Dryline Characteristics near Lubbock, Texas, Based on Radar and West Texas Mesonet Data for May 2005 and May 2006

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(Manuscript received 17 May 2007, in final form 14 November 2007)

ABSTRACT

Two months of Lubbock, Texas, radar reflectivity data and West Texas Mesonet data are examined to detect dryline finelines and to describe their thermodynamic and propagation characteristics. Before sunset the moist air mass east of the dryline was consistently denser than the dry air mass. This air density difference waned and even reversed after sunset, because of more rapid cooling on the dry side.

This study provides further evidence that the formation and propagation of the dryline convergence zone is driven by the daytime air density difference, that is, that the dryline behaves as a density current. The implication for forecasters is that the air density (or virtual potential temperature) difference across the dryline should be monitored, as a measure of dryline strength and as an additional indicator for the likelihood of convective initiation along the dryline.

1. Introduction

The dryline is a well-defined atmospheric boundary observed over the southern Great Plains, between hot dry air to the west and air of maritime tropical (mT) origin to the east. To a first order, the dryline can be viewed as the intersection of the top of a level, capped boundary layer (BL) containing the mT air mass and the sloping terrain east of the Rocky Mountains (e.g., Schaefer 1974). The dryline has clear diurnal characteristics and becomes more defined during the afternoon hours, when it tends to move eastward, at least in the absence of synoptic changes (Bluestein 1993, p. 284). The large-scale [O(100 km)] convergence along the dryline is attributed to the formation of a heat trough in the lee of the Rockies (e.g., McCarthy and Koch 1982; Parsons et al. 1991; Bluestein and Crawford 1997). The mature dryline is often a remarkably fine boundary [O(1-10 km)] with a substantial humidity jump and significant convergence over the depth of the convective BL (Miao and Geerts 2007). On account of the finescale convergence, radars [including the Weather Surveillance Radar-1988 Doppler (WSR-88D) radars] generally see the dryline as a reflectivity "fineline"

DOI: 10.1175/2007WAF2007044.1

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(e.g., Wilson et al. 1992). The radar return is generally due to insects (Russell and Wilson 1997) that congregate in regions of convergent flow, apparently because they oppose the rising motion associated with the convergence (Geerts and Miao 2005). Visible satellite imagery is of little use, since the dryline is rarely marked by a cloud line or a cloud edge.

Forecasters' interest in drylines is motivated mainly by the fact that drylines are "a major factor in the initiation of the severe thunderstorms in the central and southern United States during the spring" (Hane et al. 1993). Thus, forecasters from southwest Texas to western Kansas routinely examine WSR-88D clear-air radar reflectivity surveillance scans for the presence of a dryline fineline. Clear-air echoes are limited to 70–100 km in the lowest elevation WSR-88D scan (generally 0.5°), and the resulting radar coverage is rather limited, for instance in West Texas (Fig. 1).

The dryline tends to form on about 40% of the days, mainly in late spring (Bluestein 1993, p. 283), at an average longitude of 101°W (Hoch and Markowski 2005). This longitude is not covered well by the WSR-88D network in northwest Texas, since it falls between the Amarillo, Texas (AMA), and Lubbock (LBB), Texas, radars to the west and the Frederick, Oklahoma (FDR), and Dyess Air Force Base, Texas (DYX) radars to the east (Fig. 1). While there is considerable day-to-day variability in the dryline position depending

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FIG. 1. Map of WSR-88D radars and highways in West Texas. The gray circles, with a radius of 70 km, represent the typical clear-air coverage of the boundary layer. The preferred longitude of drylines at 0000 UTC is based on the dryline climatology by Hoch and Markowski (2005). Stations used are Amarillo, TX (AMA); Dyess Air Force Base, TX (DYX); Frederick, OK (FDR); Lubbock, TX (LBB); Midland–Odessa, TX (MAF); and San Angelo, TX (SJT).

on the synoptic conditions, on synoptically quiescent days the dryline position remains relatively close to its climatological mean position. In recent years a dense network of automated weather stations has been installed in West Texas, the West Texas Mesonet (WTM; Schroeder et al. 2005). This allows a description of the airmass contrast across the dryline in unprecedented detail.

The conventional forecast paradigm held that drylines are simply moisture boundaries without appreciable density differences (e.g., Doswell 1982, p. III-17). Yet forecasters have long been aware that in the afternoon the dry air west of the dryline can be substantially warmer than the mT air mass. Modeling work by Sun and Ogura (1979) and Sun (1987) showed that this temperature gradient drives an atmospheric density current similar to a sea breeze, and in fact Sun and Ogura

(1979) referred to the dryline as an "inland sea breeze." The density current nature of drylines was later confirmed by observational case studies (Bluestein et al. 1990; Parsons et al. 1991; Ziegler and Hane 1993; Hane et al. 1997; Bluestein and Crawford 1997; Atkins et al. 1998; Ziegler and Rasmussen 1998) and more detailed numerical simulations (e.g., Ziegler et al. 1995). However, Crawford and Bluestein (1997) describe a dryline without density current characteristics. Also, finescale convergence may be driven by other factors, for example, differential convective BL growth on opposite sides of the dryline, implying the differential downward transfer of zonal momentum into the convective BL (Hane et al. 1997). Sea breezes normally propagate inland, although their progression may be impeded by ambient offshore flow. Similarly, it depends on the ambient wind whether the dryline propagates westward ("retrogrades"). Dryline retrogression is often observed, mainly in the evening (Crawford and Bluestein 1997; Shaw et al. 1997).

The purpose of this study is to use the dense WTM network to characterize drylines in the vicinity of the Lubbock WSR-88D radar, specifically to test the hypothesis that the dryline is the result of a density difference between adjacent air masses. This study differs from previous ones in that it is not case-study based, but uses 2 months' worth of operational data.

Data sources and analysis method are described in section 2. Dryline characteristics are presented in section 3, and the density current nature of drylines is discussed in section 4.

2. Data sources and analysis method

Surveillance scan base reflectivity from the LBB WSR-88D radar data, obtained from the National Climatic Data Center (NCDC), and meteorological data from 18 WTM stations surrounding LBB were examined for 2 months: May 2005 and May 2006. The NCDC base reflectivity product generally came from the 0.5° elevation scan, although sometimes it came from a higher-elevation scan (1.3° or 1.5°), depending on the volume coverage pattern (VCP). The best fineline observations came from the operational clear-air modes (VCP31 or VCP32). Depending on a storm's development, the radar sometimes operated in other VCPs; in this case the finelines were still detectable, although less well depicted.

The Integrated Data Viewer (IDV; information online at http://www.unidata.ucar.edu/software/idv/) was used to display LBB radar and WTM station observations (Fig. 2). The IDV maps were examined at 5-min intervals for the period 1800–0600 UTC [1200–0000



FIG. 2. Example of a dryline fineline and WTM station observations at 2340 UTC 11 May 2005. The color field (with key bar on the left) is the 0.5° elevation base reflectivity (dBZ) from LBB. The conventional parts of the station data are the wind barbs, the temperature T (°C), and the dewpoint T_d (°C). The unconventional numbers, on the right of the station location, are the virtual potential temperature θ_v (K) and the mixing ratio q_v (g kg⁻¹). The station ID is shown below the station. The scale can be inferred from the range rings, shown at 40-km intervals. The red lines are county borders.

central standard time (CST); CST = UTC - 6 h] on each of the 62 days. First, radar finelines were identified. Next, the humidity contrasts revealed by the WTM data and synoptic analysis charts were used to confirm that they were drylines rather than cold fronts or outflow boundaries. Usually outflow boundaries could readily be identified in radar reflectivity animations, as they emanate from parent thunderstorms and are rather local, but this was not always the case. In some cases a "saturation point" analysis (Betts 1984) was conducted to distinguish a cold pool boundary from a true dryline. Saturation points were determined by the temperature and pressure at the lifting condensation level, and they were plotted on a skew T diagram, following the technique proposed by following Betts (1984). If, for a series of observations on the moist side

of the boundary, these saturation points were distributed fairly closely to a moist adiabat, the boundary was excluded.

Sixteen dryline days were identified in both months; that is, drylines were within view of the LBB radar on 26% of the days. Some dryline finelines formed within the LBB domain, while others moved into the area. Displays such as shown in Fig. 2 were then analyzed at 5-min intervals to determine the temperature, dewpoint, water vapor mixing ratio (q_v), and virtual potential temperature (θ_v) on opposite sides of the dryline. At each time, a pair of WTM stations was chosen, such that they were clearly located on either side of the radar fineline, but as close as possible. The distance between suitable pairs averaged 57 km (ranging from 23 to 85 km), and the dryline-normal distance averaged 26 km. This study focuses on the *differences* in q_v and θ_v between the two stations selected in each 5-min radar– WTM frame (as in Fig. 2). Only one pair per frame was used.

The speed of the dryline fineline normal to its orientation was computed using the LBB reflectivity maps at roughly hourly intervals and the "range and bearing" tool in IDV. The dryline orientation was determined at the same intervals by drawing a best-fit line through the fineline. Most finelines could be detected to a radar range of 80–100 km for the clear-air VCPs. The 11 May 2005 dryline was detectable out to 110 km (Fig. 2). In most cases the fineline had to be within ~60 km of LBB for its orientation and speed to be determinable.

The LBB reflectivity data were missing for variable periods of time on several days, including on some dryline days. Thus, it is possible that some drylines were missed. Since this study focuses on the finescale thermodynamic properties of the dryline, the presence of a radar fineline is deemed essential. Thus, the periods without LBB radar data were excluded from the dataset to be examined in this study. One exception is the 2 May 2006 dryline, for which LBB radar data were available only between 0000 and 0112 UTC 3 May; that is a fraction of the dryline lifetime. The exception was made because this was an intense, well-defined case, and because the location and propagation of this dryline at radar-data-void times could be readily estimated from the WTM data: the clear dryline passage allowed us to manually map the estimated dryline positions at various times, from which the speed and orientation were inferred. Widespread convection developed east of the dryline on this day, and a saturation point analysis indicates that after 2230 UTC the moistside air mass was part of a large cold pool of convective origin. The distribution of saturation points in Fig. 3a on the moist side of the boundary suggests that a range of equivalent potential temperature θ_e air masses is encountered before 2230 UTC, possibly due to daytime heating and/or the horizontal advection of lower- θ_e air from the east. After 2230 UTC, the saturation points roughly fall along a moist adiabat (strictly speaking, this should be a moist-virtual adiabat), indicating that the variation of saturation points is mainly due to evaporation, resulting in a more uniform θ_e (Fig. 3b). Therefore, the boundary after 2230 UTC was assumed to be an outflow boundary, and was excluded from this analysis.

The difference in air density ρ between stations ($\Delta \rho$) is expressed as a θ_v difference ($\Delta \theta_v$) as follows. Assuming that the differences are much smaller than the mean values (written with overbars), the ideal gas law implies that, for clear air (i.e., air free of liquid or ice particles),

$$\frac{\Delta\rho}{\overline{\rho}} = -\frac{\Delta\theta_{\upsilon}}{\overline{\theta}_{\upsilon}} + \frac{c_{\upsilon}}{c_{p}}\frac{\Delta p}{\overline{p}}.$$
 (1)

A scaling argument shows that the second term on the right, containing the pressure difference, Δp , has a magnitude of 10^{-4} , while the magnitude of the first term on the right often exceeds 10^{-2} , as will be shown below. Thus, the pressure perturbation term can be ignored. Also,

$$\theta_{\nu} = \theta [1 + 0.61 q_{\nu} / (1 + q_{\nu})] \cong \theta (1 + 0.61 q_{\nu}).$$

Thus, the relative density difference (or buoyancy) is

$$\frac{\Delta\rho}{\bar{\rho}} \cong -\frac{\Delta\theta}{\bar{\theta}} - 0.61\Delta q_{\upsilon}.$$
 (2)

As will be shown in section 3, typical near-dryline measurements during the afternoon indicate that the mT air mass has a lower θ_{ij} value, compared to the west side of the dryline and, thus, according to (1), it has a higher air density. This density difference is driven by temperature (the mT air mass is cooler), but it is smaller than that due to the temperature difference alone [first term on the right side of (2)], because the temperature effect is partially offset by the humidity difference across the dryline [second term on the right side of (2)]. Moist air is lighter than dry air at the same temperature and pressure, but it takes a 5–6 g kg⁻¹ mixing ratio difference to offset a 1-K potential temperature difference. The average Δq_{ν} for the 16 drylines in this study was 4.0 g kg⁻¹. The average $\Delta\theta$ was -2.3 K, and the average $\Delta \theta_{v}$ was -1.6 K.¹ These averages exclude the cases with a reverse temperature difference (cooler air on the west side), discussed below.

3. Results

a. Local dryline orientation

The humidity and air density contrast, as well as the propagation properties, are examined for the 16 drylines in this study. A variety of dryline finelines were observed; some were straight (as in Fig. 2) or arc shaped, others had one or more clear bends within the \sim 140 km LBB area of view. The most frequent mean orientation of the finelines near LBB was NNE–SSW (Fig. 4), specifically between 10° and 40° from north. The histogram in Fig. 4 reveals much variability, which is possibly related to dryline bulges and the orientation of horizontal convective rolls (HCRs). It also reveals a

¹ The difference between air masses is defined as (moist-side value) – (dry-side value). Thus, Δq_v is strictly positive, but $\Delta \theta_v$ can be of either sign.



FIG. 3. Saturation point analysis (Betts 1984) for two WTM stations (a) between 2015 and 2200 UTC 2 May 2006 (labeled as XPTS and XRLS in Fig. 2) and (b) between 0025 and 0210 UTC 3 May 2006 (labeled as XSLS and XWVS in Fig. 2). The stations were located on the moist side of the boundary during the respective periods.

secondary peak, with a NW–SE orientation (-45°) . Dryline segments may be parallel to the wind in the moist air just east of the dryline (e.g., Fig. 8 in Atkins et al. 1998), possibly merging with an HCR in the moist air such that their secondary circulations locally positively interfere. Dryline segments may also be aligned with the BL wind and HCR orientation in the dry air just west of the dryline (e.g., Buban et al. 2007). The primary and secondary peaks in dryline orientation near LBB (Fig. 4) roughly correspond with the mean wind direction on the dry and moist sides, respectively: for the 16 drylines in this study (and for the duration of the dryline fineline as seen by the LBB radar), the mean wind (u, v) was (-1.0, 6.3) m s⁻¹ for all WTM stations on the dry side and (-5.8, 6.0) m s⁻¹ for stations on the moist side, that is, a southerly wind (171°) on the dry side and a slightly stronger SE wind (136°) on the moist side. On one occasion the dryline was locally parallel with a series of HCRs embedded in SE flow on the moist side; this dryline segment later advected out of the LBB domain. The tendency for a dryline to locally line up with the wind on the moist side matters to fore-



FIG. 4. Histogram of local dryline orientations within the area of view of the LBB radar. A north-south orientation is 0°.

casters because it implies that the adjacent dryline segments are more normal to the wind. Such segments tend to have stronger convergence (Atkins et al. 1998) and thus a higher probability of convection initiation (Ziegler and Rasmussen 1998; Ziegler et al. 2007). Another possible cause for the NNW–SSE orientation of some finelines is related to the fact that these finelines were mostly located to the northeast of LBB, where the Caprock Escarpment is best developed and assumes a NNW–SSE orientation. In any event, because most drylines are roughly N–S oriented, their propagation is simply referred to as eastward (positive) or westward (negative). More precisely, a positive propagation speed is eastward or northward, by definition.

b. The 2 May 2006 dryline

One particularly vigorous dryline occurred on 2 May 2006 (Fig. 5). The dryline was located in or just ahead of a synoptic-scale trough that stretched from northern Mexico to west Kansas (Fig. 6), and was associated with strong confluence on the synoptic scale and in the vicinity of LBB (Fig. 5). Before 2230 UTC the dryline slowly propagated eastward and became stationary; after 2230 UTC it was occluded by a westward-propagating outflow boundary resulting from a series of storms to the east of the dryline (Fig. 5). Some of these were severe, including one tornadic storm at 2240 UTC at 130 km to the northeast of LBB. The fineline in Fig. 5 was not the result of any individual storm seen to the east but probably resulted from the merger of several cold pools. The first storms east of Lubbock formed

around 2000 UTC. Cloudiness and some precipitation were present as early as 1800 UTC (just before local solar noon); thus, cooling by cloud shading may have contributed to the widespread cold pool. The extent of this cold pool at 0000 UTC can be seen in Fig. 6.

The passage of the outflow boundary at Slaton, Texas (XSLS), between 0000 and 0100 UTC is obvious in Fig. 5: the temperature dropped from 33° to 26°C, the dewpoint increased from -5° to 11° C, and the wind switched from 10 kt from 260° to 25 kt from 150°. The XSLS station pressure rose by 2.6 hPa during the same period. Most of these changes occurred within 5 min. The contrast across the dryline is also broadly apparent in the soundings: on the west side of the dryline (AMA in Fig. 7a), the dewpoint was below 0°C and the convectively mixed BL reached up to 500 mb, that is, 4700 m above ground level. The momentum was well mixed as well, with 8-11 m s⁻¹ westerly flow throughout the BL (Fig. 7b). On the east side of the dryline [MAF (Midland–Odessa) in Fig. 7a], a moister, cooler BL was present. (It is not known whether this was the original mT air mass, or whether it was further cooled by convection.) The BL was within southeasterly flow, and strong northwesterly wind shear was present over the lowest 2 km (Fig. 7b). This shear vorticity was probably of solenoidal origin.

The depth of the outflow-induced cold pool near LBB on 2 May 2006 can be estimated from observations for the pressure jump ($\Delta p = 2.6$ hPa) and θ_v drop ($\Delta \theta_v = -6.6$ K) at XSLS between 0000 and 0100 UTC. Hydrostatic balance dictates that the pressure jump Δp



FIG. 5. As in Fig. 2 but for 2 May 2006 at (a) 0000 and (b) 0100 UTC 3 May 2006. XSLS is highlighted. The fineline in (a) and (b) is a westward-propagating outflow boundary, which had occluded the surface dryline around 2230 UTC.



FIG. 6. Synoptic situation of the 2 May 2006 dryline. The frontal analysis is the official version from the National Weather Service. The official analysis of the dryline and the outflow boundary is inaccurate, so these boundaries have been redrawn, based on additional surface station data, radar data, and soundings. The sea level pressure contours (2-hPa interval) are based on the North American Mesoscale (NAM) model initialization.

is proportional to the magnitude of $\Delta \theta_v$ and the depth of the cold pool (δ) on the moist side. Assuming, for simplicity, that both $\Delta \theta_v$ and θ_v are height independent up to the height δ , and that the pressure is uniform at this height ($p = p_{top}$), the hydrostatic equation can be linearized to obtain

$$\Delta p = p_{\rm top} \frac{\delta}{H} \frac{(-\Delta \theta_{\nu})}{\overline{\theta}_{\nu}} e^{(\delta/H)}, \qquad (3)$$

where *H* is the scale height of the atmosphere. This yields a cold-pool depth δ of 1.0 km. The depth corresponds with reported depths of the cooler moist air mass east of the dryline, ranging between 0.6 and 1.5 km (NSSP Staff 1963; Bluestein et al. 1990; Ziegler and Hane 1993; Hane et al. 1997; Atkins et al. 1998; Ziegler and Rasmussen 1998). The cold-pool depth estimate δ can be compared also with the observed value at MAF (Fig. 7a), although the exact depth of the cooler layer cannot be determined, because of the poor resolution of the sounding.

c. Dryline baroclinicity

With the exception of a few cases, the virtual potential temperature θ_v was lower in the mT air mass for all 16 drylines, at any time before sunset (Fig. 8). The daytime θ_v deficit averaged 1.6 K and exceeded 4.0 K in some cases, including on 2 May 2006 before the dryline occlusion. The mixing ratio difference averaged 4 g kg⁻¹ but sometimes was more than twice as large. Generally, the drylines with a larger humidity contrast also had a larger $\Delta \theta_v$ (Fig. 8); that is, they were more baroclinic. The IDV imagery suggested that drylines with a larger Δq_v and larger $\Delta \theta_v$ tended to be associated with more confluent flow, and better-defined radar finelines. These relationships have not been quantified.

The θ_v deficit on the moist side of the 16 drylines in this study generally increased slightly with time until ~1800 CST (0000 UTC), and it vanished suddenly near sunset (Fig. 9). Three drylines were observed in the LBB coverage area well after sunset. In each of these cases, a θ_v deficit developed on the *dry* side, with a magnitude of up to 6 K. This reversal is due to more



FIG. 7. (a) Skew-*T* diagram and (b) hodograph for AMA (black lines) located 12 km west of the dryline, and MAF (gray lines), located farther south and \sim 75 km east of the dryline, at 0000 UTC 3 May 2006.

rapid evening surface cooling on the dry side than in the mT air mass. It should be noted that these findings are based on surface data. The afternoon θ_v deficit probably applies over the depth of the moist-side BL, while the θ_v deficit on the dry side after sunset probably only applies to a much more shallow layer.

The evolution of Δq_v is not as well behaved as that of

 $\Delta \theta_{\nu}$. In particular, the humidity contrast did not change much around sunset (Fig. 10). It tends to be largest in the late afternoon (1800 CST), at the time that the radar fineline is best defined, and the finescale convergence strongest. The evening decrease of the humidity contrast is rather gradual.

The gradual increase in $\Delta \theta_v$ toward 0000 UTC has



FIG. 8. Scatterplot of the difference in mixing ratio vs the difference in virtual potential temperature (moist side - dry side), between stations on opposite sides of the dryline, at any time before sunset. The linear regression line is shown as well.

been documented previously for select dryline case studies using detailed aircraft data (see Figs. 13 and 14 in Miao and Geerts 2007). To our knowledge the rapid decrease of $\Delta \theta_v$ after 0000 UTC, and its sign reversal after sunset, have not been documented in the literature. As mentioned in section 1, several case studies have mentioned a θ_v deficit on the moist side of the dryline (NSSP Staff 1963; Bluestein et al. 1990; Parsons et al. 1991; Ziegler and Hane 1993; Hane et al. 1997; Atkins et al. 1998; Ziegler and Rasmussen 1998), but all these observations were taken between ~2100 and 0100 UTC, that is, before sunset. One exception can be found in Crawford and Bluestein (1997): their Fig. 13 describes a westbound (retrograding) dryline passage at 0140 UTC 12 May 1991, about the time of sunset. This passage was marked by a 1.5-K $\Delta \theta_{v}$ increase, implying that the moist side had a higher θ_{v} .

d. Dryline baroclinicity and propagation

It is well known that under synoptically quiescent conditions the dryline tends to move eastward during the daytime and westward in the late afternoon and evening (e.g., Bluestein 1993, p. 284; Crawford and Bluestein 1997). The 16 drylines in this study confirm



FIG. 9. Trend of $\Delta \theta_{\nu}$ across the drylines. The times of solar noon and sunset (both on 1 and 31 May) in Lubbock are indicated.



FIG. 10. As in Fig. 9 but for the humidity difference, Δq_{v} .

this tendency (Fig. 11).² There is much scatter though, possibly because the fair-weather drylines have not been isolated from the total sample. The 2 May 2006 event also shows this reversal in dryline motion just before the dryline became occluded (section 3b).

To test whether the dryline behaves as a density current, the "relative" dryline propagation speed $(U_{\rm rel})$ is estimated, that is, relative to the mean wind $(U_{\rm env})$ in the dryline vicinity. Here, $U_{\rm env}$ is assumed to be the average wind at the 18 WTM stations used in this study. There is some arbitrariness in this definition. One could base this average on the dry-side stations only, with the assumption that the moist side is the perturbation (the density current). But this makes the number of stations feeding the average time dependent, since the dryline moves, and imports too much westerly momentum into the average in the afternoon hours.

The component of $U_{\rm env}$ normal to the dryline is subtracted from the fixed-frame dryline propagation speed $U_{\rm fix}$; that is, $U_{\rm rel} = U_{\rm fix} - U_{\rm env}$. Laboratory experiments (summarized in Simpson 1999) suggest that the speed of a density current ($U_{\rm dc}$) can be estimated as follows:

$$U_{\rm dc} = K \sqrt{\frac{g D \Delta \theta_{\rm v}}{\overline{\theta}_{\rm v}}},\tag{4}$$

where g is the gravitational acceleration, D is the depth of the density current (m), and K is a constant; experi-

mental values for K for atmospheric density currents range between 0.7 and 1.0 (e.g., Wakimoto 1982; Mueller and Carbone 1987; Kingsmill and Crook 2003). A value of K = 0.7 is used. Some form of (4) is commonly encountered in the literature, including in atmospheric applications. Equation (4) assumes that there is no mean flow. The density current can be assumed to be Galilean invariant; that is, the effective density current speed $(U_{dc,fix})$ equals the speed of the current embedded in the ambient flow: $U_{dc,fix} = U_{dc} + U_{env}$. Based on laboratory observations, Simpson and Britter (1980) proposed an effective speed of $U_{dc,fix} = U_{dc} + 0.7U_{env}$. This reduction accounts for the effects of surface friction. In the present study, the observed $U_{\rm rel}$ is compared with U_{dc} [computed according to (4)]; thus, it is assumed that the flow is Galilean invariant.

The observed relative propagation speed $U_{\rm rel}$ of the 16 drylines in this study is plotted against $\Delta \theta_v$ in Fig. 12. Negative values of $U_{\rm rel}$ indicate westward (or southward) motion relative to the ambient flow. Also shown is the dependence of $U_{\rm dc}$ on $\Delta \theta_v$, according to (4), assuming three different values of *D*. During the daytime, $|U_{\rm rel}|$ increases with $|\Delta \theta_v|$, and the increment in $|\Delta \theta_v|$ on the ordinate for a given $|U_{\rm rel}|$ increment on the abscissa increases with $|\Delta \theta_v|$; that is, the relationship is not linear, which is consistent with (4). The least squares fit quadratic equation for all points with $\Delta \theta_v < 0$ in Fig. 12,

$$\Delta \theta_v = a + b U_{\rm dc}^2$$

(with a and b constants derived from a regression), gives a value for a = -0.019 K and b = 0.087 K s² m⁻². Thus, the relationship between the observed relative

² The number of points in Fig. 11 is much lower than in previous figures because the dryline fineline displacement could only reliably be determined at hourly intervals.



FIG. 11. Trend of the propagation speed of the dryline fineline.

dryline propagation speed and $\Delta \theta_{v}$ is indicative of density current behavior. For a given $\Delta \theta_{v}$ the observed speed generally falls between the theoretical values for D = 500 and 1000 m. The density current depth is unknown, and may not be the same as the depth of the mT air mass. It may be shallower than 1000 m, particularly during the retrograding stage. One case study in which the dryline appeared as a density current, and for which detailed measurements of D were available, is the 6 May 1995 case during the Verification of the Origin of Rotation in Tornadoes Experiment (VORTEX). Both Figs. 14 and 16 in Atkins et al. (1998) and Fig. 12 in Ziegler and Rasmussen (1998) indicate that D is about 900 m at the time of the detailed dryline observations. The 22 May 2002 dryline during the International H₂O Project (IHOP) also appears as a density



FIG. 12. Scatterplot of the dryline propagation speed relative to the ambient flow (U_{rel}) against $\Delta \theta_{v}$. Also shown are the least squares fits for the points with $\Delta \theta_{v} < 0$, and a theoretical dependence of density current speed on $\Delta \theta_{v}$ for three values of density current depth *D*.

current, with a depth of about 600 m behind the leading head (Fig. 13 in Buban et al. 2007).

Computing the relative propagation speed as $U_{\rm rel} = U_{\rm fix} - 0.7U_{\rm env}$, as in Simpson and Britter (1980), does not substantially improve the match between the observations and the theoretical estimate in Fig. 12. In any event, the relationship between the relative dryline propagation speed and $\Delta \theta_v$ shown in Fig. 12, before sunset, is a strong confirmation of the relevance of density current dynamics.

No such relationship between $U_{\rm rel}$ and $\Delta \theta_{v}$ is apparent after sunset ($\Delta \theta_v > 0$) (Fig. 12): in some cases, $U_{\rm rel}$ is negative (westbound), but generally it is near zero. As mentioned before, the surface measurement of $\Delta \theta_{\nu}$ becomes less relevant after sunset because of the shallow radiation inversion that forms, especially on the dry side. This inversion is much shallower than the moist BL depth or the density current depth. The residual dry mixed layer overlying this inversion changes only very slowly after sunset, and probably retains a higher θ_{v} than the moist BL to its east for some time after sunset. In terms of solenoidal and density current dynamics, the deeper $\Delta \theta_{\nu}$ is more important than that inferred from surface measurements. Thus, the $\Delta \theta_v$ measurements after sunset (Fig. 12) are of little significance to the dryline dynamics. It is possible that the dryline convergence zone travels westward within the residual mixed layer, decoupled from the nocturnal surface layer, as has been observed for sea breezes (Clarke 1965; Kraus et al. 1990).

4. Discussion

The analysis of this relatively small sample of drylines demonstrates that (a) the mT air mass is generally denser than the dry air mass during the formation and mature stages of the dryline, and (b) the formation of a dryline fineline convergence zone and its propagation are generally consistent with density current theory. The present analysis is consistent with a number of previous case studies in supporting these two conclusions (Bluestein et al. 1990; Parsons et al. 1991; Ziegler and Hane 1993; Hane et al. 1997; Atkins et al. 1998; Ziegler and Rasmussen 1998; Miao and Geerts 2007), but to the author's knowledge this has not previously been demonstrated with a 2-month sample of operational data in West Texas. A study of this kind has only become possible in recent years because of the WTM.

Convergence and uplift at the leading edge of a density-current-like dryline increases when $\Delta \theta_v$ is larger. Thus, ceteris paribus, the potential for lifted air to reach the level of free convection increases with $\Delta \theta_v$. In addition, the potential of convective initiation depends on how erect the updraft is above the dryline boundary (Ziegler and Rasmussen 1998; Ziegler et al. 2007), and this in turn may depend on the balance between baroclinically produced vorticity (which is a function of $\Delta \theta_{v}$) and the ambient vorticity associated with the vertical wind shear normal to the dryline.³ The present dataset lacks adequate cases and vertical structure data to test the Rotunno–Klemp–Weisman (RKW) theory. For one case, the 2 May 2006 dryline, the ambient horizontal vorticity was quite weak (see the AMA hodograph in Fig. 7b), weaker than the baroclinic vorticity (MAF hodograph in Fig. 7b). Only 1 of the other 16 drylines in the present study triggered deep convection in the vicinity of LBB.

In any event, forecasters trying to determine dryline "strength" and the likelihood of convective initiation along the dryline are encouraged to monitor the observed or modeled $\Delta \theta_v$ in West Texas, in addition to currently used indicators, such as convective inhibition. It is not coincidental that $\Delta \theta_{\nu}$ peaks at about the same time as does convective initiation along the dryline (e.g., the appendix in Bluestein and Parker 1993), that is, within a few hours before sunset (Fig. 9). Furthermore, the ratio of $\Delta \theta_{\nu}$ (computed over a distance of ~50 km and vertically averaged over the depth of the moist BL) to the dryline-normal vertical wind shear Δu (computed on the dry side over the same depth) could become a useful parameter for nowcasting convective initiation along the dryline, if this ratio exhibits a clear correlation to the occurrence of convection initiation. This would also prove that RKW theory is applicable to drylines.

One finding not revealed by previous case studies regards the rapid decrease and sign reversal of the air density difference (or $\Delta \theta_{\nu}$) across the dryline near sunset (Fig. 9), although forecasters have long been aware of a reversal of the temperature gradient across the dryline near sunset (Doswell 1982). This reversal is largely due to the more rapid evening cooling on the dry side, compared to the moist side of the dryline (section 3c). This must relate to the surface heat balance, specifically the greater net longwave radiation loss in the evening on the dry side, on account of the loweratmospheric water vapor content. Similarly, the buildup of $\Delta \theta_v$ during the afternoon (with cooler air on the moist side) probably relates to differences in the surface heat balance, in particular the dominance of the latent heat flux over the sensible heat flux on the moist

³ The vorticity balance theory for the initiation and maintenance of convection was first proposed by Rotunno et al. (1988) and therefore is referred to as the RKW theory. It was revisited by Weisman and Rotunno (2004).

Retrogression commonly occurs, especially in the mature and decaying stages of the dryline (Fig. 11), in all but two of the drylines in this study. (Both of these exceptions are marked by an approaching cold front.) The question may be asked how, in the absence of synoptic flow, a density (or gravity) current can travel uphill, westward up the slope of the Great Plains. That terrain slope α is about 100 m (100 km)⁻¹, or $|\tan \alpha| \sim 10^{-3}$. The horizontal (terrain following) surface pressure gradient force (PGF) induced by the density differences is [based on Mahrt (1982) and applied to the dryline situation]

$$PGF = \frac{g\Delta\theta_{\nu}}{\overline{\theta}_{\nu}}\sin\alpha + \frac{g}{\overline{\theta}_{\nu}}\frac{\partial(D\theta_{\nu})}{\partial x}\cos\alpha.$$
 (5)

This PGF drives the zonal flow u (Du/Dt = PGF). There may be other forces, such as the Coriolis force, friction, and the regional PGF, which is not a hydrostatic product of $\Delta \theta_{n}$ in the BL. These are ignored, for simplicity. Mahrt (1982) developed the gravity flow equations for downslope (drainage) flow but they can be applied to the local upslope flow as well. The first term in (5) drives a layer of denser air on the sloping Great Plains eastward ($\Delta \theta_{v} < 0$, $\sin \alpha < 0 \rightarrow PGF > 0$, and Du/Dt > 0, until the isentropes are level ($\Delta \theta_n = 0$). The second term in (5) is a hydrostatic pressure gradient: It is due to the greater weight of the denser fluid over a depth D. It allows for westward (upslope) acceleration if denser air occurs to the east $(\partial \theta_{\nu}/\partial x < 0)$, $\cos \alpha > 0 \rightarrow Du/Dt < 0$), assuming D is constant. For the typical slope of the Great Plains, (5) can then be written as

$$PGF = \frac{g}{\overline{\theta}_{\nu}} \left(10^{-3} \Delta \theta_{\nu} + D \, \frac{\partial \theta_{\nu}}{\partial x} \right). \tag{6}$$

For a $\Delta \theta_v$ of 2 K over $\Delta x = 20$ km (typical values found in this study) and D = 1000 m, the first term between the brackets in (6) is 100 times smaller than the second. Thus, upslope gravity flow (Du/Dt < 0) is possible, but limited of course to the point where the θ_v contours become level.

The dryline retrogression speed does seem to wane in some cases after sunset, according to Fig. 11. Two mechanisms for the termination of westbound propagation may occur. First, if $\Delta \theta_{\nu}$ does not change, the slope of the terrain imposes a limit, as discussed above. Second, $\Delta \theta_{\nu}$ does change at the surface (Fig. 9) and, more gradually, over a depth *D*, and this is sufficient to stall the westward propagation and initiate shallower westerly katabatic flow.

5. Conclusions

Reflectivity imagery from the Lubbock, Texas, WSR-88D radar and WTM station data were used to detect and describe drylines in the months of May 2005 and May 2006. Sixteen drylines were detected. The key findings are as follows:

- (i) The moist air mass east of the dryline was consistently denser than the dry air mass during the formation and mature stages of the dryline; the difference in virtual potential temperature across the dryline $(\Delta \theta_{\nu})$ typically was 1–2 K over a distance of about 50 km.
- (ii) This surface air density difference peaked in the late afternoon and rapidly weakened, and even reversed, around sunset. This weakening and reversal are attributed to more rapid evening radiative cooling at the surface on the dry side.
- (iii) The formation of a dryline convergence zone and a radar fineline, as well as the propagation of the dryline, are generally consistent with density current theory. According to this theory, the convergence and uplift along the dryline are stronger when $\Delta \theta_{\nu}$ is larger.

Acknowledgments. This work was supported by National Science Foundation Grant ATM-0444254. It benefited from the comments by two anonymous reviewers. Conrad Ziegler's comments were more in-depth and influential than those typically received from coauthors in a multiauthor paper. The data were collected by Kyle D. Wright as part of his undergraduate senior research project.

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