The Oklahoma Squall Line of 19 May 1977. Part II: Mechanisms for Maintenance of the Region of Strong Convection

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ABSTRACT

Mechanisms for maintenance of the strong convection along the leading edge of a broad squall line that occurred in Oklahoma on 19 May 1977 are investigated. The findings are based upon analysis of data from a surveillance radar, a surface mesonet, Doppler radars, proximity soundings and aircraft data, and upon the results of a two-dimensional, cloud-scale numerical simulation. The detailed results of the multiple Doppler analysis are contained in the Part I paper reporting results of research on this squall line.

It is found that at a preferred location along the squall line, an area of intense convection is maintained over a long time period. A meso-β scale organized structure, which includes an area of low pressure near the southeast edge of the intense convection and an associated area of convergence extending to the east, promotes the formation of small showers in short line segments. These showers, due to their differing motion from elements within the main line, merge with the line to the north of the mesoscale, resulting in maintenance of the strong area of convection. The observed meso-β structure on this day is believed to be made possible by a deep low-level layer of weak vertical wind shear and high water-vapor content.

At other locations along the line, the numerical simulation indicates an unsteady behavior in the maintenance of squall line convection by gust frontal convergence. Perturbations in the vertical motion field are periodically initiated by either (i) enhanced convergence at the gust front resulting from diverging downdrafts at locations farther to the west, or (ii) Kelvin-Helmholtz instability produced at the gust front head. These perturbations move westward relative to the gust front above the low-level cold air and periodically invigorate the main region of updrafts located a few tens of kilometers west of the gust front. Low-level updrafts, forced by diverging surface outflow from weak downdrafts, occasionally interact with the translating perturbations to increase their amplitude. The existence of the westward-moving perturbations is tentatively substantiated by the presence of similar structures in the analyzed Doppler wind fields. Greater time resolution in Doppler data, in combination with more comprehensive surface and upper air data ahead of squall lines of this type, would aid in confirming the reported structures.

1. Introduction

The squall line that occurred in Oklahoma on 19 May 1977 can be related to processes on a number of meteorological scales. Part I of the present work (Kessinger et al., 1987, hereafter designated KRH) includes a description and interpretation of squall line-related features on a series of scales. In addition to synoptic and meso-α influences, processes on the meso-β scale, which in this case includes both a highly convective leading edge and a nonconvective precipitation area to the rear of the convection, are investigated in Part I. A meso-γ scale consisting of the individual thunderstorms comprising the convection and in which the potential energy of the environment is converted to the kinetic energy of the line circulation is also explored in Part I. In this paper (Part II) attention is focused on the convection at the leading edge of the line, and in particular, how it is maintained during the mature stage of the line's evolution.

The most commonly held view is that forced lifting of the potentially buoyant air ahead of the line by the gust front is the primary, if not the only, mechanism...
by which the strong convection in the line is maintained. Environmental air forced upward over the more dense outflow air reaches its condensation level and begins to release latent heat, followed by rise to even greater heights due to its newly acquired buoyancy. In general, this view is probably correct and is supported by the observation that the strongest convective elements in the squall line may be traced backward to an origination point near the gust front or outflow boundary. It will be shown here, however, from evidence based upon both observation and modeling, that other influences are also significant and can strongly modulate the gust frontal convective initiation process.

It is difficult to identify in the early literature any one work that might be credited with describing the role of the gust front in maintaining squall line convection. Early conceptual models such as that of Möller (1884) described the flow quite well just above the surface and were based primarily upon visual observations. Suckstorf (1939) quantitatively explained the rain-cooling effect based upon moist adiabatic descent in heavy showers and measurement of standard meteorological variables by surface instrumentation beneath European and tropical thunderstorms. His conceptual model (probably most applicable to thunderstorms in weak shear) included outflow of cold air near the surface, with environmental warm, moist air ascending toward the middle updraft upon encountering the outflow boundary.

Using data from the Thunderstorm Project in Ohio, Newton (1950) analyzed and interpreted the two-dimensional thermodynamic and airflow structure of a prefrontal squall line existing in a strongly sheared environment. He emphasized the vertical transport of horizontal momentum by storm updrafts and downdrafts as an important factor in creating storm-scale divergence distributions favorable for storm propagation. Newton (1950) and Newton and Newton (1959) schematically depicted the storm-relative airflow within this squall line as is shown in Fig. 1. In this case, both a “squall front” (right) and a polar front (left) were identified, with strong upward motion ahead of the squall front contributing to the strong updraft near the leading edge of the line. The presence of the cold front is important, apparently because it enhances the intrusion of dry air in middle levels toward the rear of the squall line, resulting in increased evaporative cooling and augmentation of downdrafts.

Bergerson (1954) showed an example of a squall line that “formed within an originally homogeneous southerly current of tropical air.” His schematic model of the airflow, cloud, and precipitation structure, shown in Fig. 2 contains no indication of any invasion of polar air, and appears similar in a number of details to the squall line analyzed in Part I. An area of lower pressure is present at the surface just ahead of the rain area; the main updraft appears to originate near the gust front and slopes in an upshear direction. A meso-β scale region of precipitation exists to the rear of the main updraft, containing upward motion separate from the convective updrafts. An apparently stronger region of ascent is implied near the left-most edge of the meso-scale system. The meso-β scale region present in Fig. 2 and in many mature squall lines has been referred to as the “stratiform” region in a number of recent studies. The term will also be used in this paper, although it remains to be shown that such regions are totally free of convection in all cases. The resemblance of several recent conceptual models based upon Doppler wind data (Zipser and Matejka, 1982; Houze and Smull, 1982; Kessler, 1983) to Bergeron’s schematic is striking. It is also important to note that many squall lines occur from initiation to dissipation without trailing areas of stratiform precipitation. Other squall line systems undergo an evolution as outlined by Sanders and Emanuel (1977), wherein an early stage contains only convection, a mature stage includes both convective and stratiform precipitation, and a dissipating stage contains only stratiform precipitation.

Little documentation exists of factors other than gust frontal convergence that significantly influence the maintenance of strong convection in squall lines. Part of the reason for so little attention to other factors is that observation on sufficiently small scales has not been possible. Other factors were evidently important in a severe squall line reported by Boucher and Wexler (1961) that occurred in southern New England. In that case, scattered cells formed about 30 km ahead of a line of cellular echoes. With time the cells ahead intensified and became the new line while the old line dissipated. At the same time new scattered cells appeared 25 km ahead of the second line and the process of intensification and dissipation repeated. The line propagated in this manner for about 4.5 hours with the new cells forming on the average every 50 minutes or so at a distance averaging about 16 km ahead of the mature line. This sort of discrete line propagation appears to be rather uncommon.

More common is a form of discrete propagation where new cells form ahead of a mature squall line near the gust front, then grow with time to become the strong convection in the line. A number of recent studies using Doppler radar have shown that gust frontal convergence is important in initiating new convection and leads to the maintenance of squall lines (e.g., Elmore, 1982). However, there apparently has been no detailed investigation of cells forming well ahead of the squall line gust front that later merge with the line. Fankhauser (1982) and Miller and Fankhauser (1983) document a case study of a large rain and hailstorm that occurred on 22 June 1976 in northeastern Colorado. The storm was located at the southern end of a north–south convective line, and provides an excellent example of significant pre-line convective initiation. In this case a long-lived storm roughly 20 km wide by 30 km long exhibited both supercellular and multicel-
Fig. 1. Time section through a squall line passing over the Ohio Network of the Thunderstorm Project on 29 May 1947. Solid lines indicate boundaries of stable layers; dotted lines are boundaries of stable layer between strong convection and cold front with relatively dry air above it. Thin lines are wet-bulb potential temperature contours and cloudy regions are shaded. Circulation indicated by arrows is schematic. (From Newton and Newton, 1959.)

Lular characteristics. The multicellular phase began when low-level outflow became well developed and new cells intermittently developed along the southeastwardly propagating outflow boundary and moved northward into the main portion of the storm. Figure 3, taken from Fankhauser (1982), illustrates positions...
of cell origin relative to the outflow boundary over several hours of storm evolution. Cells that reached reflectivities in excess of 60 dBZ are denoted by solid circles. Note that the locations of two cells, C9 and C10, are well in advance of the boundary. It was proposed that the initiating mechanism for these two cells was an isolated area of convergence identified in an analysis of surface mesonetwork data. The confluent nature of the flow in this area of convergence was tentatively related to accelerations associated with a mesodepression located approximately beneath the storm vault.

The existence of mesoscale areas of low pressure at the surface in advance of squall lines and individual thunderstorms is well-documented. It is more difficult, of course, to determine three-dimensional structure, and in particular, the vertical extent of such mesoscale pressure fields. Fankhauser (1974) has analyzed wind data from a finescale rawinsonde network in the environment of an Oklahoma squall line and found that low pressure (pressure) values were most prominent near the 600 mb level at a distance slightly less than 10 km ahead of the line's leading edge. At low levels the trough of low heights parallel to the line was less prominent and approximately 20 km in advance of the leading edge. The derived locations and magnitudes of deviations from the mean as compared to actual fields are certainly influenced by the spatial resolution of this data. Hoxit et al. (1976) have proposed that such mesoscale lows are produced by subsident warming in the 100–500 mb layer downwind of dynamically active tops of large cumulonimbus clouds. Subsidence is produced, they claim, through compensation for storm updrafts or due to a lee wave effect. The lee wave effect apparently refers to the idea that if the thunderstorm top acts as a barrier to the flow and forces the stably stratified environmental air upward, subsidence will occur in the near downstream region as part of a near-source gravity wave response. The same authors have analyzed surface network and radar data over the National Hail Research Experiment network in northeast Colorado to show a mesolow approximately 15 km east and slightly south of the region of strongest echoes. The calculated low-level convergence northwest of the mesolow increased as the low intensified, presumably aiding in the maintenance of the area of maximum updrafts and rain production (reflectivity) in the same area relative to the storm complex.

The structure and evolution of the 19 May 1977 squall line are here examined on several time and space scales. First, over a period of several hours the influence of small cells forming ahead of the outflow boundary upon the evolution of the line as a whole is examined. Second, on a much smaller time scale and in the area of the line from the outflow boundary westward to the area of strong updrafts, two other mechanisms are investigated that may influence the maintenance of strong convection. Reflectivity data from the WSR-57 surveillance radar at NSSL are utilized to examine the evolution and structure on the larger scale. On the smaller scale both analysis of observed data and numerical experimentation with a two-dimensional convective cloud model are carried out to examine initiating mechanisms.

2. Merger of organized showers with the squall line

A remarkable feature of the 19 May squall line was the persistence of a high-reflectivity area within the line during a period of at least 3 hours. The location of this area is depicted in Fig. 4 as it approached the surface and Doppler radar networks west of Norman. Motion of this area during the 1145–1345 CST period was approximately 15 m s$^{-1}$ from 240 deg. It should be noted that the area shown in Fig. 4 was at first contained within an isolated cell and later within a solid line. Based principally upon radar data, Bluestein and Jain (1985) placed formation of severe Oklahoma squall lines within four categories. The line examined here most closely resembles those in the “broken line” formation classification, although the first several cells that formed in late morning were along the eastern edge of an area of rain that existed during the previous night. As was pointed out by KRH, motion of the line as a whole was from 270$^\circ$. Therefore, the area shown in Fig. 4, which will henceforth be termed “cluster A”, was moving northward within the line as the line as a whole moved eastward. The eastward speed of the line was noted to be about 10 m s$^{-1}$ by KRH during the time of multiple Doppler analysis and perhaps as little as 6 m s$^{-1}$ in its southern portion. In the time period
from 1145 to 1345 CST line motion was approximately 13 m s\(^{-1}\). KRH also note that individual cells within the line (smaller scale than this area) moved from about 207 deg at 10.5 m s\(^{-1}\). Thus there seem to be two scales of discrete propagation within the line, similar to that proposed by Newton and Fankhauser (1975) where the “squall line as a whole typically moves over a swath to the right of upper winds as a result of accretion and loss of clusters, while the clusters themselves move (generally to a lesser degree) to the right of the mean wind as a result of systematic accretion and loss of their component cells.” Evidence of new “cluster” growth is perhaps present in the growing activity between 1500 and 1600 CST on the southern end of the area of convection shown in Fig. 4.

The major reason for the rightward movement of cluster A (relative to cell motion) shown in Fig. 4 is that convective showers within organized lines oriented roughly perpendicular to the squall line were intermittently merging with A on its southeastern flank. Careful analysis reveals that these small cells ahead of the squall line clearly formed east of the outflow boundary and typically existed for one hour or so before merging with the line. The term “cell” is here used to denote a region of reflectivity higher than its surroundings that remains an entity that can be tracked for some tens of minutes. The minimum reflectivity associated with a cell at its start is 20 dBZ. Reflectivity levels within these cells remained low until they were very close to the line, whereupon in many cases vigorous growth occurred during merger. The result of this discrete invigoration at the same relative location along the line is that the cluster or large multicellular storm maintained a higher level of convective intensity than other portions of the line during this period.

The pre-line cells, in addition, appear to have been affected by some form of mesoscale organizing influence as evidenced by their appearance in short lines (illustrated in Fig. 5). Such a line, which will be henceforth abbreviated CASL (confluent asymptote shower line), tends to extend east-southeastward approximately 40 km from the leading edge of the squall line near the point of maximum reflectivity. Spacing of cells within the CASL varies considerably but is typically 8–12 km. Movement of the cells is approximately 15 m s\(^{-1}\) from 170 deg, which agrees remarkably well with the mean wind speed and direction in the lower 6 km or so from the Elmore City sounding at 1434 CST. The lack of strong shear in the environmental wind ahead of the line below 5 km, as well as the deep low-level moist layer, must have contributed to the longevity of the CASL cells. Merger of these cells with the line took place as they were successively overtaken by it. Figure 5 shows the positions of six of these cells, labeled A1–A6, at 10 minute intervals during the 1310–1400 CST period. New cells appeared at the eastern end as the CASL moved northward, while cells on the western end merged with the main line (cluster A) and lost identity.

Following 1400 CST new development in the CASL of Fig. 5 ceased and another short line formed 10–20 km to the east. Locations of cells within the second line during the 1330–1500 CST time period are shown in Fig. 6. Cell B2 formed first at about 1330 CST and cells B1, B3 and B4 all formed near 1350 CST. Movement of these cells was nearly the same as cells A1–A6 shown in Fig. 5. By 1420 CST the remaining cells in

![Fig. 5. Isochrones (left) of the leading edge of the 34 dBZ, reflectivity contour associated with “cluster A” from 1310–1400 CST on 19 May 1977. At 1310 changes in reflectivity at 46, 58, and 70 dBZ, are also shown. The lattice at lower right is made up of tracks (solid) of small cells ahead of the squall line and isochrones (dashed) of cell positions.](http://journals.ametsoc.org/jas/article-pdf/44/19/2866/3423969/1520-0469(1987)044_2866_toslom_2_0_co_2.pdf)
A1–A6 and the western cells in B1–B4 had merged just east of the leading edge of the squall line. In Fig. 7 the reflectivity within the squall line is shown over approximately 270 km of its north–south extent. The two cells located about 40–50 km west of Norman merging with the line in Fig. 7 have resulted from growth of cell B4 shown in Fig. 6. These two cells continue to merge and grow, forming the northeastward extension of “cell A” referred to by KRH in the 1434 CST Doppler analysis. Cells B2 and B3 in Fig. 6 continue their north-northwestward movement and are labeled “E” and “E” by KRH. Following 1430 CST CASL cells ceased to form in the area east of cluster A. Significantly, cluster A declined in intensity also, as is indicated by the decrease in strength of “cell A” in the Doppler analyses of KRH between 1434 and 1502 CST. This decrease may also be seen in the changing reflectivity pattern west of Norman in Fig. 4. There is at the same time an increase in intensity in an area of the line approximately 30 km to the south.

Prior to synthesizing all this information into a more comprehensive description, one more area of the line will be examined. In Fig. 8 the positions of two cells are plotted over a period of 50 minutes, as well as the smoothed position of the easternmost 33 dBZ_e contour within the squall line. The high-reflectivity area within the line at 1320 CST is the core of cluster A (Fig. 4). The pre-line cell with the easternmost track in Fig. 8 was initiated at least 60 km east of the squall line. At about 1350 CST a second cell formed northwest of the first, following a parallel track. The second cell formed very rapidly, attaining a reflectivity in excess of 50 dBZ by 1410 CST and prior to merging with the line. The surface network unfortunately did not extend sufficiently south and west to encompass this region. However, it is hypothesized that this cell's rapid development was enhanced by convergence associated with an outflow boundary extending southward from the eastern edge of cluster A.

Fig. 7. Reflectivity within the 19 May 1977 squall line at 1420 CST. Light stippling indicates reflectivities from 24–34 dBZ_e; no stippling, 34–46 dBZ_e; dark stippling 46–58 dBZ_e; and repeated light stippling, 58–70 dBZ_e. The dashed square indicates the area encompassed by Fig. 10.

Fig. 8. Isochrones of the leading edge (24 dBZ_e contour) of the squall line at 10 min intervals from 1320–1410, and tracks of two cells ahead of the line during the same period.
The reflectivity pattern in Fig. 8 clearly shows that at all times cluster A bulged eastward relative to the remainder of the line, resulting in a northeast-southwest orientation of the leading edge of the line immediately to its south. With the rapid growth of the strong cell shown in Fig. 8 and its subsequent merger with the line, the leading edge underwent a forward “jump” as may be surmised by examination of Fig. 7. The large cell approximately 80 km west-southwest of Norman in Fig. 7 is a result of the merger of this pre-line cell with elements of the line to its west. Indeed, the line appears to have “grown around” the merged cell into an area that subsequently became the dominant area in the line (see Fig. 4).

Finally, returning to the area along the leading edge of cluster A, surface network data is examined to help explain the structure and evolution of this portion of the line. In Fig. 9 (adapted from KRH, Fig. 16) the surface gust front position and pressure distribution are illustrated at 1434 CST. A low-pressure center is located on the gust front and is associated with a well-defined circulation in the surface winds. In the Doppler analysis of KRH at this time there is a reflectivity notch in the low levels above the northwest portion of this low-pressure center. The near surface wind field contains strong cyclonic vorticity at the notch location. Maximum upward motion at 2 km altitude is located several kilometers south of the notch within the low-pressure area and just west of the gust front. Locations of the reflectivity notch and maximum updraft aloft are indicated in Fig. 9. The notch or weak echo region extends upward into the storm and is identified aloft with a region of strong relative east-to-west flow based upon the Doppler-analyzed winds at 2 km and the winds measured by the NCAR Queen Air flying at 300 m above the surface (see Kessinger, 1983, Fig. 7.9). Extending eastward from the low-pressure center is a trough of low pressure whose identity is lost due to sparsity of data on the eastern edge of the network. Coincident with this pressure trough is an area or band of convergence oriented approximately east-west near the squall line and bounding generally southeasterly winds to its south and easterly or east-southeasterly winds to its north. An analysis of the divergence field over the PAM network is shown in Fig. 10 at 1415 and 1434 CST. An area of convergence extending eastward from the main convergence region along the squall line’s outflow boundary is present at both times. The northward component of motion of this convergence area appears to be tied to the northward movement of cluster A within the line. Examination of the time variation of wind direction from individual stations over which this convergence area passed reveals that there was a gradual rather than a sudden change in wind direction, and therefore a diffuse rather than a sharp line of convergence is implied. It is asserted here that the CASL cells, which were responsible for maintaining the strength of cluster A through successive mergers, formed within this convergence area. The locations of cells B2 and B3, as well as cell B4 (two parts are merging separately with line), are indicated in Fig. 9. Locations of the two parts of cell B4 are also shown in Fig. 10. The locations of B2 and B3 north of the convergence area are thought to result from formation of convective clouds in the convergence area followed by a gradual separation due to a larger northward motion component of the CASL cells.

In summary, then, one mechanism for maintenance of strong convection within this particular squall line is through the merger of pre-line echoes with the line in a preferred location. The pre-line (CASL) echoes exist in short lines that form within a region of convergence along a confluent asymptote, extending generally eastward from a pronounced area of low pressure in low levels. The low is located on the east flank at the southern end of a large multicellular storm within the squall line. Once formed, the CASL cells move with the lower tropospheric winds and remain weak until developing rapidly during merger while enhancing the major center of activity within the squall line.

It is not clear what process maintains the low-pressure area that is so vital to this maintenance mechanism. A possible cause for the pressure trough extending eastward from this area is the subsident warming idea suggested by Hoxit et al. (1976). Examination of the vertical velocity fields from the Doppler analysis at various heights reveals that the low-pressure center along the outflow boundary at 1434 CST is beneath the southeastern boundary of a large area of updrafts.

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**Fig. 9.** Surface ground-relative wind at PAM/SAM stations, analyzed pressure field (1 hPa interval) and locations of CASL cells near the leading edge of the squall line at 1434 CST. CASL cell positions are indicated by circled asterisks and gust front by heavy chained line. A full wind barb represents 5 m s⁻¹ and half barb 2.5 m s⁻¹. Also indicated are the low-level reflectivity notch (circled “N”) and the maximum upward motion at 2 km (circled “U”) based upon Doppler data analysis (adapted from KRH, Fig. 16).
from the surface to 8 km, and in an area without data above that. With reference to the reflectivity field at 12 km (height) the low is 17 km east-southeast of the maximum and 7 km southeast of a secondary maximum. The low-pressure area therefore appears to occur closer to the storm core than would be implied by the hypothesis of Hoxit et al. (1976).

Other examples of mesolows in advance of squall lines have been documented in past work. Using a mesoscale (hydrostatic) numerical model with convection parameterized, Fritsch and Chappell (1980) simulated the larger-scale aspects of squall line structure and evolution. Their results include a mesohigh beneath the rainy area in the line and mesolow formation due to subsidence aloft just ahead (southeast) of the line. Typical pressure differences and horizontal separation between high and low were 1.5 mb and about 100 km. In an observational study Koch and McCarthy (1982) investigated processes that contributed to the development of three mesoconvective systems near an Oklahoma dryline. They documented both “wake” lows and “inflow” lows during the course of their study. The wake lows formed to the rear of convective lines and isolated storms and were attributed to the subsidence mechanism. The inflow lows were observed to travel with the storm and therefore were presumably closely related to the existence of strong updrafts. Marwitz (1973) and Barnes (1978) have also documented the existence of these mesolows in the inflow region within 2–10 km of strong updrafts.

The existence of the low-pressure area in this case provides a favorable environment for development of CASL cells; the cells maintain the intensity of the major center of activity (large multicell) in the line, and the existence of the strong activity is in some manner related to the existence of the area of low pressure. The spatial relationship of these various features is shown schematically in Fig. 11. The center of activity in the line can therefore be shifted only by some major change at another location. In this particular case the major event that appears to have ended this cooperative interaction in and around cluster A was the development of a large cell on an outflow boundary ahead of the squall line south of the position of cluster A. The squall line advanced discretely, reforming around the new cell that later became another cluster or large multicell. This and other clusters that formed later within the line to the south had associated with them pre-line echo structures similar to those described above. The later

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**FIG. 10.** Surface divergence field over the area indicated in Fig. 7 based upon analysis of PAM/SAM data at (a) 1415 CST and (b) 1434 CST. The dashed line shows the position of the gust front (outflow boundary) and “B4” indicates CASL cell positions. Units of divergence are 10^{-5} s^{-1} (Kessinger, 1983).

**FIG. 11.** Schematic representation of the relative positions of the large multicell, the gust front, CASL cells, and mesolows present in the 19 May 1977 squall line. Low-level wind directions are drawn approximately relative to the translating large multicell.
pre-line structures often appeared more banded and contained larger cells than the previously described shower lines. Examples of some of these features are shown in the reflectivity field in Fig. 12.

3. Other mechanisms for convective maintenance found from numerical simulation and Doppler velocity analysis

The preceding description treated a mechanism for maintenance of strong convection at major centers of activity along the squall line. Maintenance at other locations is examined here through a numerical simulation of the leading convection, in combination with flow fields derived from synthesis of data from four Doppler radars. Numerical simulation offers the advantage of finer temporal and spatial resolution than was available in the observations, although a number of factors may contribute to solution inaccuracies (e.g., uncertain environmental conditions, the two-dimensional restriction, and treatment of boundaries). The two-dimensional cloud model of Hane (1973) was used to simulate the strong convection along the line's leading edge. Due to computer storage and time limitations, the model domain was restricted to a 40 km width (16 km depth) in the simulation, so that only the strong convection, and not the trailing mesoscale precipitation area, was investigated numerically. No attempt was made to simulate the complete evolution of the squall line, since it had formed many hours earlier under conditions differing from those sampled by the rawinsonde network. The expectation was that if the model environmental conditions are imposed based upon observations during the mature phase, the simulation would (after an initial start-up period) simulate certain aspects of the mature structure of the line.

For calculation of the motion field the model utilizes a diagnostic equation for the stream function and a prognostic equation for vorticity in the x-z plane. Equations for temperature, water vapor, cloudwater, and rainwater are also prognostic. The grid point separation is 400 m horizontally and vertically, and the time step is approximately 15 s. Parameterizations for microphysical processes and subgrid-scale turbulence are included also (Hane, 1973).

The model simulation in this case uses observed environmental soundings for initialization. The Fort Sill, Oklahoma (FSI) sounding at 1346 CST shown in Fig. 13 was utilized to specify the initial temperature and water vapor fields, which were horizontally uniform except for an enhanced area of vapor coinciding with a region of upward motion. The sounding shows abundant moisture through a deep layer and a relatively steep lapse rate up to 600 mb; the associated lifted index is about −6. The wind field was derived from two soundings: 1) at 1346 CST from FSI and 2) 1558 CST from TVY (40 km north of Norman). The FSI sounding was taken approximately 10 km ahead of the leading edge of the line and the TVY sounding about 30 km behind the line's leading edge. There is, in the utilization of these soundings taken at different times, an implicit assumption of steadiness in the environment of the line. The wind field ahead of the line shows a remarkably uniform structure from the surface to about 5 km with south-southeasterly winds mostly in the 12–20 m s\(^{-1}\) range. Above 5 km the winds veer rapidly to west-southwesterly and increase in strength to about 35 m s\(^{-1}\). The component of motion from these soundings (after smoothing) normal to the orientation of the line is shown in Fig. 14. It is by no means certain that the adopted wind profiles are perfectly representative of the storm environment. The fact that line motion in the simulation was 0–3 m s\(^{-1}\) toward the east, while the observed motion during this period was 6–10 m s\(^{-1}\) toward the east, leads one to question one or both of the profiles. Previous applications of this model (Hane, 1975, 1978) have resulted in good agreement between observed and modeled line speed.

As may be surmised by thoughtful examination of Fig. 14, the initial motion field contained a region of upward motion produced by horizontal convergence in the lowest 1.5 km and horizontal divergence between 1.5 and 6 km. The vertical motion and divergence fields were initially confined within an 8 km wide region containing a smaller region of nearly saturated air to hasten the onset of convection. Convergence and di-

![Fig. 12. As in Fig. 7, but at 1711 CST.](image-url)
Fig. 13. Sounding from Ft. Sill, Oklahoma at 1346 CST on 19 May 1977. Indicated are the 22.5°C pseudoadiabat, the 15 g kg⁻¹ mixing ratio line and the 30° and 40°C adiabats. Flagged wind barsbs are 25 m s⁻¹; full barb 5 m s⁻¹; half barb, 2.5 m s⁻¹.

vergence were specified such that vertical motion varied sinusoidally in the horizontal with a maximum value of several meters per second within the 8-km wide region. After the initial time all fields were allowed to vary except the horizontal motion along the lateral boundaries. The generation of horizontal and vertical motions (wave reflection) is possible when perturbations encounter these lateral boundaries. The greatest potential for this is along the left (west) boundary, which is in precipitation in both the observed and modeled cases due to the presence of the mesoscale precipitation region west of the area of convection. However, careful examination of the motion field during the simulation revealed no apparent significant effects away from this boundary.

Maximum upward motion as a function of time during the simulation is shown in Fig. 15. Maxima in upward motion are in the 8–11 m s⁻¹ range and occur regularly throughout the 100-minute simulation. Very small-scale variations on a time scale of approximately one minute represent the migration of locations of maxima through areas between grid points and should not be attributed any physical significance. The updraft develops rapidly from the initial state to a value of about 10 m s⁻¹ at 9 min and to the simulation maximum of 11.3 m s⁻¹ at 16 minutes. In this experiment cloud tops do not reach above 9 km, whereas the tops in the observed squall line generally reach about 11 km. This is believed to result from the two-dimensional restriction, which tends to lessen the strength of updrafts, as outlined by Hane (1973). In addition, based upon two-dimensional turbulence theory, Lilly (1979) asserts that in the case of turbulent flow confined to two spatial dimensions, kinetic energy propagates mostly to larger scales. Therefore, if one considers a convective line a turbulent element superimposed upon a mean flow, the result is that the line gives up energy to any existing mean shear.

The structure of the initial development at 19 minutes is shown in Fig. 16a. The updraft has a pronounced downshear tilt (with respect to the mean environmental

Fig. 14. Smoothed east–west wind components as a function of height from the Ft. Sill (FSI) sounding at 1346 CST and the KTVY tower site (TVY) sounding at 1558 CST. The TVY sounding was used to prescribe the horizontal wind west of the strong convection and the FSI sounding east of the squall line in the numerical simulation.
Fig. 16. Output vector winds and outline of rainwater field (solid lines) from the numerical simulation at (a) 19, (b) 33, (c) 40, (d) 50, (e) 61, (f) 77, and (g) 84 min. Wind vectors are ground relative and drawn so that 1 grid interval length corresponds to 7.5 m s\(^{-1}\). The model domain from 8–16 km altitude is omitted from the cross sections. In (d), the 0.5 g kg\(^{-1}\) contour (dashed) of cloudwater mixing ratio is included.
shear in the lowest 6 km) at low levels. Updraft tilt is here determined by the shape of the vertical motion field and is unaffected by storm-relative versus ground-relative flow considerations. A downdraft is forming near coordinates $x = 25, z = 3$, (25, 3) in the area of most intense rainfall. With time the updraft tilts down-shear more and more strongly with greater horizontal separation between the area of maximum updraft and the gust front. A breakdown of the more continuous updraft structure is indicated in Fig. 16b (at 33 minutes) along with strengthening of the downdraft near the ground (26, 1) and intensified westerly flow just behind the gust front (28, 0). The maximum updraft (23, 5) and area of associated precipitation (shown in outline) move westward relative to the gust front with time. A weak area of upward motion is also evident near 5 km ($x = 17$) and appears to have broken away from the major updraft area. Weak upward motion is also present at lower altitudes in this region and may result from the diverging downdraft air encountering ambient flow. The result of this weak upward motion is a westward expansion of the precipitation area with time.

At 40 minutes, maximum updrafts have weakened (see Fig. 16c and Fig. 15), and a second region of rain-induced downdraft has formed (21, 1.5) resulting from earlier rain production in updrafts located west of the initial downdraft area. The downdraft located at (26, 1) at 33 minutes is now at (27, 1) and has weakened, but horizontal motion is slowed in its vicinity, resulting in convergence and upward motion (25, 1) between this downdraft and enhanced outflow (24, 0) from the western updraft. The upward motion (25, 1) helps to reinforce the secondary surge in upward motion (25, 3), ultimately resulting in the higher vertical velocities
(Fig. 15) between 47 and 61 minutes. At 50 minutes (Fig. 16d) a relatively strong updraft (17, 5) has developed about 15 km west of the gust front, resulting from the westward movement of the updraft in Fig. 16c combined with reinforcement from low levels. The downdraft present at (21, 1.5) in Fig. 16c has remained nearly stationary (22, 1). Outflow from this downdraft toward the west has converged with ambient eastward flow to produce another low-level updraft at (19, 2). Intensification of the upper-level updraft is occurring due to merger with this low-level updraft.

At 61 minutes (Fig. 16e) the strong updraft (14, 4) is located 2–3 km farther west than at 50 minutes. Another upward motion center is located at (20.5, 3), and has developed due to convergence between westward outflow from an old downdraft (22.5, 0.5) and eastward outflow from a newer downdraft (17, 1). Additionally, a new updraft is located at (28, 3.5), having moved westward following initiation near the gust front. As is indicated in Fig. 15, at between 61 and 65 minutes the (14, 4) updraft weakens, drops its water load and produces a downdraft in a column near $x = 11$. In Fig. 16f (77 minutes) outflow from this downdraft and that from another farther east converge to force low-level upward motion (13, 1). As indicated in Fig. 15, maximum updrafts increase in the 76–79 minute time period and are sustained until about 87 minutes. The increase takes place with the development of the low-level (13, 1) updraft, which combines with the midlevel (15, 3) updraft, as is clearly indicated by the presence of a single area of upward motion above $x = 10, 11, 12$ in Fig. 16g (84 minutes). Following this time the intensity of convective circulations is weaker; wave-like perturbations originating near the gust front do not produce the deeper, more sustained convection. This decrease is perhaps related to the fact that stronger updrafts have progressed farther to the west (relative to the gust front) with time, so that downdrafts and subsequent eastward outflow must travel greater distances and undergo greater modification prior to reaching the easternmost gust front region.

In summary, in this experiment deep convection begins with development of a single updraft that tilts more strongly toward the west with time. Precipitation falls from the updraft, creating a downdraft that spreads both forward and rearward, producing enhanced convergence along the gust front and an area of weak upward motion and light precipitation to the rear. The total area of precipitation grows with time, spreading to the western boundary of the grid within one hour. Following the initial updraft development, the two-dimensional flow structure that slowly evolves is characterized by a major updraft area that slowly propagates westward with time relative to the gust front. This updraft area is periodically invigorated by the arrival of discrete perturbations in the vertical motion field, which are initiated at the gust front and travel westward above the cold outflow air. These perturbations interact with and may gain amplitude from low-level updrafts, which result from converging outflows of previous or existing downdrafts. The relative locations of the major updraft area, velocity perturbation, and a low-level updraft are shown schematically in Fig. 17. Within each of the westward-moving velocity perturbations maximum upward motion may be reached at varying distances from the gust front, and perturbations in the liquid water distribution, which lag the updraft growth by several minutes, may show a series of relative maxima at various altitudes and horizontal locations between the gust front and major updraft.

In Fig. 18 the position (in terms of distance westward from the gust front) of each of the westward-moving perturbations in the velocity field is plotted as a function of time. Six separate perturbations are identified.

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**Fig. 17.** Schematic representation of the relative positions of various features in the model squall line at a typical time beyond the simulation initial stage. Scallop line indicates cloud outline. Arrows give sense of airflow with respect to positive or negative horizontal or vertical flow, but no relation to speed is implied. Circled "P"'s mark locations of westward moving perturbations. Circled "G"'s indicate gyres or closed circulations in the low-level ground-relative flow. The circled "L" represents a low-level updraft, and circled "M" the region of strongest and sustained updrafts in middle levels.
all having been traced back to an origination point near the gust front. Based upon the figure, the westward speed of perturbations is about 7.5 m s⁻¹ and quite consistent from one perturbation to another. One may, of course, easily determine the location of all perturbations at a given time from this figure. For example, a vertical line drawn through 61 minutes in Fig. 18 indicates that perturbations should be found at 5, 12, and 18.5 km west of the gust front (Fig. 16e). Also indicated in Fig. 18 is the point along each curve where the maximum upward speed is attained. It can be seen that, in general, this point progresses farther west of the gust front as time passes, and also that the maxima are generally smaller at the later times. The latter relationship is certainly not perfect, since the intensity of maximum developments is also dependent upon the degree to which perturbations moving from east to west are maintained or augmented by phasing with the very low-level upward motion areas.

A valid question relates to the timing of perturbation initiation at the gust front. Upward motion is always present at the gust front as warm humid air overrides cooler air near the ground. Enhanced convergence cannot result from horizontal velocity gradients (eastto-west surges) in the air east of the gust front, since it is horizontally uniform. Eastward surges of cold air west of the gust front, however, could produce the periodic convergence enhancement. As downdrafts form following precipitation development both in the major updraft area and below westward-moving perturbations, flow is produced near the surface away from the point where the downdraft approaches the ground. Although most of the velocity change appears to be an increase of westward flow west of downdrafts (possibly due to downward transfer of horizontal momentum), some enhanced flow also occurs toward the east. These periodic eastward perturbations in the low-level flow are one possible cause for the upward-motion perturbations at the gust front. As air rising in these perturbations becomes saturated, latent heat is released, increasing the buoyancy and enhancing the upward motion beyond that produced kinematically. Compensating downward motion also occurs locally, suppressing for a short time any additional upward perturbations in the vicinity.

Another possible mechanism for initiation of gust front perturbations is through the production of Kelvin-Helmholtz instability at the top of the outflow. The properties of outflows have recently been investigated using high-scale numerical simulations by Droegemeier (1985). In one series of those experiments the environmental vertical wind shear was varied in the layer above the top of the outflow. A significant result was that when the gust front propagated upshear, the Kelvin-Helmholtz billows propagating backward from the outflow head were pronounced, and progress of the gust front was relatively slow due to downward mixing by billows of momentum directed oppositely to gust front motion. When the gust front propagated downshear, the Kelvin-Helmholtz instability was absent, the reason being that upward transfer of horizontal momentum by upward motion at the outflow head reduced the vertical shear near the top of the outflow below the value needed to support Kelvin–Helmholtz instability. In the downshear case the motion of the gust front was therefore faster due to the absence of billows; furthermore, the upward motion above and ahead of the gust front was much deeper due to the fact that environmental air lifted over the outflow head encountered opposing flow (due to the sense of the shear) and was forced to rise even higher.

In the squall line modeled here the gust front propagated upshear in relation to low-level environmental winds, as can be surmised by reference to Fig. 14. The behavior of the outflow seems to match well with Droegemeier's results in that perturbations were periodically produced and in that progress of the gust front toward the east was very slow. Firm conclusion that the Kelvin–Helmholtz mechanism is operative here is not possible for several reasons. First, it is unclear whether 400 m grid point separation can resolve the Kelvin–Helmholtz circulations. Second, Droegemeier's results are for experiments with shear imposed above the outflow layer, whereas the shear layer in the case modeled here extended to the ground. It would seem,
however, that the same principles apply in the extended shear layer case. Third, Droegemeier’s results were obtained in dry simulations, whereas condensation and evaporation contribute to the results reported here. This wet/dry factor would seem to have more relevance to the long-lived nature of the perturbations as they move away from the outflow boundary than to the actual formation of the perturbations. It is interesting to note that a simulation of the 22 May 1976 squall line using the same model (results not shown here) produced a gust front that propagated downslope with no perturbations present at the top of the outflow.

Another question relates to the formation of low-level updrafts west of the gust front; these features occur regularly with positive effects upon cloud growth. A logical supposition is that the air in low levels west of the gust front is not suitable for producing buoyant updrafts due to its coolness and perhaps dryness. However, it is apparent that evaporation of rain in this area maintains sufficiently high vapor content that condensational heating can aid in enhancing upward motion in these initially kinematically forced updrafts. This condensational heating is shown in Fig. 19 during the occurrence of one of the low-level updrafts. The coincidence of anomalous warming and the low-level updraft maximum appears at (19.5, 2), well displaced from the major updraft area and warming near (16.5, 4.5). The cooling near the surface resulting from cumulative evaporation of rain is also clearly shown. The gust front in this case is located about 8 km to the right of the figure. Cloud bases (not shown) associated with these low-level updrafts are quite low.

Due to the availability of wind fields based upon analysis of data from Doppler radars by K.R.H., a limited comparison of the model output with observed data can be made. In Fig. 20 the model output wind field, rain outline, and cloud cores at 50 minutes (simulation time) are shown, along with a vertical cross section through the observed squall line containing wind and reflectivity fields. The observed data is from the 1434 CST analysis time, and reveals an internal flow field that is moderate in intensity compared to that in cross sections at other locations along the line. Vertical motions within major cells in the line are strong, whereas weak upward motion occurs in the breaks between cells (not shown). Wind vectors are ground-relative in the model output, whereas about 5 m s$^{-1}$ has been subtracted from the Doppler-derived winds (must add a west wind of 5 m s$^{-1}$ to the Doppler winds to make them ground-relative). Thus, the gust front location in the observed field is very near the right border of the domain. In each case the major updraft is located about 15-20 km west of the gust front. As mentioned previously, observed vertical velocities are stronger than their modeled counterparts, and as a consequence, the model convection grows less in the vertical. It is possible that stronger westerly winds in the environment below 6 km would have allowed more rapid eastward propagation of the line and more erect (and possible more vigorous) updrafts. Both model and observed fields indicate precipitation to the west of the deep convection within a predominantly horizontal flow field. A similar shape in the precipitation distribution east of the deep convection is also evident, although features in the observed case are at greater altitudes.

In the easterly flow between gust front and main updraft areas there are perturbations in both the modeled and observed cases. Locations are at $x = -42$, and -48 in the observed case, and are clearly seen in the model output. These perturbations are also present on about the same scale in other cross sections. Less detail is present in the observations, which are available at 1 km intervals in the horizontal and vertical (400 m resolution in model output). The possibility remains that the model structure is incorrect and that these perturbations are “noise” in the observations. However, the presence of these features in both observed and modeled fields suggests that they are real. Clearly, more detailed observations are needed in future squall lines that are maintained under similar environmental conditions.

Two-dimensional modeling is limited in its ability to simulate all aspects of a squall line that contains distinctly cellular structure. The effect of two-dimensionality on the vigor of deep convective updrafts has been mentioned. However, it is also clear that there are potentially other areas where 2-D modeling has utility, as is suggested by the evolving structure found in this case in low levels west of the gust front. Preliminary efforts in 3-D simulation of squall lines have been carried out by Klemp and Weisman (1983) and Wilhelmson and Klemp (1983a,b). One experiment (Wilhelmson and Klemp, 1983b) involved utilization of a sounding from 19 May 1977, the same case as described in this paper. In that experiment the observed motion of convective cells within the line was reproduced quite

Fig. 19. Temperature anomaly (K) field at 50 minutes over a small portion of the model domain surrounding the low level updraft shown in Fig. 16d. Stronger areas of upward motion are indicated by solid arrows.
Fig. 20. Model output (upper) and observed fields (lower) in the 19 May 1977 squall line. Model output is taken from Fig. 16d, with wind vectors plotted at 400 m intervals. Observed wind vectors are at 1 km intervals and have had a small component of cell motion (~5 m s⁻¹) in this plane removed. Vectors are drawn such that one grid interval (1 km) corresponds to 15 m s⁻¹. Reflectivity is contoured at 5 dBZ intervals (solid). Width and height of domains differ between the two panels (adapted from Kessinger et al., 1982).
well, and maximum updraft speeds reached slightly more than 20 m s\(^{-1}\). Addressing the question of why in some cases lines of isolated storms form rather than solid squall lines, these preliminary 3-D efforts indicate the importance of both vertical wind shear and the geometry of the initiating mechanism.

4. Summary and comments

Three mechanisms that contribute to the maintenance of strong convection along the leading edge of the 19 May 1977 squall line have been identified based upon analysis of surveillance and Doppler radar data, surface network data, and results of numerical simulation.

i) At preferred locations short lines of individual small convective cells extend eastward from the squall line itself. The cells sequentially merge with the line, due to their differing motion, and maintain a high level of convective intensity at that point along the line.

ii) At all locations along the line, perturbations are periodically initiated at the gust front, travel westward relative to the gust front above the cold outflow, and sustain the main updraft area within the line.

iii) At all locations along the line, low-level updrafts are initiated by diverging outflow from rain-induced downdrafts in the area between the gust front and the major updraft area. These updrafts occasionally merge with and strengthen the perturbations in (ii).

The small cells mentioned in (i) appear to form in a pressure trough extending eastward from a mesoscale area of low pressure centered on the southeast edge of a cell cluster within the squall line. The cells move northward more rapidly than the pressure trough and its associated convergence (which are more tied to the cluster motion within the line). As a result, the cells merge with the line north of the low center and result in a propagative component of cluster motion toward the east or southeast within the line. The cause of the mesoscale low-pressure area is uncertain, but once formed it is apparently a stable feature, tied to the strong convection, and at the same time contributing to the initiation of convective activity in its vicinity. Shifting of the location of this mesoscale structure requires a significant event, such as the explosive development of convection elsewhere along the line.

At other locations along the line numerical simulation and wind fields derived from Doppler radar observations suggest the existence of the small-amplitude perturbations mentioned in (ii). This mechanism is really only a more detailed version of the classical concept of squall line regeneration by quasi-continuous development along the gust front. The simulation and observations suggest a more discrete regenerative mechanism, amplified in some cases by low-level updrafts that are mentioned in (iii). These low-level updrafts are initially forced by converging near-surface air currents, then augmented by buoyant forces that result from the release of latent heat.

It is clear that additional observation of squall lines of this type is needed to further investigate each of these mechanisms. The existence of abundant moisture and weak vertical wind shear in a deep layer ahead of the line appear to be essential environmental characteristics that distinguish a system with the described structure from squall lines possessing other structures. Specifically, analyses of the wind fields at time intervals of about five minutes would aid immensely in verifying some of these features. Additionally, more comprehensive surface and upper air observations of wind, pressure, temperature, and moisture fields ahead of the line would aid in the analysis of the mesoscale structure found in this case. Observations in the formative and dissipative stages of such systems are also needed.

The mesoscale area of precipitation trailing the active convection in this case has not been investigated here, but was described by KRH. An observation is that the particular wind and moisture profiles that appear to be tied to the maintaining mechanisms in this case are apparently not essential for the development of the trailing mesoscale precipitation in general. That such mesoscale areas can develop under other environmental conditions is clearly demonstrated by other case studies (e.g., Zipser and Matejka, 1982; Houze and Smull, 1982). The mesoscale and convective areas are undoubtedly interdependent features of such squall lines. The mesoscale area develops as a result of the convection, and the mesoscale area modifies the western "boundary condition" for the squall line's deep convection, changing profiles of wind, temperature, moisture, and pressure from values present when the line was developing.

Additional development of numerical modeling techniques is needed to more accurately simulate various aspects of squall lines. Larger domains must be utilized in order to include the meso-β scale precipitation area trailing the strong convection in some cases. Three-dimensional simulations are clearly required to investigate those features that do not occur uniformly along the line. Additional work is needed in comparing results of 2-D and 3-D simulations in order to better define areas where 2-D results are useful.

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