Initiation of Convective Storms at Radar-Observed Boundary-Layer Convergence Lines

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ABSTRACT

The origin of 653 convective storms occurring over a 5000 km² area immediately east of the Colorado Rocky Mountains from 15 May to 15 August 1984 was examined. Seventy-nine percent of the 418 storms that initiated within the study area occurred in close proximity to radar-observed boundary-layer convergence lines. This percentage increased to 95% when only the more intense storms (≥60 dBZ) were considered. Colliding convergence lines initiated new storms or intensified existing storms in 71% of the cases. A new storm took a median time of 24 min to grow to 30 dBZ following line collision.

The convergence lines ranged in length between ten and several hundred kilometers. Both radar and mesonet stations indicated that the primary convergence was concentrated in a zone 0.5 to 5 km in width. These lines were characterized on Doppler radar as thin lines of enhanced reflectivity between 0 and 20 dBZ and as a line of strong radial or azimuthal gradient in Doppler velocity. These lines were observed even in clear air in the absence of any clouds. The origin of many of the convergence lines was unknown and requires further study.

The most common identified origin was from convective storm outflows. Other origins were believed to be topography and differential heating.

This study, which utilized radar data, supports the findings of Purdom (1982), who utilized satellite data, which indicated that mesoscale boundary-layer convergence lines play a major role in determining where and when storms will form. These results suggest that what often appears as random thunderstorm formation (air mass thunderstorms) is usually deterministic.

Major advances now appear possible in the 0–2 h time-specific forecasts of thunderstorms. Realization of this potential will require the integration of Doppler radar to detect and monitor convergence lines, high resolution satellite data to monitor cloud growth, and surface and sounding data to estimate atmospheric susceptibility to deep convection.

1. Introduction

The initiation of convective storms by organized lines of convergence in the boundary layer has been recognized for some time. This initiation mechanism is generally not associated with what meteorologists frequently refer to as "air-mass thunderstorms" that appear to occur randomly on a warm summer afternoon in the absence of large-scale forcing. This study utilizes sensitive Doppler radars, capable of observing boundary layer winds, to show that what appears as random thunderstorm formation is actually associated with lines of convergence. The role of radar-observed convergence lines in initiating deep convection has been documented for selected cases from eastern Colorado in Wilson and Carbone (1984), Szoke et al. (1985), and Schreiber (1986). In this paper, examples and statistics are presented that are based on the 1984 convective season near Denver, Colorado.

Previous studies of boundary layer convergence initiating deep convection were based primarily on dense networks of anemometers or satellite cloud images.

Byers and Braham (1949), using a network of anemometers during the Thunderstorm Project, observed surface convergence 30 min prior to radar echo appearance. They also noted the importance of convergence associated with the land–sea breeze in the initiation of deep convection along the Florida Peninsula. Many others, including Gentry and Moore (1954), Ulanski and Garstang (1978), Burpee (1979), Cooper et al. (1982), and Watson and Blanchard (1984), have since studied the role of sea breeze-induced convergence in the initiation of Florida showers.

Purdom (1982) used satellite data to show that cloud arc lines were often generated by outflows from convective storms and other convergent wind phenomena. The intersection of these arc lines often triggered intense convection. Purdom and Marcus (1982) found that 73% of the afternoon thunderstorms in the southeastern United States developed as the result of such interactions.

Weaver (1979) and Wade and Foote (1982) showed that the organization and movement of severe storms was often influenced by strong low-level convergent features rather than by upper level winds. Koscielny et al. (1982) showed that single Doppler radar could be used to map mesoscale convergence zones. In the
case they studied, convergence zones preceded cloud and storm development by 2–3 h.

The research reported in this paper is based on data collected over the plains of Colorado immediately east of the Rocky Mountains (Fig. 1). The evolution of thunderstorms in this region has been the basis of several studies. These include radar studies by Karr and Wooten (1976) and satellite studies by Klitg et al. (1985). These studies have shown a very strong diurnal pattern of convection. Storms typically form first over the Rocky Mountains during late morning and along two ridge lines that extend eastward into the plains (the Palmer Lake Divide south of Denver and Cheyenne Ridge to the north) in early afternoon. During the afternoon, the region of maximum thunderstorm frequency progresses on to the plains and moves eastward. Toth and Johnson (1985) showed from mesonet winds that the initiation of thunderstorms over the Rocky Mountains, Palmer Lake Divide, and Cheyenne Ridge can be related to upslope confluent flow during the morning and early afternoon. They suggest that one cause of the transition to downslope flow during the afternoon appears to be associated with the propagation of thunderstorm activity to the plains.

This study will focus on thunderstorms occurring over the plains in a small area just east of the Rocky Mountains between the Palmer Lake Divide and Cheyenne Ridge. There is a very high frequency of

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**Fig. 1.** Map showing the study area topography and location of mesonet stations and Doppler radars. Contours are elevations in meters. The very rugged topography of the Rocky Mountains is not contoured. The solid arc encompasses the study area from 18 May–30 June and the dashed arc from 1 July–15 August.
storm initiation over elevated terrain to the north, west, and south of the study area. Surprisingly, the majority of storms in the study area were initiated locally—not in the surrounding regions of elevated terrain.

Research field experiments with the National Center for Atmospheric Research (NCAR) Doppler radars have shown that winds in the summertime boundary layer can be routinely monitored. During these experiments, it became apparent from viewing the radar reflectivity and Doppler velocity displays that convergence lines ranging from tens to hundreds of kilometers in length were frequently visible. When thunderstorms developed, storm locations were spatially correlated with these lines. Frequently, these convergence lines occurred in the optically clear air prior to cloud development. As will be discussed in section 2, insects are likely microwave-scattering mechanism that makes it possible to observe these winds. The Program for Regional Observing and Forecasting Services (PROFS) experimental short-period forecasting experiment in northeast Colorado during the summer of 1985 frequently used these radar-observed convergence lines to successfully anticipate where thunderstorms would first develop. Also, during the Classify, Locate and Avoid Wind Shear (CLAWS) experiment at Stapleton Airport during 1984, this same technique was used to forecast development of thunderstorms near the airport.

Using analysis of Doppler radar and mesonet data obtained during the 1984 convective season, we examine in section 4 the origin and characteristics of the convergence lines, as well as compare radar and mesonet observations of these lines. In section 5 we present statistics on the initiation of storms by convergence lines. The intent of this paper is to document, for one convective season, the observations that spatially and temporally link radar-observed boundary layer convergence lines to the initiation of deep convection. The actual meteorological processes involved in triggering the convection are not considered at this time.

2. Data

Data available to this study include those from the NCAR CP-4 and CP-2 Doppler radars, the NCAR portable automated mesonet (PAM), and the PROFS mesonet of surface stations. The location of the radars and surface stations are shown in Fig. 1.

The PROFS mesonetwork was available for the entire study period of 18 May to 15 August 1984. The CP-4 radar, 16 PAM stations and a “chase car” were available during May and June, while CP-2 was available during July and August.

The chase car was used to obtain wind, temperature, and moisture measurements across boundaries and to observe and photograph cloud development. This vehicle was directed by radio from the radar. An important aspect of this activity was the routine confirmation by the chase car meteorologists of the existence of a convergence line at locations indicated by radar and the frequent visual observations of cumulus clouds along the line. Satellite imagery would, of course, be useful for monitoring cloud development. For this purpose high resolution (1 km) rapid scan imagery is desirable. Unfortunately, sufficient amounts of these data were not readily obtainable and are not included in this study. Future studies will stress obtaining cloud information from satellite imagery and ground-based time-lapse photography.

a. Doppler radars

The CP-4 and CP-2 Doppler radars were used from 18 May to 30 June and from 1 July to 15 August, respectively. The characteristics of each radar are given in Table 1. The CP-4 radar was operated in a variety of scanning modes because it was being shared by four different research projects. The scanning generally included a full 360° azimuth scan at several different elevation angles once every 5–10 min.

The CP-2 radar is a multiple wavelength polarization diverse radar (Carbone et al., 1982). The two wavelengths have matched beamwidths of 1°. The 10-cm wavelength is Dopplerized, while the 3-cm wavelength is not. The horizontally polarized 3- and 10-cm radar reflectivity data and 10-cm Doppler velocity data were utilized from this radar. The radar was continually scanned 360° in azimuth at six elevation angles with a repeat cycle of 2.5 min.

An important aspect of this study is the ability of both CP-4 and CP-2 radars to observe the winds in the optically clear planetary boundary layer. The convergence wind lines are usually observable on the reflectivity display as thin lines of enhanced reflectivity between 0 and 20 dBZ. On the Doppler velocity display, they appear as a line of strong radial or azimuthal gradient in velocity. There are occasions when these lines are only visible with either reflectivity or velocity. It is generally easier to detect these lines on the reflectivity display because velocity detection is potentially confounded by vertical shear and orientation considerations. When using the CP-2 radar, the 3-cm wavelength was used to locate the reflectivity thin lines because there was less ground clutter contamination (see Wilson et al., 1984). Strictly speaking, a reflectivity thin line or a Doppler velocity-convergence wind signature is not proof of the existence of a convergence line. A single Doppler radar can only measure the change in wind component parallel to the beam. Changes in the wind component perpendicular to the beam are not detected. Thus a convergent line will not be apparent in the Doppler velocity field when the convergence is oriented along the radar viewing angle, i.e., along a

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1 Equivalent radar reflectivity factor (dBZ) is used when discussing clear air echoes, and dBZ is used when discussing precipitation echoes.
radial. On rare occasions, a deformation field might appear as a convergence line. Provided that the clear air return signal is $>$0 dBZ$_e$, it is rare that a convergence line is not detected or is falsely identified by viewing time-lapse displays of both Doppler velocity and reflectivity. This is discussed further in section 4c.

The scattering mechanism that makes it possible to observe clear air winds is uncertain and has been the subject of numerous investigations (Hardy et al., 1966; Battan, 1973; Doviak and Zrnić, 1984; Vaughn, 1985; Mueller and Larkin, 1985). Arguments have been presented for scattering from airborne particulates, as well as from areas with sharp gradients in the radio refractive index, primarily due to moisture inhomogeneities. We believe particulate scattering is the dominant mechanism for boundary layer clear air observation in Colorado. It is also believed the particulates are primarily insects and seeds. The following field observations support this claim.

The first is based on comparisons of $Z_e$ clear air values from the 10-cm and 3-cm wavelengths of the matched beam, dual wavelength, CP-2 radar. If the scatterers were Rayleigh targets, such as small insects, the $Z_e$ values would be equal. Backscattering from refractive index variations would cause $Z_e$ values from the 10-cm wavelength to be $\sim$ 19 dB larger than the 3-cm values. Comparisons show that the 10-cm $Z_e$ values are typically $\sim$4 dB larger than the 3-cm values, thus the backscattering is more likely particulates. The larger $Z_e$ value for the 10-cm wavelength suggests some of the targets are sufficiently large to cause non-Rayleigh scattering at the 3-cm wavelength.

The second supporting evidence for particulate scattering is from the CP-2 S-band dual polarization return from clear air. The horizontal return is 2–5 dB larger than the vertical. This is in agreement with the observations of Mueller and Larkin (1985) with an S-band radar in Illinois. They concluded that insects in flight caused the larger horizontal signal.

The third observation supporting particulates such as insects, seeds, or possibly wheat chaff is the seasonal nature of the clear air return. We have observed that the clear air return typically increases in strength throughout the convective season, being relatively weak in May and strong in August. High temperature, moisture, and stability conditions in May and June do not produce similar clear air signal strengths for the same conditions in August. This seasonal increase in signal is consistent with the observed increase in the concentration of insects, seeds, and wheat chaff as the summer progresses.

Vaughn (1985) has reviewed a large body of literature discussing insects and birds as radar targets. Particularly applicable to the reflectivity thin line is the observation by entomologists of line echoes associated with high insect concentrations. In one instance, Riley and Reynolds (1983) described a line echo associated with a dense insect concentration that arrived with a sudden and sustained increase in surface wind speeds.

b. Mesonetworks

The spacing of the PROFS stations is variable but averages $\sim$25 km. The spacing of the PAM stations is $\sim$8 km. The wind and temperature measurements are 1 and 5 min averages for the PAM and PROFS stations, respectively.

The spacing of the PROFS stations was generally too large to detect many of the convergence lines. The closer spacing of PAM stations made it possible to detect most convergence lines in that network.

3. Study design

The primary objective is to determine the importance of radar-observed boundary layer convergence lines in initiating convective storms over the Colorado plains. For the purposes of this study, a storm is defined as a radar reflectivity precipitation echo $\geq$ 30 dBZ $\sim$ 1 km AGL. A storm was declared when a new precipitation echo, not attached to an existing storm, reached 30 dBZ. New cells $\geq$ 30 dBZ that developed within or attached to an existing storm were not classified as new storms. A storm typically initiated as a single cell and evolved into a multicellular storm. Again, this was classified as one storm.

The time and place of initiation and dissipation of all storm echoes $\geq$ 30 dBZ that occurred in the study area as seen by radar were recorded, as were the maximum storm intensity and storm mergers.

All boundary layer convergence lines, which will henceforth be called boundaries, were also identified and tracked. The identification of a boundary was based solely on a radar signature of a thin line of enhanced reflectivity and/or a line of apparent convergent flow in Doppler velocity. The $\sim$1–3 km wide line is required to be $>$10 km long and present for a minimum of 15 min. While most boundaries have strong temperature and/or humidity gradients associated with them, this is not a requirement. It is believed that all boundaries were associated with convergent flow; however, it is not possible to prove this in every case. Examples of boundaries and storms are given in the next section.
An exception to the above boundary definition is horizontal roll convection (Kuettner, 1971; Lemone, 1973; Brown, 1980; Kropfli and Kohn, 1978), which is a very common planetary boundary layer feature, visible by radar during warm summer afternoons. Rolls appear as long, thin lines of enhanced reflectivity (Fig. 2) that remain almost stationary for a few hours. It is usually easy to distinguish horizontal rolls from boundaries, because the rolls occur in large numbers spaced several kilometers apart and are aligned approximately parallel to the mean boundary layer wind. For example, the lines in Fig. 2 have a 160° orientation. Inspection of the Doppler winds indicates that the winds in the lowest 1 km are 160° at 8 m s⁻¹, veering slowing to ~175° at 8 m s⁻¹ at the boundary layer top (2.5 km). Lemone (1973) has shown that the winds at the top of the boundary layer are frequently ~15° to the left of the line orientation. Frequently, a line of small cumulus clouds is associated with these rolls, but seldom were they observed in Colorado to initiate storms by themselves.

The region of the study was limited to within 40 km of CP-4 during May and June and within 50 km of CP-2 during July and August (Fig. 1). These relatively short ranges were selected to increase the likelihood of detecting convergence lines. Detection decreases with range because of the increasing height of the beam above ground and the decrease in clear air signal. An elevation angle of 0.7° was typically used to locate boundaries. However, at close ranges higher angles were used to reduce ground clutter contamination. At ranges of 40 and 50 km, a 0.7° elevation angle gives a beam height of 580 and 760 m AGL, respectively. This does not account for terrain changes. The depth of the convergence lines is usually about 1 km.

Radar clear-air detection capabilities vary considerably. Detection is poorest during cool periods. In May and June, clear air signals are generally observed out to at least 40 km, increasing to >80 km in July and August. The study area was limited to the west side by the Rocky Mountains. As will be discussed later, many storms form in the mountains; however, it is not the intent of this study to determine how these storms initiate. Barker and Banta (1985) studied the initiation of storms in the Colorado mountains. They linked preferred regions of thunderstorm development to localized terrain-forced convergence. As discussed in section 1, almost all deep convection in the study area occurred during the afternoon and early evening hours. The focus of the experiment was typically between 1800 and 0200 GMT.² It was not our intention to observe every storm and boundary that occurred during this three-month period.

Figure 3 shows the number of storms and boundaries that were recorded daily. Operations were not conducted every day for the entire convective time period of 1800–0200. Radar difficulties reduced the length of observations on three days. Operations were not conducted on 11 days because convection was not expected. It is possible that storms and boundaries occurred on some of these days. On at least two days, operations stopped early and subsequent weather occurred.³ Thus, the study should not be considered a complete history from 18 May to 15 August. A total of 653 boundaries were documented. It is estimated that this is at least 80% of the actual number that occurred.

Identification and movement of boundaries and storms were greatly facilitated by using 16-mm color motion pictures of the reflectivity and velocity images at radar elevation angles of about 0.7°, 1.6°, 2.5°, and 5.0°. The time between images was typically 2.5 to 5 min. The films were first examined to locate and track each storm at an altitude of 1 km. Second, a separate search was made to identify and track each boundary. Third, the film was again examined to determine if an apparent causal relationship existed between the initiation of a storm and the presence of a boundary. Admittedly, this third procedure was subjective, and individual cases could be disputed. However, when viewing the data in a time lapse manner, the high correlation between boundary location and storm initiation became obvious. Data presented in the following sections will substantiate this correlation.

² All times in this paper are in GMT. Local standard time (MST) is GMT - 7 h.
³ In Fig. 3, there are several days for which no data are indicated for boundaries, yet the same day indicates no storms instead of no data. This occurs because observers either watching the radar or in the chase vehicle observed no storms for the day, but it was impossible to check if boundaries occurred because radar data were not recorded for the day.
4. Boundary types and origins

a. Classification

The origin and characteristics of convergence lines are examined to the extent that synoptic scale maps, single Doppler radar, and mesonet stations can lend physical insight. In previous studies, a variety of boundary layer convergence lines have been observed, and their origins have been attributed to synoptic scale fronts, sea breeze fronts, and gust fronts.

Szoke et al. (1984) describe a quasi-stationary, north-south boundary that often develops in the general vicinity of Denver under synoptic scale southeast flow. They propose that this results from blockage of the flow by the east–west-oriented Palmer Lake Divide south of the area. They reported that this convergence line is related to numerous severe storm events.

Occasionally, a convergence line moving east from the mountains is observed in the study area. This is associated with the descent of relatively cool, gusty winds from the mountains. These boundaries, which we label “mountain outflows,” have been observed to occur (Schlatter et al., 1985) when mountain showers and cloudiness cause the potential temperature at mountain stations to drop more than 5°C below the potential temperature on the plains to the east.

Segal et al. (1984) discussed the role of horizontal temperature gradients in forming boundary layer convergence. Proposed causes of temperature gradients were spatial differences in cloud cover, soil moisture, and vegetation. Purdom (1973) and Carpenter (1982) showed examples where circumstances of persistent cloud cover next to clear skies were associated with convergence lines that initiated convective storms.

Uccellini (1975) was able to detect the presence of gravity waves in the analysis of pressure perturbation fields in the Midwest. These waves appeared to initiate convective storms or reintensify existing storms. These waves had wavelengths of 400 to 500 km and phase speeds of 35 to 40 m s⁻¹. A theoretical analysis by Lindzen and Tung (1976) indicates that gravity waves are likely to exist in atmospheric conditions similar to those reported by Uccellini (1975) and that observed scales of convective lines are consistent with gravity wave scales. It is quite possible that some of the moving convergence lines with no known origin observed in this study were gravity waves. Whether these waves might have produced a wind discontinuity in the planetary boundary layer as observed by radar is unknown.

Boundaries were classified as gust front, mountain outflow, synoptic front, Denver convergence line, unknown stationary, and unknown moving. A boundary

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FIG. 3. Daily frequency of storms and boundaries observed within the study area. The asterisks indicate data not available.
was classified as a synoptic front if the synoptic-scale weather maps showed a front in the vicinity of the study area with the same general orientation and movement of the radar-observed boundary. The radar often showed the front as multiple parallel convergence lines. Each of these multiple lines was then classified as a synoptic front. For this study, a boundary was classified as a Denver convergence line when the surface synoptic weather map indicated large-scale southeast flow and when the line was quasi-stationary and oriented basically in a north–south direction. Generally, the unknown moving boundaries originated outside of the study area at too great a distance to observe their origin with the radar. Some of these cases may have been gust fronts. It is also possible that some were gravity waves, as discussed before.

The unknown-stationary boundaries are the least understood. They were typically 20–50 km long and developed over a period of about 1 h. They showed no preferred locations and they were oriented in all directions. However, half had a northeast–southwest orientation. Some of these may be the result of differential heating as a result of persistent, localized anvil cloud cover from downstream convective activity over the mountains. Other stationary boundaries may be orographically forced similar to the Denver convergence line.

A total of 166 boundaries were observed during the 44 days, where at least one storm was observed in the study area. Insufficient radar data were available on nonstorm days to determine boundary statistics; according to the limited data available, however, there were at least 15 boundaries on nonstorm days. The actual number is undoubtedly larger.

Table 2 shows the number of boundaries for each of the above six classifications. Note that 46% of the boundaries are of unknown origin. The percentage of boundaries that initiated convection for each type is also given. When interpreting these data, it should be remembered that boundary life times vary greatly. For example, the Denver convergence line tends to be present for many hours, whereas moving gust fronts generally propagate through the study area in less than 2 h.

Sixty-three percent of the boundaries initiated storms within the study area. Some of those that did not initiate storms within the area did so elsewhere. Often, even if they did not initiate storms, they were responsible for consolidating and intensifying storms.

Figure 4 shows that the boundaries approach from all directions, the largest numbers coming from the west and northwest, and the smallest numbers coming from the north and northeast. While the mountain outflows come from westerly directions, other moving boundaries approach from all directions. Also shown in Fig. 4 are the number of boundaries that initiate storms. Those from the southeast were most likely to have storms associated with them, and those from the north and northeast were least likely.

b. Boundary examples

Examples of radar-observed boundaries and associated convection that occurred in the dense network of PAM stations are shown. Figure 5 contains examples of a mountain outflow and complex interactions of gust fronts and storms from 15 June; Fig. 8 shows a gust front and unknown stationary boundary from 20 June, and Fig. 9 shows the Denver convergence line and an unknown stationary boundary from 14 June. In all cases, the drawn boundary was observed by radar. It is then possible to compare radar and mesonet observations.

### Table 2. Classification of boundary types for the 44 days when storms occurred in the study area.

<table>
<thead>
<tr>
<th>Boundary type</th>
<th>Number (percent)</th>
<th>Percent initiate storms</th>
</tr>
</thead>
<tbody>
<tr>
<td>Gust front</td>
<td>54 (33)</td>
<td>61</td>
</tr>
<tr>
<td>Mountain outflow</td>
<td>14 (8)</td>
<td>71</td>
</tr>
<tr>
<td>Synoptic front</td>
<td>12 (7)</td>
<td>58</td>
</tr>
<tr>
<td>Denver convergence line</td>
<td>11 (7)</td>
<td>82</td>
</tr>
<tr>
<td>Unknown (stationary)</td>
<td>31 (19)</td>
<td>48</td>
</tr>
<tr>
<td>Unknown (moving)</td>
<td>44 (27)</td>
<td>68</td>
</tr>
<tr>
<td>Total</td>
<td>166 (100)</td>
<td>63</td>
</tr>
</tbody>
</table>
Fig. 5. Evolution of storms and boundaries at 30 min intervals as observed by radar and mesonet stations for 15 June. Radar-observed boundaries are indicated by solid numbered lines. Boundary 1 is of the “unknown moving” type, boundary 2 is a “mountain outflow,” and all others are gust fronts. Full wind barbs represent 5 m s$^{-1}$ and half barbs 2.5 m s$^{-1}$. Dry-bulb and dewpoint temperatures are given in °C. (a) 2030, (b) 2100, (c) 2130, (d) 2200, (e) 2230, and (f) 2300 GMT. Time series data from stations 7 and 15, indicated by an asterisk in (e) and (f), respectively, are plotted in Fig. 7.
1) 15 June

Boundary 1 in Fig. 5a was visible only as a weak gradient both radially and azimuthally in Doppler velocity. It is of unknown origin and is moving slowly northeastward. Apparently, it initiated storms B and C.

Boundary 2 originated in the mountains as cold air produced by an area of thunderstorms. Storm A (Fig. 5a) is the dissipating remnants of this activity. This boundary was visible on radar as a reflectivity thin line of 0–5 dBZ and a radial velocity gradient where the approaching flow increased from 6 to 12 m s\(^{-1}\) in a distance of 1–2 km. It was not observed farther east because of range ambiguity obscuration in the radar data (Doviak and Zrnić, 1984). Mesonet data in Fig. 5 shows that winds are typically northwest on either side of boundary 2 and that speeds increase from 3–6 m s\(^{-1}\) south of the boundary to 5–10 m s\(^{-1}\) north of the boundary. These values are similar to the radar Doppler velocities. Ambient temperature and dewpoint are lower north of the boundary.

Figure 5b shows that storm B generated a gust front (boundary 3) that moved northwest toward boundary 2. In Fig. 5c, storm B split into B1 and B2 as outflow from B1 spread in all directions. Boundary 3 apparently initiated storm D at 2145. Boundaries 2 and 3 collided at 2150 and then moved northwest as one. Storm B1, which is within a few kilometers of this collision, undergoes rapid growth and intensification as these boundaries approach and collide. After collision, the boundary was well defined on radar as a thin line (0–5 dBZ) and strong radial gradient from 4 m s\(^{-1}\) approaching ~8 m s\(^{-1}\) receding in a distance of 1–2 km. Figure 6 shows the Doppler velocity and reflectivity thin line at 2215. The line of sharp change in radial velocity from receding to approaching (black to gray) is coincident with the reflectivity thin line. Storm B2 dies very rapidly after 2230 as cold air (boundaries 3 and 4) spread out in all directions, apparently cutting off inflow to the storm. At the same time, outflow from storm D (boundary 5) and boundary 3 initiated new activity (storm E) on the north side of storm D. Originally, storm E was not connected to storm D. Storm F moved off the mountains and dissipated slowly; however, when boundary 3 passed under at ~2245, it intensified into storms F1 and F2. Boundary 6 shows on radar as a new intense downburst type outflow of 24 m s\(^{-1}\) that produced a peak gust of 24 m s\(^{-1}\) at station 15, located just north of boundary 6 in Fig. 5f.

Figure 7 shows a 3-h time series of wind speed, wind direction, temperature, and dewpoint temperature as the various boundaries cross two of the stations. The station locations are indicated in Fig. 5. The boundary locations marked on Fig. 7 are based on radar data by converting the spatial width of the reflectivity thin line to a time width based on the boundaries’ rates of movement. Small errors in time (<2 min) and position are possible in Fig. 7.

The velocity changes observed at each station in Fig. 7 are similar to those indicated by Doppler velocity. For example, as stated earlier, the Doppler velocities near the time boundary 2 passed station 7, were approaching (suggesting northwest flow), and increased from ~6 to 12 m s\(^{-1}\) from south to north across the

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**Fig. 6.** (a) Equivalent radar reflectivity factor display of boundary 3 (Fig. 5) at 2215 GMT 15 June. The boundary is visible as a north–south thin line. Reflectivities are given in dBZ by the scale at the bottom of the display. Values greater than 30 dBZ are black. The black echo east of the radar is storm B2 in Fig. 5. Range marks are at 10-km intervals. The data are for a 3.5° elevation angle and are from the CP-4 radar.

(b) Doppler velocity display corresponding to (a). The scale at the bottom indicates the velocities in m s\(^{-1}\). Gray shades are approaching values and black receding. Coincident with the thin line in (a) the velocities sharply change from receding to approaching.
boundary. This corresponds to observed changes in maximum wind at station 7 from \(\sim 310^\circ\) at 6 m s\(^{-1}\) to 330\(^\circ\) at 12 m s\(^{-1}\). For boundary 3 the Doppler velocity changed from northwesterly at \(\sim 4\) m s\(^{-1}\) to southeasterly at \(\sim 8\) m s\(^{-1}\). This corresponds very closely to what was observed at stations 7 and 15 in Fig. 7. Both radar and mesonet stations indicate that the wind changes occurred within a distance of \(\sim 1-3\) km.

2) 20 June

Figure 8 shows the evolution and interaction of a gust front and a stationary boundary of unknown or-
origin. Boundary 1 is a gust front originating from storm A. Typically, the gust front appeared as a 0 dBZ\textsubscript{e} 1–2 km wide line of enhanced reflectivity. Coincident with this line is another line of relatively strong radial gradient in Doppler velocity. The gradient was 4–9 m s\textsuperscript{−1} within a radial distance of 1–3 km.

Boundary 2 remained stationary and first appeared on radar at 2100 as a diffuse line of radial velocity gradient of \(\sim 4\) m s\textsuperscript{−1} in 2 km. Initially, this boundary was 200–500 m deep. It slowly strengthened by 2150 to 6–8 m s\textsuperscript{−1} over a horizontal distance of 1–2 km, and the depth increased to 500–700 m. This boundary did not have a reflectivity thin line.

In the 10–15 min period after the radar-located gust...
front (labeled 1) passed a station, the winds shifted to the north-northwest, the temperature decreased 3°–5°C, and the dewpoint temperature increased 1°–2°C.

The stationary boundary is less obvious in the mesonet. Dewpoints appear to be slightly less south of the boundary, and winds tend to shift from northeast to southeast when moving from north to south across the boundary. These winds are consistent with the Doppler velocities. No obvious reason for the formation of this boundary is known. This boundary is quite typical of the boundaries classified as unknown stationary.

The stationary boundary appeared to play a significant role in the development of storm E. By 2150, towering cumulus clouds were noted along the stationary boundary by observers at the radar and in the chase vehicle. At the same time, radar showed a line of −5 to +5 dBZ echo paralleling the surface convergence line. These echoes were at a height between 2.5 and 5.0 km, displaced ~4 km northwest of the convergence line. By 2208, cell E began to grow rapidly aloft as the gust front approached to within 1 km of the line of towering cumulus clouds (not yet very evident in Fig. 8d, which shows reflectivities near 1 km). After 2215, the two boundaries merged into one, as cell E rapidly developed into a 65 dBZ storm.

Storm D is more typical of the storms produced by
a moving boundary (Fig. 12); that is, it reached 30 dBZ
at a height of 1 km about 35 min after the boundary
passed the location.

3) 14 JUNE

An example of the Denver convergence line is shown
in Fig. 9 (boundary 1). The meteonet data suggest that
it first formed about 2 h prior to Fig. 9a. Radar data
first available at ~1900 indicates that the southern
part of the boundary is moving slowly northward, while
the remainder was virtually stationary. The northern
half of the boundary appears primarily as a thin line
of enhanced reflectivity (0 dBZ), and the southern half
as a Doppler velocity discontinuity. Traverses of the
boundary by the chase vehicle showed that the wind
shifted from north-northwest to south-southeast when
it crossed the boundary from west to east. Also, the
temperature increased ~2°C and the dewpoint dropped ~3°C. Skies were clear until 2040, when the
chase vehicle reported very small cumulus forming
along the boundary. By 2200 (Fig. 9b) the southern
half of the boundary continued to move slowly north-
ward. Two other north–south boundaries (boundaries
2 and 4) were observed by radar. Boundary 2, seen in
Fig. 9b, dissipated by 2300 and was replaced by
boundary 4, which formed 10 km farther west and
moved very slowly eastward. These boundaries have
the appearance of Denver convergence lines. Boundary
4 is visible as a thin line of reflectivity (0 dBZ) and
Doppler velocity discontinuity (4 m s⁻¹ in 1 km). Radar
indicated an increase in the easterly wind component
on the east side of this boundary.

At about 2230 the PAM stations indicated a con-
vergence zone forming south of the four northernmost
stations. Radar showed correspondingly stronger
northerly flow over these same four stations and a dif-
fuse thin line (boundary 3).

From 2230 to 2300 radar showed a significant in-
crease in the radial velocity gradient (4 to 10 m s⁻¹
over a 1–2 km distance) across boundary 1, in the vic-
inity where cell A eventually developed. At about
2300, observers with the chase vehicle and at the radar
noted rapid cumulus development in this same area.
This cloud area organized into one large storm that
reached 60 dBZ by 2400.

The width of the convergence zone associated with
a boundary is generally 0.5 to 5 km, as indicated by
radar reflectivity and velocity. This is consistent with
what is observed in wind time traces from mesonet
stations after accounting for boundary movement rate.

c. Boundary detection

Overlaid plots, similar to Figs. 5, 8 and 9, of Dop-
pler radar boundaries with winds from the dense PAM
network and less dense PROFS network allowed us to
subjectively evaluate the ability of each to detect and
measure the convergence lines. Similar to the evidence
presented above for 15 and 20 June, the radar and
PAM network (~8 km spacing) nearly always agreed,
within limits of the station spacing, on the presence
and location of a convergence line. The exception was
cool, stable conditions that were typified by insufficient
radar scatterers to detect the clear air winds. Fortu-
nately, these conditions are not conducive to initiating
convection by boundary layer forcing.

The PROFS mesonet (~25 km spacing) effectively
detected the approximate location of large-scale con-
vergence lines such as the Denver convergence line,
synoptic fronts, and large convective storm outflows.
In these cases the winds are affected for tens of kilo-
meters on either side of the boundary. The smaller
convergence lines, such as small-scale gust fronts and
the unknown boundaries, were not detected or were
impossible to resolve adequately. Thus the PROFS
network adequately detected less than half of the con-
vergence lines in Table 2.

One minute average wind speed values from the
PAM stations are typically less than indicated by the
Doppler velocities. However, the peak 1 s wind speed
value within the minute is nearly the same magnitude
as the radar.

Holle and Watson (1983) observed that cloud and
echo development could be related to convergence es-
timates from a mesonet with a station spacing of ~7
km. We observed on a number of occasions that lo-
calized maxima in radar-observed convergence, along
a convergence line, would immediately precede and
accompany the rapid growth of individual storms. Such
localized increases were difficult to detect with the PAM
mesonet. In addition, the actual convergence within
the primary convergence zone was underestimated by
the PAM stations because the station spacing was typi-
cally greater than the width of the zone. While the
Doppler velocities can provide considerable useful de-
tail on the spatial and temporal evolution of the con-
vergence, this is only possible when the viewing angle
is appropriate.

A more comprehensive evaluation of the use of single
Doppler radar to estimate near-surface wind conditions
requires further study.

5. Storm origins relative to boundaries

Table 3 indicates the origin of the 653 storms ob-
served during the study period. Thirty-six percent (235
storms) advected into the area. Over 90% of these were
initiated in the mountains west of the study area. The
percentage of storms advecting into the area changed
significantly from 53% prior to 15 July to only 18% af-
after 15 July. This is most likely a result of the decrease
in the westerly winds aloft during the latter period.

The remaining 418 storms formed within the area.
The locations where these echoes first reached 30 dBZ
are plotted in Fig. 10. After making allowances for the two different areas covered during the study, it is obvious that the initiation location is essentially random. However, when initiation location is considered relative to a boundary, it is not random. Figure 11 shows for each of the 418 storms the distance to the nearest boundary at the time the storm first reached 30 dBZ. The figure is divided into three sections for moving, stationary, and colliding boundaries. A boundary was classified as moving if the reflectivity thin line or velocity convergence line showed discernible persistent motion over a 15 to 30 min period prior to storm development; otherwise, it was defined as stationary.

Table 3 shows that only 21% of the storms > 60 dBZ that occurred in this study are adveeted into the area. Very notably, several severe hailstorms moved from the mountains on 13 June and caused $350 million damage in Denver. The intensity and lifetime of these storms may have been increased when they moved from the mountains over a preexisting convergence zone over the plains. The organization and intensification of storms moving over boundaries was frequently noted. Thus, boundaries are not only associated with the initiation of new storms, but also with the intensification of old storms.

In Table 3, 3% of the cases were classified as virga. This refers to echoes < 30 dBZ that adveeted from the mountains and briefly exceeded 30 dBZ while in the study area. Two percent of the storms initiated near large storms without any apparent boundary association. The remaining 9% appeared without any obvious origin feature present. In some cases it is possible that a convergence line existed and was not detectable by radar. The average maximum reflectivity of these storms was less than the other types, with the exception of the virga cases. Excluding storms that adveeted into the area, 79% of the >30 dBZ storms and 92% of the >60 dBZ storms were boundary initiated.

The are plotted in Fig. 10. After making allowances for the two different areas covered during the study, it is obvious that the initiation location is essentially random. However, when initiation location is considered relative to a boundary, it is not random. Figure 11 shows for each of the 418 storms the distance to the nearest boundary at the time the storm first reached 30 dBZ. The figure is divided into three sections for moving, stationary, and colliding boundaries. A boundary was classified as moving if the reflectivity thin line or velocity convergence line showed discernible persistent motion over a 15 to 30 min period prior to storm development; otherwise, it was defined as stationary.

It is apparent from Fig. 11 that most of the storms occur near a boundary. The shading of the bar graphs in Fig. 11 shows storms that were subjectively classified as being boundary initiated based on study of the radar movies. These are the cases listed in Table 3 as boundary origin. The unshaded cases in Fig. 11 are the virga, near-storm, and other cases listed in Table 3. Note that there is a background of 1 to 3 cases per range interval in Fig. 11a that were classified as nonboundary initiated.

If there were no causal relationship between boundaries and storm initiation, the number of storms initiated on either side of a moving boundary would be equal and not displaced to behind the boundary as shown in Fig. 11a. Behind the boundary refers to regions from which the boundary is receding. Figures 11b and 11c show storms initiated very close to stationary and colliding boundaries. This is possible only if conditions associated with boundaries caused the storm initiation, or if the spacing between boundaries was so small that a storm was necessarily near one by chance. On storm days between 1800 and 0200 GMT, an average of one boundary was present in the study area at any one time. The average increased to almost two between 2200 and 2300 GMT. Considering that the study area was 50 to 100 km wide, the chance that a boundary would be within 10 km of a location was low. The distributions in Figs. 11b and 11c could not have occurred by chance.

**Table 3.** Origin of storms occurring in the study area. See text for description of origin types.

<table>
<thead>
<tr>
<th>Origin</th>
<th>Percent of cases</th>
<th>Average maximum dBZ</th>
</tr>
</thead>
<tbody>
<tr>
<td>Boundary</td>
<td>50 (653 cases)</td>
<td>50</td>
</tr>
<tr>
<td>Advect</td>
<td>36 (133 cases)</td>
<td>46</td>
</tr>
<tr>
<td>Virga</td>
<td>3</td>
<td>36</td>
</tr>
<tr>
<td>Near storm</td>
<td>2 (2)</td>
<td>51</td>
</tr>
<tr>
<td>Other</td>
<td>9</td>
<td>42</td>
</tr>
</tbody>
</table>

**Fig. 10.** Storm initiation locations (open and closed circles). The solid arc encloses the study area from 18 May to 30 June and the dashed arc 1 July to 15 August. Contours are elevation in meters.
a. Moving boundaries

It is apparent from Fig. 11 that there was a high preference for storms to form between 0 and 20 km behind a moving boundary, within 15 km of a stationary boundary and within 5 km of colliding boundaries. This does not mean that the origin of the storm updraft was at these locations.

The tendency for storms to form behind the boundary suggested that air was being forced up and over the moving boundary. Contributing to the spread of the distribution in Fig. 11a are differences in boundary propagation velocities, time for the updraft to initiate free convection, and cloud movement velocities while growing to 30 dBZ. Discussion of the details of updraft evolution and storm growth is not possible here and remains a subject for future research. Assuming that the updraft was initiated at the leading edge of the boundary and that the clouds were not advected during the time they grew to 30 dBZ, an estimate for the time required for the updraft to produce a 30 dBZ echo can be obtained. Figure 12 shows, for moving boundaries, the frequency distribution of the time from boundary passage to when a storm reached 30 dBZ. The time difference was obtained by noting the time and place when a storm first reached 30 dBZ and subtracting that time from when a boundary passed the same location. Selection of the 115 cases used for Fig. 12 was based on the presence of an easily observable, consistently moving boundary. The distribution in Fig. 12
ranges from 8 min before the boundary reaches the location where the storm first appeared to 58 min after. The median time is 19 min after passage. This is only a crude estimate of the time distribution for a storm to grow from initial updraft to a 30 dBZ storm. Cloud advection velocities, updraft locations relative to the leading edge of the boundary, and prior history of cloudiness will affect the distribution in Fig. 12.

Other observations of the time required for an updraft to initiate a storm are few. Knight et al. (1983) studied 12 cumulus clouds in eastern Colorado. While the clouds were being scanned by radar they photographed building cumulus cloud turrets from aircraft. The most rapid development observed was 10 min from the first visual appearance of a growing cloud turret until it produced a 30 dBZ radar echo. However, the time from the first appearance of visual cloud to the first radar echo (5 dBZ) was highly variable. Typically, it required 6 min to grow from 5 dBZ to 30 dBZ. Knight et al. (1983) supported these findings with results from a simple microphysical model. Based on their visual and model calculation, a cloud could grow by the ice process from first cloud to a 30 dBZ echo in roughly 15 min. If clouds were already present, the time might be reduced to 5 min.

Although Fig. 12 and the Knight et al. (1983) observations cannot be directly compared, they are in general agreement. These results suggest that a boundary passing under existing cumulus towers could initiate a 30 dBZ storm in a minimum of 5 min. Without cloud already present, the time would more likely be 15 min or longer.

Eleven storms in Fig. 11 that are classified as boundary-initiated formed in advance of a moving boundary. Six of these also appear in Fig. 12 as becoming 30 dBZ storms before the boundary reached them. It could be argued that these are improperly classified and that a boundary had nothing to do with their initiation. However, the relatively high frequency of cases (above background) in the 0–5 km range in Fig. 11a suggests differently. A mechanism other than forced lifting along the boundary could be responsible, such as a gravity wave propagating in advance of the boundary along the mixed-layer inversion. In addition, when the convergence zone was wide and atmospheric conditions were suitably unstable, there may have been sufficient lift in advance of the visible radar boundary to initiate deep convection.

b. Colliding boundaries

Figure 11 shows that when storms formed, they were likely to be near a convergence line. While this knowledge is of importance to forecasting, precise knowledge of when and where storms will form along the boundary is highly desirable. When storms form after boundaries collide, they occur near the point of intersection (Fig. 11c) and shortly after contact. Thus, these cases offer considerable forecast information. Sixty-four of the 327 storms classified as boundary initiated were the result of colliding boundaries. While this is not a large percentage of the cases, the resulting storms averaged 54 dBZ and often resulted in major lines of thunderstorms. On 18 May two intersecting boundaries caused a series of five tornadoes (Wilson, 1986). Holle and Maier (1980) and Weaver and Nelson (1982) have also reported tornado formation following the collision of boundaries.

There were 49 cases where boundaries collided. Seventy-one percent of the time, new storms were initiated or existing storms were strengthened and enlarged. Collisions were divided into three types called “merger,” “intersection,” and “collision” and were compared with resulting convection. Merger is defined as one boundary overtaking another boundary moving in the same direction; intersection refers to those cases where the boundaries make an angle > 30° at the point of intersection, and collision refers to cases where the intersection angle is less than 30°. The results are shown in Table 4. Collision cases produced new storms or strengthened existing storms 84% of the time, as compared to 63% and 64% for the merger and intersection cases, respectively.

The likelihood of deep convection following a collision is undoubtedly related to the intensity of forced lifting and the stability of the atmosphere. A crude estimate was obtained of the stability of the atmosphere for each collision. The surface dry bulb and dewpoint temperatures were estimated using nearby mesonet stations. The upper air conditions were estimated from the 0000 Denver sounding, which often may have been modified considerably by prior convection. The air being lifted was classified as stable if either of the fol-
Table 4. Resulting convection versus the type of boundary collision. "None" indicates no new convection occurred, "intensified" indicates existing storms were intensified and enlarged, and "new" indicates at least one new storm formed. The collision type is defined in the text.

<table>
<thead>
<tr>
<th>Collision Type</th>
<th>Resulting Convection</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>None</td>
</tr>
<tr>
<td>Merger</td>
<td>3</td>
</tr>
<tr>
<td>Intersection</td>
<td>8</td>
</tr>
<tr>
<td>Collision</td>
<td>3</td>
</tr>
</tbody>
</table>

loowing two conditions occurred: 1) temperature at the convective condensation level, based on the estimated surface conditions, was colder than the sounding temperature at that level or 2) the lifted index based on the same data was >0. This procedure classified the air as stable for 71% of the cases when no new convection resulted, as compared to 31% of the cases when new storms were initiated or existing activity was strengthened. Thus, even these crude estimates of hydrostatic stability indicate the importance of knowing the stability of the air in predicting the likelihood of new convection in the vicinity of the boundaries.

The colliding boundaries provide an even better opportunity to obtain another estimate of the time required for an updraft to produce a 30 dBZ echo. Three cases (25, 28 and 30 July) were selected, each with two boundaries moving toward each other. Figure 13 shows the 25 July case. The boundaries are easily observed as enhanced northeast–southwest-oriented thin lines of enhanced reflectivity. The boundary moving from the northwest (labeled 1) was the result of cool outflow from thunderstorms over the mountains. The boundary moving from the southeast (labeled 2) is probably a gust front from the thunderstorms to the southeast. Echoes $\geq 30$ dBZ are shown as black in Fig. 13. The boundaries first collide at 2249 at 110° at 30 km (labeled A). The boundaries continue to collide both northeast and southwest of this point for the next 20 min. The boundaries then become one and continue to move northwestward, apparently because the air from the southeast is cooler and denser than from the northwest. A new line of echoes can be seen in Fig. 13f along the line of collision (labeled B). The old line of echoes has almost dissipated by this time. Besides the boundaries, other thinner lines oriented north–south are visible; these appear to be horizontal rolls. The boundary can be seen intersecting the most pronounced north–south line just beyond the 40-km range between about 60° and 90° (labeled 3). It is not known whether this line is a particularly deep horizontal roll or some other type of circulation. Figures 13d–f show that a short line of thunderstorms (labeled C) develops along this intersection with about a 20-min lag following intersection.

Figure 14 shows for each of the three colliding boundary cases, the location of the boundaries 20 min before collision, the collision line between the boundaries, and the location where each storm first reached 30 dBZ. The time from the boundary intersection at the point closest to the storm, to the storm reaching 30 dBZ varied from 5 to 34 min; the median was 24 min.

By reasons of mass continuity, the updraft necessarily starts before the boundaries actually collide. It follows that the times given are minima for the actual time from initial updraft to a 30 dBZ storm. Numerical simulations by Droegemeier and Wilhelmson (1985) of cloud development along colliding boundaries show that the updraft begins while the boundaries are still separated and increases until collision. The simulations indicate that storms will not form where the outflows first meet, but rather from $\sim 7$ km on either side of this point in the warm air where convergence is the largest. Contrary to these findings, except for storm 1 on 30 July, the first storms to form in Fig. 14 form very near the first point of contact.

6. Conclusion

This study supports many of the findings in Purdom (1982), which suggest that mesoscale boundary layer convergence lines play a major role in determining when and where storms will form. Thus, what appear as random occurrences of thunderstorms, actually may be the result of one of the following: (i) the creation of a particular intense or sustained updraft associated with a convergence-line-related kinematic feature, (ii) collision of convergence lines, (iii) intensification of existing clouds when a boundary passes under them, (iv) a boundary encountering a localized relatively unstable air mass, or (v) the interaction of a boundary with a range of hills or mountains. Although storm initiation locations are in close proximity to a convergence line, specifically where and when they will occur along the line is often unknown. This is particularly the case when boundary collisions are not involved. Occasionally, Doppler radar will show a localized area of increased convergence immediately preceding rapid storm growth. Also, monitoring of cloud growth along, and in advance of, a convergence line was sometimes useful in identifying regions where subsequent storm formation was likely.

Location and time-specific 0–2 h forecasts of thunderstorm development now appear possible. The integration of Doppler radar data is required in order to detect and monitor convergence lines, high resolution satellite data are required to monitor cloud growth, and surface and sounding data are required to estimate atmospheric susceptibility to deep convection. Limited
Fig. 13. Equivalent radar reflectivity displays depicting the collision of two boundaries on 25 July and the subsequent initiation of a line of thunderstorms. Reflectivities are given in dBZ, by the scale at the bottom of the display. Values greater than 30 dBZ are solid black. Precipitation echoes are then black and the gray shading represents clear air return. The boundaries appear as northeast-southwest oriented thin lines. The north-south line of >30 dBZ echo just west of the radar is the front range of the Rocky Mountains. Range marks are in 20-km intervals. The data are for an elevation angle of 0.9° and are from the horizontally polarized X-band portion of the CP-2 radar. Major features are indicated by arrows and are described in the text. Times are (a) 2226, (b) 2242, (c) 2252, (d) 2302, (e) 2317, and (f) 2333 GMT.
Fig. 14. Storm initiation locations and times relative to boundary collision locations and times for three cases: (a) 25 July, (b) 28 July, and (c) 30 July.
real-time forecast experiments conducted with PROFS during the summer of 1985 indicate that there is great potential for this technique.

This study showed that 79% of the storms (≥30 dBZ) initiating over the plains of eastern Colorado were in association with convergence lines observed by radar. This number increases to 95% for storms ≥ 60 dBZ. Colliding boundaries initiated storms or intensified existing storms near the point of collision in 71% of the cases. The more direct the collision and less stable the air, the higher the likelihood of new convection.

It is not known how transferable these results are to other geographical locations. Field experience with the NCAR Doppler radars in Montana, Oklahoma, Kansas, and Illinois indicate that clear-air convergence lines are also observed in these locations. However, it is our impression that the frequency of convergence lines in the Midwest is less than in Colorado, and they tend to be associated with larger-scale weather systems and larger-scale outflow boundaries.

To routinely detect boundaries during the convective storm season, the radar must be capable of detecting a −5 dBZ signal at 50 km in Colorado. The primary problems hampering observation of boundaries are due to ground clutter and radar range ambiguities. Automatic detection of these boundaries, as planned for the NEXRAD or Terminal Doppler Weather Radar programs, will require successful removal of these data artifacts. Low sidelobe antennas, narrow beamwidths, and short wavelengths all contribute to less obscuration from ground clutter. This was particularly obvious during this study in comparing the low sidelobe X-band data to normal sidelobe S-band data from CP-2.

Observational and numerical studies are needed to determine details of updraft initiation and resulting cloud growth caused by moving, stationary, and colliding boundaries. Understanding of boundary formation and maintenance mechanisms is also needed. This is particularly the case for the unknown stationary lines that were observed. Observational equipment to conduct the above studies exists. These include Doppler radar and lidars, dense mesonets, instrumented aircraft, rapid-scan satellite imagery, mobile radiosondes, and photographic equipment.

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