Analysis of a Dryline during IHOP: Implications for Convection Initiation

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ABSTRACT

A detailed analysis of a dryline that formed on 22 May 2002 during the International H2O Project (IHOP) is presented. The dryline was classified as a null case since air parcels lifted over the convergence boundary were unable to reach the level of free convection preventing thunderstorms from forming. A secondary dryline associated with a distinct moisture discontinuity developed to the west of the primary dryline. The primary dryline exhibited substantial along-frontal variability owing to the presence of misocyclones. This nonlinear pattern resembled the precipitation core/gap structure associated with cold fronts during one of the analysis times although the misocyclones were positioned within the gap regions. Radar refractivity has been recently shown to accurately retrieve the low-level moisture fields within the convective boundary layer; however, its use in forecasting the initiation of convection has been restricted to qualitative interpretations. This study introduces the total derivative of radar refractivity as a quantitative parameter that may improve nowcasts of convection. Although no storms developed on this day, there was a tendency for maxima of the total derivative to be near regions where cumulus clouds were developing near a convergence boundary.

1. Introduction

One of the major challenges during the summer months is predicting the initiation of deep, moist convection (Olsen et al. 1995; Fritsch and Carbone 2004). Forecasters often have difficulty assessing, under weak synoptically forced conditions, the potential of rising parcels of air to reach the level of free convection (LFC) and then maintaining positive buoyancy as they pass this level. Improvements in these short-term forecasts, however, resulted when organized lines of convergence that form in the boundary layer were successfully detected using remote sensing techniques (e.g., Purdom 1976, 1982; Wilson and Schreiber 1986; Wilson and Mueller 1993; Wilson et al. 1998). Purdom (1982) provided evidence that 93% of the storms forming over the southeastern United States were associated with arc clouds that were associated with convergence lines. Wilson and Schreiber (1986) showed that a majority of thunderstorms (80%) developed near boundary-layer convergence zones. In spite of these major findings, it is also known that the presence of a prominent convergence line is not a sufficient condition for convection to develop (e.g., Rotunno et al. 1988; Stensrud and Maddox 1988; Crook 1996; Ziegler and Rasmussen 1998; Crook and Klem 2000; Richter and Bosart 2002; Cai et al. 2006; Markowski et al. 2006).

Stensrud and Maddox (1988), Richter and Bosart (2002), and Cai et al. (2006) have shown that larger-scale descent can prevent convection initiation even when large convective available potential energy (CAPE) and conditionally unstable environments exist. Rotunno et al. (1988) have proposed that the balance of horizontal vorticity generated by the ambient low-level shear with the baroclinically produced vorticity owing to the cold pool to the rear of a boundary can alter the orientation of the updraft. Vertically erect updrafts might be more likely to produce deep convection. Ziegler and Rasmussen (1998) have suggested that air parcels that are lifted from the boundary layer must reach the lifted condensation level (LCL) and the LFC prior to leaving the mesoscale updraft to form deep convection.
Additional challenges have been suggested based on results from numerical simulations. Lee et al. (1991) and Crook (1996) have shown that relatively small changes in mixing ratio (e.g., 1 g kg\(^{-1}\)) can have a significant effect on the developing convection. These findings are challenging to address since comprehensive measurements of moisture have been difficult to collect in field studies. Moreover, variations of moisture over distances of only a few kilometers, as suggested by the modeling studies, have been shown to commonly occur within the convective boundary layer (e.g., Weckwerth et al. 1996; Weckwerth 2000; Murphey et al. 2006).

The present study focuses on a dryline that developed on 22 May 2002 during the International H\(2\)O Project (IHOP; Weckwerth et al. 2004). The dryline is a convergence line that frequently forms over the southwestern United States as hot dry air from the Mexican plateau collides with moist air flowing northward from the Gulf of Mexico. Its association with deep (severe) convection has been noted for a number of years (e.g., Rhea 1966; Schaefer 1974, 1986; Bluestein and Parker 1993). The dryline on 22 May was well sampled by a number of platforms during IHOP and has been extensively examined (Demoz et al. 2006; Weiss et al. 2006; Buban et al. 2007). The boundary that developed on this day also generated interest since it was a null case; that is, a boundary that was not associated with the initiation of convection.

A unique aspect of this study is an airborne Doppler radar that was used to collect kinematic data along this frontlike boundary. The aircraft flew a box pattern around the dryline with the along-boundary legs ~100 km long. These long legs allowed for an accurate assessment of both the along-frontal variability and the mean characteristics of the dryline. Detailed observational studies have already shown that substantial variability along convergence boundaries can exist (e.g., Kingsmill 1995; Friedrich et al. 2005; Arnott et al. 2006; Markowski et al. 2006; Murphey et al. 2006). Demoz et al. (2006) provided high-temporal resolution data of the vertical structure of this dryline but it was primarily focused on measurements made from a single site. Weiss et al. (2006) analyzed the vertical structure of the dryline with unprecedented detail but only in a plane normal to the boundary. Buban et al. (2007) examined a 30 km \(\times\) 30 km section of the dryline using ground-based Doppler radars. Their study was largely based on a Lagrangian analysis of thermodynamic variables. The analysis depends on advecting in situ data collected along one-dimensional transects using parcel trajectories derived from the multi-Doppler radar analysis. Local conservation following the motion is also assumed (Ziegler et al. 2007).

Another aspect of this study is to examine further the utility of radar refractivity in providing thermodynamic information that can be used to improve convective forecasts. Radar estimates of refractivity can be derived by measuring the apparent fluctuations in the range of fixed ground targets. Variations of refractivity are primarily owing to changes in water vapor under typical warm summertime temperatures (Fabry et al. 1997; Fabry 2004, 2006). Weckwerth et al. (2005) have shown the excellent agreement between radar refractivity and surface moisture gradients; however, most past studies have used refractivity in a qualitative manner for improving forecasts (e.g., Cai et al. 2006; Demoz et al. 2006; Fabry 2006; Buban et al. 2007).

Section 2 discusses Electra Doppler radar (ELDORA) and S-band dual-polarization Doppler radar (S-Pol), which are two of the primary platforms used in this study. Surface and ground-based radar analyses are presented in section 3. Vertical cross sections based on a series of soundings through the dryline and an airborne Doppler radar analysis are shown in sections 4 and 5, respectively. A summary and conclusions are presented in section 6.

2. ELDORA and S-Pol

One of the primary platforms used in the study is a 3-cm airborne Doppler radar (ELDORA) operated by the National Center for Atmospheric Research (NCAR) and flown on board a Naval Research Laboratory (NRL) P-3. ELDORA is equipped with two antennae that scan fore and aft of the normal to the aircraft. Data collected by the radar can be synthesized into a dual-Doppler windfield using the fore–aft scanning technique (FAST; Jorgensen et al. 1996). Critical to this study was ELDORA’s ability to detect echoes and Doppler velocities within the clear air (Wakimoto 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). Research flight plans required the P-3 to fly at low levels (~400 m above ground level (AGL); hereafter, all heights are AGL except where indicated) and parallel to the thin lines. The thin line was positioned within 2–3 km from the aircraft and resulted in a flight track that resembled a box-type pattern ~100 km long. The radar scanning parameters used during IHOP are shown in Table 1 and a discussion of the radar methodology is presented in the appendix. For more information regarding ELDORA’s hardware and design, the interested reader is referred to Hildebrand et al. (1994, 1996).
Measurements of radar refractivity during IHOP were obtained using S-Pol, an S-band multiparameter radar (Lutz et al. 1995). Refractivity values can be derived by identifying ground targets (e.g., towers, buildings, poles) that surround the radar site and quantifying the small variations in the return phase from stationary targets caused by changes in the index of refraction in the intervening atmosphere (Fabry et al. 1997; Fabry 2004). High correlation between radar refractivity and refractivity calculated from low-flying aircraft and surface mesonets was shown by Weckwerth et al. (2005). Buban et al. (2007) showed close agreement between two-dimensional water vapor fields derived from radar refractivity with analogous fields calculated from in situ data. The maximum range for retrieved estimates of refractivity from S-Pol data collected during IHOP was between 40 and 60 km.

3. Surface and radar analyses

An upper-level trough was situated over the western United States at the 500-mb level (not shown). At the surface, there was a cyclone over southeastern Colorado resulting in southerly flow over the south-central United States (Fig. 1). These conditions are typical for a well-defined dryline to develop (e.g., Schultz et al. 2007). The dryline is located at the leading edge of a gradient of mixing ratio and a convergence boundary in the surface wind-field analysis at 2100 UTC over the Oklahoma and Texas panhandles (Fig. 1).

Surface analyses at 2000, 2100, 2200, and 2300 UTC superimposed on visible satellite images are presented in Fig. 2. A dryline in Fig. 2a was determined by a combination of the surface mixing ratios and also placing the boundary near the western edge of the cumulus cloud field that was apparent in the satellite image. The beginning of the flight track flown by the P-3 is shown by the dotted line (Fig. 2a). This westbound flight leg was intended to provide an initial survey of the near surface conditions in order to identify any convergence boundaries in the region. There were several fine lines (i.e., enhanced lines of radar reflectivity) seen in the surveillance scan of S-Pol radar reflectivity that are highlighted by the black arrows in Fig. 3a. The middle fine line was located on the eastern edge of a dry air mass as shown by the mixing ratio values and the low values of radar refractivity (N values < 255). The eastern fine line is diffuse; however, it denotes the incipient stages of the primary dryline that formed on this day. The western fine line is not a focus in this study and becomes less defined with time. An enlargement of Fig. 3a is shown in Fig. 4a. There is a drop in mixing ratio (>6 g kg\(^{-1}\)) from just east of S-Pol to the western edge of the region shown in Fig. 4a. This horizontal gradient of moisture is also apparent in the analysis of N (color plot and isopleths).

Surface confluence across the dryline increases at 2100 UTC as the winds began to back within the moist air (Fig. 2b). The moisture discontinuity associated with the primary dryline also increases in response to this enhanced confluence based on the isopleths of N (Fig. 4b). Another boundary (denoted by a dashed line) has been drawn on the analysis at 2100 UTC (Fig. 2b) and is indicative of the formative stages of a second dryline (also noted by Buban et al. 2007). At this time, the secondary dryline is accompanied by a weak horizontal gradient of moisture in the N-field in the western section in Fig. 4b.

Miao and Geerts (2007), however, showed that no measurable virtual potential temperature gradient existed across the secondary boundary. Weiss et al. (2006) also

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<th>TABLE 1. ELDORA scanning mode.</th>
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<td>Antenna rotation rate (° s(^{-1}))</td>
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<td>Number of samples</td>
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<td>Pulse repetition frequency (PRF) (Hz)</td>
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estimated that the secondary dryline formed around this time. The existence of double drylines has been well documented in the literature (e.g., Hane et al. 1997; Crawford and Bluestein 1997; Hane et al. 2001). It should be noted that the drier air that is suggested by the low refractivity values in the southeast corner of Fig. 4b is believed to be in error (e.g., Buban et al. 2007; also compare this plot with the one shown in Fig. 4a). In situ data collected at flight level (to be discussed later) reveals that the southeastern air mass is still moist. Possible reasons for these errors include nonstationary targets and swaying vegetation (Fabry 2004; Weckwerth et al. 2005). These errors are transient and not predictable based on known meteorological features (Weckwerth et al. 2005; Buban et al. 2007).

Cumulus clouds were prevalent throughout the triangular region bounded by the two drylines at 2200 UTC (Fig. 2c). The fine line associated with the primary dryline has become more prominent (Fig. 3c; note the eastern arrow) and in the enlargement (Fig. 4c), which is consistent with the presence of strong surface convergence (Wilson et al. 1994). Dry air now extends much further to the east by comparing the positions of the $N = 254$ isopleth at 2100 and 2200 UTC (Figs. 4b,c). The
eastern edge of this dry air is the location of the secondary dryline and can be identified in the reflectivity pattern as a developing fine line (the western arrow in Fig. 3c). The P-3 aircraft was in the early stages of flying a box pattern around the primary dryline as shown by the dotted line in Fig. 4c (also shown in Figs. 2c, 3c).

The cumulus cloud activity between the two drylines is more prominent in the satellite imagery at 2300 UTC (Fig. 2d). In addition, the fine lines associated with the primary and secondary drylines are easily identifiable in the S-Pol surveillance scans (Figs. 3d, 4d). The intersection of the drylines, which resembles an inverted

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**Fig. 3.** Surface observations for (a) 2000, (b) 2100, (c) 2200, and (d) 2300 UTC 22 May 2002 superimposed onto (left) radar reflectivity and (right) radar refractivity images. Virtual potential temperature (K) and mixing ratio values (g kg\(^{-1}\)) are plotted. The track of the P-3 is shown by the dotted line. The black arrows in (a) and (b) denote the positions of three fine lines in radar reflectivity. Sounding locations are shown on the plots at 2200 and 2300 UTC. The region denoted by the gray square is enlarged in Fig. 4. The black rectangle at 2200 and 2300 UTC represent the location of the dual-Doppler analysis shown in Fig. 8. Wind vectors are plotted using the following notation: barb = 5 m s\(^{-1}\), half barb = 2.5 m s\(^{-1}\). Color scales for dBZ and N values are shown. Gray areas for refractivity denote values of N less than 245.
V-shape, is located ∼30 km northeast of S-Pol and is shown in the refractivity analysis where the two horizontal gradients of the $N$-field merge (Fig. 4d). Further confirmation that the dry air located in the southeastern section of the refractivity field in Fig. 4d is erroneous was provided by the in situ data collected at flight level by the P-3. The flight-level data collected east of the boundary shown in the figure revealed moist air in this region (not shown). The primary dryline regresses by a few kilometers by 0000 UTC; however, the secondary dryline has regressed by ∼20 km at this time (Fig. 4e). As a result, the intersection point of the two drylines has moved farther to the north.

4. Vertical cross section based on sounding data

A series of dropsondes deployed by a jet flying at ∼500 mb between 2215 and 2233 UTC was combined with three balloon soundings released at 2150, 2235, and 2247 UTC to produce a vertical cross section oriented
The dryline was believed to be nearly stationary and evolving slowly during this time period. Examination of the sounding spacing in Figs. 4c,d suggests that it would be difficult to distinguish the primary and secondary drylines in the vertical cross section since the boundaries are close to each other in this region and are of order 0.5–1 km in width (Buban et al. 2007; Miao and Geerts 2007).

Fig. 4. Surface analysis for (a) 2000, (b) 2100, (c) 2200, (d) 2300 UTC 22 May 2002 and (e) 0000 UTC 23 May 2002 superimposed onto (left) radar reflectivity and (right) radar refractivity images. Black lines superimposed on both plots are contour values of $N$. The track of the P-3 is shown by the dotted line. Virtual potential temperature (K), mixing ratio values (g kg$^{-1}$), and radar refractivity ($N$) are plotted. The location of S-Pol is shown by the gray star. The location of the analysis region is shown by the gray box in Fig. 3. Wind vectors are plotted using the following notation: barb = 5 m s$^{-1}$, half barb = 2.5 m s$^{-1}$. Color scales for dBZ and $N$ values are shown. Gray areas for refractivity denote values of $N$ less than 245.
Indeed, the virtual and equivalent potential temperature analyses (Fig. 5) reveal a single maritime air mass extending east of the dryline.

The head of the maritime air mass reaches ~620 mb near the dryline but the strong capping inversion is lower and located at 790 mb approximately 100 km to the east (Fig. 5). Time series of S-Pol surveillance scans suggest that the primary dryline was regressing slowly to the west during the time that the soundings were deployed (this westward movement was also shown by Buban et al. 2007). The analysis of virtual potential temperature suggests that cooler air is being modified by surface heating as it moves westward, producing a superadiabatic layer at low levels. The top panel in Fig. 5 is
a plot of the calculated CAPE and convective inhibition (CIN). The CAPE values were low to the west of the dryline but rose within the moist air until reaching a maximum of \( \sim 1300 \) J kg\(^{-1}\) for the 2220 UTC sounding. The lowest CIN value was 19 J kg\(^{-1}\) (sounding at 2215 UTC) but was at a location 30–40 km east of the dryline. The dotted line in the middle panel of Fig. 5 identifies the height of the LFC. The lowest level for the LFC was calculated to be \( \sim 615 \) mb for the 2215 UTC sounding. Accordingly, the stability analysis suggests that it would be difficult to initiate convection without significant lifting provided by a convergence boundary.

Figure 6 shows the adiabatic frontogenesis calculated from the equation in two dimensions as follows:

\[
F = \frac{d}{dt} \left( -\frac{\partial \theta}{\partial y} \right) = \frac{\partial v}{\partial y} \frac{\partial \theta}{\partial y} - \frac{\partial \omega}{\partial y} \frac{\partial \theta}{\partial p}, \tag{1}
\]

where the term \((\partial u/\partial y)(\partial \theta/\partial x)\) has been assumed to be small and is therefore neglected. The \(y\) direction points to the denser, moist air (see inset in Fig. 7). The two terms represent the frontogenetical effects of confluence and tilting, respectively. Past studies on drylines have calculated frontogenesis based on the moisture fields rather than temperature (e.g., Anthes et al. 1982; Buban et al. 2007). Both the frontal boundary and maritime air mass to the east were well defined in the present case. Accordingly, Eq. (1) was deemed appropriate to use to determine the frontogenetic effects. Separate frontogenesis calculations were made using the virtual potential temperature and mixing ratio fields instead of potential temperature. In both cases, the results were similar and would not have changed the conclusions reached in this paper.

Low-level convergence produced frontogenesis below 800 mb with a maximum \(>8 \) K (100 km\(^{-1}\) (3 h)\(^{-1}\) near the ground (Fig. 6a). A region of frontolysis followed by frontogenesis further to the east exists aloft between 700–800 mb. This couplet was created by the tilting term owing to descent with a minimum located at the top of the mixed layer \(\sim 100 \) km to the east of the dryline (Fig. 6b). Updrafts at the leading edge of the dryline can also be identified in Fig. 6b. The diabatic effect on frontogenesis was qualitatively estimated by analyzing the low-level flight level data collected by the P-3 on either side of the dryline. The dry air was warming relative to the moist air during this period. Accordingly, inclusion of the diabatic term would have augmented the frontogenesis across the dryline.

5. Airborne dual-Doppler analysis

a. Horizontal structure

As previously noted, the segment of the dryline that was targeted by the P-3 was located in eastern Oklahoma and northern Texas panhandles (Figs. 2, 3, and 4). The aircraft initially performed an east–west survey pattern.
before initiating a series of flight transects oriented approximately 100°–190° (Fig. 7). The northbound (southbound) legs were intended to fly on the dry (moist) side of the primary dryline. These legs were ~100 km in length in order to collect kinematic data that would reveal the along-frontal variability and the mean structure of the boundary. The aircraft collected Doppler radar data on the boundary for ~3 h. The tracks in Fig. 7 show
that the main dryline did not move substantially during the period under investigation.

1) 2148–2203 UTC

The first 100-km leg flown by the P-3 was used to determine the orientation of the boundary and to position the aircraft as close as possible to the dryline. The second leg was flown between 2148 and 2203 UTC when the aircraft was flying toward the south, parallel and to the east of the dryline (Fig. 8a). The airborne Doppler radar was able to detect clear air returns out to a range of 10–15 km on this day. The thin line associated with the dryline was associated with substantial variability with pockets of maximum reflectivity ~8 dBZ. Updrafts >2 m s⁻¹ are apparent along the northern segment of the dryline and appear to develop near pockets of high reflectivity as shown by Christian and Wakimoto (1989) and Wilson et al. (1994). The southerly flow was not associated with an abrupt shift in wind direction across the dryline even though the horizontal convergence was strongest along the boundary (not shown).

Also apparent in Fig. 8a are maxima in vertical vorticity or misocyclones. Misocyclones developing along convergence boundaries have been noted by a number
of investigators (e.g., Carbone 1982; Mueller and Carbone 1987; Wakimoto and Wilson 1989; Kingsmill 1995; Markowski and Hannon 2006; Arnott et al. 2006; Marquis et al. 2007). One theory for the development of these circulations is horizontal shearing instability (e.g., Miles and Howard 1964; Carbone 1983; Lee and Wilhelmson 1997); however, Marquis et al. (2007) have cautioned that it can be difficult to verify this initiating mechanism if misocyclones exist at the beginning of the radar observations. Another possible theory is that the intersection of horizontal convective rolls (HCRs) with a convergence boundary can produce small-scale misocyclones (e.g., Wilson et al. 1992; Atkins et al. 1995). Buban et al. (2007) have suggested that the latter mechanism may have produced misocyclones on this day and speculate that strong vertical shear-induced horizontal vortex lines could be tilted up into the base of the misocyclone. Buban et al. (2007) also suggest that the core vortex line in a misocyclone upwells from the near-ground level. There have been examples shown in the literature when updrafts were located immediately upstream or downstream from misocyclones (e.g., Kingsmill 1995; Murphey et al. 2006). However, there have been other studies that have shown no general relationship suitable for all misocyclone sizes (e.g., Marquis et al. 2007). No obvious relationship between the updrafts and the vorticity extrema can be identified in Fig. 8a. The maximum vertical vorticity values associated with the misocyclones reported in the present study are much weaker than those shown by other investigators using ground-based radars (e.g., Arnott et al. 2006). These differences are primarily owing to the higher spatial resolution afforded by the latter platform.

The radar refractivity plot reveals two regions characterized by strong moisture gradients that were also noted in Fig. 4 (black arrows in Fig. 8b). The western and eastern gradients are associated with the secondary and primary drylines, respectively. As previously mentioned, the apparent drying revealed in the refractivity plot near the southern part of the aircraft track is false. Buban et al. (2007) also noted this false drying located southeast of S-Pol as well as a dry bias in the northern region (see Fig. 4c). Flight-level data (not shown) collected at low levels did not show a decrease in mixing ratio from 2200 to 2204 UTC while the aircraft was flying on the eastern side of the primary dryline. As a result, the questionable refractivity data were removed during the editing process. The secondary dryline was not associated with a fine line at this time, consistent with the radar observations shown in Fig. 4. Weckwerth et al. (2005) suggest that refractivity gradients instantaneously reflect moisture gradients while a fine line in radar reflectivity requires that a concentration of insects accumulate within the convergence zone before they are detected by the radar (e.g., Achtemeier 1991).

The locations of cumulus clouds were determined by carefully superimposing visible satellite images at high resolution onto the radar reflectivity fields at 700 and 2500 m (Figs. 8c,d). The former and latter reveal the relationship of the clouds with the fine line and echoes aloft, respectively. The clouds are developing over the fine line associated with the primary dryline. Clouds are also developing in the triangular-shaped region between the two drylines (also shown in the satellite images; Fig. 2c). The echoes aloft (Fig. 8d) are a result of echo plumes that are associated with updrafts that form within the convective boundary layer (e.g., Wakimoto et al. 2004; Miao et al. 2006). Several of the cell-shaped echoes aloft are accompanied by clouds, which is consistent with past studies (e.g., Christian and Wakimoto 1989; Knight and Miller 1993, 1998; Arnott et al. 2006; Markowski et al. 2006; Buban et al. 2007; Ziegler et al. 2007). It is possible that some of the clouds are not being detected at this time since they are smaller than the scales resolvable in the satellite image.

The combination of the high temporal resolution of refractivity data recorded by S-Pol and the dual-Doppler wind synthesis provides a unique opportunity to estimate the total derivative of radar refractivity ($\frac{DN}{Dt}$). Positive (negative) values of $\frac{DN}{Dt}$ would suggest moistening (drying) of an air parcel. Accordingly, this quantity...
may have utility for nowcasting of convection initiation. Fabry (2006) estimated the local change of $N$ (see his Fig. 2). Large changes in this quantity, however, could be attributed to the movement of the boundary and may not significantly improve a forecaster’s ability to determine when convection might initiate.

The total derivative was calculated by determining local changes of $N$ and the advective term ($-\mathbf{V} \cdot \nabla N$). The winds based on the dual-Doppler synthesis and the distribution of $N$ from S-Pol surveillance scans were used to estimate the advective terms. The local tendency of $N$ was estimated by comparing different surveillance scans of refractivity. The elapsed time between scans was chosen to be the same as the dual-Doppler analysis period which varied between 9 and 15 min. The plot of $DN/Dt$ is shown in Fig. 8d. The dash–dot line in the figure denotes the limits of the analysis area. There is a tendency for maxima of $DN/Dt$ to be near regions where clouds were developing during this analysis time. Surface fluxes of moisture or mixing are examples of processes that could result in extrema of $DN/Dt$ following a near-surface parcel. These moist air parcels could highlight favorable regions for cloud development if they were collocated with regions characterized by positive vertical velocities. In this regard, the close proximity of the $DN/Dt$ maxima to a convergence line would be consistent with the visual appearance of clouds.

Fabry et al. (1997) and Weckwerth et al. (2005) have shown that at a temperature of 18°C a change in 1 N unit can be caused by either a temperature or dewpoint temperature change of 1° or 0.2°C, respectively. It is important to assess whether the values of $DN/Dt$ shown in Fig. 8d are reflective of moistening/drying versus temperature changes. The minimum contour of $DN/Dt$ plotted in Fig. 8d is $10 \times 10^{-3}$ s$^{-1}$. If it is assumed that the time interval between consecutive dual-Doppler wind syntheses is 10 min, then this minimum value translates to a change of $N$ of 6 units. This corresponds to either a temperature change of 6°C or a dewpoint temperature change of 1.2°C. The analysis shown in Figs. 8a, 9a, 10a, and 12a do not appear to support such large changes in temperature. However, it is consistent with the type of moisture gradients that are evident in the figure.

2) 2245–2254 UTC

The analysis of the flight leg flown from 2245 to 2254 UTC is shown in Fig. 9. The misocyclone pattern along the dryline has become better defined by this time (Fig. 9a). The refractivity gradient along the secondary dryline is more prevalent (Fig. 9b, highlighted by the western black arrow) and a second fine line has developed (Fig. 9a) as was shown in the S-Pol analysis presented in Fig. 4. The air mass between the two drylines is approximately represented by the refractivity values that are shaded green. Clouds have formed throughout this region between the drylines with the largest clouds located above the primary dryline (Fig. 9c). Cellular echoes at 2500 m are located above both fine lines (Fig. 9d). Many of these echoes are associated with clouds. The $DN/Dt$ field during this time is shown in Fig. 9d. There is good agreement between the cloud locations and the $DN/Dt$ maxima similar to the results shown during the previous analysis time.

3) 2321–2333 UTC

The reflectivity structure of the primary dryline developed a distinct along-frontal pattern during the 2321–2333 UTC flight leg. Striking in the analyses shown in Figs. 10a,c is the resemblance of the reflectivity pattern to the precipitation core/gap structure documented along cold fronts (e.g., Hobbs and Biswas 1979; James and Browning 1979). The angle between the dryline and the primary axes of the echo cores was ~20°. There is a strong tendency for the misocyclones to be positioned in the gap regions between the echo cores (Fig. 10a). This relationship suggests that the circulations are contributing to the observed reflectivity structure. It should be noted that past studies of cold fronts have shown that the vorticity maxima tend to be associated with the precipitation cores (Hobbs and Persson 1982; Wakimoto and Bosart 2000). Kawashima (2007) has argued that the collocation of the vorticity maxima with the precipitation cores (i.e., locations of maximum convergence) suggests that horizontal shear instability is not the likely mechanism for producing the variability observed along cold fronts even though it has

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Fig. 8. Airborne Doppler radar analysis for 2148–2203 UTC 22 May 2002. (a) Dual-Doppler wind syntheses, vertical velocity (red lines, m s$^{-1}$), and vertical vorticity (black lines, $10^{-3}$ s$^{-1}$) superimposed onto radar reflectivity at 700 m AGL. Surface stations are plotted with virtual potential temperature (K), mixing ratio (g kg$^{-1}$), and radar refractivity (N). (b) Dual-Doppler wind syntheses, vertical velocity (red lines, m s$^{-1}$), and vertical vorticity (black lines, $10^{-3}$ s$^{-1}$) superimposed onto radar refractivity at 700 m AGL. Black arrows represent the locations of the two drylines. Surface stations are plotted with virtual potential temperature (K), mixing ratio (g kg$^{-1}$), and radar refractivity (N). (c) Cloud locations based on visible satellite imagery superimposed onto radar reflectivity at 700 m AGL. Two sounding locations are indicated on the plot. (d) Cloud locations based on visible satellite imagery superimposed onto radar reflectivity at 2500 m AGL. Positive and negative values of total change of radar refractivity are plotted as solid red and dashed red lines, respectively. Dashed–dot line denotes the limits of the analysis. Flight track of the P-3 is shown by the dashed line.
FIG. 9. Same as Fig. 8, but for 2245–2254 UTC. (b) The western and eastern black arrows denote the radar refractivity gradients associated with the secondary and primary drylines, respectively.
Fig. 10. Same as Fig. 8, but for 2321–2333 UTC. (b) The black arrow denotes the location of a misocyclone that is described in the text.
been the prevailing theory that has been advanced by a number of investigators (e.g., Parsons and Hobbs 1983). Instead, horizontal shear instability should lead to forced convergence between vorticity extrema as has been shown by Lee and Wilhelmson (1997) and suggested by the analysis shown in Fig. 10. In addition to the reflectivity and misocyclone pattern, the along-frontal variability associated with the primary dryline can be shown in a time plot of the in situ measurement of the mixing ratio recorded at flight level (Fig. 11). There is a several grams per kilogram variation of mixing ratio as the aircraft flew parallel to and west of the dryline. The dry and moist air was predominantly located in the gap and echo core regions, respectively. There is also a tendency for the moist (dry) air to be located north (south) of the misocyclone, which is consistent with the results shown by Buban et al. (2007) and also shown for a different case during IHOP by Murphey et al. (2006). The black arrow in Fig. 10b indicates a misocyclone that is associated with a wave pattern in the refractivity field that suggests an advection of moisture in a manner consistent with the previous discussion.

The refractivity plot continues to delineate the positions of the two drylines (Fig. 10b) although the secondary dryline has retrograded since the previous analysis time. The secondary dryline is difficult to identify in the reflectivity field at low levels (Figs. 10a,c); however, it is prominent at higher levels (Fig. 10d). The opposite is true for the primary dryline (i.e., it is difficult to locate at 2500 m). These observations are also suggested in the results presented by Buban et al. (2007). This difference in echo structure in Fig. 10d suggests that the echo plumes/updrafts for the secondary dryline are stronger, which is supported by the more vigorous cloud development identified in the satellite image (Figs. 10c,d). There is a tendency for clouds to be located where the total derivative of $N$ is large; however, the relationship is not as apparent as in the previous two analysis times.

4) 2342–2357 UTC

The primary dryline was in the early stages of propagating to the west during the last transect by ELDORA (2342–2357 UTC). The retrogression of the line resulted in a more organized updraft structure and enhanced cloud development (Figs. 12a,b,c). The precipitation core and gap structure observed earlier is less apparent in the radar fine line even though the misocyclones are still prevalent along the entire boundary.

The secondary dryline also continued its westward movement during this time. The refractivity gradient along the secondary dryline exhibits a wavelike pattern (~15 km) after appearing more linear during the previous synthesis time (cf. Figs. 10b, 12b). This along-frontal variation in moisture is well matched with the echo structure at 2500 m. This can be seen with the superposition of the echoes with intensity >0 dBZ (shaded yellow) onto the refractivity plot (Fig. 12b). This is another example of the utility of radar refractivity in locating finescale details of moisture gradients associated with convergence boundaries and its relationship with cloud development. Clouds are, in general, positioned near regions of $\frac{DN}{Dt}$ extrema (Fig. 12d). It is also apparent that regions largely devoid of clouds are located in areas with weak positive or negative values of $\frac{DN}{Dt}$.

The usefulness of $\frac{DN}{Dt}$ fields to facilitate nowcasts of convection initiation will require further testing although the results presented here show promise even though no thunderstorms were observed on this day. There is a suggestion that $\frac{DN}{Dt}$ maxima near convergence lines (i.e., regions of enhanced positive vertical motion) could help pinpoint locations that are

![Fig. 11. In situ measurements of mixing ratio (red line) at flight level for 2321:30–2333:55 UTC superimposed on radar reflectivity data at 700 m AGL. Positive and negative values of vertical vorticity are shown as solid and dashed black lines, respectively.](image-url)
FIG. 12. Same as Fig. 8, but for 2342–2357 UTC. (d) The 0-dBZ contour for the echoes at 2500 m AGL have been superimposed onto (b) the refractivity plot and are shaded yellow.
favorable for storms to develop. Forecasters should be able to derive this variable even in the absence of dual-Doppler wind syntheses. Horizontal wind fields at low levels used to estimate advection can be approximated by tracking the movement of clear air echoes (e.g., Tuttle and Foote 1990); tracking echoes and adding a constraint using the radial velocity momentum equation (e.g., Laroche and Zawadzki 1995); or by combining meteorological observations, and a numerical boundary layer model and its adjoint (Sun and Crook 2001). In this regard, the $D\xi/Dt$ analysis is seen to be superior to combining the low-level convergence field with an $N$ analysis. The latter technique requires the identification of small pockets of convergence (i.e., updrafts) that can only be accurately resolved using a dual-Doppler wind synthesis. These detailed wind syntheses would often be unavailable to a forecaster.

b. Vertical structure

The long legs flown by the P-3 provided an opportunity to properly assess the along-frontal variability of the dryline. The voluminous amount of wind data collected during these transects were also used to construct mean vertical cross sections oriented perpendicular to the primary dryline. The horizontal grid spacing of the Doppler wind syntheses was 600 m, as described in the appendix. Although the number varied slightly for each analysis time, ~105 cross sections were averaged for each flight leg (Fig. 13). The mean structures shown in the figure should be largely devoid of the variability along the dryline.

The mean vertical cross section for the analysis shown in Fig. 8a is presented in Fig. 13a. The thin line echo can be seen in the top panel with maximum reflectivities greater than 4 dBZ. Peak updrafts over the dryline were greater than 1.5 m s$^{-1}$. There is a head structure with weak downdrafts to the east of the dryline. There is a slight veering of winds with increasing height through the maritime air mass. Mean cyclonic vorticity greater than $0.5 \times 10^{-3}$ s$^{-1}$ associated with the misocyclones extends up to 1 km. Shallow, easterly flow can be seen in the component of horizontal flow perpendicular to the dryline ($u'$) east of the dryline (Fig. 13a, bottom panel). Enhanced return flow can be seen above the maritime air mass. A rotor circulation generated by the solenoidal circulation is located east of the fine line. Miao and Geerts (2007) state that no circulation existed across the dryline during this time. This inconsistency is likely the result of the analysis of a single cross section in their study versus the average of numerous cross sections presented in this article. In addition, a rotor circulation is evident in the analysis presented by Weiss et al. (2006).

The structure during the period when the dryline was beginning to move westward is shown in Fig. 13b. The easterly flow within the maritime air mass has increased in magnitude and depth. The peak updrafts

FIG. 13. Mean vertical cross sections perpendicular to the dryline for (a) 2148:35–2203:14 and (b) 2342:00–2357:01 UTC 22 May 2002. (top) Radar reflectivity (dBZ) is plotted as gray lines with values greater than 2-dBZ shaded gray. Positive values of vertical vorticity ($10^{-3}$ s$^{-1}$) are plotted as thin black lines. Vertical velocities and component of horizontal flow in the plane of the cross section are plotted as black arrows. (bottom) Positive and negative values of vertical velocity (m s$^{-1}$) are plotted as black and dashed lines, respectively. Component of horizontal flow (m s$^{-1}$) perpendicular to the boundary is plotted. Positive (easterly) and negative (westerly) flow are shown by the gray and dashed gray lines, respectively. The rotated coordinate system is shown in the inset in Fig. 7.
have increased (>3 m s⁻¹) and shifted to the western edge of the fine line. The rotor circulation is better defined and deeper at this time, consistent with the results presented by Miao and Geerts (2007). Mean surface-based trajectories were calculated using the dual-Doppler wind syntheses from 2321 to 2333 UTC and 2342 to 2357 UTC (and a backward time difference scheme) by releasing air parcels at the lowest grid point in a line perpendicular to the dryline. The maximum height that surface-based parcels would attain (indicated by the black line in the middle panel in Fig. 5) is far below the height of the LFC suggesting that convection initiation would be unlikely on this day. In contrast, Buban et al. (2007) suggest parcel trajectories reaching a higher level. They showed that the air parcels from both west and east of the dryline were associated with insufficient CAPE to initiate convection even if they attained heights between 2.5 and 3.5 km. The present study used the wind field averaged along the length of the dryline in order to examine the mean lift along the boundary. This would underestimate the maximum heights of individual parcels that are captured in the study by Buban et al. (2007). In addition, the dual-Doppler syntheses do not include the three-dimensional winds from the surface to the top of the boundary layer, which would preclude deep trajectories. Surface-based CAPE estimates may also underestimate the maximum attainable parcel altitude versus a layer-averaged value at low levels.

The resemblance of a retrograding dryline to a density current has been documented before (e.g., Parsons et al. 1991; Ziegler et al. 1995; Murphey et al. 2006; Sipprell and Geerts 2006). The solenoidal-generated circulation is believed to be the primary mechanism for the generation of horizontal vorticity within the head of the density current (e.g., Sun and Ogura 1979; Ziegler et al. 1995; Sipprell and Geerts 2006; Miao and Geerts 2007). Sipprell and Geerts (2006) estimated the vorticity generation owing to baroclinic effects using in situ data collected by an aircraft. Buban et al. (2007) used a Lagrangian analysis of thermodynamic variables based on high resolution dual-Doppler wind syntheses and in situ data collected by a mobile mesonet, aircraft, and sounding data to document the solenoidal generation of the horizontal vorticity. In the present case, the changes in horizontal vorticity (ξ) and the solenoidal effects in the vertical plane perpendicular to the primary dryline were estimated using the thermodynamic and wind data from the soundings presented in Fig. 5.

A band of negative vorticity < −8 × 10⁻³ s⁻¹ is located near the capping inversion (Fig. 14a) and is in response to the strong vertical wind shear between the maritime air mass and the dry, westerly flow aloft. The strongest shear (largest negative values of horizontal vorticity) is located ~100 km to the east of the dryline. The spacing of the sounding data is insufficient to fully resolve the rotor circulation (the position of the center of the rotor is shown by the cross in Fig. 14).

The horizontal vorticity equation can be simplified as follows:

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

where B is the buoyancy, the flow is assumed frictionless, and there are no along-frontal variations. It is also assumed that tilting of vorticity is small (Sipprell and Geerts 2006; Buban et al. 2007). The solenoidal term is shown in Fig. 14b and reveals that most of the air mass east of the dryline and beneath the capping inversion was generating negative horizontal vorticity with a minimum ≈ −2 × 10⁻⁶ s⁻². Buban et al. (2007) presented values that are approximately 5 times larger concentrated within a narrow 1-km zone using the higher-resolution data contained in their Lagrangian analysis. Their background values to the east of the dryline were ~1 × 10⁻⁶ s⁻², consistent with the mesoscale analysis presented in this study. The solenoidal term is positive aloft and to the east of the dryline owing to the reversal of the horizontal temperature gradient to the rear of the “head” of the maritime air mass. The advective terms were calculated (not shown) in order to assess the tendency of ξ (Fig. 14b). The tendency is negative at low levels (<1.5 km) but there are several regions where minimum values are < −2 × 10⁻⁶ s⁻².

One of the areas located near the 2150 UTC sounding at ~800 mb is nearly collocated with the rotor circulation resolved by the dual-Doppler wind synthesis. Accordingly, the coarse resolution sounding analysis illustrates the importance of the solenoidal generation of horizontal vorticity and updrafts at the leading edge of the maritime air mass similar to past studies. Frontogenesis would also promote positive vertical motion in the warm sector; however, this was insufficient to initiate convection.

6. Summary and conclusions

A detailed analysis of a dryline that formed on 22 May 2002 during IHOP was presented. An airborne Doppler and a ground-based multiparameter radar were the primary platforms used in this study. The former flew long legs (~100 km) at low levels that were parallel to the boundary in order to collect data that would resolve the along-frontal variability of the dryline. In addition, averaging of numerous cross sections of the Doppler wind syntheses revealed the mean vertical
structure of the dryline. The S-Pol radar provided continuous surveillance scans of the dryline and was equipped with the capability to record refractivity. Variations of radar refractivity are primarily caused by changes in water vapor under typical summertime conditions.

The 22 May dryline was classified as a null case since no convection initiated along the boundary on this day. The absence of deep convection is supported by the analysis of a series of soundings that revealed the height of the LFC at 620 mb. This level was significantly greater than the maximum height for air parcels lifted from the boundary layer based on the mean dual-Doppler wind syntheses. Another study of this case showed that air parcels were associated with insufficient CAPE to initiate convection. The formation of a secondary dryline located to the west of the primary dryline was also documented on this day. The secondary dryline was apparent in the reflectivity data as a fine line and in the refractivity analysis as a distinct moisture discontinuity.

The kinematic structure of the primary dryline revealed substantial along-frontal variability. Indeed, the
reflectivity pattern during one of the analysis times was remarkably similar to the precipitation core/gap structure documented along cold fronts. Circulations associated with misocyclones were positioned in the gap regions and contributed to the observed echo pattern. The misocyclones were advecting moist (dry) air, as shown in the refractivity analysis and the in situ data collected at flight level, in the regions north (south) of the circulations. Similar distortions of the moisture fields by misocyclones have been shown in past studies.

A pronounced maritime air mass with a strong capping inversion was identified in the vertical cross section through the dryline. The dryline was undergoing frontogenesis of the potential temperature field. The horizontal circulation within the head at the leading edge of the maritime air mass was shown to be primarily driven by solenoidal effects. The largest values of horizontal vorticity were located ~100 km to the east of the dryline and were in response to the strong vertical shear across the inversion.

The positions of cumulus clouds using high-resolution satellite images were superimposed onto the S-Pol reflectivity data. The clouds were forming over the primary and secondary drylines and within the air mass between the two boundaries. Many of the clouds were collocated with clear air echoes aloft (2500 m). These echoes were associated with updraft plumes that formed within the convective boundary layer and were often located at the leading edge of strong gradients of refractivity.

Past studies of radar refractivity have focused on its ability to accurately assess the low-level moisture fields; however, its use in forecasting has been restricted to qualitative interpretations. This study introduced the total derivative of radar refractivity (\(DN/Dt\)) in order to quantitatively assess its ability to improve the prediction of the initiation of convection. There was a general tendency for maxima of \(DN/Dt\) to be near regions where cumulus clouds were developing along a fine line. The collocation of these maxima near a convergence line where favorable vertical velocities exist may help identify preferred regions for convection initiation. Although dual-Doppler wind syntheses were used to calculate \(DN/Dt\), current techniques are available to estimate the horizontal wind field in the boundary layer using single-Doppler radar data. Accordingly, it should be possible for forecasters to make good estimates of this derivative.

While the locations of the cumulus cloud field can be obtained currently from high-resolution satellite images, an analysis of \(DN/Dt\) provides important information on the evolving water vapor field following an air parcel. It should also be noted that cirrus clouds can

often mask the cumulus cloud field at lower levels limiting the utility of forecasting convection initiation using satellite imagery alone (e.g., Weckwerth et al. 2005). In addition, high values of \(DN/Dt\) near a convergence line might precede the appearance of clouds on a visible satellite image. Unfortunately, the coarse temporal resolution of the dual-Doppler wind syntheses did not allow for an accurate assessment of the latter.

The forecaster also requires knowledge of whether lift is sufficient to carry parcels to the LFC (it was insufficient in the present case), the existence of midlevel entrainment that may inhibit convective growth, and whether the vertical circulation associated with deep convection is surface-based (e.g., Banacos and Schultz 2005). The main constraining factor in deriving \(DN/Dt\) is the short range over which refractivity data can be collected (40–60 km). In spite of this limitation, the observations presented in this paper are quite promising and suggest that future studies should continue to focus on the utility of this variable for nowcasting of deep, moist convection.

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APPENDIX

Analysis Methodology

The radar data were edited and the aircraft motion was removed from the velocity fields using the SOLO II software package (Oye et al. 1995). The data were then corrected for navigation errors using a technique developed by Testud et al. (1995). The along-track and sweep-angle resolution for ELDORA during IHOP was ~600 m and 1.5°, respectively, based on the information presented in Table 1. The sweep angle resolution led to an effective sampling in the vertical of ~300 m for the maximum distances from the radar used in the present study.

The radar data were 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 set at 400 m AGL. No time–space adjustment was applied to the data since the dryline was quasi-stationary during the observational period.

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. 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 reduced in the mean vertical cross sections that were created. All wind fields presented in this article are ground relative.

The vertical structure of the dryline was reconstructed by averaging \(\sim 105\) cross sections for each flight leg. 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 fine line was positioned in the same location before averaging the data. Therefore, the kinematic features shown in the figures are presented in a fine-line relative frame of reference.

The calculation of the omega field shown in Fig. 6 was based on the sounding wind field and assumed no variation in the along-front direction. The latter assumption was partially assessed by examining the dual-Doppler wind syntheses. The along- and across-front contributions to the total convergence was determined from the syntheses. The partitioning revealed that the along-front component was very small. Accordingly, neglecting this contribution would not significantly alter the derived vertical velocity field.

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