-wave trough. As shown by this example, operational numerical weather prediction models can identify the upper level vorticity signals, but cannot pinpoint the location of the surface cyclogenesis. The need for a monitoring of satellite imagery and surface reports is apparent.

This case also shows that “thundersnow” can occur over land in midwinter mesoscale disturbances. Strong forcing aloft appears to be the key to the development of thunder and lightning in an Arctic-like environment in which surface temperatures are below 0°C.

Despite its track through the heart of the midwestern United States, the mesoscale vortex identified here has

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**FIG. 12.** A DMSP visible photograph depicting snow cover at 1129 UTC 6 February 1984.

**FIG. 13.** Cross-sectional analysis of potential temperature (K) at 1200 UTC 5 February 1984.
many features in common with the so-called “polar lows” identified in earlier studies (Harrold and Browning 1969; Mansfield 1974; Mullen 1982; Smart and Carr 1986). These similarities include development on the cyclonic-shear side of a long-wave trough, strong positive vorticity advection associated with a 500 mb short-wave trough, upstream tilt of the geopotential heights, conditional instability in the lower troposphere (Fig. 10), a southeastward track, small (~200 km) diameter, and a 9-mb deepening of the surface mesoslow during 12 h. The rate of deepening satisfies Forbes and Lott’s (1985) intensification criterion for polar mesoscale vortices.

However, the term “polar lows” encompasses systems having a wide variety of evolutionary, structural and dynamical characteristics. The midwestern mesoslow differs from most “polar lows” in its development over a continental region far from a major surface source of heat and moisture. The complex system of frontal boundaries identified here is also atypical of previously studied systems (Rasmussen and Lystad 1987, p. 802).

There is no simple structural or evolutionary “model” even for maritime polar lows. For example, the one maritime polar low measured intensively by research aircraft (Shapiro et al. 1987a) formed in a synoptic-scale environment different from that in which many other polar lows have developed over the Norwegian Sea. The satellite imagery of Shapiro et al. indicated that their case was merely one of a family of five polar-low cloud systems in the Norwegian Sea region during the last several days of February 1984.

Perhaps the most intriguing features of the present analysis are the extremely high potential vorticities and horizontal temperature gradients in the midtroposphere. While stratospheric extrusions have been documented in large-scale synoptic cyclogenesis (Uccellini et al. 1985) and arctic outbreaks (Shapiro et al. 1987b), their association with wintertime mesocyclone development has not been documented. Midtropospheric fronts have, however, been noted in previous mesoslow analyses (e.g., Reed 1979; Smart and Carr 1986). The role played by the upper-air dynamics in the mesocyclone development is unclear. Because of the sparsity of upper-air measurements in the vicinity of developing polar lows, the frequency with which stratospheric extrusions accompany these systems is also unknown. The analysis presented here suggests that upper-level dynamics merit further investigation in observational analyses and model studies of developing polar lows.

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APPENDIX

Formulation of Diagnostic Quantities

Frontogenesis was evaluated quantitatively using Petterssen’s (1956) expression for the rate of change of the gradient of potential temperature,

$$\frac{d}{dt} |\nabla \theta| = \frac{1}{2} |\nabla \theta| \left[ \left( \frac{\partial u}{\partial x} - \frac{\partial w}{\partial y} \right) \sec 2a \cos 2b - \nabla \cdot \mathbf{v}_h \right],$$

where $\theta$ is the potential temperature, $u$ the wind speed, $a$ the angle between the $x$-axes and the axis of dilatation, $b$ the angle between an isentrope and the axis of dilatation and $\mathbf{v}_h$ the horizontal wind vector. The first of the two terms on the right-hand side is the deformation of the wind field, while the second is the divergence. When $b$ is less than 45 deg, the deformation term will contribute to frontogenesis.

The kinematic vertical velocities were computed using the following expression:

$$\omega(P) = \omega(P_0) - \int_{P_0}^P (\nabla \cdot \mathbf{v}) dP,$$

where $\omega(P)$ is the vertical velocity at the pressure $P$, $P_0$ is a reference pressure level, and $\nabla \cdot \mathbf{v}$ is the divergence of the wind field between two pressure levels. The wind fields (e.g., Fig. 5) were obtained from a Barnes-type analysis in which the rawinsonde winds from ten stations were used as input at each grid point. The vertical velocity was assumed to be zero at the reference level $P_0 = P_{100}$, and a constant correction was applied to the divergence profile in order to ensure that $\omega(100 \text{ mb}) = 0$. 

Fig. 14. Cross section of potential vorticity ($10^{-5}$ K mb s$^{-1}$) along 39°N between 95°W and 88°W at 1200 UTC 5 February 1984.
The adiabatic vertical velocities were computed by assuming that the potential temperature, \( \theta \), is conserved following a parcel:

\[
\omega = - \left[ \frac{(\partial \theta / \partial t) + V(\partial \theta / \partial s)}{\partial \theta / \partial P} \right],
\]

where \( V \) is the horizontal wind speed, \( \partial \theta / \partial t \) local change of potential temperature on an isobaric surface, \( \partial \theta / \partial s \) local variation of \( \theta \) along a streamline, and \( \partial \theta / \partial P \) vertical gradient of \( \theta \) at a specific time and horizontal location.

In the computations leading to Fig. 9, \( \partial \theta / \partial t \) was approximated by the 24-h change between 0000 UTC 5 February and 0000 UTC 6 February as deduced from the radiosonde soundings. The \( \partial \theta / \partial P \) was calculated from the raw radiosonde data for 1200 UTC February 5. The advection term was computed from the objective analyses of \( \theta \) and wind velocity for 1200 UTC 5 February.

REFERENCES


