FINE-SCALE VERTICAL STRUCTURE AND DYNAMICS OF SOME DRYLINE BOUNDARIES OBSERVED IN IHOP

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Abstract

Several radar fine-lines, all with a humidity contrast, were sampled in the central Great Plains during the 2002 International H₂O Project (IHOP). This study primarily uses aircraft and airborne mm-wave radar observations to dynamically interpret the presence and vertical structure of these fine-lines as they formed within the well-developed, convective boundary layer. In all cases the fine-line represents a boundary-layer convergence zone. This convergence sustains a sharp contrast in humidity, and usually in potential temperature, across the fine-line. The key question addressed herein is whether, at the scale examined here (~10 km), the airmass contrast itself, in particular the horizontal density (virtual potential temperature) difference and resulting solenoidal circulation, is responsible for the sustained convergence and thus the radar fine-line.
1. **Introduction**

Clear-air weather radar surveillance scans often reveal “fine-lines” during the warm season. These lines correspond with zones of sustained convergence and rising motion within the convective boundary layer (CBL) (Wilson et al. 1994; Geerts and Miao 2005). The CBL is best developed in the early afternoon, and is populated with columns of rising, buoyant air [‘thermals’, Stull (1988, p. 461)]. Thermals tend to penetrate higher in fine-line convergence zones, especially if the CBL is weakly capped (e.g., Karan and Knupp 2006; SG07). Therefore, under suitable conditions, thunderstorms are most likely to initiate along fine-lines (e.g., Wilson and Schreiber 1986; Wilson et al. 1992; Wilson and Megenhardt 1997; Koch and Ray 1997). The organization and behavior of the ensuing deep convection may be strongly affected by the triggering fine-line.

In late spring 2002, several fine-lines were studied in detail as part of the International H$_2$O Project (IHOP), conducted in the central Great Plains, with the specific objective to improve understanding of convective initiation. While most of the IHOP fine-lines triggering thunderstorms were synoptic or gust fronts rather than drylines (Wilson and Roberts 2006), most fine-lines sampled by the IHOP armada (described in Weckwerth et al. 2004) were drylines, or at least carried a significant, sustained humidity gradient. Drylines commonly form in broadly convergent flow in West Texas, particularly in spring, and can occur under synoptically quiescent conditions (e.g., Schaefer 1974). Most of the IHOP dryline cases were observed farther north, and in synoptically active situations.

The objective of this study is to describe and dynamically interpret the meso-$\gamma$-scale (2-20 km) fine-line vertical structure and convergence. On the meso-$\beta$ scale (20-200 km) or larger, drylines result from differential surface fluxes and the advection of distinct airmasses. The
convergence of these airmasses is in part an isallobaric response to daytime pressure falls over West Texas (e.g. Crawford and Bluestein 1997; Bluestein and Crawford 1997), due mainly to the regional slope of the terrain, and a larger surface sensible heat flux on the drier west side. These mechanisms are well-understood (e.g., Schaefer 1974; Sun and Ogura 1979; Sun and Wu 1992; Jones and Bannon 2002) and operate on a larger scale than examined here.

The scale over which structures and gradients are examined here is O(10 km), and the focus is on the cross-fine-line vertical circulation and associated thermodynamic contrast in the afternoon. The key question to be answered is whether the density contrast in the CBL is responsible for meso-γ-scale convergence and fine-line formation. The null hypothesis is that the presence and strength of convergence lines is independent of the density contrast. This question can regard any radar fine-line, but this study focuses on humidity boundaries, i.e. drylines.

Much attention has been devoted to the fine-scale horizontal structure of several IHOP fine-lines, esp. the structure and evolution of misocyclones (Arnott et al. 2006; Markowski and Hannon 2006; Murphey et al. 2006; Xue and Martin 2006; Marquis et al. 2007). This along-line variability is not examined here, but we are aware that it modulates the flow, convergence and vertical motion along the fine-line, and thus may largely explain the difference between successive cross-sections across the line.

The methodology is summarized in section 2. Several case studies are presented in section 3. Section 4 synthesizes and dynamically interprets findings from these case studies plus two VORTEX (Verification of the Origin of Rotation in Tornadoes Experiment) cases.

2. Methodology

a. Data sources and processing
This study primarily uses data collected aboard the University of Wyoming’s King Air (WKA) aircraft. We examine data from 30-40 km long transects across fine-lines at elevations ranging between 150 and 1600 m above ground level (AGL), in particular the gust-probe winds along the flight track (roughly normal to the fine-line) as well as water vapor mixing ratio ($r$) and potential temperature ($\theta$). Water vapor is measured by a rapid-response Lyman-alpha probe, which was calibrated by a slow-response chilled-mirror dewpoint sensor.

The WKA carried the 95 GHz (W-band) Wyoming Cloud Radar (WCR). The WCR simultaneously operated two fixed antennas, primarily in profiling (up/down) or vertical-plane dual-Doppler (VPDD) modes (Fig. 2 in Geerts et al. 2006). In the profiling mode, radar data portray the vertical structure of the CBL over its entire depth except for a ~220 m deep blind zone centered at flight level. The VPDD mode allows an estimation of the along-track two-dimensional (2-D) air circulation below the aircraft at a resolution of ~30 m. The extraction of the echo motion from the Doppler velocity measured from a moving platform, and the dual-Doppler synthesis, are discussed in Damiani and Haimov (2006).

Several other IHOP data sources were used. Low-elevation-angle reflectivity from various scanning ground-based radars were used to place the WCR transects in the context of radar fine-lines. Data from both aircraft dropsondes and ground-based balloon sondes, collected in close proximity to the fine-lines, were used to determine the CBL depth and airmass contrast.

b. Density-gradient driven circulations

Radar fine-lines are a manifestation of sustained linear convergence within the CBL (Wilson et al. 1994; Geerts and Miao 2005). The central hypothesis of this paper is that this convergence primarily results from a horizontal difference in buoyancy and thus in virtual
potential temperature ($\theta_v$). The horizontal scale of this convergence and $\theta_v$ difference ($\Delta \theta_v$) needs to be determined. Positive $\theta_v$ anomalies are buoyant, i.e. they tend to rise spontaneously, at least if they are rather small-scale (e.g., Houze 1993, p. 225; Doswell and Markowski 2004). Buoyant ascent implies a horizontal vorticity dipole (viewed in a vertical cross section), and roughly circular features (viewed on a map) with diameter of $\mathcal{O}(1 \text{ km})$, depending on the CBL depth (e.g., Stull 1988, p. 461). These features should be present irrespective of the spatial organization of thermals, i.e. whether or not the thermals are aligned by horizontal convective rolls (HCRs).

A horizontal $\theta_v$ gradient also drives a convergent solenoidal circulation that may lead to a density current, whereby the less dense air is forced over the denser air. In this case the buoyancy-induced horizontal vorticity is a single extreme, although under suitable ambient shear, a vorticity dipole may be present across the boundary (e.g., Rotunno et al. 1988). As will be illustrated below, such singular extreme or dipole has a horizontal scale larger than that of thermals. Also, features tend to be elongated on a map since, to a first order, the secondary circulation is 2-D. The circulation produces a radar fine-line, which separates the airmasses of different density.

The slope of this fine-line echo, and of the associated updraft, in a vertical plane, is evidence of wind shear ($\frac{\partial u}{\partial z}$), which can be large-scale or local. If it is large-scale, all echo/updraft plumes should be tilted. A tilted echo/updraft plume at the boundary only, surrounded by upright thermals, is evidence of local baroclinicity at that boundary. The Lagrangian change ($D/Dt$) of horizontal vorticity ($\eta$) due to baroclinicity can be estimated as:

$$\frac{D\eta}{Dt} = \frac{g}{\theta_v} \frac{\partial \theta_v}{\partial x}$$

(1)
where $x$ is the boundary-normal direction [in the cases examined here, pointing (south)east],

$$
\eta = \frac{\partial u}{\partial z} - \frac{\partial w}{\partial x},
$$

$w$ vertical velocity, and $g$ gravitational acceleration. In the cases examined below, the length scale is greater than the height scale (the depth of the CBL $h$) by a factor of two or more, thus wind shear dominates $\eta$. In what follows we assume along-boundary uniformity ($\partial/\partial y = 0$). To start, we treat (1) as an initial value problem and examine the development of shear $(u_h - u_o)/h$ within an initially motionless CBL as a result of a steady gradient $\Delta \theta_v/\Delta x$, during a period $\Delta t$. Initially advection by the baroclinically-generated flow can be ignored, thus (1) can be integrated vertically and in time to obtain:

$$
u_h - u_o = \frac{g\Delta t}{\bar{\theta}_v} \int_0^h \frac{\Delta \theta_v}{\Delta x} dz
$$

where $\bar{\theta}_v$ is the average $\theta_v$ over $0<z<h$. The $\Delta \theta_v$ driving this circulation is largest near the ground (where it is referred to as $\Delta \theta_{v,o}$) and decreases with height, as illustrated in Fig. 1 for two dryline cases described in the literature, and as confirmed in section 3a below.

The CBL depth $h$ is defined as the level where potential temperature starts to increase. Experimentation has yielded these objective criteria, for sounding data: $h$ is at the base of the layer where $\frac{d\theta}{dz} > 0.5K/200m$ and $\frac{d\theta}{dz} > 1.0K/1000m$ and nowhere within the 1000 m layer does $\theta$ decrease by over 0.2K. The CBL depth on the denser side of the dryline ($h_{de}$) often is considerably less than the depth on the lighter side ($h_l$) [e.g., Fig. 17 in Atkins et al. (1998)]. In the two dryline cases shown in Fig. 1, $h_l$ (on the dry side) was 2-3 times larger than $h_{de}$. Flight levels higher than shown in Fig. 1 were not flown, but linear extrapolation of $\Delta \theta_v$ data shown in Fig. 1 suggests that $\Delta \theta_v$ nearly vanishes at $h_{de}$. The ratio of $\Delta \theta_v$ at $h_{de}$ to $\Delta \theta_{v,o}$ is defined as $R_{\theta_v}$, i.e.,
\[ \Delta \theta_v \equiv \Delta \theta_{v,o} \left( 1 - (1 - R_\theta) \frac{z}{h_{de}} \right) \]  

(2)

For the cases shown in Fig. 1 and in section 3a, \( R_\theta \) averages 0.37, ranging between 0.20-0.51.

The above integral can be solved over \( 0 < z < h_{de} \) using (2), and the shear can then be estimated as:

\[
\frac{u_{ho} - u_o}{h_{de}} \equiv \frac{(1 + R_\theta) g \Delta t \Delta \theta_{v,o}}{2 \Theta_v} \frac{\partial \Delta \theta}{\partial x}
\]

(3)

This approximation is useful only in the initial stages (\( \Delta t \) small), before frictional dissipation and advection become important. Later, an equilibrium develops, in which the steady-state secondary circulation (a form of kinetic energy) tries to destroy the horizontal \( \Delta \theta_v \) (a form of potential energy) that is maintained by advection and/or surface heat fluxes. The convergence associated with this circulation can be estimated as follows, starting with (1):

\[
g \frac{\partial \theta_v}{\partial x} = \frac{D \eta}{D t} = u \frac{\partial^2 u}{\partial x \partial z} + w \frac{\partial^2 u}{\partial z^2}
\]

That is, the tendency term is ignored in the total derivative. The above relationship can be integrated over \( h_{de} \):

\[
\int_0^{h_{de}} \frac{g}{\Theta_v} \frac{\partial \theta_v}{\partial x} \, dz = \int_0^{h_{de}} u \frac{\partial^2 u}{\partial x \partial z} \, dz + \int_0^{h_{de}} w \frac{\partial^2 u}{\partial z^2} \, dz
\]

We now apply the chain rule twice, and assume 2-D incompressible continuity (\( \frac{\partial u}{\partial x} + \frac{\partial w}{\partial z} = 0 \)) and \( w=0 \) at \( z=0 \) and \( z=h_{de} \), to modify the right side of the above equation as follows:

\[
= \int_0^{h_{de}} u \frac{\partial^2 u}{\partial z \partial x} \, dz + \int_0^{h_{de}} \frac{\partial u}{\partial x} \frac{\partial u}{\partial z} \, dz
\]

\[
= \int_0^{h_{de}} u \frac{\partial^2 u}{\partial z \partial x} \, dz + u \frac{\partial u}{\partial x} \int_0^{h_{de}} \frac{\partial u}{\partial z} \, dz
\]

\[
= \int_0^{h_{de}} u \frac{\partial^2 u}{\partial z \partial x} \, dz + u \frac{\partial u}{\partial x} \mid_0^{h_{de}} - \int_0^{h_{de}} u \frac{\partial^2 u}{\partial z \partial x} \, dz
\]
The first and third terms cancel, thus
\[
\left[ \frac{\partial u^2}{2\partial x} \right]_{h_{ve}} - \left[ \frac{\partial u^2}{2\partial x} \right]_{o} = \frac{g}{\Theta_v} \int_{0}^{h_{ve}} \frac{\partial \theta}{\partial x} \, dz
\]
(4)

Observations indicate that the boundary-normal confluence is strongest near the surface and, by extrapolation, vanishes near \(h_{de}\). The average ratio of \(\Delta u\) at \(h_{de}\) to \(\Delta u_o\), defined as \(R_u\), is only 0.17 for the cases illustrated in Fig. 1 and section 3a. Thus \(\Delta u\) vanishes at a slightly lower level than \(\Delta \theta_v\). For simplicity we use the same linear expression (2) for the profiles of \(\Delta u\) and \(\Delta u^2\), but with a slope \(R_u\). Then a finite difference expression of (4) becomes:
\[
- \Delta u^2_o (1 - R_u) \equiv \frac{2\Delta \nu g}{\Theta_v} \int_{0}^{h_{ve}} \frac{\partial \theta}{\partial x} \, dz
\]

or:
\[
- \Delta u_o \equiv \frac{1}{1 - R_u} \frac{\Delta \nu g}{u_o \Theta_v} \int_{0}^{h_{ve}} \frac{\partial \theta}{\partial x} \, dz
\]
(5)

In the absence of ‘ambient’ flow, \(u_o\) merely is the horizontal component of the solenoidal circulation. Using (2), (5) can be approximated as follows:
\[
- \Delta u_o \equiv \frac{(1 + R_u) g h_{de} \Delta \theta_{v,o}}{2(1 - R_u) u_o \Theta_v}
\]
(6)

where the subscript ‘o’ refers to measurements near the surface, in practice below 0.5\(h_{de}\). The magnitude of the confluence \(\Delta u_o\) in (6) is affected by the rates of decline of \(\Delta \theta_v\) and \(\Delta u\) with height in the CBL affect, and by \(u_o\). In general, \(R_u\) should be close to zero, since diffluent return flow should occur in the upper CBL, however the density contrast can be found over a greater depth. In practice, \(u_o\) is calculated as the average low-level flow on the ‘dense’ side of the boundary, minus the ‘ambient’ flow (defined as the average low-level boundary-normal flow in nearby soundings on opposite sides of the boundary), and is positive towards the boundary. Note
that $u_o$ is in the denominator of (6), thus when it is small, as with some weak boundaries, the uncertainty of $u_o$ can make confluence estimates based on (6) unreliable.

Expression (6) fundamentally is an application of the Bjerknes circulation theorem (Bjerknes 1898), and relates the steady-state low-level cross-boundary confluence ($-\Delta u_o$) to $\Delta \theta_{v,o}$ at corresponding horizontal scale. The term $-\Delta u_o$ is similar to convergence, but it does not imply a scale ($\Delta u_o$ is not divided by $\Delta x$), and the along-boundary component ($\partial v/\partial y$) is ignored. A similar expression has been obtained by Grossman et al. (2005), but they start from a circulation integral, and ignore the variation of $\Delta \theta_v$ with height.

Another relationship between confluence and density contrast can be obtained from density current theory. The boundaries examined herein may not have a sufficient $\Delta \theta_v$ to drive a classic atmospheric density current, as documented for instance for thunderstorm-generated gust fronts (e.g., Mueller and Carbone 1987) or cold fronts (e.g., Wakimoto and Bosart 2000); nevertheless the theory may apply. Laboratory experiments suggest that the speed of a density current ($U_{dc}$) can be estimated as follows (Simpson and Britter 1980):

$$U_{dc} = K \sqrt{g D_{dc} \Delta \theta_v / \theta_v} + b U_{env}$$  (7)

This equation is based on the horizontal momentum conservation equation (e.g., Houze 1993); $U_{dc}$ bears similarity to the shallow water wave speed. Some form of (7) is commonly encountered in the literature, including in atmospheric applications. In (7) $K$ corresponds with the Froude number; experimental values for $K$ range from 0.7 to 1.0 or higher for atmospheric density currents (e.g., Wakimoto 1982; Mueller and Carbone 1987; Kingsmill and Crook 2003). $U_{env}$ is the ambient low-level flow normal to and ahead of the density current. Laboratory work by Simpson and Britter (1980) indicate that $b=0.70$. Numerical work by Liu and Moncrieff
(1996) suggest that \( b = 0.74 \), ranging between 0.71 and 0.78. The density current depth \( (D_{dc}) \) is assumed to equal \( h_{dc} \), although this is uncertain (see Section 4.d below). The \( \theta \)v difference normally is measured on the ground, thus \( \Delta \theta_{v,o} \) is used in (7).

The movement of a boundary provides poor evidence for the density-current nature of the boundary (e.g., Smith and Reeder 1988). Firstly, the estimation of the observed motion of the boundary is complicated by along-line variability. Secondly, the estimation of the theoretical boundary motion using (7) has several uncertainties, including the determination of the “ambient” low-level flow \( U_{env} \). This determination is complicated by the ambient wind shear and variable CBL depth.

We can however measure the flow confluence \((-\Delta u)\) with some accuracy, and it can be compared with the following estimate from density current theory. First, observations have shown that the feeder flow \( (U_{max}) \) exceeds the density current speed by about 40% (Simpson et al. 1977),

\[
U_{max} \equiv 1.40U_{dc}
\]

(8)

Thus the confluence normal to a density current, \(-\Delta U_{dc}\), can be inferred from (7) and (8), assuming \( K = 0.85 \) and \( b = 0.74 \):

\[
-\Delta U_{dc} = U_{max} - U_{env} \equiv 1.19 \sqrt{\frac{gD_{dc}\Delta \theta_{v,o}}{\theta_v}} + 0.04U_{env} \equiv 1.19 \sqrt{\frac{gh_{dc}\Delta \theta_{v,o}}{\theta_v}}
\]

(9)

The neglect of the term \( 0.04U_{env} \) implies an uncertainty \(<1\) ms\(^{-1}\), less than other sources of error in this estimation. Both (6) and (9) relate the steady-state near-surface cross-boundary confluence to \( \Delta \theta_{v,o} \) over the same horizontal scale. The main difference is that in the case of a solenoidal vortex (6), \( |\Delta u| \propto h_{dc}\Delta \theta_{v,o} \), while in the case of a density current (9), \( |\Delta u| \propto \sqrt{h_{dc}\Delta \theta_{v,o}} \).
Even though the term $\Delta x$ does not appear in (6) nor in (9), the horizontal scale over which these differences are computed clearly matters. The question of scale was raised by Atkins et al. (1998) and is further explored here. WKA wind and thermodynamic data are averaged over either $\Delta x = 3$ km or $\Delta x = 10$ km on either side of the leading edge of the boundary along all available low-level flight legs (below $0.5h_{de}$). The averaging is necessary in order to reveal small differences (the signal) embedded in much larger perturbations of convective origin (the noise). The choice of these scales is based on the following. A distance of 3 km should exceed the diameter of most coherent eddies (thermals) in the fair-weather CBL: aircraft data (e.g., Kaimal et al. 1976) and WCR vertical profile data (Miao et al. 2006) indicate that in the mid-CBL the diameter of most eddies is $1.5h$ or less. However HCRs generally have wavelengths longer than 3 km. And 10 km is about the maximum flight track distance generally (but not always) available on either side of the boundary in IHOP. The distance between the centers of the averaging domains on either side of the dryline equals the averaging distance $\Delta x$, because we do not assume a buffer zone around the boundary. The third and largest scale is the boundary-normal distance between soundings. In the cases examined here, this distance ($\Delta x$) ranges from 15 to 62 km. Sounding data differences are computed based on averages between the ground and $0.5h_{de}$.

The theory is tested first for a vigorous but shallow cold front observed on 24 May 2002 in west Texas (Xue and Martin 2006; Karan and Knupp 2006). This cold front had all characteristics of a density current (Geerts et al. 2006), large $\Delta \theta_{v,o}$ values (3-6 K over scales of 3-19 km), and strong convergence ($-\Delta u$) at the head (8-11 ms$^{-1}$ over the same scales) (Table 1). Both the solenoidal vortex equation (6) and the experimental density current equation (9) overestimate the low-level confluence at all scales, compared to observed values (Table 1).
3. Fine-line observations

a. 22 May dryline

On 22 May 2002 the IHOP armada studied a dryline fine-line near S-POL in the Oklahoma Panhandle. Two fine-lines developed: a more defined one to the east (the “primary dryline”) and a less apparent line to the west [Fig. 2; Fig. 3 in Weiss et al. (2006a)]. These lines merged north of the IHOP operations. The humidity and temperature contrasts between the moist and dry airmasses were considerable [Fig. 3, as well as Fig. 4 in Demoz et al. (2006)]; the airmass between the two drylines had intermediate properties. The soundings shown in Fig. 3 indicate that the moist-air $\theta_v$ was 4.5K lower than that on the dry (Table 1). Even larger $\theta_v$ differences across the dryline complex existed at the surface (Fig. 2). The dry-side CBL was remarkably deep, and about twice as deep as the moist-side CBL depth ($h_{de}$), too deep for the CBL depth to be inferable from WCR profiles (Miao et al. 2006): the upper ~1 km of this CBL was below the temperature threshold of most insects. But east of the eastern dryline, WCR-inferred CBL heights corresponded well with the thermodynamic CBL height (Fig. 4).

Clear-air reflectivity imagery, for instance from S-POL, indicate that the convergence zone of the primary dryline had a sharply-defined western edge and a more fuzzy eastern edge [Fig. 2; top panels of Fig. 2 in Demoz et al. (2006); Fig. 4a in Weiss et al. (2006a)]. This contrast is most apparent at greater radar range, where S-POL sampled the upper CBL. That suggests that lofted insects were carried eastward over moist-air wedge. In other words, westerly shear across the dryline (Fig. 3) tilted the dryline updraft eastward and caused an eastward trailing echo anvil.

This eastward tilt of both fine-lines is apparent in transects of WCR reflectivity, for instance at 2220 and 2243 UTC (Fig. 4). The echo anvil is most obvious east of the eastern dryline at 2243 UTC (Fig. 4h). The air was moister and cooler on the east side of both lines, but
the eastern line, with higher reflectivity values, had a larger temperature and humidity contrast, as well as more convergent flow. The latter is evident from the gust probe wind normal to the dryline (Fig. 4b and f), and from the WCR vertical velocity field (Fig. 4c and g). The flight-level along-track confluence resulted from slight changes in the direction of a strong wind, mainly across the eastern dryline (see wind barbs in Fig. 4b and f): the wind in the CBL generally blew at ~20 ms\(^{-1}\) from the SSW, along the dryline, but east of the dryline it was more southerly and stronger. Especially the eastern line was marked by strong updrafts, up to ~ 5 ms\(^{-1}\). The dryline echoes penetrated well above \(h_{de}\) [Fig. 4d and h, and Fig. 9 in Weiss et al. (2006b)], notwithstanding the low temperature near the echo tops (12ºC at 2700 m AGL).

The \(\theta_v\) gradient across the primary dryline is about 1.0K over 3 km, and 1.3K over 10 km (Table 1). No measurable local \(\theta_v\) gradient exists across the secondary western dryline. (Only 3 km averages are available for this boundary.) The 10 km \(\Delta \theta_v\) across the primary dryline is only 30% of the \(\Delta \theta_v\) across 62 km (Table 1), implying a broader (meso-\(\beta\)) \(\theta_v\) gradient.

The updraft tilt is consistent with the \(\theta_v\) gradient across the primary dryline, i.e., it appears to be due to baroclinically generated horizontal vorticity. The weak \(\theta_v\) gradient across the western dryline may explain why this line gradually disappeared, mostly between 2220 and 2320 UTC [e.g., Fig. 2 in Demoz et al. (2006)].

A cross-dryline circulation was well-established at 2326 UTC (Fig. 5). A 6.7 ms\(^{-1}\) strong rear-to-front current (highlighted in the red box in Fig. 5b) advected low-\(\theta_v\) air towards the leading edge, which was tilted about 45º from vertical. Because the primary dryline was quasi-stationary at this time, the flow shown in Fig. 5b can be thought of as relative to the boundary, which resembled a density current head. At flight level (~1,540 m AGL) this head was intercepted between 7,300<x<8,200 m (Fig. 5b); air in this region had \(r\) and \(\theta\) values between
those characterizing the dry and the moist airmasses at lower levels (Fig. 4e). This confirms observations by Ziegler and Hane (1993) that the head is a mixing zone. One mixing mechanism must be the recirculation of flow into the head: to the rear of the head, a strong downdraft mixes air into the rear-to-front current (Fig. 5b). At the time of the WKA crossing, the air in the head was generally rising but negatively buoyant (Fig. 5a). This negative buoyancy and the strong downdraft behind the head are additional indications of density current dynamics.

Farther east of the primary dryline the moist air was sampled intermittently at flight level (not shown), which roughly corresponds with \( h_{de} \) (Fig. 4d and h, Fig. 1). Thus the CBL top had undulations, possible trapped lee waves behind the density current head (e.g., Weckwerth and Wakimoto 1992, Jin et al. 1996). The WCR reflectivity profiles reveal no evidence of breaking Kelvin-Helmholtz (KH) waves. Such waves were observed in the case of the 24 May cold-frontal density current, marked by a thrice stronger \( \theta_v \) gradient over 10 km (Table 1). In that case the minimum value of the Richardson number (\( Ri \)) at the interface between denser and lighter air was \(~0.1\) (Geerts et al. 2006). The moist-side temperature and wind profiles on 22 May (from the sounding shown in Fig. 1), filtered to a vertical resolution of 100 m (as done in Geerts et al. 2006), yield a minimum \( Ri \) of 0.4 just above \( h_{de} \). KH instability occurs when \( Ri<0.25 \) (e.g., Miles and Howard 1964; Mueller and Carbone 1987). Thus the \( \theta_v \) deficit of the 22 May dryline density current appears insufficient, or rather the resulting shear at the interface is too weak, to generate breaking KH waves.

The horizontal vorticity within the \(~1\) km diameter solenoid highlighted by black streamlines in Fig. 5b is about \( 2\times10^{-2} \) s\(^{-1}\). An earlier VPDD transect, at 2201 UTC (1.4 hours earlier), showed no coherent circulation across the dryline; at that time the dryline plume was essentially upright. Flight-level data from the primary-dryline crossings below \( 0.5h_{de} \) show a
steady increase of $\Delta \theta$, over 3 km from 0.69K to 1.31K between 2222 UTC and 2356 UTC. The mean $\theta_v$ gradient (1.03K/3km, Table 1) can produce the observed horizontal vorticity in 0.75 hours, according to (3). Thus the circulation at 2326 UTC (Fig. 5b) could have been generated in the time available, and may have become rather steady. Even during the short time period between 2220 and 2243 UTC an increase in echo tilt of the primary dryline is apparent (Fig. 4d and h), and, correspondingly, an increase in $\theta_v$ gradient and in confluence. (It is possible that this difference between the two transects in Fig. 4 is not a uniform temporal change but rather due to along-line variability advecting into the geographically-fixed track from the south.) Between 2300 UTC, the primary dryline began to retrograde (move westward), at a speed of 2-3 ms$^{-1}$ by 00 UTC.

Vertical variations of the airmass difference across the 22 May primary dryline are shown in Fig. 6. In all transects the moist airmass is sufficiently cooler, such that it is also denser. Both $\Delta r$ and $\Delta \theta_v$ decrease with height, as for the VORTEX drylines (Fig. 1), but the level where the extrapolated $\Delta r$ and $\Delta \theta_v$ values reach zero is lower, only slightly above $h_{dc}$. The flow is convergent in all transects except one$^2$ (Fig. 6c), and $\Delta u$ also tends to be strongest at low levels. The dry air rises relative to the moist air in most transects, especially at low and middle levels (Fig. 6c). The slopes of $\Delta \theta_v$ and $\Delta u$ are used in the estimation of $R_\theta$ and $R_u$ [eqn (6), Table 1].

The dryline convergence zone (DCZ) can be several km wide yet the term humidity “discontinuity” is appropriate at the scale of the coherent eddies captured by the WCR. The rapid change in $r$ is evident in Fig. 4a and e, and the horizontal gradient $\Delta r/\Delta x$ averaged over 300 m (Fig. 6d) across the dryline is almost 10 times as large as that averaged over 3 km (Fig. 6a), and

$^2$ The one exception (at 2201 UTC, at a flight level of 1470 m) could be discounted because, even though there is a small humidity difference (0.6 g kg$^{-1}$), the CBL is shallower at this time, the $\Delta \theta_v$ values are insignificant, both at flight-level (0.23K) and on the ground (<0.5K), and the density current circulation is not yet present, as mentioned above.
nearly the entire humidity change is concentrated in 300 m. Yet the $\theta_v$ change is less discontinuous: $\Delta \theta_v/\Delta x$ over 300 m (Fig. 6d) is only 3 times larger than that over 3 km (Fig. 6b). The presence of a humidity jump sharper than the $\theta_v$ change generally applies to the other IHOP cases studied here; data for the two VORTEX cases shown in Fig. 1 do not allow this fine-scale comparison.

The observed cross-dryline confluence ($-\Delta u_o$) is compared to the one expected assuming a steady-state solenoidal circulation (6) and a density current (9) (section 2b) in Table 1. Both theoretical estimates are high (as for the 24 May cold front), and far higher than observed at a length scale of 62 km, indicating that (6) and (9) do not apply at a larger scale in the presence of a large-scale $\theta_v$ gradient. This agrees with the dryline study of Atkins et al. (1998): they find that density current theory can be applied at 5 km, but not at 50 km.

In summary, the moist airmass just east of the 22 May primary dryline developed a $\theta_v$ deficit of sufficient magnitude to drive solenoidal circulation with sloping updraft and an echo plume trailing towards the moist air. WCR transects and flight-level data support the notion of a density current, with a well-defined leading head and low-level rear-to-front flow. Density current characteristics are evident in all WKA/WCR transects except the earliest one, notwithstanding significant along-line variability and an along-line flow far stronger than the solenoidal circulation.

b. 19 June dryline

On 19 June 2002 the IHOP armada examined a synoptic-scale frontal shear zone in NW Kansas (Murphey et al. 2006; SG07). Weak southwesterly flow converged at a prefrontal dryline with a southerly low-level jet confined in the moist airmass (Fig. 7). Deep convection erupted
along the dryline near 2130 UTC, in particular in the vicinity of misocyclones that modulated the moisture and vertical velocity field (Murphey et al. 2006; Marquis et al. 2007). These misocyclones resulted from strong horizontal windshear across the dryline, far stronger than on 22 May. This is evident by comparing the flight-level wind barbs in Fig. 4 to those in Fig. 8. Sipprell and Geerts (2007) (referred to as SG07) have shown that the dryline updraft and echo plume tilted towards the west in an early phase, when the dryline progressed eastward, and later their slope reversed, towards the moist side, while the dryline became quasi-stationary. The change in slope of the updraft was accompanied by a circulation reversal. Soundings were collected on both sides of the dryline, in relative proximity to the dryline, both in the early and later phases (Fig. 7). Initially $\theta_v$ was slightly lower on the dry side, which is unlike any dryline reported on in this study, but has been recorded before (Fig. 4 in Crawford and Bluestein 1997). In the course of ~1.5 hrs, between 1:30 to 3 after local solar noon, the CBL warmed and deepened on the dry side, resulting in a lower mixing ratio, due to the entrainment of dry free-tropospheric air into the deeper CBL. Yet on the moist side the stable layer above the CBL subsided, and the CBL temperature and humidity changed little. As a result, the humidity and mainly temperature contrasts across the dryline increased, and the net effect was that air on the moist side became denser.

In both phases the updraft and echo slopes are dynamically consistent with the observed buoyancy gradient across the dryline: in the early phase air on the dry side is denser (Fig. 8a), but as the dry side warms faster than the moist side, the $\theta_v$ gradient has a reversed sign in the late phase (Fig. 8d). The transition from a westward-tilted boundary, with denser air on the dry side, to an eastward-tilted dryline with denser air on the moist side is examined further in Fig. 9, using WKA data within the CBL. Some drying occurs during the early afternoon, but the humidity
difference persists, thus the reversal of the \( \theta \) gradient is entirely due to differential warming: the dry-side CBL warms more than the moist-side CBL, due to advection and/or surface sensible heat flux. The \( \Delta \theta \) is small both in the early and the late phases, less than 0.5K at the 3 km scale (Table 1), and less than 1.0K at the 10 km scale and according to the sounding pairs. So the resulting solenoidal circulation should be weak.

Indeed, weak solenoidal circulations, weaker than the one observed on 22 May (Fig. 5b) have been documented in both phases (SG07), although an updraft/downdraft dipole is not always apparent (e.g., Fig. 8e). Indirect evidence for the secondary circulation and its reversal arises from the WCR reflectivity profile across the dryline. The higher reflectivity on the moist (denser) side of the 22 May primary dryline (Fig. 2, Fig. 4h) was interpreted as the result of a solenoidal circulation. In the early traverses the WCR reflectivity peaks near the humidity jump that marks the dryline and continues to be high to the west (Fig. 9d), suggesting a solenoidal circulation with upper-CBL transport to the west; during the last two traverses the lowest reflectivity values are found to the west, and higher values are found near and to the east of the dryline. The WCR vertical velocity pattern is consistent with this (Fig. 9e): in the early (late) phase rising motion tends to occur just east (west) of the dryline, however this signal difficult to discern in a vigorously-convective BL.

Low-level WKA dryline traverses indicate that the confluence is 6-7 ms\(^{-1}\) over 3 km for both phases (Table 1). This is stronger than what can be expected from baroclinicity at this scale (6-7 ms\(^{-1}\)). This applies at other scales as well (except for the 10 km scale in the early phase, Table 1). It is remarkable that it applies even to the largest scale: the \( \Delta u \) between the two late-phase soundings, 38 km apart, is above the two theoretical \( \Delta u \) estimates. As mentioned before
(section 3a), the density current confluence at such scale tends to be an overestimate in the presence of a background $\theta_v$ gradient.

Thus some other mechanism may have contributed to fine-scale convergence leading to the 19 June dryline fine-line. Additional evidence for this arises from the fact that at some time between the two phases illustrated in Fig. 8, $\Delta \theta_v$ must have been zero for the reversal to occur. Yet throughout the period Doppler-on-Wheels (DOW)-3 reflectivity and velocity data indicate that the dryline remained well-defined and that the cross-dryline flow was convergent. Horizontal-plane WCR velocity data for an along-dryline flight leg at 2040-2047 UTC indicate strong convergence at $< 1$ km scale [Fig. 11 in SG07].

It is not clear what mechanism explains the rather strong fine-scale convergence during all phases of the 19 June dryline. One possibility is that it was driven by buoyancy relative to both adjacent airmasses. Such buoyancy, with a $\theta_v$ magnitude of $\sim 1$ K, was evident from a dropsonde transect (Fig. 17b in SG07) and, on a smaller scale, from a stepped WKA traverse (Fig. 12a in SG07), both just prior to CI. It is evident also on a scale of a few km, although the magnitude is only $\sim 0.5$K at 2135 UTC (Fig. 8d). This buoyancy may explain why mainly ascending motion occurs at this time, rather than a solenoidal circulation (Fig. 8e). The flight-level wind field (shown below Fig. 8d) suggests strong vertical vorticity near the dryline between 1.2$< \chi <$ 2.0 km, roughly collocated with the main updraft in the WCR transect. At this time the WKA intersected an intensifying misocyclone, labeled ‘J’ in Fig. 6 in Marquis et al. (2007).

A second possibility is downward transfer of momentum towards the dryline, on either side of the dryline, although evidence presented in section 4b argues against that process. Thirdly, larger-scale dynamics contributed throughout the period, in particular the synoptic scale wind shear associated with the decaying cold front and the west-bound ageostrophic acceleration
towards the high plains where daytime surface heating caused surface pressure falls (Murphey et al. 2006). The acceleration of the moist-side southerly flow is quite apparent from the WKA data between 2004 to 2135 UTC (Fig. 8) and from several soundings (e.g., Fig. 7). Still, it is not clear how this leads to fine-scale convergence.

The largest $\Delta \theta_v$ value occurred in the last WKA flight leg and probably continued to increase afterwards. An animation of the 0.5° elevation scan of the Goodland Weather Surveillance Radar (WSR)-88D radar shows that after 2135 UTC, the dryline accelerated towards the northwest. Thunderstorms erupted just east of the dryline and convective outflow may have enhanced the buoyancy deficit on the moist side, and thus the westward propagation of the dryline, although thunderstorm outflows do not fully explain this retrogression, since the dryline also retrogressed in the gap region between thunderstorms.

c. 24 May dryline

On 24 May a dryline was sampled in the Texas Panhandle by the IHOP armada before it was intercepted by a well-defined cold front (Weckwerth et al. 2004; Wakimoto et al. 2006; Xue and Martin 2006). The differences in $h$, $r$, and $\theta$ between the moist and dry airmasses were not as large as on 22 May (Fig. 10). The low-level flow was convergent (Fig. 10). The along-line flow was far weaker than on 22 May, but, as on 22 May, strong (south)westerly momentum prevailed above the CBL (Fig. 10). The dryline propagated at $\sim$2 ms$^{-1}$ to the east [Fig. 7 in Ziegler et al. (2006); Geerts et al. (2006)]. According to dropsonde data collected further east in the moist air, a mesoscale $\theta_v$ gradient was present with denser air to the east.

The WKA sampled two radar fine-lines. A first line, at least 80 km long (Wakimoto et al. 2006), appeared well defined in terms of a humidity contrast, wind shift, echo plume, and deep
updraft, for instance at 1947 UTC (Fig. 11, left), yet where the WKA sampled it, its $\Delta \theta_v$ was tiny, <0.30 K over 10 km (Table 1). The higher humidity on the moist side cancels the lower temperature there, yielding a rather uniform $\theta_v$ in the cross section (Fig. 11a,b). A uniform $\theta_v$ is consistent with the erect dryline updraft (Fig. 11c). At low levels the dryline plume was buoyant (Fig. 11b), and WCR echoes towered some 500 m above the surrounding echo tops along the 34-km long leg which is only partially shown in Fig. 11 (left).

At about 2006 UTC, a well-defined cold-frontal density current intercepted this dryline fine-line. Some 16 km to the east a second dryline formed in the background $r$ and $\theta_v$ gradient fields. This line appeared on ELDORA radar reflectivity maps [Fig. 16b in Wakimoto et al. (2006)], and, just barely, on the Amarillo WSR-88D low-elevation scan [Fig. 6e in Xue and Martin (2006)]. WKA data indicate that this eastern dryline separated the moist airmass ($9<r<10$ g kg$^{-1}$) from an even moister airmass ($12<r<13$ g kg$^{-1}$) (Fig. 11e). A ~5 km wide zone was convergent and buoyant at low levels (Fig. 11f), and the updraft was relatively weak and broken up in several cores (Fig. 11g). Both $\Delta r$ and $\Delta \theta$ were larger than across the first (western) dryline, resulting in $\Delta \theta_v \sim 0.5$K, about twice as large as across the western line (Table 1). A sounding pair (Fig. 10) collected about the same time as the WKA transect shown in Fig. 11 (right) reveals a $\Delta \theta_v$ of 0.7K over 27 km (Table 1). This $\Delta \theta_v$ is still small compared to 22 May, but large enough to produce a slight echo plume slope on the eastern side of the convergence zone (Fig. 11h, near x=8 km), where the $\Delta \theta_v$ is largest.

Shortly after the WKA traverse shown in Fig. 11 (right), the cold front intercepted the eastern dryline as well, and deep convection erupted some 20 km further east, possibly along a third dryline boundary, for which no WKA data were collected.
d. 18 June dryline

In the afternoon of 18 June a dryline developed over southwest Kansas in a region of strong mesoscale humidity contrast (Fig. 12). The CBL was relatively shallow on the moist side, and more than twice as deep on the dry side (Fig. 13). Linear echoes suggestive of HCRs formed in the CBL both ahead and behind the boundary [Fig. 12, as well as Fig. 10 in Karan and Knupp (2006)]. A line of cumulus clouds developed around 2130 UTC, initially mostly west of the primary dryline, and for some time the dryline itself was marked by a line of cumuli. The WKA crossed the primary dryline nine times between 1836-2139 UTC (the “early phase”), and, after refueling, three more times between 2326-0002 UTC (the “late phase”).

The early phase is marked by a rather weak but steady \( \Delta \theta_v \). The available low-level transects reveal \( \Delta \theta_v / \Delta x \) values of 0.3K/3km and 0.8K/10 km (Table 1), and only small differences between flight legs. A much larger \( \Delta \theta_v \) occurred between the two soundings shown in Fig. 13 (\( \Delta x = 32 \) km)(Table 1). The strong dependence of \( \Delta \theta_v \) on scale in the early phase confirms the existence of a mesoscale \( \theta_v \) gradient, evident in the surface data (Fig. 12a). The WCR echoes were weak, but they revealed some evidence of a solenoidal circulation, as plume anvils spread towards the moist air (not shown). Confluence near the dryline was quite weak, consistent with the small \( \Delta \theta_v \) values (Table 1). Baroclinic forcing associated with the 1.9K \( \Delta \theta_v \) in the sounding pair suggests a much larger confluence than observed. In reality this \( \Delta \theta_v \) was not concentrated at one point. WKA transects reveal at least one small discontinuity in \( r \) in addition to the primary dryline, making it difficult to define the primary dryline from WKA data alone. The \( r \) jumps were often collocated with a minor \( \theta_v \) change and a WCR echo plume.

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3 There is a stable kink in the \( \theta \) profile at 1 km AGL on the dry side, however according to the criteria listed in section 2b, this stable layer is too weak to mark the CBL top. CBL depth estimates from signal-to-noise values of a 915 MHz profiler located on the dry side of the dryline (Karan and Knupp 2006) confirm that \( h_l \) is about 2.3 km.
Some rather weak deep convection erupted during the WKA’s refueling break. During the late phase, some rain locally reached the ground along the flight track, contaminating the clear-air echoes, and possibly also the dryline contrast (both $r$ and $\theta_v$). The rain was very light however, occurred on both sides of the primary dryline (Fig. 12b), and most evaporated before reaching the ground, without developing a measurable cold pool. Thus Table 1 includes statistics for the late phase. Two dryline boundaries were present in the late phase, about 17 km apart. Only the primary (eastern) one can be seen in Fig. 12b. Its $\Delta\theta_v$ values again show a strong scale dependence (Table 1). The largest scale (15 km) is derived from a pair of soundings. This sounding pair (not shown) is similar to the one shown in Fig. 13, except that warming and drying occurred on both sides, and that the spatial differences are smaller. This may be because the dry-side sounding was released at just 1-2 km from the primary dryline.

The secondary dryline has $\Delta\theta_v$ values about half as large, but is more confluent (Table 1). This is incongruent with baroclinic theory (section 2b), which predicts a proportionality between $\Delta\theta_v$ and $\Delta u$. Thus the theory overestimates confluence near the primary line at all scales, and underestimates it near the secondary one.

In the early phase, the dryline retrograded slowly ($\sim 2$ ms$^{-1}$), while during the late phase the primary dryline retrograded at $\sim 9$ ms$^{-1}$. This acceleration is not due to baroclinicity alone. The combined $\Delta\theta_v$ and confluence of both late-phase dryline boundaries are about twice that of the early-phase dryline, i.e. the dryline complex did intensify. But station data do not reveal an increase in the meso-$\beta \theta_v$ gradient during this period of acceleration: at both times in Fig. 12, the $\theta_v$ gradient across the Oklahoma Panhandle was $\sim 3$K/100 km. Also, the late-phase primary dryline properties, over a scale of 10 km (Table 1), support a density current speed of only 5 ms$^{-1}$, according to (7) (with $U_{env}=0$ ms$^{-1}$), about half the observed retrogression speed. This suggests
that an ambient wind established from the southeast. Surface observations on the moist side confirm this (Fig. 12b).

4. Discussion

a. $\theta_v$ gradients across drylines, and fine-scale convergence

Several observational studies have documented a $\Delta \theta_v$ across the dryline, with values of about ~1.0K (ranging between 0.5-1.9K) over a distance of ~10 km (Table 2). The May 22 dryline $\Delta \theta_v$ is similar, but all others in this study have $\Delta \theta_v < 1.0K$ over 10 km (Table 1). The other drylines in this study also generally are weaker in terms of humidity contrast and confluence than the 22 May IHOP, the 6 May 95 VORTEX, and the 7 June 94 VORTEX drylines (Table 1) and, unlike these three cases, occurred under clear synoptic forcing and/or outside the main geographic area of drylines. In all cases listed in Table 2b, the moist airmass is the denser one. The 19 June case, with a reversal in the sign of $\Delta \theta_v$ and slope of the dryline updraft (SG07), is by no means a classic dryline.

Even in a boundary layer with vigorous convective motions, relatively small horizontal $\theta_v$ differences drive a solenoidal circulation in which the less-dense air (usually the dry air) rises over the denser air. The sharp humidity contrast found along the dryline (e.g. NSSP Staff 1963, Fig. 4, Fig. 8, Fig. 11) is believed to be the direct result of the low-level confluence associated with the solenoidal circulation. This convergent circulation can be sustained as long as some meso-$\beta$ $\theta_v$ gradient is present.

The null hypothesis of this study, that the presence and strength of convergence lines is independent of their air density contrast, can be rejected by the ensemble of boundaries analyzed herein. A clear correlation exists between $\Delta u$ and $\Delta \theta_v$ (Table 1), although exceptions exist, in
particular on 18 and 19 June. Baroclinic theory predicts confluence values similar to those observed, although this depends on the scale selected (Fig. 14). At larger scales the theory tends to yield an overestimate, but at all scales the linear regression slopes of predicted vs observed confluence, at matching scales, is close to 1:1, at least for the solenoidal confluence equation (6), where $|\Delta u| \propto \Delta \theta_{s,o}$ (Fig. 14). The predicted:observed $\Delta u$ slope is smaller for the density current equation (9) at all scales (Fig. 14), i.e., the confluence at weak (strong) $\theta_v$ boundaries is over-(under-)estimated. This suggests that the power in the proportionality, 0.5 $[|\Delta u| \propto \sqrt{\Delta \theta_{s,o}}$ in eqn (9)], is too small.

The solenoidal forcing is believed to be the leading mechanism of linear, irregular (non-wave-like), fine-scale convergence over flat terrain. It should apply generally, not just in the geographic domain of drylines. Fine-lines of often unclear origin are frequently found in the CBL during the warm season (e.g., Wilson and Schreiber 1986; Koch and Ray 1997). *Fine-scale convergence lines (drylines or other radar fine-lines) are expected to form whenever the meso-$\beta$ $\theta$, gradient (over $\Delta x=25 \text{ km or longer}$) exceeds some threshold. Our study suggests that this threshold is relatively small, $<1.0K/25\text{km}$ in some cases.*

It is conceivable that on days with a large east-west gradient in daytime surface buoyancy flux over the southern/central Great Plains, this threshold is exceeded in several locations, thus several roughly parallel dryline boundaries may form in the afternoon, as observed in this study (on 22 May and 18 June) and elsewhere (e.g., Hane et al. 1997; Hane et al. 2002). Multiple fine-lines are common in West Texas (C. Weiss, pers. comm.). Closely-spaced ancillary lines may coalesce into a single line, possibly by differential propagation related to differences in $\Delta \theta_v$, and the absorption of weaker boundaries by stronger ones (Kingsmill and Crook 2003).
b. **Differential CBL depth, westerly momentum transfer, and dryline formation**

The higher surface sensible heat flux west of the typical dryline formation zone results in a deeper CBL on the dry side, and this, in addition to the elevated topography to the west, allows westerly shear to produce convergence when westerly momentum is mixed vertically in the CBL (e.g., Hane et al. 1993). This process applies to the meso-β scale. It may also produce fine-scale convergence, especially in the presence of a CBL depth discontinuity (Hane et al. 1997).

The necessary conditions for this process are a deeper CBL on the west side ($h_{li} > h_{de}$) and “westerly” shear (or, more precisely, dryline-normal shear) in the $h_{li}$ growth layer. The first condition is generally satisfied for the cases examined here, the second one is not, at least not for the cases for which suitable soundings are available. On 22 May 2002 strong westerly shear and momentum are present above the CBL top, but they did not increase between the two soundings, in fact it weakened somewhat near $h_{li}$ (Fig. 15a) during 2235-2329 UTC, a period during which $\Delta \theta_v$ across the primary dryline increased (section 3.a, and section 4.e below). Earlier, the dry-side CBL may have ingested westerly momentum from aloft as it deepened.

A more conclusive validation necessitates a time period of diurnal $h_{li}$ increase: then at time 1 westerly shear is needed in the layer of CBL growth, and at time 2 increased westerly momentum is expected throughout the CBL. On 19 June 2002 (Fig. 15b) there is westerly shear above $h_{li}$ in the early phase, but in the $h_{li}$ growth layer, there is no westerly momentum on the dry side, and by the late phase, westerly momentum has not increased in the CBL. The strongest evidence that vertical momentum transfer in the deepening CBL does not drive momentum changes on the dry side of the developing dryline comes from the 6 May 1995 VORTEX case (Fig. 15c). Between 2013 and 2300 UTC, westerly momentum decreases, even though there is some westerly shear in the growth layer.
Even if differential vertical momentum transfer does occur, it may be a consequence of the development of a \( \text{meso-} \gamma \Delta \theta_v \). The evidence presented here indicates that it is the latter that drives the circulation, the fine-line, and the discontinuity in CBL depth.

c. Solenoidal circulation and mixing in the dryline convergence zone

Horizontal mixing does occur within the DCZ (e.g., Ziegler and Hane 1993; Atkins et al. 1998; Karan and Knupp 2006; Weiss et al. 2006b), especially when the 10 km \( \Delta \theta_v \) becomes large enough for a density current to develop with leading head and trailing turbulent wake. A first estimate of the DCZ width is given by the circulation around the density current head, about 2 km wide on 22 May (Fig. 5).

A clear circulation did not occur in the other, lower-\( \Delta \theta_v \) cases in this study. Even on those days a circulation may have been present, but obscured by far stronger convective eddies. To extract evidence of a circulation, we examine the vertical profiles of the 3 km difference of WCR vertical velocity (\( \Delta w \)) and reflectivity (\( \Delta Z \)) across the dryline (Fig. 16). The 3 km average is chosen to capture the solenoidal flow and reduce the convective ‘noise’. Fig. 16 only includes the ‘classical’ drylines cases, with a denser moist side. Rising motion of the less dense air relative to the denser air implies negative \( \Delta w \) values (Fig. 16b). A tilt in the updraft plume towards the denser air would imply an increase in mean \( \Delta w \) with height. These features generally are observed on 22 May and on the three other days (Fig. 16b), although there is much scatter, and the difference from one flight leg to the next is large.

Inspired by the echo anvil spreading over the denser, moist air on 22 May (Fig. 2, Fig. 4) and over the cold post-frontal air on 24 May (Geerts et al. 2006), we examine the difference in reflectivity. Reflectivity may have more temporal continuity than vertical velocity, since it
depends on some time integral of vertical motion (Geerts and Miao 2005). The solenoidal circulation implies near-zero ∆Z values near the surface and positive ∆Z in the middle and upper CBL, due to the spreading of the insect plume over the wedge of denser air. This is observed for May 22, but in the other three cases the ∆Z profiles are widely scattered (Fig. 16a). This suggests that a clear secondary circulation is present only on 22 May, and that in general mixing in the DCZ is less a result of a steady secondary circulation and more due to transient eddies (thermals) or horizontal variations (misocyclones).

The humidity trace generally shows a sharper discontinuity at the dryline than the θ_v trace, and θ_v is more likely than r to continue to decrease into the dense-air wedge, in the cases examined here (e.g., Fig. 4). These two characteristics have been documented in other dryline studies (NSSP Staff 1963; Parsons et al. 1991; Ziegler and Rasmussen 1998; Atkins et al. 1998; Karan and Knupp 2006). We believe that this is because the meso-β θ_v gradient is essential for the circulation to occur, while the meso-β humidity gradient, a common characteristic of dryline environments [e.g. Fig. 11 in Ziegler and Hane (1993)], is more circumstantial. For all drylines listed in Table 1 (except the anomalous early phase on 19 June), and all scales, the potential temperature term (which makes the moist side more dense) is 2.8 times larger than the water vapor term (which makes the moist side less dense) in the expression for ∆θ_v. Thus to a first order water vapor can be regarded as a passive tracer.

d. A dryline as a density current

The typical ∆θ_v is small across a dryline, compared to that across cold fronts and gust fronts (Table 2), yet often large enough for the dryline boundary to assume density current characteristics (e.g. Ziegler and Hane 1993; Atkins et al. 1998). The critical 10 km ∆θ_v for
density current characteristics to appear in a CBL may be close to 1.0K, a value exceeded only on 22 May and the two VORTEX cases in Table 1. This value may depend on ambient shear. Florida sea breezes occur with an average $\Delta \theta_v$ of only 1.1K and a shallow marine CBL (Atkins and Wakimoto 1997). A sea breeze was maybe the first atmospheric phenomenon to be compared to a laboratory density current (e.g. Clarke 1955). In many cases the dryline may well behave as an “inland sea breeze” (Sun and Ogura 1979; Sun 1987; Bluestein and Crawford 1997) at the meso-$\gamma$ scale.

In principle a solenoidal circulation occurs over the depth of the $\theta_v$ difference, which varies between $h_{de}$ and $h_l$ (Fig. 1, Fig. 6). If $\Delta \theta_{vl}$ is sufficient and sufficiently long-lived for a density current to develop, its depth $D_{dc}$ is not $h$, but rather a function of ambient shear normal to the dryline and $h$ (Xue et al. 1997). For instance the 24 May cold-frontal density current (Table 1) occupied only $\sim 37\%$ of the warm-side $h$. Published $D_{dc}$ values vary widely (Table 2). Some studies report $D_{dc}$ values less than $h$ (Atkins and Wakimoto 1997) or even less than $h_{de}$ (Atkins et al. 1998). The uncertainty in flow depth implies an uncertainty in the theoretical low-level confluence (Table 1). The fact that density current theory generally overestimates confluence (Fig. 14) may be due to an overestimate of $D_{dc}$. Another factor is that the feeder flow strength (8) probably is the maximum value in the rear-inflow current (Simpson et al. 1977), while WKA and sounding data represent averages.

Retrogression occurred in later stages in all cases studied here, except of course on 24 May, when a cold front moved in. In some cases, as on 19 June, changes in $\Delta \theta_{vl}$ may explain or contribute to dryline retrogression. Dryline propagation is affected by the ambient low-level flow, which is variable and may be diurnally modulated. In most papers where the dryline has been characterized as a density current, retrogression is observed (Table 2), although Crawford
and Bluestein (1997) report on retrogressing drylines without density current characteristics. Larger-scale factors contribute to this retrogression (e.g., Bluestein and Crawford 1997; Shaw et al. 1997), but on the scale of the density current, the hydrostatically higher surface pressure in the moist, cooler air drives ageostrophic flow westward, even if that direction locally is upslope (Mahrt 1982).

e. Diurnal dryline trend

It is remarkable that notwithstanding the clear synoptic “forcing” in all but one case in this study, and the range in geographic locales (from NW Texas to NW Kansas), a diurnal trend is apparent in the cross-dryline differences (Fig. 17). The humidity contrast does not change much, but -Δθ tends to increase with time towards 00 UTC. The net result is that -Δθ, tends to increase (Fig. 17b); this is consistent with an increase in low-level confluence, although such trend is not clearly present (Fig. 17d). At a scale of 3 km, the gust probe data do not reveal a trend towards dry air rising over the moist air (Fig. 17c).

A similar trend emerges from previous dryline studies (Table 2b), although detailed data are often missing. The 7 June 1994 (Ziegler and Rasmussen 1998) and the 6 May 1995 (Atkins et al. 1998; Ziegler and Rasmussen 1998) dryline case studies appear to be amongst the best available. In these studies flight-level water vapor was measured by a chilled-mirror dewpoint sensor on the NOAA P-3 aircraft. This is a slow-response sensor; thus it displayed an hysteresis when the aircraft encountered a sharp moisture gradient (Ziegler and Hane 1993). Overshooting (undershooting) occurs when the probe experiences a sudden humidity increase (decrease). Overshoots (undershoots) often appear as sharp spikes (exponential decays) in the time series of r. The width of the spikes varies from 10 to 20 seconds, and the exponential adjustment can take
30 seconds (~3.3 km along-track). For this reason, no reliable 3 km averages can be taken. One cannot accurately correct bad data, so we removed them. For overshoots, we determined the first and last stable values at each side of the spikes. For undershoots, we eliminated 30 seconds of data from the last trustable point at the moist side. Averages of any variable were then computed from the 1 Hz NOAA P-3 data over 10 km from either side of the best-guess dryline location, excluding the flagged times.

The resulting differences for the two VORTEX cases (Fig. 18) are generally larger than the 3 km differences for the IHOP drylines (Fig. 17). The 10 km $\Delta r$, $\Delta \theta_v$, and $\Delta u$ values (~5.2 g kg$^{-1}$, ~1.9 K, and ~8.3 ms$^{-1}$ respectively) are significantly larger than on 22 May 2002 (Table 1). But the diurnal trends of $\Delta r$, $\Delta \theta$ and $\Delta \theta_v$ are similar. Dryline-normal confluence $\Delta u$ tends to increase as well, consistent with the $\Delta \theta_v$ trend (Fig. 18d). Notwithstanding the larger $\Delta \theta_v$, there is still no evidence for ascent of dry air over moist air at a scale of 10 km (Fig. 18c).

*The diurnal trend of $\Delta \theta_v$ in both IHOP and VORTEX drylines suggests that $\Delta \theta_v$ is largely driven by regional differences in surface heat fluxes,* which are strongly diurnally modulated. The difference in surface buoyancy fluxes probably peaks before 00 UTC (~5 hrs 20 min after local solar noon), but the sign of the difference may be maintained until at least 00 UTC. The regional $\theta_v$ gradient then increases at finer scales by baroclinic circulations. Thus dryline characteristics probably are strongly affected by regional differences in surface heat fluxes. We hope to examine this further in a future campaign that includes regional flux estimates.

5. **Conclusions**

The fine-scale kinematic and thermodynamic properties of several radar fine-lines observed in IHOP are examined. All boundaries are marked by a sharp humidity contrast, and
thus they are referred to as drylines. Significant along-line horizontal variability exists in some cases, but the focus of this study is on the vertical structure. The main conclusions are as follows:

- All fine-lines are convergent and the convergence strength generally is related to their meso-$\gamma$ scale air density contrast, expressed in terms of $\Delta \theta_v$.
- Two methods, one based on a steady solenoidal circulation and the other on density current theory, are used to estimate the meso-$\gamma$ scale convergence $\Delta u$ near a baroclinic boundary. The solenoidal confluence is proportional the CBL depth and $\Delta \theta_v$, while the density current confluence is proportional to their square root. The former yields a slope of predicted vs. observed confluence close to 1:1 for the cases examined here. Both methods tend to overestimate confluence, especially for well-defined boundaries (large $\Delta \theta_v$). The optimal scale at which $\Delta \theta_v$ and $\Delta u$ are measured appears relatively small: both methods tend to overestimate $\Delta u$ more when $\Delta \theta_v$ is computed at distances $\Delta x$ of 10 km or more.
- One boundary, the 22 May primary dryline, has a clearly sloping reflectivity and updraft plume, a rear-inflow current, a secondary circulation in the vertical plane across the dryline, and other density current characteristics. The moist airmass is denser, with a $\Delta \theta_v$ of 1.3K over 10 km, decreasing with height in the CBL.
- Other fine-lines, observed outside the classic domain of dryline occurrence and/or under synoptically disturbed conditions, are all are marked by some low-level $\Delta \theta_v$, but less than 1.0K over 10 km. Some slope in the echo plume, towards the denser airmass, is generally present, but WCR data reveal no solenoidal circulation. Thus, especially in weak $\Delta \theta_v$ cases, mixing in the DCZ appears to be dominated by convective activity and possible along-line variability, not a secondary circulation. In some cases the fine-scale confluence exceeds that
expected from baroclinicity alone, in particular in the late phases of 18 and 19 June, when
deep convection broke out.

- In most cases a meso-β scale $\theta_v$ gradient is present. A threshold gradient may exist for the
development of one or more radar fine-lines, concentrating the $\Delta \theta_v$. Limited observations
herein suggest that this threshold is rather low, 1.0K/25 km or even less. Similarly, fine-line
boundaries may assume density-current properties once their $\Delta \theta_v$ exceeds a threshold, about
1.0K over 10 km.

- This study does not lend credence to the idea that differential vertical transfer of boundary-
normal momentum, associated with differences in CBL depth across a dryline, produces fine-
scale convergence.

- This study reveals an unexpected diurnal trend of the meso-γ scale $\Delta \theta_v$ across drylines in both
IHOP and VORTEX. We speculate but do not demonstrate that results from a diurnally
varying meso-β scale $\Delta \theta_v$ that is largely driven by differential surface heat fluxes.

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research. Sam Haimov was responsible for the WCR data collection. Rick Damiani developed
the WCR Doppler velocity analysis package. Conrad Ziegler provided the NOAA P-3 data for 7
June 1994 and 6 May 1995 and advised us on the use of the data. Margaret LeMone reviewed an
early draft of the document.
References


FIGURE CAPTIONS

**Fig. 1**: Differences in along-track (approximately cross-dryline) wind (grey symbols) and $\theta_v$ (black symbols) across two drylines documented during VORTEX, and best linear fit (dashed lines). (a) 6 May 1995 (Atkins et al. 1998; Ziegler and Rasmussen 1998); (b) 7 June 1994 (Ziegler and Rasmussen 1998). The differences are defined as moist ($x>0$) minus dry ($x<0$), and are computed from 10 km averages on either side of the dryline, for flight legs at various levels. The data were collected at 1 Hz frequency aboard the National Oceanic and Atmospheric Administration (NOAA) P-3 aircraft. The CBL depths listed are derived from Fig. 17 in Atkins et al. (1998) for (a) and Fig. 9 in Ziegler and Rasmussen (1998) for (b).

**Fig. 2**: Radar reflectivity and surface station observations near the dryline (DL) on 22 May 2002. The reflectivity field, ranging from -5 dBZ (black) to +30 dBZ (white), is from the 0.5º scans of the Amarillo (AMA) WSR-88D and S-POL radars. $\theta_v$ (K) and $r$ (g kg$^{-1}$) are plotted instead of the conventional temperature and dewpoint. A full barb is 5 ms$^{-1}$. The time shown in the upper left applies to the station data. Radar data were collected within a few minutes of this time.

**Fig. 3**: Profiles of $r$ and $\theta$ derived from two near-simultaneous soundings released on opposite sides of the 22 May 2002 split dryline. Grey (black) lines and symbols refer to the moist (dry) side.

**Fig. 4**: WKA observations of the 22 May dryline (DL) system, along two transects, on the left around 2220 UTC (flight level ~590 m AGL), and on the right around 2243 UTC (flight level ~160 m AGL). The horizontal axis is distance along the flight track; the reference point ($x=0$) is
arbitrary but it is the same in both transects. In this and other transects, the moist side is on the right. Flight-level data are shown on top (panels a, b, e, and f). This includes the along-track (dryline-normal) wind, and the horizontal wind barbs (full barb=5 m s$^{-1}$). Along-track flow convergence occurs when the blue trace descends from left to right in b) and f), and the main convergence belt is highlighted by the opposing blue arrows. Panels c) and g) show WCR vertical velocity (positive values indicate ascent), and panels d) and h) show corresponding WCR reflectivity in the vertical plane above the aircraft. In the WCR plots the vertical axis (height) is exaggerated somewhat (see “aspect ratio”); this makes the tilt of plumes less apparent.

Fig. 5: (a) WKA flight-level data and (b) WCR vertical-plane dual-Doppler wind field across the 22 May primary dryline around 2326 UTC. The vector field is visually enhanced by objectively-drawn streamlines (black lines with arrows). The vertical velocity field is shown in color. The nadir antenna -15 dBZ contour, enclosing the strongest echoes, is shown in green.

Fig. 6: Profiles of the difference between moist and dry air on 22 May, based on in situ WKA data at various flight levels between 2207-2350 UTC. The difference [moist-dry] is based on 3 km averages, on opposite sides of the primary dryline, which was marked by a clear humidity gradient. The top of the vertical axis is $h_{dc} \equiv 1.6km$ (Fig. 3). Black + symbols refer to the lower axis and grey stars to the upper axis. The dryline-normal wind ($u$) and vertical air motion ($w$), shown in (c), are obtained from the gust probe. In (d) the maximum gradient at the dryline is shown, over $\delta x=300m$, for 100-m filtered values. Dashed lines indicate linear best fits.
**Fig. 7**: As Fig. 3, but for the 19 June 2002 dryline. (a) early phase ($\theta_v$ higher on the moist side). (b) late phase, nearly 1.5 hours later ($\theta_v$ higher on the dry side). Note the correspondence of the x-axes in (a) and (b).

**Fig. 8**: Flight-level data and WCR vertical transects above and below flight level for the 19 June 2002 dryline, during the: (a-c) early phase (2004 UTC); (d-f) late phase, 1.5 hours later. The flight-level gust probe wind is shown as barbs above the WCR vertical velocity panel. The direction of the solenoidal circulation is shown in (b).

**Fig. 9**: Composite of WKA measurements across the 19 June dryline, illustrating temporal change. The lines become lighter with time. Flight levels range between 150 – 900 m AGL, all well within the CBL on both sides of the dryline. All data apply to flight level, except the WCR reflectivity and vertical velocity. These data are an average of the first 7 range gates (210 m) from both the nadir and zenith antennas. (For the 150 m AGL flight legs, the first 14 gates from the zenith antenna are used.) All data have been averaged in along-track segments of 200 m.

**Fig. 10**: As Fig. 3, but for the 24 May 2002 dryline.

**Fig. 11**: As Fig. 4, but for the western (on the left) and eastern (on the right) drylines on 24 May 2002 in the western Texas Panhandle. The flight level is ~200 m AGL. The height dimension of the WCR images is slightly exaggerated (aspect ratio ~1:1.45). A cold front is present just to the left of the image domains of both (d) and (h), some 6 and 8 km to the NW of the drylines, respectively.
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Fig. 14: Summary of boundary-normal confluence observations and predictions listed in Table 1. The black symbols apply to the density current equation (9), the grey ones to the solenoidal equation (6). Linear regression curves and mean values are shown in black and grey for the respective equations. The mean values are also expressed as a percentage of the observed mean, between brackets. The top panel shows 3 km averages, the middle 10 km averages, and the bottom values are derived from soundings.

Fig. 15: Profiles of dryline-normal momentum on the dry side of the dryline at different times, on (a) 22 May 2002, (b) 19 June 2002, and (c) 6 May 1995, from soundings released at the times indicated, all within 30 km from the dryline.

Fig. 16: Vertical variation of the difference in (a) WCR reflectivity, and (b) WCR vertical velocity, across the dryline, on four days studied here. The difference is defined as [moist-dry] and is based on 3 km averages on either side of the dryline. Black symbols apply to the strong case (22 May), grey symbols to the other days. The solid line is the mean of all profiles. The profiles are based on seven flight legs on 22 May, and 16 legs on three other days, at various
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TABLE CAPTIONS

**Table 1**: Analysis of the horizontal $\theta$, difference ($\Delta \theta_v$) and confluence ($-\Delta u$) for several fine-line boundaries, all with a humidity gradient, and comparison with theoretically estimated confluence values. The latter either assumes a steady solenoidal circulation [$-\Delta u_o$, see equation (6) in section 2b], or a density current [$-\Delta U_{dc}$, see equation (9)]. The factor $R_\theta$ and $R_u$ in (6) are set at 0.37 and 0.17 respectively; these are averages for the profiles of $\Delta \theta_v$ and $\Delta u$ on 22 May 2002, 6 May 1995, and 7 June 1994. The differences are scale-dependent, and three different scales are assumed: flight-level WKA data are averaged over either 3 or 10 km on either side of the boundary (defined as the location of strongest humidity gradient), or the distance between two soundings on opposite sides of the boundary is used, if available. Only flight legs below 0.5$h_{de}$ are used. Sounding data ("sonde") are averaged between the surface and 0.5$h_{de}$. The distance between soundings is calculated normal to the boundary. $u_o$ is the mean boundary-normal flow on the dense side, minus the ‘ambient’ low-level flow (positive towards the boundary).

**Table 2**: A comparison between drylines (b) and other boundaries interpreted as density currents (a). Only observational studies are listed. $\theta_v$ differences over distances of 10 and/or 25 km are estimated from published data or graphs. The lowest flight leg is used where aircraft data are available. Also listed are: in (a), the estimated density current depth $D_{dc}$; and in (b), the time or period of observations, $h_{de}$ and reported dryline motion. Negative values of dryline motion are westbound (retrogression).
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<th>$\Delta \theta_v$</th>
<th>$h_{de}$</th>
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64
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<th>$\Delta U_{dc}$ (9)</th>
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<td>NOAA P-3</td>
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<td>7.0</td>
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Table 2: A comparison between drylines (b) and other boundaries interpreted as density currents (a). Only observational studies are listed. $\theta_v$ differences over distances of 10 and/or 25 km are estimated from published data or graphs. The lowest flight leg is used where aircraft data are available. Also listed are: in (a), the estimated density current depth $D_{dc}$; and in (b), the time or period of observations, $h_{de}$ and reported dryline motion. Negative values of dryline motion are westbound (retrogression).

a. Non-dryline boundaries interpreted as atmospheric density currents

<table>
<thead>
<tr>
<th>reference</th>
<th>phenomenon</th>
<th>$\Delta \theta_v$ (K) over ~10 km</th>
<th>$D_{dc}$ (km)</th>
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<td>Geerts et al. (2006)</td>
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<td>Mueller and Carbone (1987)</td>
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<td>Atkins and Wakimoto (1997)</td>
<td>6 Florida gust fronts</td>
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<td>Kingsmill and Crook (2003)</td>
<td>10 Florida gust fronts</td>
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<td>18 Florida sea breeze fronts</td>
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b. Drylines in West Texas / Oklahoma

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<th>$h_{de}$ (km)</th>
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<td>1.9</td>
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<td>see Fig. 18</td>
<td>1.8</td>
<td>2-3 K</td>
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