Doppler Radar Study of the Trailing Anvil Region Associated with a Squall Line

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ABSTRACT

The kinematic structure and reflectivity distribution within a region of widespread precipitation associated with a summertime midlatitude (Illinois) squall line, as revealed by an analysis of Doppler radar data, are presented and discussed. The squall line moved in a southeasterly direction while active convection forming on its leading edge moved in a northeasterly direction. Decaying thunderstorms and their anvils merged to form the extensive region of stratiform precipitation which trailed the squall line. An extension of the VAD (Velocity Azimuth Display) method, or the EVAD (Extended VAD) method, has been developed for the analysis of single-Doppler radar data. In contrast to the VAD method, which requires knowledge of the particle fall velocities (or assumptions regarding it) to calculate the divergence of the horizontal wind, the EVAD method yields the vertical distributions of both the particle fall speed and the divergence. Also presented are results from a multiple-Doppler (MDOP) analysis of data from three radars. A function-fitting technique has been used for the MDOP analysis which filters out motions of high wavenumbers and yields the mesoscale motions. The MDOP method yields the horizontal distributions of the winds, their divergence and the vertical air velocity, while the EVAD method yields their horizontal averages.

The EVAD method gave weak (<10 cm s\(^{-1}\)) ascending air motion below about 1-km height, descent (peak magnitude about 25 cm s\(^{-1}\)) between approximately 1 and 5-km height and ascent (peak approximately 35 cm s\(^{-1}\)) above 5-km height. The vertical air velocity calculated by the MDOP method also shows ascent below about 1-km height. Above that height, the MDOP method gave descending motion in the eastern part of the trailing anvil (adjacent to and behind the squall line convection), coinciding with a region of low and dissipating reflectivities. Further to the rear, in the western part of the anvil, the vertical air motion calculated by the MDOP method is qualitatively similar to that found by the EVAD method. It is speculated that the descending motion in the eastern part of the trailing anvil may be part of the subsidence associated with a new line of convection which formed ahead of the old line as it dissipated.

The divergence field calculated by the MDOP method shows a banded structure especially pronounced at the middle and higher levels. The component of the horizontal wind relative to the squall line and transverse to it, calculated by the MDOP method, suggests the following: above approximately 3-km height, air entered the trailing anvil at its front and ascended towards its rear; at heights between approximately 3 and 6.5 km, environmental air entered the rear of the anvil and descended towards its front in a height interval from near the ground to 3 km; below the latter flow regime, the air moved from the front to the back of the anvil in a layer whose thickness increased towards the rear of the anvil.

1. Introduction

Long-lived squall lines are often accompanied by an extensive region of stratiform precipitation and cloudiness. The stratiform region frequently trails the squall line and apparently is a result of merger of the anvils from deep cumulonimbus clouds occurring along the squall line. As such, the stratiform region is often referred to as the trailing anvil region and the stratiform cloudiness and precipitation are referred to as the trailing anvil cloud and the trailing anvil precipitation respectively. For the sake of brevity, we shall often omit the adjective "trailing" in referring to these regions.

Tropical squall lines of this kind were studied by Zipser (1969, 1977), Houze (1977), and others. Their studies showed the base of the stratiform clouds to be near the melting level, in contrast to the lower bases of the convective clouds along the squall line. The stratiform cloud layer was observed to extend about 6 km above the melting level. Below the melting level,
the air was markedly subsaturated but contained precipitation falling from above. Zipser (1977) and Houze (1977) postulated that mesoscale ascent occurs in the deep cloud layer and mesoscale descent occurs below it in the subsaturated air. The existence of the mesoscale ascent was postulated to account for the observation that large quantities of condensed water are contained in and fall from the anvil cloud. Light widespread precipitation is indeed observed to occur from the anvil cloud for several hours after the cessation of the deep convection along the squall line. The existence of the descent below the cloud base was deduced from the divergent nature of the wind field in that region, and because the wet-bulb potential temperature in that region was found to be characteristic of the air in the midtroposphere outside the squall-line system.

Earlier, Newton and Newton (1959; see their Fig. 6), in their study of a midlatitude squall line, had conjectured that mesoscale descending air motions occurred throughout the depth of the region of widespread precipitation behind the squall line rather than in its lower layers alone.

Recently, Gamache and Houze (1982, 1983) have reported on studies of mesoscale air motion and water budget of a tropical squall line. Using a technique of analysis of rawinsonde data, similar to that used for midlatitude squall-line systems (to be discussed later), they confirmed the existence of mesoscale ascending and descending motions in the upper and lower layers of the anvil. They also deduced that the mesoscale updraft generated 25–40% of the condensate in the stratiform cloud, the remainder being supplied by horizontal transfer of condensate from the cumulonimbus towers of the convective regions.

Midlatitude squall systems having a structure similar to the tropical squall system described above, were studied by Sanders and Paine (1975), Ogura and Liou (1980), and others. These authors used serial rawinsonde data and compositing techniques to construct an instantaneous wind cross section, oriented perpendicular to the squall line, assuming the wind (and thermodynamic) structure to be steady in a frame of reference translating with the squall line. Ogura and Liou deduced mesoscale ascending motion of several tens of centimeters per second in the upper regions of the stratiform cloud and descending motions of similar magnitudes in the lower levels.

Here we report on observations of an extensive trailing anvil region associated with a squall line observed in Illinois during project NIMROD (Northern Illinois Meteorological Research on Downbursts; Fujita, 1979, 1981). This project utilized a triple-Doppler radar network and other observing instruments. Doppler radar gives practically instantaneous observations of radial wind components throughout the echo volume with unprecedented resolution in space. We have used these observations to deduce practically instantaneous fields of horizontal and vertical winds in the anvil region. It was not necessary to resort to compositing techniques or to assume stationarity of system-relative wind fields. The horizontal and vertical winds have been deduced both by single-Doppler and multiple-Doppler radar methods. We first present results derived from the single-Doppler radar measurements. We have extended the VAD (velocity-azimuth display) method to yield accurate divergence of the horizontal wind so that the vertical air velocity can be found accurate to perhaps 10 cm s\(^{-1}\). We then present the results of our multiple-Doppler analysis. In the past, multiple-Doppler radar methods have been applied mostly to convective clouds. As is well known, the vertical air velocity deduced in these cases generally has an uncertainty of a few to several meters per second. Here, we have applied an analysis method which we believe has yielded mesoscale vertical air motions of high accuracy comparable to that of our extended VAD method.

The single- and multiple-Doppler radar analyses have given vertical distributions of the horizontal divergence and vertical air velocity which exhibit many similarities to the corresponding profiles suggested by Zipser and Houze and calculated by Ogura and Liou. In addition, we find a systematic horizontal variation of the vertical air velocity and a banded structure in the field of the horizontal divergence viewed on constant height surfaces.

2. Data used in this study

As already mentioned, the data used in this study were collected during project NIMROD conducted in Illinois during May and June 1978. The principal observing network (Fig. 1) consisted of three Doppler radars: the CP3 and CP4 radars of the National Center for Atmospheric Research and the CHILL (University of Chicago-Illinois State Water Survey) radar. The CP3 and CP4 are C-band (approximate wavelengths 5 cm) radars, each with a beam width of approximately 1 degree and a variable range resolution, typically about 200 m. The CHILL radar is an S- and X-band dual wavelength (wavelengths approximately 10 and 3 cm) radar, but in this study only data from the S-band component of the radar were used. The beamwidth of this radar is also approximately 1 degree and its range resolution is 150 m. Each of the radars has on-line processing and recording of reflectivity and mean Doppler velocity data in each range bin. The radars were operated in a coordinated scan mode, with a volume scan time of 3 min or less when observing convective clouds. In the observation of the trailing anvil region, the observing strategy was altered to allow for an extended VAD analysis as described in section 4.

Supplementary data came from several sources. Rawinsondes were launched every 30 min from the CP3 radar site. One of these coincided in time with the detailed radar observation of the trailing anvil region. A network of 27 surface stations, consisting of NCAR's
Portable Automated Mesonet (PAM), was situated in a $70 \times 80$ km area covering and surrounding the triple-Doppler radar network. These stations provided measurements of precipitation, wind, and thermodynamic quantities near the surface at 1 min intervals. A network of raingages, operated by the Illinois State Water Survey, was distributed more extensively over northern Illinois and Indiana and recorded rainfall averages over 5 min intervals.

The extended VAD analysis has been performed on the CP3 radar data collected during a 9 min time period starting at about 2345 CDT.\(^1\) Multiple-Doppler radar analysis has been performed using data from all three radars, for the same 9 min time period. The time period of these analyses will be put in context of the squall system's life history in section 3.

3. Description of the squall line

Satellite imagery shows that the convection that was to constitute the squall line began to intensify rapidly at approximately 1630 on 17 June 1978. The squall line formed approximately 250 km northwest of the observational network along a slow-moving cold front that stretched from the Great Lakes to the Texas Panhandle and along which several weak pressure depressions moved northeastward. As the convection intensified, the line became more continuous and by 2000 upper level cloud shields had merged to form a cloud band over 2000 km in length extending from northeast to southwest and 70 to 200 km in width (Fig. 2a).

The squall line moved approximately southeastward through northern Illinois at a speed of about 15 m s\(^{-1}\). This was considerably faster than the motion of the cold front, with the result that the squall line and the region of widespread stratiform cloud and precipitation behind it passed through the observing network between 2100 on 17 June and 0300 on 18 June, well before the cold-frontal passage at 0700 on 18 June.

The large-scale meteorological situation in which the squall line formed is shown in Fig. 3. At the 850-mb level (Fig. 3a), a tongue of relatively warm air was being advected northeastward through Illinois and the lower midwest ahead of a well-defined frontal trough. At the 700-mb level (Fig. 3b), as well as the 500-mb level (not shown), the flow above the warm tongue was west southwest and cold air advection was occurring well ahead of the surface cold-frontal position. The squall line, therefore, formed in a zone where differential horizontal temperature advection was destabilizing the atmosphere, where ascent on the eastern side of a short-wave trough and frontal zone was probably occurring, and the vertical shear of the horizontal wind was southeastward, all being conditions favorable for the formation of the squall line.

A detailed study of a time sequence of radar reflectivity pictures was performed to study the motion of the squall line and its constituent reflectivity cores. Briefly, the results of this study are as follows: Individual cores typically had lifetimes of 30 to 40 min and

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\(^1\) All time designations are central daylight time.
FIG. 2. (a) Satellite infrared image showing the squall line at 2000 CDT fairly early in life. (b) As in (a) but at 2330 CDT showing increase in width of cloud shield.
proximately 75% of the forward motion of the squall line. The remaining 25% was due to a small southeastward component of translation of the individual thunderstorm cores themselves. The southeastward propagation of the line combined with the generally northeastward motion of the individual cores was apparently an important factor in the merger of decaying thunderstorms and thunderstorm anvils into the large mesoscale area of stratiform cloudiness and precipitation behind the line. The squall line entered the observational network at about 2200 on 17 June. Figure 4a is a PPI display (elevation angle 4.5 degrees) of equivalent radar reflectivity factor at 2215 that shows the squall line at a stage when the stratiform precipitation to the rear of the active convection is just starting to become extensive. By 2345, the line of convection had weakened considerably, and the area of stratiform precipitation had enlarged, extending behind the line for a distance of about 140 km (Fig. 4b). Satellite imagery at approximately this time (Fig. 2b) also shows an increase in the width of the associated cloud shield. Radar reflectivity data indicate that at about this time the line of active convection became detached from the region of stratiform precipitation at all heights leaving a gap of about 20 km. This was due to dissipation of the leading edge of the original line of convection and formation of a new line of convection ahead of it. The line of cellular convection shown in the southeastern part of Fig. 4b is the new line of convection. It should be emphasized that the extensive stratiform region had formed to the rear of the earlier line while it was still active, remained attached to it throughout its life, and survived its dissipation. Light precipitation from the stratiform cloud continued to fall for about two hours after the dissipation of the line, suggesting that mesoscale ascent was perhaps providing additional condensation.

The precipitation rate, wind speed, wind direction, and the wet-bulb potential temperature measured at two of the surface (PAM) stations are shown in Fig. 5. Figure 5a is for a station over which a reflectivity core associated with the convective line happened to pass. The burst of heavy rainfall occurred shortly after the characteristic wind shift and thermodynamic changes signaling the passage of the squall gust front. Because of their discrete nature, the reflectivity cores did not pass over many of the PAM sites. Figure 5b is for one such site. This station did not experience the burst of heavy precipitation but nonetheless received prolonged rainfall from the region of stratiform echo trailing the squall line.

As stated in section 2, the single- and multiple-Doppler analyses were done on data collected during a 9-min period starting at 2345, during which each of the radars performed a complete volume scan. Thus the analysis of the stratiform region has been done for a time period approximately one hour after the dissipation of the line of convection with which it was as-

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Fig. 3. Synoptic analysis for 1900 CDT 17 June 1978 for (a) 850 mb; (b) 700 mb. Latitude (N) and Longitude (W) are shown, also height contours (dam), solid lines, and isotherms (°C), broken lines.
the study of downbursts during project NIMROD, most radar scans were limited to low elevation angles when severe convection was occurring in or around the radar network. Therefore, the entire depth of the convective line was not observed at ranges close enough to provide adequate vertical resolution. During the 9-min time period of this study the radar scans were such that the stratiform region, as well as the new convective line, were observed throughout their entire depths. However, no attempt was made to perform multiple-Doppler radar analysis of the new convective line because of contamination of that line by ground targets in the Chicago downtown area and because the volume scan time of 9 min, although suitable for the study of mesoscale phenomena, is excessive for the Doppler analysis of convective storms.

**FIG. 4.** PPI display of CP3 radar reflectivity factors. Maximum range from CP3 radar (+) is 108 km. (a) For 2215 CDT, elevation angle 4.5 deg. (b) For 2346 CDT, elevation angle 3.5 deg. The maxima in reflectivity factor at about 50 km range to the north and southwest are associated with the melting band. The square is the 120 × 120 km area of multiple-Doppler radar analysis. (Note there is a scale change between panels a and b.)

Associated. A multiple-Doppler analysis of the convective line itself was not attempted because the data were not suitable for this purpose. Because of the emphasis on

**FIG. 5.** PAM traces for two sites showing rainfall rate, wind speed and direction, and wet-bulb potential temperature. Panel (a) is for a site over which the core of a thunderstorm passed; (b) is for a site over which a core did not pass and the rain was from the extensive region of stratiform precipitation.
4. Analysis of single-Doppler radar data

a. The VAD method

Lhermitte and Atlas (1961) showed that in a situation of widespread homogeneous precipitation the horizontal wind speed and direction can be determined from observation of the radial velocity \( V \) as a function of the azimuth at a constant elevation angle (the so-called VAD or Velocity Azimuth Display). Caton (1963) showed that the divergence of the horizontal wind may be obtained from VAD observations. Browning and Wexler (1968) showed that the deformation of the horizontal wind also may be obtained from VAD data, and presented a systematic derivation of the VAD analysis method. Their main results are summarized here.

The radial component of the particle velocity is given by

\[
V = (u \sin \beta + v \cos \beta) \cos \alpha + W \sin \alpha,
\]

where \( \beta \) is the azimuth angle measured clockwise from the north, \( \alpha \) the elevation angle, and \( u, v, W \), the components of the particle velocity along the \( x \)-eastward, \( y \)-northward, and \( z \)-upward axes respectively. If for a given range gate 1) \( u, v, W \) do not change over the observation period, 2) \( u, v, \) are linear functions of \( x \) and \( y \) and 3) \( W \) is independent of position over the VAD circle, then equation (1) may be rewritten as:

\[
V = a_0 + a_1 \sin \beta + b_1 \cos \beta + a_2 \sin 2\beta + b_2 \cos 2\beta,
\]

where

\[
a_0 = \text{DIV} \frac{r \cos \alpha}{2} + W \sin \alpha,
\]

\[
a_1 = u_0 \cos \alpha,
\]

\[
b_1 = v_0 \cos \alpha.
\]

Here DIV is the divergence of the horizontal wind, and \( u_0 \) and \( v_0 \) are the values of \( u \) and \( v \) at the center of the VAD circle, whose radius is denoted by \( r \). The coefficients \( a_2 \) and \( b_2 \) are related to the deformation of the horizontal wind and will not be needed here. In widespread precipitation, the vertical precipitation velocity is essentially due to the terminal fall speed \( V_T \), i.e., \( W = w - V_T \approx -V_T \); hence (3) may be rewritten as

\[
a_0 = \text{DIV} \frac{r \cos \alpha}{2} - V_T \sin \alpha.
\]

In order to determine the parameters of the wind field, we first take \( V \), for a given range, over the VAD circle, and determine the coefficients \( a_0, a_1, a_2, b_1, \) and \( b_2 \) by Fourier analysis or least-squares fitting procedures. From the coefficients, \( a_1 \) and \( b_1 \), the horizontal wind speed and direction can be directly computed. However, the divergence of the horizontal wind can be determined from \( a_0 \) only if \( V_T \) is known or the elevation angle is so small that the term \( V_T \sin \alpha \) can be neglected in Eq. (6). According to Browning and Wexler (1968), inhomogeneities in the particle fall speed can lead to errors in the determination of DIV. For these reasons, Browning and Wexler recommended that, for the determination of DIV, \( \alpha \) should be kept small—specifically \( <27^\circ \) (9\(^\circ\)) in snow (rain). Obviously, the upper limit for \( \alpha \) should also depend upon the accuracy desired in the determination of DIV.

b. The extended VAD (EVAD) method

Our extension of the VAD method pivots on a transformation of Eq. (6) to read

\[
\frac{2a_0}{r \cos \alpha} = \text{DIV} - 2V_T \frac{h}{r^2}
\]

where \( h \) is the height corresponding to the horizontal range \( r \) and elevation angle \( \alpha \). The effects of earth curvature and the curvature of the ray path have been ignored in writing (7), although the equation can be easily generalized to account for these. We assume now that the DIV and \( V_T \) are functions of the height only, i.e., they are horizontally uniform. Consider the VAD scans for many paired values of range and elevation angle such that the resulting VAD circles lie in a narrow interval of height over which the DIV and \( V_T \) may be regarded constant in height as well. If these conditions are satisfied, then a plot of \( 2a_0/(r \cos \alpha) \) vs \( h/r^2 \) for these VAD scans, should be a straight line according to Eq. (7). Conversely, if the plot is a straight line, it is likely that the conditions mentioned above are satisfied. Both the DIV and \( V_T \) may be determined, from the intercept and the slope of the straight line obtained by least-squares fit to the data. Our extension of the VAD method does not require knowledge of the particle fall speed or restriction of the radar scans to low elevation angles. On the contrary, it is important for the success of the method that \( h/r^2 \), or equivalently the elevation angle, should have a sufficient range of variation to enable a proper fit of the data to Eq. (7). This is also physically clear from the fact that, for a given height interval, observations at low elevation angles give \( a_0 \)'s largely influenced by the DIV, observations at high elevation angles give \( a_0 \)'s largely determined by particle vertical motion, and \( a_0 \)'s from observations at intermediate elevation angles are influenced by both DIV and \( V_T \). Hence observations over a sufficient range of variation of elevation angles will be needed to determine both DIV and \( V_T \).

c. Comparison with other extensions of the VAD method

The extended VAD method presented here may be compared with extensions of the VAD method pro-
posed by others. We shall consider here the work of Easterbrook (1975), Waldteufel and Corbin (1979), and Koscielny et al. (1982).

Instead of considering the radial velocity data for one range gate along a complete VAD circle as proposed by Browning and Wexler (1968), Easterbrook used all the radial velocity data in a sector on a conical surface bounded by two azimuths and two range circles. Under the assumption of a linear wind field, he showed that the data could be used to determine the divergence of the horizontal wind provided the elevation angle was low enough to neglect the contribution of the precipitation vertical motion to the Doppler velocity. However, the horizontal wind could not be determined by this method without independent knowledge of the vertical vorticity.

Waldteufel and Corbin (1979) suggested processing data within a volume (VVP—Volume Velocity Processing) rather than within a sector on a conical surface. All the radial velocities in a certain volume are considered. The volume need not be centered over the radar. Waldteufel and Corbin made the assumption of linear variations of the particle velocity components in the volume. However, they extended the set of parameters to be determined to include the vertical derivatives of $u$, $v$, $W$, and the horizontal derivatives of $W$, in addition to $u$, $v$, $W$, and the horizontal derivatives of $u$, $v$ considered in the VAD analysis by Browning and Wexler. The method is, in principle, capable of determining the horizontal divergence without the need to confine the observations to low elevation angles or independent knowledge of $V$. Waldteufel and Corbin applied the method to actual observations and showed that $u$, $v$ and their vertical derivatives determined by the method are consistent with each other. However, the accuracy of determination of $W$ and, in particular, its vertical derivative was not good. The authors concluded that the determination of $W$ and its height derivative are so sensitive to departures from the assumption of linear variations and to small-scale irregularities in $u$, $v$, and $W$, that the estimates of those parameters cannot be relied upon.

Koscielny et al. (1982) considered a method similar to that of Waldteufel and Corbin. They performed an error analysis for the estimates of the parameters of the wind field. In an application of the method to actual data, Koscielny et al. used data at two elevation angles (0.4 and 0.8 degrees) to calculate the horizontal divergence in the clear air boundary layer over sectors 30 km in radial extent and 30 degrees in azimuthal extent. Because of the low elevation angles and the nature of the meteorological situation, the contribution of the vertical velocity of scatterers to the Doppler velocity could be safely neglected. Therefore, this work does not constitute a test of the success of the VVP method in separating the divergence and the vertical velocity of the scatterer in (3).

The EVAD method described here is thought to have certain advantages over the VVP method of Waldteufel and Corbin (1979). First, in the EVAD method, $u$ and $v$ are calculated for each VAD circle, and several VAD circles, with different elevation angles, usually exist for any narrow height interval. Therefore, the homogeneity and steadiness of the wind field can be checked. In this step the functions on which the radial velocities are fitted form an orthogonal set. By contrast in the VVP method, the functions on which the radial velocities are fitted may be dependent on each other. Secondly, the error of fit of $2a/(r \cos \alpha)$ against $h/r$, by linear regression, provides a built-in test of the validity of some of the assumptions of the EVAD method, namely, the uniformity of $W$ and $\nabla W$ in the horizontal. Third, and most importantly, the EVAD method is robust compared to the VVP method insofar as the determination of the $\nabla W$ and $V$ are concerned. This robustness arises because of the use of radial velocity data over complete circles in the EVAD method. In the VVP method, however, data over sectors may be used. When data over a sector are used, the determination of $\nabla W$ is critically dependent upon the validity of the assumption of linear variations of velocity fields.

On the other hand, the $a_0$ determined by a full circle of radial velocity data is related to the average $\nabla W$ over the circle by Eq. (6), even if the velocity field is not linear, provided the $V$ is taken to be the average fall velocity over the circumference of the VAD circle. In other words, Eq. (6) is not adversely affected by irregular variations in $\nabla W$ and $V$ when a full circle of data are used.

5. Results of the extended VAD (EVAD) analysis

a. Details of the analysis

The data from the CP3 radar were used for the EVAD analysis. This radar was completely embedded in the region of stratiform precipitation, which appeared fairly homogeneous on the PPI reflectivity pictures. The CP4 radar had somewhat broken VAD coverage, especially to its north and east. Similarly, the CHILL radar also had broken VAD coverage. As already mentioned, the squall line was separated from the region of the stratiform precipitation by a gap of about 20-km width. The CHILL radar was situated in the vicinity of this gap.

The VAD data of the CP3 radar consisted of 21 complete azimuthal scans at elevation angles ranging from 0.5 to 84.5 degrees. The range gate spacing was 210 m and the average spacing between successive azimuths of data was 0.7 degrees. The latter implies overlapping coverage in the azimuthal direction, since the half-power beam width is approximately 1 degree. The 21 azimuthal scans were performed in 9 min starting
at 2345. Immediately following the VAD scans, 5 min of data at vertical incidence were collected.

As a first step in the analysis, the radar data (both velocity and reflectivity) were reformatted. The reformatting was done by elevation angle and range gate number. After the reformatting, the velocity data for a complete VAD circle could be conveniently accessed by range gate and elevation angle. The Doppler velocities were then unfolded by comparing them with the radial component expected from the winds measured by the rawinsonde. This resulted in correct unfolding of about 90% of the folded radial velocities. Incorrectly unfolded velocities were detected and then correctly unfolded or rejected by the subsequent analysis.

Next, the Doppler velocities for a given VAD circle were fitted according to (2) by a least-squares method and the coefficients $a_0$, $a_1$, $b_1$, $a_2$, and $b_2$ determined. The standard error of the fit was calculated. Each of the Doppler velocities was then reexamined and those that differed from the fit by more than twice the standard error were flagged. A second fit was then performed excluding the flagged velocities. The differences between the Doppler velocities and the second fit were then examined to see if any of the flagged Doppler velocities could be made to lie within a predetermined range of the fit (2) by unfolding. Such Doppler velocities were then ("correctly" unfolded and) unflagged and a third fit performed to determine the coefficients in (2). This was taken as the final fit to the Doppler velocities. If no Doppler velocity was unflagged, then the second fit was the final fit. This procedure helped to identify and remove spurious Doppler velocities in addition to helping correct errors in the initial unfolding of the Doppler velocities. For example, a small sector in which the Doppler velocities were consistently flagged at low elevation angles was found to have blocking of the radar beam due to obstructions near the radar.

The VAD analysis was limited to range gates whose horizontal distance from the radar was less than 40 km. Upwards of 2500 individual VAD circles were analyzed, each of which resulted in a set of values consisting of the coefficients $a_0$, $a_1$, $b_1$, $a_2$ and $b_2$, the error of fit, and other parameters, arranged by elevation angle and horizontal range. These data were then classified into narrow height intervals. The height intervals were variable, averaging 500 m in depth, but deeper near the cloud top where data density was low, and shallower in and around the melting band and near the surface to account for the sharper variations in those regions. In calculating the heights, a four-thirds earth curvature correction was applied. The coefficients $a_1$ and $b_1$ were used to determine the horizontal winds. Winds in the same height interval, but for different VAD circles, were compared for consistency, and good agreement was found between the winds for the different circles. A linear least-squares fit was then performed, as suggested by (7). The standard error of the fit was determined to locate and reject spurious values of $a_0$. The final fits yielded values of the DIV and $V_T$ as a function of the height.

b. Winds and thermodynamic structure

The vertical profiles of wind speed and direction obtained from the VAD analysis and from a rawinsonde launched from the CP3 radar site at 2356 are shown in Fig. 6b. Very good agreement is seen between the two sets of winds, which gives us confidence in both techniques of wind measurement.

The vertical profiles of the temperature and dew point determined by the rawinsonde are shown in Fig. 6a. The thermal structure resembles that found behind tropical squall lines described by Zipser (1977) and behind an Oklahoma squall line described by Ogura and Liou (1980). Below about 4.5 km, the sounding is consistent with the occurrence of evaporation of precipitation and unsaturated descent. (See also section 5e.) Above 4.5 km, the atmosphere is saturated with respect to ice. Between 4.0 and 8.7 km and above 11.0 km, the atmosphere is stable. Between 8.7 and 11.0 km, it exhibits slight conditional instability. The layer from the surface to about 1 km is very stable. The wet-bulb potential temperature in this layer is 292 K. In the layer from 1 to 4 km, it is generally between 290.5 and 291.5 K. This contrast supports Zipser's (1977) conclusion that the air in these two layers has different origins.

c. Hydrometeor fall speed and reflectivity factor

The vertical profile of hydrometeor fall speed obtained from the EVAD method is shown in Fig. 7. The profiles of average Doppler velocity and reflectivity factor, measured while the radar was in the vertically pointing mode for 5 min immediately following the VAD scans, are also shown. The two sets of fall speeds are in surprisingly good agreement; especially noteworthy is the agreement in the melting region between the altitudes of about 3 and 4 km. The agreement supports our method of determining the DIV and $V_T$ from the composite parameter $a_0$ and implies that the assumptions implicit in the EVAD method were satisfied in this case. In the following we discuss some of the physical processes that were probably responsible for producing the observed vertical profiles of reflectivity and particle fall speed. Strictly speaking, the validity of this discussion requires the assumptions of homogeneity and steadiness; these should hold, at least approximately, in our case considering the duration and stratiform nature of the precipitation.

From 11 km down to 6.5 km height, the fall speeds generally increased from 1 m s$^{-1}$ to 2 m s$^{-1}$. These fall speeds are typical of snow. The increase of fall speed
and the concomitant increase in the radar reflectivity factor imply the occurrence of particle growth by some or all of the following processes: deposition, riming, and aggregation. Between 6.5 and 4.2 km height, the increase of fall speed was arrested and the rate of increase of the reflectivity factor also diminished. The former may be due to the fact that large aggregates had formed by the 6.5 km height, and the fall speed of large aggregates is known to be a weak function of their mass. Another possible reason for the diminution in the rate of increase of radar reflectivity in this height interval will be discussed in section 5e.

The fall speeds increased rapidly from about 2 to 9 m s\(^{-1}\) in the melting region and then decreased downwards to approximately 7.5–8 m s\(^{-1}\). The reflectivity factor increased sharply from about 30 to 40 dBZ and then fell again to 30 dBZ. This implies considerable particle aggregation above, and breakup below the peak of the reflectivity curve (Lhermitte and Atlas, 1963). The peak fall speed of about 9 m s\(^{-1}\) is unusually high for a melting band situation. It implies the existence of rather massive particles in the form of aggregates or rimed particles; these particles could have been derived from the debris of the intense convection. The fall speed of 7.5–8 m s\(^{-1}\) in the lowest levels is consistent with increasing air density and continued particle breakup. However, the magnitude of the fall speed is rather high for the reflectivity factor observed (30 dBZ) in these levels. Empirically, a fall speed of 6.0–6.5 m s\(^{-1}\) is expected for raindrops for a reflectivity factor of this magnitude. The deviation could be due to a nontypical particle size distribution over the radar and within 40 km of the radar since the fall speeds from the vertical incidence measurements agree with those deduced from the EVAD analysis. Another possible reason for the deviation could be the presence of partially melted ice particles. It is also possible that the calibration of the radar is in error by several decibels. Differences in reflectivity factor values measured by the three radar were noted; however, no attempt was made to obtain an absolute calibration because little use is made of the absolute reflectivity factors in this study.
data suggest a cloud top height of 12.7 km. We have extrapolated the EVAD-determined divergence profile to this height as shown by the dashed curve.

Below 3.5 km height, the wind changed rapidly to a strongly divergent field having a maximum divergence of about $4 \times 10^{-4}$ s$^{-1}$ at 1.0 km height. In the 270 m nearest to the ground, the wind field again became convergent. This result, which was not observed in previous studies, is supported by the multiple Doppler analysis (sections 7a and 7b) and by an objective analysis of the wind data from the PAM network, which also yielded a convergent wind field, although of much smaller magnitude, viz., $3 \times 10^{-5}$ s$^{-1}$. This is consistent with the results of a few individual VAD analyses at shallow elevation angles and close ranges that suggests, in fact, that the convergence reached a maximum between 50 and 100 m above ground and decreased rapidly below.

It is perhaps appropriate at this point to comment on an aspect of the EVAD method of divergence calculation. In view of the “excellent” agreement between the $V_r$-derived by the EVAD method and that observed at vertical incidence, one may question the need for the EVAD method. One could simply take the $q_0$ determined by the VAD method, and the fall speed from the vertical incidence measurement and determine DIV

d. Divergence of the horizontal wind

The vertical profile of the horizontal divergence obtained from the EVAD analysis is shown in Fig. 8. The divergence values determined for the various layers have been joined by straight-line segments. It is seen that the divergence values which are derived from independent measurements, join more or less smoothly, lending further credence to the EVAD analysis method. Disregarding the large convergence in the very lowest layers near the surface for the moment, the maximum convergence ($1.5 \times 10^{-4}$ s$^{-1}$) occurred at 3.5 km height, i.e., in the melting region. Above 3.5 km height the wind field was generally convergent up to about 9.9 km, above which it became divergent. The EVAD method yielded the divergence profile up to a height of 11.1 km. Above this, the radar data were spotty and the EVAD method could not be applied reliably. However, occasional VADs covering the whole VAD circle but with a rather low density of data points suggested a divergent horizontal wind when a VAD analysis was performed and the divergence determined by applying a correction for particle fall speed. Infrared satellite

![Fig. 7. Vertical profiles of hydrometeor fall speed from the EVAD analysis (solid line) and from measurements in the vertical pointing mode, VPM (+). The profile of radar reflectivity factor from the VPM measurements is shown by the dashed curve. The EVAD analysis is based on data collected during 2345–2354 CDT and the vertical pointing data were collected during 2354–2359 CDT.](image)

![Fig. 8. Vertical profile of divergence of the horizontal wind from the EVAD analysis. The dashed line is an extrapolation of the profile to the satellite estimated cloud top. The data for the EVAD analysis were acquired during 2345–2354 CDT.](image)
from Eq. (6). Two points must be made in this connection. First, the fall speed determined by the EVAD method is representative of an average value over the VAD circles used, and it is this average value that needs to be inserted for $V_T$ in equation (6). The vertical incidence measurements give the $V_T$ in a small volume directly over the radar and may not be representative of the particle fall speed over the VAD circle or circles. Secondly, small deviations do exist between the EVAD-derived $V_T$ and the vertical-incidence derived $V_T$. These differences are generally less than 1/2 m s$^{-1}$, except in the melting region. Although such difference may be small for most considerations, they should not be neglected when it is desired to obtain DIV with high accuracy. In short, the virtue of the EVAD analysis is that it yields the $V_T$ most appropriate for the calculation of the DIV.

e. Vertical air velocity

The vertical air velocity was determined by numerical integration of the anelastic continuity equation:

$$\text{DIV} + \partial (\rho w)/\partial z = 0. \quad (8)$$

In this equation $\rho$ is the air density which was determined from the rawinsonde ascent. The three curves shown in Fig. 9 correspond to three different boundary conditions. Curve A uses $w = 0$ at the ground and the measured divergence profile up to the radar-cloud top (solid curve, Fig. 8). Curve B uses the boundary condition $w = 0$ at the radar-cloud top and the same divergence profile as curve A. Curve C uses the boundary condition $w = 0$ at the satellite observed cloud top height and the measured divergence profile extrapolated to the satellite cloud top (dashed portion of curve, Fig. 8). It is seen that the integrations starting from the radar and the satellite cloud top give a small residual $w$ of $\pm 5$ cm s$^{-1}$ near the ground. There are two main uncertainties in the determination of the vertical air velocity profile, viz., 1) uncertainty in the cloud top and the magnitude of the vertical air velocity there, and 2) uncertainty in the divergence profile very close to the ground. Although, the low elevation angle VAD scans provided a measure of velocity at low heights near the radar, these data may have been contaminated by ground clutter. However, it is difficult to see how such contamination could result in the large convergence deduced by the EVAD method, except through a systematically asymmetrical distribution of the contamination as a function of azimuth angle.

Considering the three vertical air velocity curves, curve B can perhaps be rejected since it does not take account of the presence of cloud known to exist above the radar echo top (i.e., above the level at which the integration was started in this case). Curves A and C are consistent with each other. If curve A were extended in accordance with the dashed portion of the divergence curve in Fig. 8, the vertical air velocity would become zero at a height somewhat below the satellite-estimated cloud top. The maximum difference between curves A and C is about 15 cm s$^{-1}$ and occurs at the higher levels. If the vertical air velocities were normalized to the ground level in a momentum sense, i.e., if $w$ were multiplied by $p/p_0$ where $p$ and $p_0$ are the air densities at the height of $w$ and at the ground, respectively, then the difference between the two curves would be less than about 5 cm s$^{-1}$.

Curve C has two points in its favor. First, the vertical air velocity become nearly zero at the stable 0°C isotherm level expected in association with the melting level. Secondly, the top of the layer of pronounced downdraft near 4.5 km marks the division between the subsaturated and the saturated air on the sounding in Fig. 6a. It is known that the evaporation of precipitation in a downdraft generally leads to subsaturation (Kamburova and Ludlam, 1966; Leary, 1980); saturation can be maintained only if cloud drops are present in adequate concentrations. The saturated region on the sounding is consistent with the upward air motion in curve C. (Note that curve B has downward air motion well into the region of saturation; this is another reason for rejecting it.) It may also be noted that the region
of near-zero vertical air velocity on curve C correlates well with the height interval over which the rate of increase of radar reflectivity with decreasing height diminished in Fig. 7. Thus, it is possible that this feature of the reflectivity profile is related to a decrease in the rate of growth of particles by the condensation (or deposition) of water vapor.

The profiles of divergence and vertical air velocity are remarkably similar to those postulated or deduced from indirect evidence by Zipser (1969, 1977) and Houze (1977) for the region of stratiform precipitation behind tropical squall lines. They are also similar to the divergence and vertical air velocity profiles behind an Oklahoma squall line deduced by Ogura and Liou (1980) by composing a time sequence of radiosonde observations. In all cases, the vertical air motion is upward in the higher levels, and downward in the lower levels. However, our calculations do not depend upon assumptions of system-relative steadiness, and the profiles presented are practically instantaneous.

A significant difference between our results and those of others is the weak ascending motion, in the lowest layer in the present case. This weak ascent is due to the intense convergence near the surface. This convergence is also supported by the results of the multiple-Doppler analysis (section 7). This corroborative evidence, coupled with the constraints imposed by the rawinsonde observations and the observed cloud top, make the conclusion that ascending motions exist in the lowest layers seemingly unavoidable. Also noteworthy in this connection is the thermodynamic evidence (section 5b) suggesting that the air in the lowest 1 km may have had an origin different from the air immediately above it.

6. Analysis of multiple-Doppler (MDOP) radar data

a. Introductory

Using data from a single Doppler radar, the VAD and EVAD analysis methods yield the vertical distributions of the horizontal averages of the wind field and its kinematic properties over the analysis domain. In this situation, we also have triple-Doppler radar data which can be used to obtain information about the horizontal structure of the wind field. Because the single- and triple-Doppler data relate to the same time period and cover common volumes of space, triple-Doppler analysis can also be used to check further the validity of the EVAD analysis method.

MDOP analysis is commonly applied to convective storms. The determination of the vertical air velocity in these cases typically has an uncertainty of a few to several meters per second. In the present situation our goal is to compute vertical air velocity accurate to 10 cm s\(^{-1}\) or better. Because of this requirement and the spatial distribution of the radar observations, it was necessary to perform averaging over relatively large areas. The averaging is justified by the mesoscale nature of the phenomenon of interest.

The method of MDOP analysis applied in this study is essentially based on the methods of MDOP analysis of convective storms developed at the Laboratory for Atmospheric Probing, University of Chicago, but with certain modifications appropriate to the mesoscale nature of the phenomenon of interest. The major steps in the analysis consist of 1) data storage, 2) editing and unfolding of Doppler velocity data, 3) reflectivity calibration, and 4) determination of wind fields. The following very brief discussion relates to one volume scan of radar data.

b. Data storage method

An analysis volume is first selected. In this case, a volume 120 km \(\times\) 120 km \(\times\) cloud depth was selected. The 120 \(\times\) 120 km area is shown by the square in Fig. 4b. The radar data lying within the volume are sorted into narrow height intervals and transferred to different "height" files of a random access medium. The area is then divided into a 120 \(\times\) 120 grid with a spacing of 1 km in each direction. Each grid point is sequentially numbered and each data point is tagged with an identification number equal to the number of the nearest grid point. The data in each height file are then rearranged in order of ascending identification number. A directory for the location of data, by identification number, is then set up. After this step, the data lying in the neighborhood of any point can be conveniently accessed. Some of the subsequent operations performed on the data would either be very cumbersome or all but impossible without this initial reorganization of the data. It may be pointed out that all the data are preserved up to this point. No editing or smoothing of data fields has been performed and the "height" intervals used in sorting the data do not in any way predetermine the height resolution of the analyzed fields.

c. Doppler velocity editing and unfolding

The next step in the analysis is the editing and unfolding of the Doppler velocity data. As mentioned in subsection 6b, the Doppler velocity data for the EVAD analysis were successfully unfolded by comparison of the measured Doppler velocities with the radial velocity expected from the rawinsonde measured winds. This method did not succeed as well for the bigger volume used in the multiple-Doppler analysis. This is due to a larger variation of the wind across the MDOP analysis domain compared to that across the EVAD analysis volume. We could perhaps still use the method by varying the "rawinsonde" wind across the analysis volume. Instead, we have used a more powerful method designed for application to the case of convective storms with smaller-scale variability.
The unfolding methods described in the literature usually unfold Doppler velocities for each radar individually by using automatic or interactive methods (Carbone et al., 1980). Here, we consider all radars simultaneously in an automated objective unfolding method. The process is started by estimating a plausible range of values for the wind vector, and selecting several "assumed" wind vectors in that range. For each assumed wind vector, the expected Doppler velocity is calculated, and the observed Doppler velocity is "unfolded" to lie as close as possible to the expected Doppler (see Fig. 10). The deviation between the expected Doppler and the unfolded Doppler is determined. This is repeated for all the observed Doppler velocities in a small volume over which the actual wind vector may be anticipated to be fairly uniform. The mean square deviation of the unfolded radial velocities about the expected Doppler velocity is computed for each radar and for the radars taken in various combinations. The assumed wind vector which minimizes the mean square deviation should be close to the actual wind vector and is used to unfold the Doppler velocity in the small volume under consideration; this wind vector can, of course, vary from one small volume to the next. Large values of the minimum mean square deviation indicate regions of large shear, or spurious or noisy data. The last condition results in a mean square deviation approaching that of a uniform distribution of Doppler velocities over the unambiguous velocity interval of the radar. Such data are examined and rejected or unfolded using an interactive graphics procedure. In this case, the objective method worked successfully on almost 100% of the data. A very useful output of this method is the field of wind vectors that minimizes the mean square deviation. It is a first approximation to the actual wind field. It can be used to locate regions of sharp shear in the horizontal and the vertical. In an extension of the method outlined here, it is proposed to recognize automatically regions of large shear and use the information to unfold Doppler velocities which are presently rejected by the method because of a large mean square deviation.

d. Reflectivity calibration

Most radars in meteorological research give different reflectivity factors when looking at the same target. In order to compare reflectivities, it is necessary to inter-calibrate the radars. In the present case, this was accomplished as follows: The average reflectivity factor was determined at each of the 120 × 120 grid points (section 6a) for each radar. The average reflectivities for pairs of radars were plotted against each other and regression curves determined. The curves enabled the reflectivity measured by any of the radars to be reduced to the reflectivity that would have been measured by any of the other radars. It is necessary, of course, that the reflectivities measured by the different radars be well correlated. In the present case, the reflectivities were reduced to what would have been measured by the CP3 radar. This procedure produces an internally consistent set of reflectivities and enables rather complete reflectivity maps to be constructed. Figure 11 shows the distributions of reflectivity factor at heights of 2.0 and 3.0 km above the ground obtained by this method. If only one of the radars had been used to construct the reflectivity maps, the contours would not have covered the area shown. These constant-height reflectivity distributions are discussed later.

e. Determination of wind fields

The vertical precipitation velocities \( W = w - V_P \) are small throughout the trailing anvil region except below the melting layer where the radar elevation angles are generally small. Therefore, we cannot directly determine \( W \) with any precision from the triple-Doppler radar measurements. Hence, only the horizontal com-

![Fig. 10. Schematic illustrating the method of unfolding Doppler velocities.](image)
profiles of the reflectivity factor and precipitation fall speed (Fig. 7) to establish a relationship between particle fall speed and radar reflectivity factor. Different relationships were used above and below the melting region. The analysis volume was then divided into a number of overlapping boxes or subvolumes, each approximately 15 × 15 km in the horizontal and a thickness selected according to the spatial distribution of the data. For each box, u and v were assumed to be linear functions of the Cartesian coordinates x, y, z. The functions were determined by least-squares methods. This method of recovering the wind components was earlier used by Heymsfield (1979). Thus the horizontal wind and its divergence (and other kinematic properties) were determined in each box. The vertical air velocity was determined by further horizontal averaging of the divergence, and integrating the anelastic continuity equation.

In certain subvolumes, only one radar provided data. For these subvolumes, the MDOP method degenerated into a VVP method. The winds and divergence calculated for these subvolumes were inconsistent with those for neighboring subvolumes and so these single-Doppler velocity data were rejected although the associated reflectivity measurements (adjusted to ensure internal consistency) were retained.

7. Results of the multiple Doppler (MDOP) analysis

a. Vertical profiles of divergence

Figure 12a shows the vertical profile of divergence of the horizontal wind determined by the EVAD method and four of the profiles determined by the MDOP method. The MDOP-determined profiles are for the volumes, based on the four areas (each approximately 15 × 15 km, as mentioned in the previous section and designated by A1, A2, D13, and D14 on the curves), which together most nearly cover the volume analyzed by the EVAD method. There is a general agreement between the five divergence profiles but the divergence profiles determined by the MDOP method show more variability—presumably because they represent averages over smaller volumes than the EVAD analysis. It is believed that at least some of the variability between the four MDOP-determined profiles is real. The divergence profiles determined by the MDOP method do not extend as high as that from the EVAD method since adequate dual- or triple-Doppler coverage did not exist at the higher levels.

Figure 12b is a comparison between the EVAD divergence profile and the average divergence profile for the four MDOP subvolumes. The agreement between the EVAD- and MDOP-derived profiles is considerably improved giving further support to the validity of the EVAD analysis method. It is seen that the strong convergence nearest the surface is also found by the MDOP method not only in the average curve, but also in three
of the four constituent curves. The peak divergence in the EVAD-determined profile, found at a height of 1 km, has a larger magnitude than the corresponding value for the MDOP-derived average curve. This is probably due to the larger smoothing implied by the lower vertical resolution of the MDOP method. Note, however, that area A1 exhibits the same peak divergence as the EVAD profile. The differences between the two profiles in Fig. 12b are likely due to a small mismatch of the two analysis volumes and the unique spatial weightings of the two curves as determined by the distribution of radar data points in space and time.

**b. Horizontal distributions of divergence and vertical air velocity**

The horizontal distribution of divergence determined by the MDOP method is shown in Fig. 13 for four selected heights. The divergence values ($10^{-5}$ s$^{-1}$) for each of the MDOP analysis subvolumes are shown and regions of convergence are hatched. At the lowest height (0.5 km), convergence predominates while the reverse is the case at the 1.0 km height, corresponding to the sharp change in the divergence profile in Fig. 12a below 1 km height. Similarly, the predominance of convergence at 4 and 7 km heights is consistent with the vertical profiles of divergence in Fig. 12.

The field of divergence is not horizontally uniform, but instead exhibits a banded appearance which is especially pronounced at the 7-km height. The bands are oriented approximately north–south and thus are only very roughly parallel to the approximate northeast–southwest orientation of the squall line. The scale of the bands is approximately 60 km and one can anticipate a similar structure in the vertical air velocity field. The reason for the banded structure is not known, but it may be noted that in his numerical simulation of mesoscale unsaturated downdrafts, Brown (1979) found an alternation of upward and downward air motions, which he attributed to internal gravity waves. The alternating upward and downward motions would translate into patterns of alternating convergence and divergence.

The vertical air velocity in the individual columns, obtained by integrating the divergence profiles, was found to vary considerably in the horizontal; this is not surprising in view of the differences between the divergence profiles shown in Fig. 12a. A problem in calculating $w$ for each of the columns was that, in some cases, the divergence profiles were not continuous in the vertical due to spatial limitations of the data dictated by the radar scans. Therefore, it was decided to determine average profiles of the vertical air velocity. The square area, shown in Fig. 4b, was divided into two halves (each 120 km in the north–south direction and 60 km in the east–west direction). The average divergence profile over each area was determined; together with the average profile over the entire area and the EVAD-determined profile, they are shown in Fig. 14a.

The vertical air velocities for the left-half (western) and the right-half areas and that determined by the EVAD method (curve A, Fig. 9) are shown in Fig. 14b.
To keep the figure uncluttered, the average \( w \) for the entire MDOP analysis area is not shown; it will fall between the left-half and right-half curves. Each of the profiles shows a small upward vertical air velocity in the lowest layers extending from the surface to about 1 km height. The peak magnitude of this vertical air velocity is about 10 cm s\(^{-1}\). At higher levels there are significant differences between the vertical air velocities in the two halves.

The left half of the analysis area shows a profile of \( w \) similar to the curve determined by the EVAD method. Above the shallow layer of weak ascent near the surface, downward motion extends to a height of about 6 km, reaching a peak magnitude of about 20 cm s\(^{-1}\). Above 6 km, the motion is again ascending and rapidly reaches a maximum value of about 35 cm s\(^{-1}\) at a height of 9 km. At this height, \( w \) still shows a tendency to be increasing. It is reasonable to assume,
however, that if the MDOP analysis could have been carried to higher levels, $w$ would have attained a maximum, and then would have declined to zero as in curve C of Fig. 9. Based on the pattern of the vertical air velocities and reflectivity factors, it is reasonable to assume that the ascending motion in the upper region played a role in the maintenance of the stratiform cloudiness.

In the right half area, immediately rearward of the leading convective region, downward air motion of 10–20 cm s$^{-1}$ extends from about 2 km height to the top of the analysis at about 9 km height. Reference to Fig. 4b shows that the downward air motion is consistent with the reflectivity distribution. The intense squall line convection marked by cellular structure and high reflectivities is situated to the east of the MDOP analysis area. To the west of the squall line, a region of clearing and patchy echoes with low reflectivities extends into the right half of the analysis area. These features are consistent with the downward air motion which would have a dissipative effect on the clouds.

The cause of the predominantly downward air motion in the right half of the analysis region is not known, but we can speculate about it. As mentioned in section 3, the line of active convection in the analysis region is not known, but we can speculate about it. As mentioned in section 3, the line of active convection formed ahead of it during the course of the observations. (The line in Fig. 4b is the new line.) The initiation of new convection is due to convergence at the leading edge of the gust front. Perhaps the gust front had marched so far ahead of the old line, as the air from the convective-scale downdrafts accumulated and spread out near the ground, that the new convection was not able to merge with the old line. This led to the dissipation of the old line and formation of the new one. The downdraft in the right half of the MDOP analysis area, and to the rear of the new line, may be part of the compensating subsidence associated with the new line of convection or lingering descent associated with the old convection line.

A symptom of the dissipation of the old convection line may also be seen clearly in the constant altitude reflectivity maps of Fig. 11. The band of relatively high reflectivities running approximately northeast–southwest along the diagonal of the analysis area is a remnant signature of the old convection line. Lower reflectivities prevail to the southeast between the band and the new line of convection. To the northwest of the band the reflectivities decrease regularly towards the rear of the anvil. This is probably a simple kinematic consequence of precipitation particles from decaying thunderstorms along the convection line having been carried to the rear with the heavier more reflective particles falling nearer to and the lighter less reflective particles falling farther from the rear of the line.

c. Wind distribution transverse to the squall line

The MDOP analysis gave the wind velocity distribution on constant-height surfaces. We have used the
data to calculate the average horizontal winds in narrow bands running parallel to the northeast-southwest diagonal (i.e., approximately parallel to the squall line) of the analysis area shown in Fig. 4b. The results of this calculation were used to obtain Fig. 15a which shows the vertical cross section of the component of the average horizontal wind transverse to the squall line and relative to it. The abscissa in this figure is distance along the northwest-southeast diagonal of the analysis area. The origin of the abscissa is the projection of the location of the CP3 radar on the diagonal. (The location of the radar may be seen in Fig. 4b, 11 or 13.)

The length of the diagonal is $120\sqrt{2}$ km, i.e., approximately 170 km but the cross section itself spans only about 150 km since the averaging of horizontal winds could not be carried right up to the ends of the diagonal. Because of the orientation of the averaging direction, the averaging length decreases as one moves from the center of the diagonal in either direction; the least averaging length occurs at the ends of the cross section. However, the number of radar data points involved is very large even at the ends, particularly at the southeast end of the cross section. To put this wind cross section in the context of the reflectivity distribution, the height cross section of the reflectivity along the northwest-southeast diagonal is shown in Fig. 15b. No averaging was performed in this case since the intention is simply to put the wind cross section and the reflectivity distribution in context. An indication of the melting band may be seen in Fig. 15b. The highest reflectivities again are associated with the “old” convective line as in the constant height representations of Fig. 11. The center of the new convective line is situated at an abscissa value of about 90 km, outside the range of the plot of Fig. 15.

The wind distribution in Fig. 15a shows three layers with distinct flow regimes. In the top layer, there is inflow at the front of the trailing anvil above 3 km height, and outflow at the rear above 6.5 km height. In the middle layer, environmental flow enters the rear of the anvil, between heights of 3 and 6.5 km, and there is outflow at the front of the anvil near the ground to a height of about 3 km. In the lowest layer, there is flow towards the back of the anvil; the thickness of this layer increases from near zero at the front to about 3 km at the rear of the anvil. The three layers are separated by two surfaces of zero motion.

The flow field depicted in Fig. 15a is practically instantaneous. If system-relative stationarity, two-dimensionality and conservation of horizontal momentum are assumed, then we can conclude that 1) there is inflow into the front of the anvil above 3-km height which ascends to above 6.5 km by the time it reaches the back of the anvil; 2) the wind enters the back of the anvil in the height range 3–6.5 km and descends toward its front in a layer extending from the ground to about 3 km height; and 3) in the lower layers air ascends in the anvil region in a layer which deepens with distance to the rear of the anvil where it reaches a height of 3 km above the ground. If it is also assumed that the winds do in fact ascend and descend along the contours in Fig. 15a, then the average intensity of ascending and descending motions can be calculated. This calculation yields an average downdraft of 30 cm s$^{-1}$ at middle levels and an upward motion of similar
magnitude at higher levels. These magnitudes are very similar to those calculated by the EVAD and MDOP analyses. This suggests that the assumptions made may have some validity. However, the profiles of the vertical air velocities in the two halves of the analysis area, presented earlier, cannot be understood completely in terms of winds ascending and descending along the contours of Fig. 15a. Hence, the assumptions of twodimensionality, system-relative steadiness, and momentum conservation can be only partly true.

The foregoing description of ascending and descending motions in the anvil region is qualitatively very similar to those of Sanders and Paine (1975; their Fig. 5) and Ogura and Liou (1980; their Fig. 13). We can compare our results with those of the latter authors. In their Fig. 13, $x = 0$ corresponds to the location of the squall line, while in our case the corresponding point is at an abscissa value of 90 km in Fig. 15a. Thus the portion of their cross section for $x < -10$ km should be compared with our Fig. 15a. The qualitative features of the two figures are very similar, except that the descending motion does not form a continuous layer in the case of Ogura and Liou. A point of difference that may be relevant here is that in our case a line of convection decayed and a new line formed.

It may be repeated that Sanders and Paine, as well as Ogura and Liou, assumed system-relative stationarity to calculate winds from rawinsonde observations. We obtained practically instantaneous winds but were forced to assume stationarity to interpret the air flows inside the anvil region. Clearly a more satisfactory solution would be to perform a time series of analyses of the type reported here so that conclusions could be drawn without assuming stationarity.

8. Summary and discussion

Doppler radar data have been used to study the region of stratiform precipitation falling from the trailing anvil accompanying a midlatitude squall line. The squall line moved at approximately 15 m s$^{-1}$ in a southeasterly direction. Individual thunderstorms formed up to 20 km ahead of the leading edge of the squall line and moved in a generally northeasterly direction. Dissipating thunderstorms and their anvils merged to form the region of stratiform precipitation. The leading line of active convection dissipated and a new line formed ahead of the dissipating line during the course of the observations.

One of the objectives of this study was to determine the mesoscale vertical air motions to an accuracy of 10 cm s$^{-1}$ or better. This led to the development of some new techniques for the analysis of the Doppler radar data. The VAD method is well known in radar meteorology as a method for the determination of horizontal winds and vertical air velocity in situations of widespread homogeneous precipitation. However, accurate vertical air velocities are difficult to determine by this method because of the contaminating effect of the particle fall velocity on the determination of the divergence of the horizontal wind, and the difficulty of correcting accurately for that contamination. We have described an extended VAD (EVAD) method which does not require knowledge of the particle fall speed. Rather, it determines both the particle fall speed and the horizontal divergence from a series of VAD observations at a number of elevation angles. Because of the necessity for using a series of VAD observations, the EVAD method imposes a more stringent requirement on steadiness of the phenomenon as compared to the VAD method which requires, in principle, only one VAD scan. In the present instance the series of VAD observations took 9 min; however, the required sequence could have been completed in 5 min or perhaps less without sacrificing the accuracy of the analysis.

The application of the EVAD method to the present data yielded particle vertical velocities in excellent agreement with those determined by measurements at vertical incidence. This lends support to the EVAD analysis method and implies that the divergence calculated by the method was highly accurate. The accuracy of the vertical air velocity found by integrating the continuity equation is obviously contingent upon the accuracy of the divergence. However, because of the integration, the vertical air velocity may suffer an offset (and errors may amplify) if knowledge of the divergence profile is incomplete at the boundaries. In radar observations, such incompleteness usually occurs near the ground and the cloud top. The incompleteness near the cloud top due to low reflectivities can perhaps be eliminated (in future experiments) by the use of more sensitive radars. However, the incompleteness near the ground is of a more fundamental nature, being due to "seeing" limitations because of earth curvature and electromagnetic wave propagation effects. Ground clutter contamination is also a problem, but this can be overcome with proper equipment and data processing techniques.

In the past, multiple Doppler radar methods have been usually applied to convective storms. An uncertainty of a few to several meters per second usually exists in the determination of the vertical air velocity by the method. To obtain accurate determinations of mesoscale horizontal wind fields and divergence, and thereby the vertical air velocity, we used the Doppler velocities in relatively large volumes to determine $u$ and $v$ as linear functions of the Cartesian space coordinates by least-squares fitting techniques. The fitting procedure leads to a filtering of motions of high wave number. We found very good agreement between the divergence profile determined by the single-Doppler EVAD method and an average divergence profile calculated by the MDOP method over the approximate domain of the EVAD analysis. There was a correspondingly good agreement between the vertical air velocities determined by the EVAD method and that.
determined by integrating the divergence found by the MDOP method. It is to be noted that whereas the EVAD method gives horizontal averages of quantities over its analysis domain, the MDOP method also yields the horizontal variation of the quantities over its anal-
ysis region.

The vertical air velocity profile determined by the EVAD method showed weak (peak magnitude about 10 cm s⁻¹) ascending motions below a height of 1 km, descending motions (peak about 25 cm s⁻¹) from about 1 km to 5 km height and ascending motions (peak about 35 cm s⁻¹) from about 6 km height to the cloud top over the EVAD analysis domain. With the exception of the ascending motion in the lowest layers, this profile of vertical air motion is remarkably similar to that deduced or hypothesized by Zipser (1977) and Houze (1977) for tropical squall lines, and calculated by Ogura and Liou (1980) for a squall line observed in Oklahoma. The latter authors used compositing techniques on a time sequence of radiosonde data, collected over a period of about 7 hours, to calculate the winds, divergence and vertical air velocity across the squall line. Their technique required the assumptions of steadiness in a frame of reference attached to the squall line over the period of the data collection. The profiles calculated here are free of such assumptions and represent practically instantaneous distributions. The weak ascending motion in the lowest levels is thought to be real and not the result of uncertainties in the data or the analysis methods. Its existence is supported by both the EVAD and the MDOP analyses.

The pattern of the average horizontal component of the wind velocity transverse to the squall line, and relative to it, determined by the MDOP method, is similar to the corresponding patterns calculated by Sanders and Paine and Ogura and Liou. The air enters the trailing anvil from the front and apparently ascends along a sloping surface. At its rear, the trailing anvil is overtaken by environmental air at middle levels, which apparently descends as it moves toward the front of the system. Thus these two flow regimes are separated by a surface of zero relative velocity, which slopes upward with increasing distance toward the system's rear. In the lowest layers of the trailing anvil region, the air again apparently moves from the front to the back in a layer that increases in depth toward the rear of the system. The two lower flow regimes are also separated by a surface of zero relative velocity, which slopes upward toward the system's rear.

The vertical air velocity profiles calculated by averaging the divergence over the left (western) and right (eastern) halves of the MDOP analysis volume show important differences. The two profiles agree with the EVAD profile in showing weak ascending motions up to a height of about 1 km. Above that height, the right-half area nearest to the leading convective region shows descending motion through the depth of the analysis region. This descending motion correlates well with the region of weak reflectivities found between the dissipating squall line and the new line of active convec-
tion that formed ahead of it. The left-half region has a vertical air velocity distribution essentially in agreement with the results of Ogura and Liou, namely mesoscale downdraft below and mesoscale updraft above. A signif-
ificant difference is the occurrence of the weak asc-
cending motions in the lowest levels in our case. Also our results are practically instantaneous depictions of the field and did not assume system-relative station-
arity.

The horizontal variation of vertical air velocity found here was not suggested by previous workers since their data and analysis techniques precluded detailed analysis of horizontal variations. Similarly, previous efforts have not revealed details of the variation of the divergence field. The distribution of divergence calculated here has a pronounced banded pattern on horizontal surfaces, especially at the middle and higher levels.

Some speculation has been offered on the possible reasons for the horizontal variations of the vertical air velocity. The descending motion in the eastern half of the MDOP analysis volume has been ascribed to compensating subsidence associated with a newly formed line of convection. More extensive observations are needed to test this hypothesis and also the reasons for the banded structure of the divergence field. In the present study, radar observations enabled us to compute practically instantaneous distributions of hori-
zontal winds, divergence, and vertical air velocity for a nine-minute period. A time sequence of such distributions, over a larger area, in conjunction with detailed thermodynamic information, is needed to clarify the physical mechanisms.

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