Evolution of a Surprise Snowfall in the United States Midwest

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ABSTRACT

On Friday, 10 December 1982, a modest, yet unforecasted snowfall occurred in a band extending from the Ohio Valley states eastward to western New York State. Aside from this case study representing a crucial forecasting problem, the scientific issues suggested by the results of this examination are especially intriguing. The precipitation was associated with neither a surface cyclone nor an obvious surface front. Although the precipitation began in the vicinity of quasi-geostrophic ascent, the details of the precipitation pattern are better explained by the atmosphere’s susceptibility to moist slantwise convection. Additionally, the ascent associated with this precipitation event during its later stages in Illinois was part of an elevated thermally direct frontal circulation. The relatively strong ascent on the warm side of this frontal circulation was likely assisted by the low moist symmetric instability in the same region.

The synoptic-scale flow pattern played a role in the evolution of this precipitation through quasi-geostrophic ascent, weakened environmental moist symmetric stability, and geostrophic frontogenetic flow. However, in the western part of the precipitation band, the moisture responsible for the precipitation onset is shown to have been transported from the Texas Gulf Coast into the Midwest by a low-level wind maximum. The depth of this moist layer ranged from 20 to less than 150 mb, and its horizontal extent was about 200 km—dimensions which are substantially smaller than synoptic scale. The limited depth of the moist layer may have contributed to this precipitation event being missed by the operational Limited-Area Fine-Mesh Model (LFM), which has only six tropospheric layers, averaging 150 mb in depth.

1. Introduction

During the late evening hours of 9 December 1982, unforecasted light snow, freezing rain, and sleet broke out over northern Illinois. By the following morning (1200 UTC 10 December 1982), a band of precipitation extended eastward from the Mississippi River valley into western New York State. Figure 1 shows this precipitation pattern, along with the apparent suddenness with which this band appeared.

Though the 12-h precipitation amounts associated with this band were modest (up to 10 mm), the 12–24 h operational Model Output Statistics (Klein and Lewis, 1970), forecasted measurable precipitation probabilities ranging from less than 5% in Ohio and western New York to 15% in western Illinois. These forecasts are guided by the operational Limited-Area Fine-Mesh Model (LFM). Although the LFM has more difficulty in forecasting precipitation amounts than in predicting measurable precipitation events (Bosart, 1980), we have chosen to study this unforecasted precipitation episode because of its interesting forecast and scientific implications.

The objective of this paper is to investigate the meteorological details of this unforecasted precipitation. We will show diagnostic calculations germane to this event, and we hope to stimulate further research into such a phenomenon. Section 2 contains a synoptic overview, and includes the surface, satellite, and radar observations of the precipitation. Sections 3 and 4 describe the vertical motions and frontogenetic forcing, respectively, during this event. The fifth section assesses the susceptibility of the atmosphere to moist slantwise convection. Section 6 examines the moisture source of this precipitation event. The concluding discussion is presented in section 7.

2. Synoptic overview

LFM analyses of sea level pressure and 1000–500 mb thickness for the 24-h period beginning 1200 UTC 9 December 1982 are shown in Fig. 2. At 1200 UTC on the 9th, the middle west region of the United States (U.S.) is dominated by a 1045 mb surface anticyclone centered in northeastern Iowa. The area of strong lower tropospheric warm advection northwestward from this high moves eastward into the northern Mississippi River valley by 0000 UTC 10 December, just a few hours prior to the commencement of precipitation. By 1200 UTC 10 December, the Great Lakes region is dominated by a surface trough south of the deepening 994 mb Hudson Bay surface cyclone. The 1034 mb anticyclone continues to decay and move eastward at this time.

Analyses at 300 and 850 mb are shown in Figs. 3, 4 and 5 for the 24-h period beginning at 1200 UTC 9 December. Figure 4a shows locations and three-letter

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identifiers of stations discussed in the text. These station names and WMO numbers are summarized in Table 1.

The 300-mb flow in the Mississippi Valley region is apparently unperturbed and mostly zonal at 0000 UTC 10 December—just prior to the onset of precipitation. This Midwest flow strengthens with time in response to the passage to the north of a southeastward-moving trough located in northern Idaho and western Montana at 1200 UTC 9 December, and which extends from northern Minnesota through the Dakotas at 1200 UTC 10 December. However, the 300-mb flow remains generally zonal through 1200 UTC 10 December. The 300 mb jet core extends from eastern South Dakota through northern Michigan, well to the north of the precipitation area. At 0850 mb, winds in the Midwest are associated with the eastward-moving and weakening anticyclone centered in Oklahoma at 1200 UTC 9 December. By 0000 UTC 10 December, as the trough associated with the Hudson Bay low approaches, warm advection exists from the Great Lakes region southwestward into eastern Nebraska and Kansas at 0000 UTC 10 December. The 850 mb warm advection pattern dominates the eastern Great Lakes region, including Michigan, Ohio, New York, and Pennsylvania, at 1200 UTC 10 December.

FIG. 1. Conventional radar summaries valid (a) 2335 UTC 9 December 1982, (b) 0535 UTC, and (c) 1135 UTC 10 December. Maximum echo top heights are indicated by underlined numbers (km). Movement of precipitation areas is indicated by wind pennants (one long barb = 5 m s⁻¹, one half barb = 2.5 m s⁻¹). Movement of cells is indicated by arrows with speed (m s⁻¹).
The 0000 UTC 10 December surface observations (Fig. 6) show the precipitation to extend from the South Texas Gulf Coast northward to southwest Oklahoma. An area of 5–10 m s\(^{-1}\) southerly winds and warm advection extends from western Oklahoma northeastward into Wisconsin. By 0600 UTC, the just-developed precipitation discussed in Fig. 1 and extending from Kansas into western Illinois, is associated with south-southwesterly flow and is occurring mainly as light rain and freezing rain. Little evidence exists to document a surface front in the vicinity of the precipitation band, although a much more prominent 850-mb temperature contrast does exist through Illinois (Fig. 4b). At 1200 UTC, precipitation extends eastward of the Fig. 6 domain into extreme western New York and northwestern Pennsylvania. Figure 7, showing the accumulated 6-hourly precipitation from 0600 UTC through 1200 UTC 10 December, illustrates clearly the narrow (≤200 km) band of precipitation extending from western Illinois into western New York. This figure is based upon the hourly precipitation data archived by the National Climatic Center. That the data points are about 30 km apart allows us to define subsynoptic scale precipitation patterns.

The Peoria surface report time section (Fig. 8) is typical of those stations in the path of this mesoscale precipitation band. Precipitation fell continuously for about 11 hours, with 7 mm of precipitation and a snow accumulation of 50 mm. Essentially steady pressure falls, and temperature and dewpoint rises exist prior to and during the early part of the precipitation event. Clouds observed were stratiform with a few observations of stratocumulus. Winds veer from southerly to southwesterly by the end of the period in response to the approach of the surface trough seen in Figs. 2 and 6. After this surface trough passage at 2100 UTC 10 December, winds continued to veer to northwesterly, pressures rise, and the air cools and dries in response to an approaching cold anticyclone seen in Alberta, Canada (Fig. 2c).

The satellite image (Fig. 9) at 2330 UTC 9 December shows a high cloud mass extending from Mexico northward into Colorado and western Kansas. The first medium-gray enhancement shows cloud top temperatures below −32°C (Corbell et al., 1976). Cloud bands emanate eastward from this main cloud shield, with the leading edge of the band extending into north central Illinois (shown by the arrow) responsible for the initial clouds seen at Peoria by 2100 UTC 9 December (Fig. 8). These cloud bands, which end in north central

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**Fig. 2.** LFM analyses of sea level pressure (solid lines) and 1000–500 mb thickness (dashed lines) for (a) 1200 UTC 9 December, (b) 0000 UTC, and (c) 1200 UTC 10 December 1982. Sea level pressures are indicated in mb (hundreds and thousands digits omitted) and thicknesses are shown in decameters. Heavy solid lines A, B and C are cross-sections used in symmetric stability calculations of Figs. 20 and 21.
Fig. 3. (a) 300 mb analysis for 1200 UTC 9 December. Heights (solid lines) are shown every 12 dam, temperatures (dashed) are shown each 4°C. Wind pennants follow the standard shown in Fig. 1. Solid circles show a temperature-dewpoint difference of ≤5°C. Temperatures (°C), dewpoint depressions (°C), and heights (dam) are plotted for each station. Dewpoint depressions are not reported for stations at which temperatures are less than -40°C. X's denote relative humidity observations less than 20%. M's denote missing observations. (b) As for (a), except for 850 mb, and the height contours are shown every 6 dam. The heights are plotted in m, with the thousands digit omitted.
Illinois at 2330 UTC 9 December (Fig. 9a), are approximately 50 km wide. Precipitation echoes were first observed by the Marseilles, IL radar (100 km northeast of Peoria) four hours later at 0330 UTC 10 December, in the vicinity of Peoria. By 0530 UTC, the cloud bands extend as far east as central Ohio. The lower, warmer cloud shield covers southern Iowa, northern Missouri, and west-central Illinois. This region corresponds closely to the eastern part of the precipitation shield seen in Fig. 1b. At 1130 UTC (Fig. 9c), colder and deeper clouds cover most of northern Illinois, and are associated with continuing precipitation there (Fig. 1c). High cloud bands extend into east-central Ohio and eastern New York.

An examination of the National Weather Service WSR-57 10 cm radar logs for sites affected by this event (Des Moines, IA; Monett, MO; Marseilles, IL; Kansas City, MO; St. Louis, MO; Evansville, IN; and Pittsburgh, PA) shows that all radar echoes were of VIP (Video Integrator and Processor) level 2 or less, with the majority of reports showing VIP level 1. Some of these VIP 2 echoes were likely associated with excessive backscatter due to melting, since numerous surface reports showed a mixture of frozen and melted precipitation in the vicinity of these echoes. However, some precipitation reports in northern Illinois, south-central Iowa, and northern Missouri showed hourly amounts ranging up to 2.5 mm, which is a moderate precipitation rate, corresponding to the VIP 2 intensity threshold. The echo tops at about 500 mb are consistent with the notion that this precipitation was associated with the low- and middle-level clouds previously de-
scribed for northern Illinois at 0530 UTC 10 December (Figs. 1b and 9b).

3. Vertical velocity diagnosis

So that we may understand the synoptic-scale dynamics associated with this event, we have computed quasi-geostrophic vertical velocities (ω) according to the expression:

$$\sigma \nabla^2 \omega + f^2 \frac{\partial^2 \omega}{\partial p^2} = -2 \nabla \cdot \mathbf{Q},$$

(1)

where

$$\sigma = -\frac{R}{P} \left( \frac{P}{P_0} \right) \frac{\partial \theta}{\partial p}$$

$$\mathbf{Q} = \left[ \frac{-R}{P} \left( \frac{\partial u_g}{\partial x} \frac{\partial T}{\partial x} + \frac{\partial u_g}{\partial y} \frac{\partial T}{\partial y} \right) \right] - \frac{R}{P} \left( \frac{\partial u_g}{\partial y} \frac{\partial T}{\partial x} \right) + \frac{R}{P} \left( \frac{\partial v_g}{\partial y} \frac{\partial T}{\partial y} \right) \left( \frac{R}{P} \frac{\partial \mathbf{V}_g}{\partial x} \cdot \nabla T_v - \frac{R}{P} \frac{\partial \mathbf{V}_g}{\partial y} \cdot \nabla T_v \right)$$

- $R$ gas constant for dry air
- $P$ pressure
- $\theta$ virtual potential temperature
- $P_0$ 1000 mb
- $C_p$ specific heat at constant pressure
- $\kappa$ ($=R/C_p$)
- $T_v$ virtual temperature
- $\mathbf{V}_g$ geostrophic wind vector
- $u_g$ geostrophic wind component in the east ($x$) direction
- $v_g$ geostrophic wind component in the north ($y$) direction
- $\nabla$ horizontal gradient operator

This form of the quasi-geostrophic $\omega$-equation, obtained from the work of Hoskins et al. (1978) and Hoskins and Pedder (1980), is convenient to use because areas of geostrophically-forced ascent and descent can be viewed on the basis of the respective convergence.
and divergence of $Q$-vectors computed on constant pressure surfaces.

The forcing in (1) is computed using the regional subjective analyses of temperature, dewpoint, and geopotential height of 850, 700, 500 and 300 mb. The computations are performed using centered-finite differencing on a 1° latitude-longitude grid. The subjectively analyzed data are manually interpolated to this grid. To insure that small-scale noise is eliminated, a five-point smoother (defined as one-eighth of the sum of four times the grid point value plus each of the adjacent grid point values) is applied to each grid point. The vertical velocity is constrained to vanish at 1000 and 100 mb.

Figure 10 shows the results of these quasi-geostrophic calculations at 700 and 500 mb. At 0000 UTC 10 December, a broad band of 700 mb ascent extends from eastern Nebraska into northeastern Wisconsin, while weak descent is maximized in southern Illinois. The 500 mb pattern has a slightly larger areal coverage of ascent maximized in northwestern Iowa. By 1200 UTC 10 December, the strongest ascent is found in western Iowa and in Ontario north of Lake Huron at both 700 and 500 mb. The area in Iowa is about 200 km north of the heaviest precipitation band accumulated by this time in northeastern Kansas (Fig. 7). Unlike the 500 mb pattern, both 700 and 850 mb (not shown) suggest ascent extending without interruption from western Iowa through northern Indiana and southern Michigan into western New York at 1200 UTC 10 December.

Kinematically-computed vertical motions at 700 mb for 0000 and 1200 UTC 10 December are shown in Fig. 11. These are derived from objective analyses of eastward and northward winds on a one degree lati-
TABLE 1. Station names, WMO numbers, and three-letter identifiers of locations discussed in the text.

<table>
<thead>
<tr>
<th>Station</th>
<th>WMO number</th>
<th>Three-letter identifier</th>
</tr>
</thead>
<tbody>
<tr>
<td>Amarillo, TX</td>
<td>72363</td>
<td>AMA</td>
</tr>
<tr>
<td>Buffalo, NY</td>
<td>72528</td>
<td>BUF</td>
</tr>
<tr>
<td>Dodge City, KS</td>
<td>72451</td>
<td>DDC</td>
</tr>
<tr>
<td>Green Bay, WI</td>
<td>72645</td>
<td>GRB</td>
</tr>
<tr>
<td>Huntington, WV</td>
<td>72425</td>
<td>HTS</td>
</tr>
<tr>
<td>Maniwaki, Quebec</td>
<td>71722</td>
<td>YMW</td>
</tr>
<tr>
<td>Monett, MO</td>
<td>72349</td>
<td>UMN</td>
</tr>
<tr>
<td>Nashville, TN</td>
<td>72327</td>
<td>BNA</td>
</tr>
<tr>
<td>North Platte, NE</td>
<td>72562</td>
<td>LBF</td>
</tr>
<tr>
<td>Peoria, IL</td>
<td>72532</td>
<td>PIA</td>
</tr>
<tr>
<td>Pittsburgh, PA</td>
<td>72520</td>
<td>PIT</td>
</tr>
<tr>
<td>Salem, IL</td>
<td>72433</td>
<td>SLO</td>
</tr>
<tr>
<td>Topeka, KS</td>
<td>72456</td>
<td>TOP</td>
</tr>
</tbody>
</table>

The analysis used is based upon a Barnes (1973) scheme in which weighting functions of each of the nearest ten rawinsonde winds at a given gridpoint are computed. The exponential weighting function decreases from one at zero distance to 0.1 at the average distance of the ten stations from the grid point. The horizontal divergence of these winds is integrated upward from the surface to 100 mb, where the corrected values of ω are constrained to be zero. A constant divergence correction (O'Brien, 1970) is made at each pressure level. The resulting kinematic vertical velocity calculations at 700 mb are shown in Fig. 11. At 0000 UTC, vigorous ascent extends from Lake Superior southwestward to northern Kansas. The southern edge of this ascent extending from northern Kansas to northwestern Illinois corre-
Fig. 6. Regional surface maps for (a) 0000 UTC, (b) 0600 UTC and (c) 1200 UTC 10 December. Wind convention is as described in Fig. 1. Isobars (mb) are based upon altimeter settings converted to mb. Temperatures (°C) are shown by dashed lines. Weather symbols are conventionally plotted. Fronts are indicated with the conventional symbolism.
Fig. 7. Precipitation amounts for the period 0600 UTC through 1200 UTC, 10 December 1982. First contour is 0.25 mm, representing the threshold of measurable precipitation. Contours of 5, 10 and 15 mm are also shown. The heavy solid boundary encloses the precipitation data domain.

responds to the cloud band activity previously described for Fig. 9a, and to the axis of maximum precipitation to begin 6 h later (Fig. 7). This ascent maximum shifts southward to southern Missouri and southern Illinois by 1200 UTC 10 December (Fig. 11b), so that all of

Fig. 8. Time section of hourly surface reports for Peoria, Illinois, plotted as a function of sea level pressure, beginning with the 2000 UTC 9 December observation. Visibilities (parentheses) are indicated in km. Wind directions are plotted at the end of each pennant (tens of degrees). Wind plotting convention is as shown in Fig. 1. Conventional weather symbols and comments are shown. Temperatures (solid line) and dewpoints (dashed line) are shown in °C.

Fig. 9. GOES MB-enhanced satellite images for (a) 2330 UTC 9 December, (b) 0530 UTC and (c) 1200 UTC 10 December 1982.
Illinois except for its extreme northern section is covered with ascent. This is consistent with the fact that measurable precipitation broke out later in the day in southern Illinois. However, weak descent is seen throughout western Pennsylvania and New York. In summary, these kinematic analyses are generally consistent with the quasi-geostrophic calculations (Fig. 10), except that the magnitudes are greater.
4. Frontogenetic forcing

Since the precipitation has a banded structure, we consider in this section the roles that frontogenetic forcing and the resulting thermally direct vertical circulations may have played in this case.

Surface observed and geostrophic wind frontogenesis values for 0000, 0600 and 1200 UTC 10 December are shown in Fig. 12. These geostrophic calculations are helpful, because we can examine the larger-scale features' effect upon this mesoscale precipitation event. The calculations are based upon the expression used by Petterssen (1956), among others:

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

where \( u \) and \( v \) are the wind components along the \( x \) and \( y \) axes, respectively, \( a \) is the angle between the dilatation and \( x \)-axis, \( b \) is the angle between the isentropic and the dilatation axis, \( \mathbf{V}_h \) is the horizontal wind vector, and the rest of the symbols are defined as in (1). This expression is equivalent to the horizontal confluence term in Miller's (1948) two-dimensional frontogenesis equation. Geostrophic frontogenesis is computed from (2) by assuming the winds to be geostrophic and using the resulting geostrophic dilatation axis. The calculations were carried out from subjective surface analyses of altimeter setting, wind speed, wind direction, and virtual temperature. The computations are performed on a 1° latitude–longitude grid, as in the previous section.

At 0000 UTC, the most pronounced observed surface frontogenesis (Fig. 12) is in eastern South Dakota where both a cold front and surface trough exist (Figs. 2b and 6a). This same area is mainly frontolitic geostrophically. At 0600 UTC, surface geostrophic frontogenesis now is most pronounced along an axis extending from Nebraska to eastern Wisconsin. Weaker observed frontogenesis lies along this same axis. The most intense frontogenesis seen in Nebraska is associated with the southeastward-moving cold front (Fig. 6b). This frontogenesis axis lies slightly to the cold side of the newly-formed precipitation band which extends into northern Illinois at this time (Figs. 1, 6b, and 7). By 1200 UTC 10 December, strong geostrophic frontogenesis closely follows the surface cold front seen in Fig. 6c. The weaker observed field at 1200 UTC follows this same course except the frontogenesis band is shifted southward from central Wisconsin into northern Illinois.

We consider frontogenetic forcing above the surface by noting that \( Q \), as defined in (1), is proportional to the rate of change of potential temperature gradient moving with the horizontal geostrophic velocity (Hoskins and Pedder, 1980):

\[ \frac{d}{dt} |\nabla \theta|^2 = \left( \frac{P_0}{P} \right) \left( \frac{P}{R} \right) Q, \quad (3) \]

where all symbols are defined as in (1).

Thus, the geostrophic frontogenesis function, \( G \), may be defined as

\[ G = \frac{d}{dt} |\nabla \theta|^2 = 2 \left( \frac{P_0}{P} \right) \left( \frac{P}{R} \right) \mathbf{Q} \cdot \nabla \theta. \quad (4) \]

This calculation is displayed for 700 and 500 mb at both 0000 and 1200 UTC 10 December in Fig. 13. At 0000 UTC, the most pronounced frontogenesis appears
in the northwest corner of the domain, and is associated with the surface cold front. However, frontogenesis does extend southeast of the cold front into northwest Iowa. This frontogenesis substantially diminishes at 500 mb, and the rest of the domain is mainly frontolytic. By 1200 UTC 10 December, the area of 700 mb frontogenesis continues to move southeastward into eastern Wisconsin and northwestern Illinois in association with
the surface cold front. At 500 mb, this frontogenesis is tilted to the northwest and is less intense than that found at 700 mb.

Since we would expect to see a thermally direct frontal circulation (Sawyer, 1956) develop in response to frontogenetic forcing, we display vertical cross sections of meridional and vertical winds to confirm this association. These wind components are taken from the objective horizontal wind analysis and resulting kinematically-computed vertical motions (Fig. 11) found in the previous section. The cross sections displayed in Figs. 14 through 16 are also mapped in Figs. 12 and 13 to facilitate comparison with the frontogenetic forcing. We find a thermally direct circulation extending from the westernmost longitude of Fig. 11 (102°W) eastward to 92°W, in the vicinity of southeastern Minnesota and northeastern Iowa. The cross section along 95°W (Fig. 14) shows the center of this frontal circulation to be between 44 and 45°N, and near 650 mb, with the most intense rising branch of this circulation located between 42 and 43°N. The southern (rising) portion of this frontal circulation corresponds closely to the east–west oriented cloud bands (Fig. 9a), which extend from north-central Nebraska through northern Iowa and into northwestern Illinois. Also, an axis of 700 mb geostrophic frontogenesis does extend along these cloud bands to about 94°W (Fig. 13), which is also located within the elevated thermally-direct circulations. The areal coverage of the thermally-direct circulation at 700 mb at 0000 UTC 10 December is mapped onto the domain of 700 mb geostrophic frontogenesis (Fig. 13a).

By 1200 UTC 10 December, the 90°W cross section (Fig. 15) shows a thermally-direct frontal circulation centered in Illinois between 41° and 42°N, and at 700 mb. The ascent on the warm side of this circulation appears associated with the precipitation in north–cen-
tral Illinois (Fig. 1c), and with the clouds seen in the northern part of the state (Fig. 9c). The most intense 700 mb ascent occurs on the warm side of the axis of maximum geostrophic frontogenesis at this level (Fig. 13) from 37° to 40°N. Other cross sectional displays reveal this thermally direct frontal circulation to extend as far east as 88°W, centered at 42°N. This location corresponds to the easternmost penetration of the 500 and 700 mb geostrophic frontogenesis at 1200 UTC 10 December (Fig. 13). These results suggest this thermally direct circulation is effected by large-scale frontogenetical forcing, as the areal coverage of this circulation (Fig. 13c) suggests.

The precipitation and cloud bands at 1200 UTC 10 December (Figs. 1c and 9c) do extend eastward of northern Illinois and the region of middle-tropospheric frontogenesis (Fig. 13) into extreme western New York. The role of slantwise convection in this precipitation will be discussed in the next section.

While thermally direct vertical circulations appeared to have played a role in the precipitation evolution in Illinois westward to Nebraska, the strong subsidence seen in eastern Indiana and western Ohio at 0000 UTC 10 December 1986 may have played an indirect role in the precipitation evolution east of Illinois. The cross sectional display of winds at 87°W (Fig. 16) shows a deep thermally-indirect vertical circulation centered between 42° and 43°N, at about 700 mb. This location above extreme southeastern Lake Michigan is located directly downstream of the 300 mb jet shown in Fig. 4 to be north of Peoria. At 800 mb, in the middle of this indirect circulation, the potential temperature increases by as much as 10°C in 50 mb. Thus, the horizontal potential temperature gradient will be enhanced by this thermally indirect circulation in the absence of significant frontogenetic forcing by horizontal wind deformation. The areal coverage of this thermally indirect circulation is also shown by Fig. 13a to be in the midst of weak geostrophic frontolitic forcing, as we should expect.

Indeed, a comparison of the 1000–700 mb thickness fields for 0000 and 1200 UTC 10 December (Fig. 17), shows a strengthening of the mean meridional horizontal temperature gradient in Illinois, Wisconsin, Michigan, and Indiana. Evidently, the frontogenetic effects of the thermally indirect circulation dominated the weaker geostrophic frontolysis. A consequence of this enhanced eastward thermal wind is that the 700 mb wind increased. This enhanced low level jet in turn steered the precipitation band into western New York state by 1200 UTC 10 December. This low-level jet, and its associated moisture, will be discussed later in this paper.

5. Thermodynamic structure and susceptibility to slantwise convection

The Peoria, Salem, and Topeka soundings for 0000 UTC 10 December are shown in Fig. 18. All stations show strong stability, and veering winds in the lower troposphere. Especially prominent is the relatively high moisture content between 680 and 810 mb at Peoria, four hours prior to the onset of precipitation. By 1200 UTC 10 December (Fig. 19), strong lower tropospheric static stability still exists at Peoria and Salem, while both soundings are now mainly saturated with lapse rates slightly less than moist adiabatic. Middle tropospheric cooling at these two stations since 0000 UTC is also evident, illustrating the effects of the pronounced ascent described earlier. The Buffalo sounding at the eastern edge of the precipitation shield shows nearly saturated and weakly stable conditions above 715 mb. The strong hydrostatic stability, dry air, and veering winds at Buffalo are similar to those found at 0000 UTC for Salem and Peoria.

Our observations of 50 km cloud bands from the satellite imagery (Fig. 9) and a relatively strong 1-h precipitation accumulation of 2.5 mm motivate us to consider whether conditional symmetric instability played a role in this precipitation event, since our arguments for frontogenetic forcing only apply to a part of this precipitation band. Bennetts and Hoskins (1979) have suggested this dynamical instability as an explanation for these frequently-observed precipitation bands. Emanuel (1983a) has shown that the concept of parcel instability can be applied to conditional symmetric instability (just as parcel instability is used to consider moist convection), and has called this "slantwise" moist convection. The fundamental parcel instability in this case results from a combination of un-
FIG. 18. Soundings of temperature, dewpoint and winds for 0000 UTC 10 December 1982. The stations are Peoria (72532), and Salem (72433), Illinois, and Topeka (72456), Kansas. Abscissa shows temperature (°C), and ordinate in $P^0.282$. The heavy solid line indicates the equivalent potential temperature line of 321 K.

FIG. 19. As in Fig. 18, except for 1200 UTC 10 December 1982 and for the stations of Peoria and Salem, Illinois, and Buffalo, New York (72528).
stable gravitational forces acting vertically and unstable centrifugal forces acting horizontally. Hence, the parcel accelerations are directed slantwise. Emanuel has further demonstrated that a local assessment of the atmosphere's susceptibility to moist slantwise convection can be made by considering cross sections of $M$ and $S$ in a meridional moist baroclinic flow associated with a geostrophically-balanced base-state environment. These are defined as

$$M = v + fx,$$

and $S$ as potential temperature in an unsaturated atmosphere or equivalent potential temperature in a saturated atmosphere. The variable $v$ is the northward geostrophic velocity, $x$ the eastward position, and $f$ the Coriolis parameter. It can be shown for two-dimensional, inviscid flow that when the $M$-surface slopes are shallower than those of $S$, parcel instability for moist slantwise convection exists (see Emanuel, 1983a, and Sanders and Bosart, 1985).

To assess this parcel instability, a cross section normal to the 1000–500 mb thermal wind (shown as A in Fig. 2b) is seen in Fig. 20. The orientation of this thermal wind closely approximates the 1000–300 mb thermal wind in all cross sections considered. The $M$-surfaces are constructed from the geostrophic winds found in the mandatory-level height analyses, and the equivalent potential temperatures were found from the sounding data. The equivalent potential temperature is used in this examination, because we are focusing upon areas in which the atmosphere is saturated. Strong static stability is indicated through much of the lower troposphere except in the lowest 1 km in the northern part. Noteworthy is the saturated and marginally unstable region between 600 and 700 mb in the vicinity of the cloud band, where the subsequent precipitation occurred. A second area of marginal instability is seen between 200 and 300 mb. Figure 21 illustrates two additional cross sections normal to the 1000–500 mb geostrophic shear vector (lines B and C in Fig. 2c) at 1200 UTC 10 December. Clearly, neutral or unstable conditions exist through a deep layer in an area nearly coincident with the cloud (Fig. 9c) and precipitation bands (Fig. 1c) in the western section of Green Bay through Nashville, while the unstable area over and to the north of Buffalo corresponds well to the easternmost section of the precipitation band. The results suggest a correspondence between an atmosphere neutral or unstable to slantwise convection and the precipitation band.

Our examination of the frontogenetical forcing in the previous sections has indicated that forced frontal circulations seen in Fig. 13 are located to the west of the conditional symmetrically unstable air found in cross sections A and C. We suggest that the cloud bands and precipitation occurring in regions along these cross sections are the result of moist slantwise convection. It appears that the leading edge of the precipitation bands east of the frontal circulations are associated exclusively with this instability. As the area of forced thermally-direct circulations moves eastward (Fig. 13), these circulations act synergistically with the slantwise convection.
instability to produce locally strong bands of precipitation. This interaction may have occurred in the precipitation area along cross section A by 0600 UTC 10 December, as an interpolation of movement of these thermally-direct circulations suggests they had moved into this area by 0600 UTC. This synergistic interaction is also applicable to cross section B at 1200 UTC 10 December, as will be discussed. The larger scales of precipitation seen west of the Mississippi River (approximately 90°W, see Figs. 1c and 9c) at this time, are likely due to the dominant quasi-geostrophic ascent and frontogenetic forcing mechanisms (Figs. 10; 13c, d).

As can be inferred from the precipitation analysis (Fig. 7) and by the satellite images (Fig. 9), the accumulated precipitation appeared in a band of less than 200 km width. The precipitation band widths seen by the Marseilles, IL radar during the onset of the precipitation at Peoria (Fig. 8) ranged from 50 to 100 km. A typical configuration of the precipitation band during this event, as seen by the Marseilles radar, is shown for 0638 UTC 10 December in Fig. 22. Peoria, in the midst of this 50–100 km band, was experiencing snow and ice pellets. The bands are closely aligned with the tropospheric thermal wind vector. Similar observations of more intense precipitation pulses have been found
by Sanders and Bosart (1985) in a symmetrically-unstable atmosphere for a much more synoptically-active case. A similar pattern of echoes is seen on the radar films from Marseilles through 1200 UTC 10 December. The Buffalo radar films also showed precipitation bands of similar or slightly smaller width by 1200 UTC 10 December.

The symmetrically-unstable air found in cross section B (Fig. 21a) above Peoria may have contributed to the enhanced ascent seen on the warm side of the apparent thermally-direct frontal circulation at 1200 UTC 10 December seen in Fig. 15. Such a synergistic relationship between frontogenesis and weak symmetric stability has been discussed observationally by Sanders and Bosart (1985) and theoretically by Emanuel (1985). This result is also consistent with that found by Hsie et al. (1984) in their moist frontogenesis numerical simulations.

6. An examination of the low-level moisture source

The nearly saturated, high moisture content air found over Peoria at 0000 UTC 10 December (Fig. 18) between 680 and 810 mb figures prominently in the precipitation outbreak. This air is located in an area of ascent (Fig. 11a), and is unstable with respect to moist slantwise convection (Fig. 20). To examine the origin of this moisture, the 297 K isentropic surface is studied. Winds, pressures, and Montgomery stream functions are computed on the 297 K isentropic surface for the times 0000 UTC 9 December through 1200 UTC 10 December. High mixing ratio air (greater than 5 g kg\(^{-1}\)) penetrates as far north as Dodge City, KS (72451) at 0000 UTC 9 December, and moist south-easterly flow is pronounced over west Texas at this time (not shown). Figure 23 shows that at 1200 UTC 9 December, mixing ratios in excess of 5 g kg\(^{-1}\) are observed from southeast of Amarillo, TX to as far north as North Platte, NE (72562). A rapid progression of the moist air eastward to Peoria, IL at 0000 UTC 10 December (where the eastern edge of 50 km wide cloud bands are observed; e.g., Fig. 9a) and to western New York and Pennsylvania by 1200 UTC 10 December is indicated. The trajectory shown in Fig. 22 was constructed using the energy budget method of Danielson (1961). Clearly, the moist air seen at Peoria at 0000 UTC 10 December can be tracked to the moist, south-easterly flow found in west Texas at 0000 UTC 9 December. The anticyclonic gyre of this trajectory is evident through 0000 UTC 10 December, and it closely follows the high mixing ratio area as it progresses rapidly eastward into western Pennsylvania at 1200 UTC 10 December. The moist tongue seen at 1200 UTC 10 December (Fig. 23c) is clearly associated with the precipitation band seen at this time (Fig. 1c).

While the movement of the moist tongue at the 297 K level provides one useful precursor to this precipitation event, its presence exists only in narrow layers. Figure 24 shows this relatively thin layer of high mixing ratio (greater than 5 g kg\(^{-1}\)) at both Dodge City (between 762 and 830 mb) and North Platte (just above 800 mb) at 1200 UTC 9 December; both soundings are representative of those found in the moist tongue seen in Fig. 23a. The winds advecting this moist tongue from 0000 UTC 9 December through 0000 UTC 10 December are all low-level maxima located near the top of a stable layer—the 20 m s\(^{-1}\) southwest wind at North Platte (1200 UTC 9 December, Figs. 23a and 24), the 25 m s\(^{-1}\) northwest wind at Peoria (0000 UTC 10 December, Figs. 18 and 23b), and the 18 m s\(^{-1}\) Pittsburgh wind (1200 UTC 10 December, Fig. 23c). This preferential advection of moisture, by the low-level wind maximum into an area of ascent which is susceptible to moist slantwise convection (northern Illinois, see Figs. 11a and 20) by 0000 UTC 10 December played a primary role in the initiation of this unforecasted precipitation.

7. Conclusions

The unforecasted snowfall of 10 December 1982, which accumulated in a narrow band from northwestern Illinois to extreme western New York (Fig. 7), was associated with neither an obvious surface front nor a cyclone. Melted precipitation amounts along this band were generally less than 10 mm. Although this precipitation event was clearly minor in terms of amount, the fact that it was unforecasted, and that it
Fig. 23. Plotted winds (convention as in Fig. 1), mixing ratios (dashed lines, g kg$^{-1}$), and pressures (solid lines, mb) on the 297 K isentropic surface for (a) 1200 UTC 9 December, (b) 0000 UTC and (c) 1200 UTC 10 December. Heavy line indicates the trajectory described in the text, with six-hourly positions (labeled as time/date) shown beginning with 0000 UTC 9 December through 1200 UTC 10 December.
occurred in bands ranging from 50 to 200 km wide, makes this case an important one to study.

Although the precipitation did occur in an area of quasi-geostrophic ascent (Fig. 10), and in lower tropospheric warm advection (Figs. 4, 5 and 6), the scale of this ascent is much too large to explain the location of this event. An examination of frontogenetic forcing as a mechanism to explain the precipitation showed that, at 0000 UTC 10 December, the most consequential frontogenesis extended to central Iowa at 700 mb (Fig. 13). The associated thermally direct frontal circulation (Sawyer, 1956), seen in Fig. 14, shows its ascent band in the region where cloud bands are occurring in northern Iowa (Fig. 9a). This elevated frontal circulation has its base at approximately 850 mb (Fig. 14). This frontal circulation, however, can only be detected as far east as 92°W at 0000 UTC 10 December, and does not explain the cloudiness extending into northern Illinois at this time (Fig. 9a). An examination of the atmosphere's susceptibility to moist slantwise convection at 0000 UTC 10 December (Emanuel, 1983a) reveals that the atmosphere in the vicinity of Peoria, IL at approximately 700 mb is neutral or unstable to this convection (Fig. 20). This symmetric instability may explain the rapid onset of precipitation in Illinois by 0330 UTC 10 December. The banded structure of the precipitation echo (and its width in the range of 50–100 km, Fig. 22) is similar or slightly smaller than the intense precipitation pulses observed by Sanders and Bosart (1985) in a case of heavy snowfall along the east coast of the U.S. By 1200 UTC 10 December, precipitation bands of similar scale to what had been observed at 0000 UTC, were observed as far east as western New York State. Moist symmetric instability is observed through this band in western New York State, and in Illinois (Fig. 21), and appears to extend eastward of the frontogenetical forcing (Figs. 12 and 13) and frontal circulations (Fig. 15). Our evidence does show that the ascent on the warm side of the thermally-direct frontal circulation may be aided by weak symmetric stability (Figs. 15 and 21a). Such behavior has been suggested observationally by Sanders and Bosart (1985), and theoretically by Emanuel (1985).

The moisture associated with the initial impulse of this precipitation has been found to originate in Texas and to follow a low-level jet into the vicinity of Peoria, IL by 0000 UTC 10 December (Fig. 23). The same location and elevation at which this moisture and saturated air exists is also symmetrically unstable (Fig. 20) to saturated parcel displacements. Since the overall synoptic environment is characterized by ascent at 0000 UTC 10 December (Fig. 10), it is probably no coincidence that precipitation broke out in this area by 0330 UTC 10 December. The strong thermally indirect vertical circulation seen in Indiana and Michigan.
at 0000 UTC 10 December (Fig. 16) downstream of the 300-mb jet (Fig. 4) likely enhanced the meridional lower tropospheric thickness gradient (Fig. 17) during the next 12 h, and enhanced the zonal 700 mb wind (Fig. 23). This, in turn helped to steer the moisture and precipitation band rapidly toward the east into western Pennsylvania and New York State by 1200 UTC 10 December.

Though frontogenesis is seen as far east as Illinois at 1200 UTC 10 December, a surface front is absent from where the precipitation band is observed. It is hypothesized that the strong low-level hydrostatic stability (Figs. 14 and 15) acts to inhibit any vertical circulations which develop as a consequence of frontogenesis. This may be why no surface front is seen with the precipitation band. The frontal circulations seen in our study appear elevated (Figs. 14 and 15). Since a front is normally associated with a trough, its vorticity-producing mechanism, in the form of surface convergence, is minimized as a result of the strong static stability inhibiting vertical motions.

The reason this case was poorly forecasted is likely to be at least two-fold. The LFM-forecasted vertical velocities (not shown) were substantially weaker than observed, though approximately in the correct location. Additionally, crucial moisture transport for this case was accomplished in a narrow layer substantially less than 50 mb deep (see North Platte at 1200 UTC 9 December, Figs. 23 and 24). The narrow vertical extent of the 0000 UTC 10 December Peoria moisture (Fig. 18) and symmetric instability (Fig. 20) are also evident. The LFM vertical resolution with six tropospheric layers is insufficient to capture these features. The new Regional Analysis and Forecast System at NMC which includes an improved vertical resolution model (Hoke et al., 1985) may improve forecasts of such precipitation as described in this paper.

Physical processes such as the frontogenetic forcing identified in this paper, and its relationship to Emanuel’s (1983b) moist symmetric instability deserve further investigation. Recently, Sanders and Bosart (1985) examined these processes in the context of a more synoptically-active case of heavy snowfall along the United States East Coast, and found extremely intense pulses of heavy-snowfall embedded in the broader frontal circulations. Our study has emphasized the importance of moist symmetric instability in a synoptically weak environment in the precipitation evolution. We have also observed elevated frontal circulations associated with some of this precipitation. More cases of a similar nature need to be examined to determine the general significance of moist slantwise convection and elevated frontal circulations in these frequently-observed precipitation bands.
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