Analysis of Convergence Boundaries Observed during IHOP_2002

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ABSTRACT

An analysis of six convergence boundaries observed during the International H2O Project (IHOP_2002) is presented. The detailed kinematic and thermodynamic structure of these boundaries was examined using data collected by an airborne Doppler radar and a series of dropsondes released by a jet flying at ~500 mb. The former and latter platforms were able to resolve the meso-γ- and meso-β-scale circulations, respectively. Convection initiated on three of the days while no storms developed in the regions targeted by the mobile platforms on the other days (referred to as null cases). The airborne radar resolved the finescale structure of four drylines, a cold front, and an outflow boundary on the six days. Horizontal profiles through radar-detected thin lines revealed “bell-shaped distributions” and there appeared to be a seasonal dependence of the peak values of radar reflectivity. The echo profiles through the fine line in May were, in general, greater than those plotted for the June cases. There was no apparent relationship between the intensity of the low-level updraft and convection initiation. The strongest updraft resolved in the dual-Doppler wind synthesis was associated with a null case. There was also no relationship between the strength of the moisture discontinuity across the boundaries and convection initiation.

The three days during which storms developed were all associated with two convergence boundaries that were adjacent to each other. The two boundaries collided on one of the days; however, the boundaries on the other two days were approximately parallel and remained separated by a distance of 5–15 km. The total derivative of the horizontal vorticity rotating along an axis parallel to the boundary was calculated using dropsonde data. The horizontal gradient of buoyancy was the largest contributor to the change in vorticity and revealed maximum and minimum values that would support the generation of counterrotating circulations, thus promoting vertically rising air parcels. These updrafts would be more conducive to convection initiation. The null cases were characterized by a low-level vorticity generation of only one sign. This pattern would support tilted updrafts. The results presented in this study suggest that it is not necessary for two boundaries to collide in order for thunderstorms to develop. Solenoidally generated horizontal circulations can produce conditions favorable for convection initiation even if the boundaries remain separate.

1. Introduction

There have been important advances in short-term forecasts (nowcasts) of thunderstorm initiation during the warm season. These advances are critical as illustrated by Olsen et al. (1995). They highlighted the pronounced drop in our predictive skill during the summer months when the precipitation totals are the greatest. The improvements in our understanding of thunderstorm formation are largely attributed to the recognition that storms frequently develop near boundary layer convergence zones that are often detected by Doppler radars and satellite imagery (e.g., Purdom 1976, 1982; Wilson and Schreiber 1986; Wilson and Mueller 1993; Wilson et al. 1998). Indeed, Wilson and Schreiber (1986) showed that 80% of thunderstorm initiation occurred near surface convergence zones just east of the Rocky Mountains.

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in Colorado. It is also known, however, that the existence of a convergence boundary does not imply that convection will develop even when large convective available potential energy (CAPE) and conditionally unstable environments exist (e.g., Stensrud and Maddox 1988; Richter and Bosart 2002; Cai et al. 2006).

There have been a number of individual case studies that have examined the detailed structure of convergence boundaries and their relationship to convection initiation (e.g., Murphey et al. 2006; Buban et al. 2007; Miao and Geerts 2007; Weckwerth et al. 2008). However, there have been few systematic attempts to perform a comprehensive analysis of a number of convergence boundaries using analogous datasets. Such an analysis would result in generalized conclusions concerning the thermodynamic and kinematic characteristics of the boundaries and their relationship to thunderstorm formation. Hane et al. (2002) compared the characteristics of boundaries but was primarily restricted to using in situ data collected at flight level. Karan and Knupp (2006) and Harrison et al. (2009) studied a series of convergence boundaries but only at a one location along the thin line. As a result, the life cycle and the along-boundary variability could not be accurately assessed. Marquis et al. (2007) examined the kinematic structure of several boundaries but focused on the characteristics of mesocyclones using ground-based dual-Doppler analysis. Miao and Geerts (2007) examined the vertical structure of drylines primarily based on the analysis of data collected by the Wyoming King Air.

The current study presents airborne dual-Doppler wind syntheses and thermodynamic analyses based on soundings for six convergence boundaries observed during the International H2O Project (IHOP_2002; Weckwerth et al. 2004). The aircraft flew a box pattern around the

![Surface analysis for different dates and locations](image-url)
boundaries at low levels with along-boundary legs \(\sim 100\) km long. The Doppler radar data collected allowed for an assessment of both the along-boundary variability but also the mean characteristics normal to the convergence zone over an extended region. The flight legs also resulted in a dataset with analogous spatial resolution at the meso-\(\gamma\) scale so that direct comparisons between the case studies could be made. In addition, the kinematic and thermodynamic structure of all of these boundaries was well documented using data from a series of dropsondes deployed by a jet flying at \(\sim 500\) mb. The spatial resolution of the sounding data at the meso-\(\beta\) scale was comparable, which facilitated comparisons between the cases.

A description of IHOP_2002 and the primary datasets used in this study are discussed in section 2. Section 3 presents surface analyses and the flight track of the research aircraft. Single- and dual-Doppler analyses of the convergence boundaries are shown in section 4. Section 5 presents the kinematic and thermodynamic structure across the convergence boundaries based on dropsonde data. A summary and discussion are presented in section 6.

2. IHOP_2002 and the primary data

One of the main objectives of IHOP_2002 was to document the three-dimensional water vapor distribution in the lower troposphere to better understand the processes that lead to the initiation of deep convection. The field phase took place during the spring and summer of 2002 over the southern Great Plains and brought together many mobile platforms. These platforms were necessary to sample a number of convergence boundaries over an extensive geographic region. Intensive observation periods (IOPs) for 6 days are presented in this study. Convection initiation occurred on 24 May, 10 June, and 19 June. No storms developed in the area targeted by the research platforms on 22 May, 11 June, and 12 June (hereafter, referred to as null cases). The latter has also been referred to as convection initiation failure by Markowski et al. (2006). Detailed analyses are presented
on four drylines, a cold front, and an outflow boundary on the six days.

The datasets collected by two platforms are highlighted in this study—an airborne Doppler radar and dropsondes deployed from an aircraft. A 3-cm airborne Electra Doppler Radar (ELDORA) is operated by the National Center for Atmospheric Research (NCAR) and is flown on board a Naval Research Laboratory (NRL) P-3. ELDORA is equipped with two antennas that scan fore and aft of the normal to the aircraft. Dual-Doppler wind fields can be synthesized with these data using a fore-aft scanning technique (FAST; Jorgensen et al. 1996). Convergence boundaries often appear as radar-detectable thin lines even in the absence of precipitation particles (e.g., Wilson and Schreiber 1986; Hane et al. 2002). Fortunately, ELDORA is capable of detecting echoes and Doppler velocities within the clear air (e.g., Wakimoto et al. 1996).

The nominal research flight track required the P-3 to fly between 400 and 700 m above ground level (AGL; hereafter, all heights are AGL except where indicated) and parallel to the thin lines. The aircraft flew rectangular box-like patterns within 2–3 km on either side of the thin line. The flight tracks were ~100 km in length. The flight tracks were challenging to execute and required continuous adjustments in the heading owing to the non-linear nature of many of the thin lines. The flight pattern flown by the P-3 for the convergence boundaries studied during IHOP_2002 resulted in spatial and temporal resolutions of the Doppler wind syntheses that were analogous. (The radar scanning parameters are shown in Table A1 and the radar methodology is presented in the appendix.) For more technical information regarding ELDORA, the interested reader is referred to Hildebrand et al. (1994, 1996).

The kinematic and thermodynamic structure of the convergence boundaries in a vertical plane on a larger spatial scale was revealed by a series of dropsondes that were deployed by a jet flying at ~500 mb. The orientation of the flight track was approximately perpendicular to the convergence boundary. The rapid deployment of
FIG. 2. Flight track for the P-3 on 22 May, 24 May, 10 Jun, 11 Jun, 12 Jun, and 19 Jun 2002. The inset represents the rotated coordinate system that was used to create mean vertical cross sections for the dual-Doppler analyses shown in Fig. 5. The star denotes the location of S-Pol or a WSR-88D radar.
FIG. 3. Surface observations for (a) 2200 UTC 22 May, (b) 2000 UTC 24 May, (c) 2100 UTC 10 Jun, (d) 2200 UTC 11 Jun, (e) 2000 UTC 12 Jun, and (f) 2100 UTC 19 Jun superimposed onto low-level surveillance scans of radar reflectivity. Temperature and dewpoint temperature (°C) are plotted. The track of the P-3 is shown by the dotted line. Sounding locations are shown on the figures. Symbols denoting fronts, drylines, and outflow boundaries are partially drawn so they do not obscure the thin line features on the radar reflectivity plots. Wind vectors are plotted using the following notation: barb = 5 m s\(^{-1}\) and half barb = 2.5 m s\(^{-1}\).
the dropsondes from the aircraft resulted in an elapsed time between the first and last sounding of <25 min.

3. Surface analysis and the flight track

Surface analyses for the six cases are shown in Fig. 1. The IHOP_2002 field operations plan for convection initiation studies proposed that the intensive observing region was centered on the box pattern flown by the P-3. Accordingly, the mobile ground-based facilities and locations where dropsondes were released were concentrated near the center of the aircraft flight track.

A well-defined dryline that developed ahead of a cold front was the focus of the IOP on 22 May. This type of dryline formation is typical in the region (e.g., Schultz et al. 2007). A secondary dryline also formed to the west of the primary dryline later in day (Fig. 1a) as noted by several investigators (e.g., Demoz et al. 2006; Geerts et al. 2006; Weiss et al. 2006; Buban et al. 2007; Miao and Geerts 2007; Wakimoto and Murphey 2009). An opportunity to examine a dryline that formed a “triple point” with a cold front where three air masses converge (e.g., Reed and Albright 1997; Weiss and Bluestein 2002) occurred on 24 May (Fig. 1b). ELDORA collected Doppler radar data on the dryline and the intersection of the boundary with a cold front that was at the leading edge of a Canadian air mass. The initiation of an intense squall line was documented by several investigators (e.g., Wakimoto et al. 2006; Xue and Martin 2006; Ziegler et al. 2007) and can be seen near the bottom half of the satellite image at 2000 UTC (Fig. 1b). A section of a cold front over southwest Kansas was well sampled by the IHOP_2002 observing platforms on 10 June (Fig. 1c). An interesting aspect of this case was a nearby dryline that developed to the southeast and approximately parallel to the front as documented by Friedrich et al. (2008a,b). Conditions became favorable for thunderstorm development as isolated cells initially formed at 2000 along the cold front and north of the Oklahoma–Kansas border (Fig. 1c). The dryline on 11 June was associated with the weakest kinematic discontinuity observed during the experiment; however, there was a substantial moisture gradient across the boundary as shown by Cai et al. (2006; see Fig. 1d). No storms developed along the dryline on this day. An approximate west–east-oriented boundary along the Oklahoma–Kansas border was produced by outflow from a mesoscale convective system on 12 June (Fig. 1e). A dryline intersected the outflow just east of a circulation associated with mesolow in the Oklahoma Panhandle. A cold front extended from the mesolow and merged with the dryline in the western Texas Panhandle (not shown). Although convection initiated along the outflow boundary (Fig. 1e), it occurred at the far eastern edge of the observational area defined by the P-3 flight track (Markowski et al. 2006; Weckwerth et al. 2008). This case is classified as a null case since no convection developed along the primary region of the boundary sampled by the P-3. The final case analyzed is a dryline that developed parallel to and just east of a cold front in northwest Kansas on 19 June (Fig. 1f). The first legs flown along the dryline by the P-3 occurred under clear skies. Subsequently, a strong squall line developed as shown in Karan and Knupp (2006) and Murphey et al. (2006).

The P-3 flight tracks illustrating the long flight legs along the convergence boundary for all six cases are shown in Fig. 2. Some of the flight tracks in the figure that deviate from this pattern were a result of temporary mechanical problems with the airborne radar (10 June) and flying survey patterns over an extensive region while waiting for the convergence boundary to become well defined (11 June). The insets in the figure denote the rotated coordinate system that was used to create the mean vertical cross sections based on the dual-Doppler wind syntheses presented in the next section.

4. Single- and dual-Doppler analyses

Surface analyses for a single time superimposed onto the low-level surveillance scans recorded from a nearby Doppler radar are shown in Fig. 3. Enhanced lines of echo (i.e., thin lines) developing in the clear air can be seen on all missions. Representative examples of the dual-Doppler wind syntheses are shown in Fig. 4. The dryline on 22 May was well sampled by ELDORA based on numerous flight legs by the P-3. A convergence pattern can be seen in the surface analysis and the dual-Doppler wind synthesis (Figs. 3a and 4a) and the strong moisture discontinuity across the dryline is evident (Fig. 3a). There is a pronounced along-boundary variability in the echo pattern of the thin line resolved by ELDORA (Fig. 4a). The maximum radar reflectivity is >6 dBZ. The intersection of the dryline and the cold front on 24 May results in an echo pattern that resembles an “inverted V” (Fig. 3b). The cold front–dryline intersection is easily identified in the ELDORA analysis and the aircraft collected in situ measurements at flight level near the triple point (Fig. 4b). There is a pronounced cyclonic shift of the dual-Doppler winds across the dryline.

The cold front and the approximate position of a dryline on 10 June is shown in Fig. 3c. The existence of the dryline was shown in a series of analyses presented by Friedrich et al. (2008a,b). Note the surface station reporting a southwesterly wind direction and a dewpoint temperature of only 5.5°C located between cold front and dryline in the southwest corner of Kansas in Fig. 3c.
indicates the intervening dry air mass between the two boundaries. The thin line accompanying the cold front can be identified in the surveillance scan (Fig. 3c) and was the primary boundary sampled by the P-3 (Fig. 4c). Thunderstorms were initiating along the cold front at 2100 (Fig. 3c). The gray lines in Fig. 4c indicate the positions of two horizontal convective rolls (HCRs). The intersections of HCRs with convergence boundaries have been hypothesized to be locations where convection may initiate (e.g., Wilson et al. 1992; Atkins et al. 1995). The maximum values of radar reflectivity along the thin line occur near the intersection points.

The dryline that developed on 11 June was associated with the weakest kinematic discontinuity observed during IHOP_2002. The moisture contrast across the boundary, however, was substantial (Fig. 3d). The fine line based on ELDORA data was apparent (Fig. 4d), but was not as well defined as the other cases presented in this paper. The horizontal convergence and derived updrafts based on the wind syntheses (to be shown later) were weak and the wind shift across the dryline was not distinct (Fig. 4d). The P-3 flew along an outflow boundary and a triple point formed by the intersection of the boundary with a dryline on 12 June (Fig. 3e).
North–south-oriented banded structures in the radar reflectivity plots shown in Figs. 3e and 4e that emanate from the outflow boundary thin line have been hypothesized to be internal gravity waves (Weckwerth et al. 2008). A shift of the winds from southerly to easterly in the air masses located south and north of the outflow boundary, respectively, is apparent in Figs. 3e and 4e. The P-3 collected data on a dryline that formed parallel to and east of a cold front in northwestern Kansas on 19 June. Strong northwesterly and southerly flow in the postfrontal and pre-dryline air masses, respectively, were apparent on this day (Fig. 3f). There was relatively dry air associated with westerly winds within the air mass between the two boundaries. A thin line associated with the dryline is shown in Fig. 4f.

The long flight legs flown by the P-3 provided an opportunity to reconstruct the mean vertical structure of the six boundaries at the meso-γ scale by averaging individual cross sections from the dual-Doppler wind syntheses. The averaging was effective in removing the along-line

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**FIG. 4. (Continued)**

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variability that existed. Examples of the mean vertical cross sections for each of the cases are presented in Fig. 5. Average profiles at low levels (700 m) across the convergence boundaries for several kinematic variables are shown in Figs. 6a–f. Figures 6g,h depict the mean in situ profiles at flight level of mixing ratio and virtual potential temperature, respectively, based on several passes through the boundary at approximately 700 m.

To the authors’ knowledge, a detailed comparison of the reflectivity structure of thin lines associated with convergence boundaries with comparable spatial resolution has only been attempted by Harrison et al. (2009). Visual inspection of Figs. 4 and 5 suggests that the widths of the thin lines, based on radar reflectivity, are highly variable. This is also supported by the approximate “bell-shaped distributions” presented in Fig. 6a. In addition, there appears to be a seasonal dependence of the peak values of radar reflectivity. The echo profiles through the fine lines collected in May are greater than those plotted for the June cases. This would suggest the presence of either smaller and/or fewer scatterers in late spring as a result of insects since this is the primary source of echo return in the clear air (e.g., Wilson et al. 1994; Geerts and Miao 2005). Note that the echo profiles (except for 12 June) are asymmetric with higher background values in the cool, moist versus the dry air masses (see Fig. 6g). These observations should be viewed with some caution owing to the limited number of cases presented. Other factors such as the impacts of different geographic locations of the boundaries, type of boundary, and the characteristics of the boundary layer air on either side of the boundary should also be considered.

The results from Wilson et al. (1994) suggest a relationship between the peak value of radar reflectivity and strong horizontal convergence and updrafts. Harrison et al. (2009) showed that the average updraft velocity increases with increasing radar reflectivity within the thin lines. The results presented in Fig. 6a suggest that this relationship is more complex than previous presented. The strongest updraft (Figs. 6e and 7) and horizontal convergence (Fig. 6c) occurred on 22 May; however, the maximum echo intensity across the fine line is less than the 24 May case when the updrafts and horizontal convergences were weaker. It might be expected that the peakedness of the radar reflectivity profiles would be greater for the cases exhibiting strong horizontal convergence. This does not appear to be the case since the 22 May plot in Fig. 6a has a relatively “flat” distribution even though the updrafts were the strongest observed along a thin line during IHOP_2002. A broader distribution of radar reflectivity is indicative of a wider thin line. Strong convergence could still exist since insects may not be able to stay in strong updrafts and, instead, spread out with the upper-boundary layer divergent flow as suggested by Miao et al. (2006).

The strongest and weakest mean peak updrafts occurred on 22 May and 11 June, respectively (Figs. 6e and 7). These were also the days that experienced the strongest and weakest horizontal convergence (Fig. 6c) highlighting the strong relationship between these two variables. Comparing the updraft profiles for the six cases (Fig. 7) reveals the weak relationship between positive vertical velocity and convection initiation. Although the null case on 11 June was associated with the weakest updrafts, the most intense updrafts were observed for another null case on 22 May. This suggests that other factors are important in determining whether deep convection will develop. The fine lines were all associated with cyclonic vorticity (Fig. 6b) largely a result of the horizontal shear of the component of the flow parallel to the boundary (u, Fig. 6d). For example, the change of u across the fine line on 19 June is \( \pm 3.25 \text{ m s}^{-1} \) over a distance of 5 km (estimated from Fig. 6d) or \( 6.5 \times 10^{-4} \text{ s}^{-1} \). The value is close to the \( \sim 8.0 \times 10^{-4} \text{ s}^{-1} \) maximum value of vertical vorticity estimate from Fig. 6b.

The horizontal vorticity along an axis that is parallel to the thin line (Figs. 6f and 8) reveals substantial variability for the six cases. The 19 June case is associated with counterrotating horizontal circulations on either side of the convergence boundary that approximately balance (Figs. 5f and 8f). This situation has been described by Rotunno et al. (1988) as producing updrafts that are vertically erect and, therefore, favorable for the initiation of convection. In contrast, the horizontal vorticity pattern accompanying the 22 May convergence boundary (a null case) is dominated by negative values suggesting a strongly tilted updraft over the denser air mass (Figs. 5a and 8a). The horizontal vorticity pattern on 24 May (a day when convection initiated) is similar to 19 June although the positive and negatives values of vorticity are weaker (Figs. 6f and 8b). The profile of horizontal vorticity on 10 June resembles the null case on 22 May even though convection initiated on this day (Figs. 6f and 8c). Accordingly, the latter case suggests that the observed balance of horizontal vorticity at small scales does not appear to distinguish null days from days when convection initiation occurs.

Average horizontal profiles of mixing ratio and the virtual potential temperature across the boundaries based on in situ measurements collected at flight level are shown in Figs. 6g,h, respectively. The average profiles derived from data collected during in situ penetrations of the thin line are less robust than the mean vertical profiles based on the ELDORA wind syntheses. This is owing to the fewer transects though the boundary by the aircraft (typically three to six penetrations for each case). The flight level data were smoothed using a 15-s
FIG. 5. Mean vertical cross sections perpendicular to the thin line for (a) 2148–2203 UTC 22 May, (b) 1915–1926 UTC 24 May, (c) 1930–2005 UTC 10 Jun, (d) 2121–2133 UTC 11 Jun, (e) 2044–2050 UTC 12 Jun, and (f) 2104–2116 UTC 19 Jun 2002. (top) Radar reflectivity (dBZ) is plotted as gray lines with values greater than 2 dBZ shaded gray. Positive (negative) values of vertical vorticity (10⁻³ s⁻¹) are plotted as thin black (dashed) lines. Vertical velocities and component of horizontal flow in the plane of the cross section are plotted as black arrows. (bottom) Positive (negative) values of vertical velocity (m s⁻¹) are plotted as black (dashed) lines. Component of horizontal flow (m s⁻¹) perpendicular to the boundary is plotted. Positive (westerly) and negative (easterly) flow are shown by the gray and dashed gray lines, respectively. Plotted vectors are the total horizontal wind. The rotated coordinate system is shown in the inset in Fig. 2.
Fig. 6. Profiles across the thin line based on (a)–(f) the mean dual-Doppler wind syntheses at 700 m AGL or (g), (h) in situ data collected at flight level. (a) Echo intensity, (b) vertical vorticity, (c) horizontal divergence, (d) horizontal shear, (e) vertical velocity, (f) horizontal vorticity, (g) mixing ratio, and (h) virtual potential temperature for 22 May, 24 May, 10 Jun, 11 Jun, 12 Jun, and 19 Jun 2002.
running average. There is no apparent relationship between the strength of the moisture discontinuity and convection initiation. The strongest moisture gradient during IHOP_2002 was associated with the 11 June null case (Fig. 6g). Moreover, thunderstorms developed along the boundary on 10 June even though the moisture gradient was relatively weak (Fig. 6g). The virtual potential temperature gradients were comparable except for the 24 May and 19 June cases (Fig. 6h). The former exhibits a reverse temperature gradient even though the moist air mass was denser on a larger scale on this day as depicted by the sounding data (Fig. 9b). Atkins et al. (1998) also noted that the virtual potential temperature gradient measured within 10 km of a dryline could be different than the larger-scale dryline environment (also discussed by Geerts et al. 2006). It should be noted that Geerts et al. (2006) and Miao and Geerts (2007) did document very small decreases of virtual potential temperature (~0.25 K) in the moist air mass using the Wyoming King Air measurements at flight levels lower than those flown by the P-3 (e.g., 400 and 170 m versus the 700 m flown by the P-3).

5. Vertical cross-sectional analysis based on dropsonde data

An important component of the IHOP_2002 dataset was the numerous and rapid deployment of dropsondes...
Fig. 9. Thermodynamic analysis through the convergence boundaries on (a) 22 May, (b) 24 May, (c) 10 Jun, (d) 11 Jun, (e) 12 Jun, and (f) 19 Jun 2002. (top) Calculated CAPE and CIN values for a surface-based parcel for each sounding. (bottom) Vertical cross section perpendicular to the dryline of virtual potential temperature (black lines, K) and mixing ratio values (gray lines, g kg$^{-1}$). Values of mixing ratio $>10$ g kg$^{-1}$ are shaded gray. The height of the LFC is shown by the thick dotted line. The rectangular box represents the airborne dual-Doppler region. The locations of the soundings are shown by the dashed lines. Wind vectors are plotted using the following notation: flag = 25 m s$^{-1}$, barb = 5 m s$^{-1}$, and half barb = 2.5 m s$^{-1}$. The release points of the soundings are shown in Fig. 3.
Fig. 9. (Continued)
Fig. 9. (Continued)
by a jet flying at midlevels. Analysis of the dropsonde data resolved the kinematic and thermodynamic fields that were apparent at the meso-β scale. The flight tracks were approximately perpendicular to the convergence boundaries (see Fig. 3). The typical elapsed time to complete the leg was <25 min.

The moist air mass is capped by a strong inversion to the east of the dryline on 22 May (Fig. 9a). The plot of surface-based CAPE and convective inhibition (CIN) is shown in the top panel. The CAPE values were low to the west of the dryline and reached a maximum ~75 km to the east of the boundary. The minimum in CIN was also located near the dryline. The lowest level of free convection (LFC) was estimated to be ~4.2 km for the 2214:46 UTC sounding.

The series of dropsondes on 24 May captured the cold front as it was beginning to undercut the moist air east of the dryline. The combined solenoidally driven circulations by the cold front and the dryline would tend to enhance the updrafts in the vicinity of the merger between these two boundaries (Fig. 6f). Note that the 2 g kg\(^{-1}\) isopleth of mixing ratio protrudes upward near the 2037:36 UTC sounding. This denotes the location where deep convection subsequently initiated and was near low values of the LFC.

The two boundaries apparent on 10 June (Friedrich et al. 2008a) can be identified in the vertical cross section presented in Fig. 9c (note the 317-K isentropes in the analysis). The juxtaposition of these boundaries suggests the presence of counter-rotating horizontal circulations produced by solenoidal effects that approximately balanced on this day even though it was not apparent in the dual-Doppler analysis shown in Figs. 5c and 6f. Thunderstorms developed along the cold front and were supported by the upward bulge in mixing ratio and virtual potential temperature between 2 and 3 km in Fig. 9c. The initiation of storms above the cold front is not surprising since there was low CIN (Fig. 4c) and enhanced convergence along the boundary due to the southwesterly flow in the environment.

Two air masses are evident in the vertical cross section on 11 June (Fig. 9d). The frontal and dryline boundaries are approximately delineated by the 317-K isentrope and also suggested by the 10 g kg\(^{-1}\) isopleths of mixing ratio. The CAPE attains a maximum near the surface front, decreases to a minimum between the front and dryline, and then rises monotonically starting at the location of the dryline. The values of CIN reach a maximum in the intervening air between the two air masses. The greatest potential for deep convection to develop was along the dryline if the updrafts could overcome the CIN.

The cool and moist air mass accompanying the easterly flow behind the outflow boundary on 12 June is evident in the cross section presented in Fig. 9e. Markowski et al. (2006) hypothesize that the lack of a persistent, spatially continuous corridor of mesoscale ascent along the outflow boundary and associated moisture upwelling led to the failure of convection initiation on this day. The cold front and dryline are both apparent in the vertical cross section on 19 June (Fig. 9f). There is westerly flow associated with drier and warmer air between the two cooler air masses. As previously mentioned, the horizontal vorticity profile (Fig. 6f) would help promote updrafts that are erect rather than tilted leading to an increased probability of the initiation of convection. A plume of moisture can be identified in the mixing ratio data at the leading edge of the dryline from the sounding at 2122 UTC similar to the results shown on 10 June (Fig. 9c). The initiation of convection was aided by the southwesterly flow between the two boundaries that resulted in enhanced convergence along the dryline as well as high CAPE and low CIN values (Fig. 9f).

Favorable conditions for the thunderstorm development on the six days studied in this paper might be expected to be both a function of the height of the LFC and the CIN estimated from a sounding representative of the rising air parcels. The plot of LFC versus CIN (Fig. 10a), however, does not suggest any obvious relationship between these two variables. The plot of CAPE versus CIN (Fig. 10b) suggests that days when convection initiates are associated with higher CAPE values.

Figure 11 depicts the adiabatic frontogenesis for five of the cases examined in this paper calculated from the equation in two dimensions as follows:

\[
F = \frac{d}{dt} \left( -\frac{\partial \theta}{\partial y} \right) = \frac{\partial w}{\partial y} \frac{\partial \theta}{\partial y} + \frac{\partial w}{\partial \zeta} \frac{\partial \theta}{\partial \zeta},
\]

where the term \((\partial u/\partial y)(\partial \theta/\partial x)\) has been assumed to be small and is therefore neglected. This assumption was partially supported by examining the dual-Doppler wind syntheses. The gradient of \(u\) along the boundary was very small. Frontogenesis could not be calculated on 11 June owing to the missing wind data on several of the dropsondes. The \(y\) direction used in the frontogenesis calculations is noted in each panel in Fig. 11 and points toward the denser air mass of the primary boundary sampled by ELDORA. Caution must be used in interpreting positive and negative values of frontogenesis when the leading edge of two air masses are in juxtaposition as was the case on several days examined during IHOP 2002. For example, the cold front and the dryline on 10 June were both experiencing frontogenesis at low levels. The negative values accompanying the dryline (Fig. 11) is owing to the direction of the \(y\) axis chosen to point into the postcold frontal air mass.
The two terms in Eq. (1) represent the frontogenetical effects of confluence and tilting, respectively. The plots shown in Fig. 11 reveal that frontogenesis was occurring on all of the days analyzed during the experiment and does not appear to be a distinguishing factor on days when convection initiates versus null cases. The largest values of frontogenesis at low levels occurred on 22 May when no storms formed along the dryline.

The wind data from the dropsondes was used to estimate the horizontal vorticity along an axis parallel to the convergence boundary (Fig. 12). Missing wind reports prevented a similar analysis on 11 June. The spacing of the sounding data is insufficient to resolve the detailed rotor circulations identified in the Doppler wind syntheses shown in Fig. 5. The vertical velocity was calculated using the sounding wind field and assuming no variation in the along-boundary direction is also plotted on the figure. Two counter-rotating circulations at low levels are apparent in the cross sections on 10 and 19 June (the circulations on 10 June were not evident in higher-resolution analysis presented in Fig. 6f). The axes of the updraft on these two days are positioned between these two circulations suggesting that these rotations would promote air parcels that would rise vertically. The other three days do not exhibit an obvious pattern that would lead to a similar conclusion.

The horizontal vorticity equation can be simplified as follows:

\[
\frac{D\xi}{Dt} = \frac{\partial B}{\partial y},
\]

where \( B \) is the buoyancy and the flow is assumed frictionless and there are no along-frontal variations. It is also assumed that tilting of vorticity is small as shown by Sipprell and Geerts (2007) and Buban et al. (2007). The effects of stretching were calculated (not shown) and were much smaller than the solenoid effects consistent with the results from Ziegler et al. (1995).

The horizontal gradient of buoyancy \( \frac{\partial B}{\partial y} \) superimposed onto the vertical velocity is presented in Fig. 13. These analyses suggest that the solenoidal effects within the boundary layer associated with adjacent air masses were supporting the generation of counter-rotating circulations on a larger scale (i.e., meso \( \beta \)) than shown in Fig. 6. These circulations could lead to upright vertical motion on 24 May, 10 June, and 19 June and an increased probability of thunderstorm development. Simple calculations show that a value of horizontal buoyancy gradient of \( \pm 1.5 \times 10^{-6} \text{ s}^{-2} \) would require \( \sim 44 \text{ min} \) to generate \( \pm 4 \times 10^{-3} \text{ s}^{-1} \) horizontal vorticity. This is a reasonable amount of elapsed time for the development of low-level circulations with magnitudes comparable to those shown in Figs. 6f and 12. It is likely that the computed horizontal buoyancy gradients are underestimating the values near the boundary on a smaller scale. The null cases (22 May, 11 June, and 12 June) were characterized by low-level vorticity generation of only one sign, which would support tilted updrafts. It should be noted that there were adjacent air masses on 11 June; however, unlike the cases on 24 May, 10 June, and 19 June, the solenoidal effects did not support the generation of two counter-rotating horizontal circulations (Fig. 13).

The model of a “balance” of horizontal vorticity across a density current on the meso-\( \gamma \) scale advanced by Rotunno et al. (1988) assumes that baroclinically generated horizontal vorticity along a boundary is opposite to the horizontal vorticity associated with the line-normal wind.
shear over the depth of the cooler air mass (the vertical cross sections shown in Figs. 9c,f reveal no substantial line normal shear in the ambient air on 10 and 19 June). The present case suggests that a similar situation can occur when two boundaries are adjacent to each other at the meso-β scale. This may not be a rare event since it occurred in the three most prominent cases of convection initiation during IHOP_2002. It should be noted that the 24 May event was associated with the collision of two boundaries. This has been a well-documented scenario that can lead to

Fig. 11. Vertical cross section through the convergence boundary on 22 May, 24 May, 10 Jun, 11 Jun, 12 Jun, and 19 Jun. Frontogenesis \([F; \text{black lines, } K (100 \text{ km})^{-1} (3 \text{ h})^{-1}]\) and virtual potential temperature (gray lines, K). The cases are presented such that the days when storms (no storms) developed in the analysis region are plotted on the left-hand (right-hand) side of the figure. Missing wind reports on several soundings prevented a frontogenesis analysis on 11 Jun. The direction of the y axis used in the adiabatic frontogenesis calculations is shown.
the development of intense convection (e.g., Droegemeier and Wilhelmson 1985; Wilson and Schreiber 1986; Harrison et al. 2009). Convection on 24 May also initiated well to the south of the triple point (Fig. 1b), where the effects of the collision would not be felt. It is not known what caused convection to develop in this region.

The cases on 10 and 19 June, however, were characterized by two convergence boundaries that were approximately parallel and did not collide (Friedrich et al. 2008a,b; Murphey et al. 2006). The two boundaries appeared to be as close as 5–10 km on 10 June (Friedrich et al. 2008a,b). The separation between the cold front and the dryline on 19 June was difficult to accurately measure since the front was not associated with a thin line or sharp discontinuity in the radial velocity pattern. However, it is believed that the minimum separation distance between these two boundaries was 5–15 km although Sipprell and Geerts (2007) suggest that it could have been larger (30–50 km).

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**FIG. 12.** As in Fig. 11, but for horizontal vorticity ($\xi$) along an axis parallel to the convergence boundaries (black lines, $10^{-3}$ s$^{-1}$) and vertical velocity (gray lines, cm s$^{-1}$). Values of vertical velocity $>$12 cm s$^{-1}$ are hatched. Values of horizontal vorticity $>$8 $\times$ 10$^{-3}$ s$^{-1}$ or $<$−8 $\times$ 10$^{-3}$ s$^{-1}$ are shaded gray. Vertical velocities were calculated using the sounding wind field and assuming no variation in the along-front direction.
6. Summary and discussion

An analysis of six convergence boundaries observed during IHOP_2002 was presented. Convection initiated along the boundaries on three of the days. Dual-Doppler wind syntheses allowed for a detailed comparison of the kinematic structure of all of the boundaries. The horizontal profile of the radar reflectivity through the boundary at low levels revealed that the peak echo values were, in general, greater in May than in June suggesting a seasonal dependence. There was no apparent relationship between the intensity of the low-level updraft and

\[ \frac{\partial B}{\partial y} \times 10^6 \text{ s}^{-2} \]

Fig. 13. As in Fig. 12, but for horizontal gradient of buoyancy \((\partial B/\partial y)\) perpendicular to the dryline (black lines, \(10^{-6} \text{ s}^{-2}\)) and vertical velocity (gray lines, cm s\(^{-1}\)). Vertical velocities were calculated using the sounding wind field and assuming no variation in the along-front direction. Gray arrows represent the generation of horizontal vorticity as a result of the solenoid effects.
convection initiation at the meso-γ scale. The strongest updraft was associated with a null case. There was also no relationship between the strength of the moisture discontinuity across the boundaries and convection initiation. The three days when storms developed were associated with high CAPE values. The CIN and LFC values did not appear to be distinguishing factors on these six days.

The three days that storms developed were all associated with two convergence boundaries that were adjacent to each other. The two boundaries collided on the one of the days; however, the boundaries on the other two days were approximately parallel and remained separated by a small distance (5–15 km) although the separation on 19 June could have been greater. The total derivative of the horizontal vorticity rotating along an axis parallel to the boundary was calculated using dropsonde data. The horizontal gradient of buoyancy was the largest contributor to the change in vorticity and revealed maximum and minimum values that would support the generation of counter-rotating circulations that would promote vertically rising air parcels at the meso-β scale. These updrafts would be more conducive to convection initiation. The null cases were characterized by low-level vorticity generation of only one sign. This pattern would support tilted updrafts. An important remaining question is how often are two boundaries parallel to each other and in close proximity during convection initiation as was observed for the IHOP_2002 cases presented in this paper.

The schematic model proposed by Rotunno et al. (1988) illustrating the balance of horizontal vorticity across a density current at the meso-γ scale. The model assumes that baroclinically generated horizontal vorticity along a boundary is opposite to the horizontal vorticity associated with the line-normal wind shear at low levels. The vorticity produced by the low-level wind shear is replaced by solenoidally generated vorticity created by the adjacent convergence boundary. Convection initiated along the boundary that experienced enhanced convergence owing to the surface flow of the intervening air mass.

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APPENDIX

Radar Methodology

The radar data was edited and the aircraft motion was removed from the velocity fields using the SOLO II
software package (Oye et al. 1995). The data was then corrected for navigational errors using a technique developed by Testud et al. (1995). The along-track and sweep-angle resolution for ELDORA during IHOP_2002 was 550 m and 1.5°, respectively, based on the information presented in Table A1. This led to an effective sampling in the vertical of 250 m at a distance of 10 km from the radar. The data for all six case studies were subsequently interpolated onto a grid with a horizontal and vertical grid spacing of 600 and 300 m, respectively. A Cressman filter (Cressman 1959) was applied during the interpolation process with a radius of influence of 600 m in the horizontal and 450 m in the vertical. The lowest level was chosen to be 400 m AGL.

The synthesis of the radar data was performed using the Custom Editing and Display of Reduced Information in Cartesian Space (CEDRIC; Mohr et al. 1986). A three-step Leise filter (Leise 1982) was applied to the Doppler wind field. This type of filtering removes wavelengths less than 4.8 km. No correction for hydrometeor fall speeds was applied since the echo returns were collected in the clear air and were relatively weak. Vertical velocities were derived from an upward integration of the horizontal convergence field using the anelastic continuity equation. Errors in the calculation of vertical velocity accumulate using upward integration and are expected to be largest at the top. An estimate of the vertical velocity below the lowest grid level is based on the scheme proposed by Nelson and Brown (1987). The maximum errors associated with the vertical velocities are estimated to be less than 1–2 m s⁻¹ (Wilson et al. 1994). These errors are significantly reduced in the mean vertical cross sections that were created. All wind fields presented in this paper are ground relative.

The vertical structure of the convergence boundaries was reconstructed by averaging dual-Doppler cross sections for each flight leg. The number of cross sections varied by case study but ranged between 80 and 110. These mean cross sections needed to account for the nonlinear nature of the fine line. Accordingly, adjustments were made to the individual cross section so that the maximum radar reflectivity associated with the thin line was positioned in the same location before averaging the data.

Therefore, the kinematic features shown in the figures are presented in a thin-line relative frame of reference.

References


