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A Dryline in Southeast Wyoming. Part I: Multiscale Analysis Using Observations and Modeling on 22 June 2010

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ABSTRACT

A first observational and modeling study of a dryline and associated initiation of deep convection over the high plains of southeastern Wyoming is presented. Radar and station measurements show that the dryline is a well-defined convergent humidity boundary with a modest density (i.e., buoyancy) gradient. Its development, intensity, and movement are regulated by the terrain, diurnal land surface and boundary layer processes, and synoptic-scale evolution. At least one of the thunderstorms that emerged from the dryline became severe. Weather Research and Forecasting Model (WRF) simulations accurately reproduce measured aspects of this dryline, as well as the timing and location of convection initiation. The WRF output is used further to characterize the dryline vertical and horizontal structures and to examine convection initiation processes. A dryline bulge over a local terrain ridge appears to be an essential ingredient in convection initiation on this day: just north of this bulge the surface convergence and buoyancy gradient are strongest, and deep convection is triggered. In this region especially, the WRF simulation produces horizontal convective rolls intersecting with the dryline, as well as small cyclonic vortices along the dryline. In fact, the primary storm cell initiates just downwind of one such vortex. Part II of this study describes the finescale vertical structure of this dryline using airborne Raman lidar data.

1. Introduction

a. Background

A mesoscale nonfrontal boundary that separates two air masses with distinctly different humidity is commonly referred to as a dryline (DL; Markowski and Richardson 2010). The transition across a well-defined DL typically is meso-γ (2–20 km) in scale, whereas its length is on the meso-α (200–2000 km) scale (Ziegler et al. 1997).

DL formation east of the Rocky Mountains is tied to the sloping terrain and depends on synoptic conditions. DLS generally form due to lee troughing, resulting in moist-air advection northwestward from the Gulf of Mexico and dry-air advection eastward from the intramountain region (Schaefer 1974b, 1986). A well-defined surface convergence zone forms during the daytime, with a humidity contrast and wind shift (Schultz et al. 2007). The DL boundary can be viewed at the intersection of the shallow moist boundary layer (BL) with topography rising to the west (Fig. 2.43 in Bluestein 1993). The meso-γ-scale convergent flow leads to a pronounced dewpoint gradient, sometimes >10 K km⁻¹, as frontogenetic flow superimposes on large-scale flow (Markowski and Richardson 2010). This frontogenetic flow is driven by an air density (or buoyancy, or virtual potential temperature) difference that is affected by mesoscale surface heat fluxes and boundary layer mixing depth (Ogura and Chen 1977; Geerts 2008), which explains why the DL is strongly diurnally modulated and is best defined in the afternoon (Schaefer 1974b).

Because of zonal (east–west) differences in surface sensible heating and zonal differences in the depth of the shallow moist layer associated with the sloping terrain, the DL usually displays a diurnal pattern of zonal movement under weak synoptic forcing (Schaefer 1974a, 1986): DLS tend to progress toward the east during the day, due to BL mixing, and retrogress to the west from the late afternoon into the night, initially due to the BL density difference (Schultz et al. 2007).

The interest in studying DLS mainly stems from their ability to initiate deep convection, which may lead to the development of severe thunderstorms. Convection
initiation (CI) along DLs is difficult to predict and has been the main motivation for several field campaigns (e.g., Hane et al. 1997; Ziegler and Rasmussen 1998; Weckwerth et al. 2004) and modeling studies of the DL (e.g., Ziegler et al. 1997; Shaw et al. 1997). Although it is generally understood that mesoscale mass and moisture convergence along DLs explains CI, the ability of embryonic deep convection to be sustained beyond its level of free convection (LFC) depends on more complex factors such as vertical wind shear and small-scale along-DL variability (McCarthy and Koch 1982; Koch and McCarthy 1982; Hane et al. 1997, 2002; Murphey et al. 2006; Xue and Martin 2006a,b). These issues as well as uncertainties about DL strength and movement make it difficult to predict if, when, and where thunderstorms will develop.

The characteristics and climatology of DLs in the U.S. southern Great Plains (SGP) have been thoroughly studied. SGP DLs are most frequently observed during the spring and early summer, near an average longitude of 101°W (Hoch and Markowski 2005). DLs are not limited to the SGP, however. The term dryline has been used for boundaries farther north along the Rockies’ eastern flanks, as far north as Alberta, Canada (Hill 2006; Taylor et al. 2011), and elsewhere in the world (Weston 1972; Schaefer 1986). McCaul and Blanchard (1990) mention that SGPs DLs may continuously extend into Wyoming (WY) and Nebraska (NE). However, DLs in WY and NE have hardly been studied. This serves as motivation for this study, a case study of a DL in southeast WY (SE WY).

b. Drylines in southeast Wyoming

The meridional extent and movement of DLs along the eastern flanks of the central Rockies is strongly affected by the details of the terrain (Fig. 1). Westward DL propagation along the Colorado (CO) Front Range (FR) is generally blocked by the high terrain, as the moist layer is too shallow to scale the steeply ascending terrain. A cool, humid, and shallow air mass often resides over the CO high plains in spring, with continuous advection of dry, potentially warmer air from the west above this shallow layer. The latter air mass is heated by the elevated terrain below, and becomes decoupled from the surface along a well-defined surface humidity boundary at some elevation along the rather steep eastern flanks of the FR. This boundary may be quasi-stationary in CO, but commonly propagates across the more gradually sloping high plains of SE WY. Such a propagating baroclinic boundary usually is not analyzed on surface charts, and if it is, it is identified as a front. In those cases where the boundary’s movement is primarily controlled by diurnal surface heating and BL mixing, such a boundary should be analyzed instead as a DL, according to a preliminary climatological study of DLs in SE WY (P. Bergmaier 2013, unpublished manuscript, hereafter PB).

That study finds that DLs in SE WY typically form as a postfrontal air mass settles over the Great Plains. Thus, the synoptic environment of DLs in SE WY is different from those in the southern high plains: SE WY is generally north of the climatological lee low pressure center east of the central Rockies, whereas the southern high plains are south of it. DLs in the SGP tend to become best defined in the warm sector, as subsiding westerly flow aloft produces a lee trough (Hoch and Markowski 2005). They tend to follow the surface pressure trough as a cold front approaches (e.g., see Hane et al. 2001). In contrast, DLs in SE WY may form under quiescent conditions in the wake of a postfrontal high. In other cases, as in the present case study, DLs form within a lee trough extending from Texas (TX) northward into SE WY or western NE. In either case, weak upslope flow is present to the east of the developing DL in SE WY.

No DL was analyzed in SE WY for this case (22 June 2010) on the National Weather Service (NWS) Hydro-meteorological Prediction Center (HPC; http://www.hpc.ncep.noaa.gov) surface analysis (Fig. 2). Yet, as shown below, a mesoscale humidity boundary developed and propagated across the high plains of SE WY on this day. Deep convection initiated along this DL, and at least one storm evolved into a severe thunderstorm.
NWS forecasters referred to the developing boundary as a DL in their area forecast discussion (http://mesonet.agron.iastate.edu).

This study presents a multiscale analysis of a DL on 22 June 2010 in SE WY. Section 2 provides a brief overview of the data sources. This study progressively zooms in: section 3 focuses on synoptic and mesoscale observations to investigate the characteristics of the DL environment. Section 4 validates a high-resolution numerical simulation and analyzes the model output in terms of DL structure and CI. Bergmaier et al. (2013, manuscript submitted to Mon. Wea. Rev., hereafter Part II) then describe the finescale structure of the DL, using data from an airborne Raman lidar.

2. Data sources

The synoptic overview is based on 12-km North American Mesoscale Model (NAM) data (http://nomads.ncdc.noaa.gov), Storm Prediction Center (SPC) plots (http://www.spc.noaa.gov), and soundings from the University of Wyoming (http://www.weather.uwyo.edu). Geostationary Operational Environmental Satellite-13 (GOES-13) data were provided by the University of Wisconsin–Madison (http://www.ssec.wisc.edu). Weather Surveillance Radar-1988 Doppler (WSR-88D) data were obtained from the National Climatic Data Center (NCDC) Next Generation Weather Radar (NEXRAD) data inventory search (http://www.ncdc.noaa.gov), while MesoWest surface station data were obtained from the University of Utah (Horel et al. 2002; http://mesowest.utah.edu). One-minute Automated Surface Observation System (ASOS; ftp://ftp.ncdc.noaa.gov) data were obtained from NCDC. The Advanced Research core of the Weather Research and Forecasting Model (WRF-ARW; http://wrf-model.org) (Skamarock et al. 2008) was initialized using 12-km NAM data.

3. Synoptic and mesoscale observations

a. Synoptic environment

The 22 June 2010 DL in SE WY was best defined between 1800 and 0000 UTC; thus, we examine the synoptic environment in that period. A DL (scalloped line) was analyzed in a lee trough from eastern CO to the south, but this DL did not extend from the CO low along a trough to the northwest into SE WY (Fig. 2). The CO lee depression was associated with a low-amplitude 500-hPa short-wave trough progressing eastward over WY and CO (Figs. 3a,b). A weak 250-hPa jet streak was present downstream of the 500-hPa short wave at
1800 and 0000 UTC. The intensification of the lee low toward 0000 UTC (Figs. 3a,b) is due to diurnal surface heating rather than to synoptic changes. The surface winds became more cyclonic around this surface lee low, drawing moist air from the central Great Plains, as well as relatively drier air from the southwest. A 48-h backward-trajectory analysis (not shown) shows that BL air just west (east) of the DL at 0000 UTC 23 June arose from within the BL in Arizona (Oklahoma). This convergent zonal flow $u$ in SE WY amplified the gradient in the surface specific humidity $q_v$ (Figs. 3c,d). The coincident gradients in $u$ and $q_v$ extended southward from WY into TX, but they were strongest in SE WY at 0000 UTC. At 0000 UTC, the absolute values of the $u$ and $q_v$ gradients along the southeast border of WY were about 2 m s$^{-1}$ (100 km)$^{-1}$ and 3 g kg$^{-1}$ (100 km)$^{-1}$, respectively, according to the 12-km NAM data. Given the analysis of Fig. 3, this DL in SE WY is deemed to occur under a synoptically active environment. Under these conditions, the upper-level flow plays a significant role in the formation and motion of the DL (Hane 2004; Schultz et al. 2007). This will be revisited later.

The convective potential on 22 June 2010 is examined in Fig. 4. The 12-km NAM indicates that there was moderate surface-based convective inhibition (CIN) in SE WY ($\sim$200 J kg$^{-1}$) at 1800 UTC, mostly vanishing toward 0000 UTC (Figs. 4a,b). The surface convective available potential energy (CAPE) increased dramatically across the convergence zone in SE WY, especially
at 0000 UTC (Figs. 3d and 4b), when CAPE reached a local maximum of ~2000 J kg$^{-1}$ in SE WY. Farther east, CAPE values reached 2900 J kg$^{-1}$ according to the 0000 UTC North Platte, NE (LBF), sounding, but CIN remained significant there due to a capping BL inversion (Fig. 4d). On the west side of the surface convergence zone [e.g., in Riverton, WY (RIW)], both CAPE and CIN were essentially absent because of a lack of low-level moisture (Fig. 4c). Penetrative cumulus (Cu) convection would be possible, but the convection would be benign. The LBF sounding (Fig. 4d) and SPC analysis (not shown) indicate a region of strong 0–6-km vertical wind shear from SE WY into NE. If the capping inversion at LBF at 0000 UTC was flat, then the shallow moist layer would intersect the terrain about 10 km west of Cheyenne in SE WY (location shown in Fig. 5). A much drier elevated mixed layer is present above the capping inversion at LBF, between 730 and 625 hPa (Fig. 4d). This is the residual of a surface-based convective BL over the elevated terrain of CO, advected northeastward and becoming decoupled from the surface. In summary, Fig. 3 supports the presence of a DL in SE WY, with a high potential for deep convection and even severe thunderstorms from far eastern WY eastward across NE, according to Fig. 4.

b. Dryline identification and the radar fineline

In the warm season, when insects are present in the BL, convergent DLs may be observed using radar data at a low elevation angle (e.g., Geerts 2008). The radar...
echoes are due to scattering by insects that have con-
gregated in regions of convergent flow. Because the
convergence zone is spatially thin, it is commonly re-
ferred to as a radar "fineline" (Russell and Wilson
1997). The reflectivity of clear-air echoes is generally
weak, at most 20 dBZ (Rinehart 1997). The fineline
depth depends on the depth of the convective BL, typ-
ically 1–2 km (e.g., Miao and Geerts 2007).

Figure 5 shows the 0.5° base reflectivity from the
Cheyenne, WY (KCYS), and Denver, CO (KFTG),
WSR-88Ds at 2000 UTC. The radar return indicates
a fineline that extends from SE WY southward along
the FR in northern CO. Two distinctly different air
masses converge toward the fineline at 2000 UTC (Fig. 5).
Station dewpoint temperatures $T_d$ are lower on the
west side of the fineline ($T_d \approx -9^\circ$C; 264 K), as
compared to the eastern air mass ($T_d \approx 13^\circ$C; 286 K).
The large humidity gradient across this fineline suggests
that it is a DL. This fineline is also evident farther south
in the 0.5° base reflectivity maps for the Pueblo, CO
(KPUX), and (discontinuously because of the larger dis-

cance) Amarillo, TX (KAMA), radars (not shown). In
northern CO, the fineline appears along the eastern
flanks of the FR (Fig. 5), where the cooler moist air
intersects the terrain. We are confident this boundary
can be called a DL, as it does not remain stationary.
Although there is some uncertainty as to the continuity
of the DL boundary in CO, the surface analysis (Fig. 2)
indicates that the DL is present in southeastern CO and
near the TX–New Mexico (NM) state line.

No thunderstorms were present at this time (2000 UTC),
so this fineline was not a cold pool boundary. A sheared
cumulus congestus started to produce precipitation just
northwest of KCYS (red arrow in Fig. 5) at this time.
This incipient storm (which we call storm 1) is evident
in visible satellite imagery (not shown), and from the
forward camera aboard a research aircraft (see Part II).
The fineline reflectivity was most intense along the
Laramie Range (LR; Fig. 1), reaching values up to
20 dBZ. The fineline in SE WY corresponds well with
a line of confluence indicated by the KCYS 0.5° Doppler
velocities (not shown), and, more coarsely, the Meso-
West station wind barbs, as there is a distinct wind shift
between the southwest winds at Laramie, WY (KLAR),
and the east winds at KCYS (Fig. 5). It is not by chance
that the location and meridional orientation of the
fineline agrees well with the topography of the LR. In SE WY, topography has a strong influence on the location of the fineline, given the relatively steady terrain rise from 1400 m near the WY–NE state line to 2400 m across the LR.

The DL echo diminishes at a range of 75 km northwest of KCYS, where the boundary layer coverage ends as the 0.5° beam becomes too elevated. Furthermore, it is difficult to determine the DL location given the scarcity of station observations in SE WY, compared to west TX, for instance.

c. Dryline characteristics and motion

The dry air mass to the west of the FL was less dense than the moist air mass, as the virtual potential temperature \( \theta_v \) was 2–5 K higher (Fig. 5). This positive \( \theta_v \) difference between KLAR and KCYS (locations shown in Fig. 5) developed around 1600 UTC, peaked around 2100 UTC, and vanished at 0012 UTC, when the DL progressed eastward across KCYS (Fig. 6). If the \( \theta_v \) gradient across the DL was of significant depth, then the gradient suggests a solenoidal forcing and a frontogenetic “inland sea breeze” circulation pattern with the less dense dry air rising over the moist air mass (Ogura and Chen 1977; Sun and Ogura 1979; Sun and Wu 1992). This frontogenetic circulation is relatively finescale, \( O(10 \text{ km}) \) (Miao and Geerts 2007; Sipprell and Geerts 2007), and brings the moist mixed layer in close proximity to the dry air west of the DL.

Terrain is important; stations on the west side of the DL are generally at higher elevations than are those on the east side in Fig. 5. The convergent flow between KLAR and KCYS may be influenced by a thermally forced mountain–plain circulation pattern (Zardi and Whiteman 2012) driven by surface heating over the LR. Such circulation is observed across individual mountain ranges under light wind conditions (Banta 1984). On a larger scale, there is confluent flow into the Rocky Mountain lee trough that appears to coincide with the DL, according to mesoscale analyses (Fig. 5 and Fig. 7) and the KCYS meteogram (pressure trend in Fig. 6b). Such coincidence is common in the SGP (e.g., Hane et al. 2001).

The DL passage at KCYS around 0012 UTC is accompanied by a sustained drop in humidity and wind shift from easterly to westerly (Fig. 6b), and is coincident with an increase in station pressure. Thus, the eastward movement of the DL was not only influenced by stronger westerly flow behind, but also by the eastward progression of the 500-hPa short-wave trough (Fig. 3). The brief spike in \( \theta_v \) at KCYS just after the DL passage is a significant anomaly in the diurnal temperature cycle and is an indication of residual baroclinicity across the DL. The positive \( \theta_v \) difference across the DL usually disappears in the evening (Geerts 2008). A second sustained wind shift from northwest to northeast, heralding a cooler, moister air mass, starts at 0340 UTC (Fig. 6b). This change was due to the passage of an outflow boundary from a thunderstorm northwest of KCYS, and not a retrograding DL.

The DL remained quasi stationary between 1600 and 2000 UTC (Fig. 6a), as the radar fineline continued to become better defined 5–10 km east of the crest of the LR, from where the terrain continuously slopes down to the east. As will be shown in Part II, the moist air mass remained too shallow to advance westward across the LR. Between 2000 and 0100 UTC, the DL moved and accelerated eastward. Isochrone analysis of the DL position after 0100 UTC was impossible, as the fineline had disappeared and because of the low station density. A mesoscale DL bulge emerged west of KCYS between 2300 and 0000 UTC (Fig. 6a). Between 2300 and 0000 UTC, the DL moved eastward at \( 5 \text{ m s}^{-1} \) in this bulge, but only half as fast at north and south locations marked in Fig. 6a. The bulge formed over the higher terrain of the Cheyenne Ridge (see Fig. 1), where the moist layer was shallower, and thus it is possible that the moist layer mixed out more rapidly to give way to dry westerly flow. Indeed, meso-\( \beta \)-scale bulges often occur due to local terrain or land surface aspects (Hane et al. 1997; Hane 2004). East of the FR, the fineline only started to progress eastward and thus became diagnosable as a DL around 0000 UTC. The average DL speed between 2000 and 0100 UTC (\( \sim 2 \text{ m s}^{-1} \)) was small compared to the \( u \) component of the surface environmental wind at 0000 UTC on the west side of the DL (\( \sim 8 \text{ m s}^{-1} \); Fig. 3d).

On a finer scale, the eastward DL progression had some discontinuities, as suggested by the ASOS 1-min \( q_v \) values as the DL passed by KCYS (not shown). Depressions in \( q_v \) were associated with positive \( \theta_v \) anomalies and bursts of westerly flow (\( u > 0 \)), suggesting a rather shallow moist layer with intermittent mixing down of westerly dry air aloft. The discontinuous drying upon DL passage is due to either vertical or horizontal eddy motions in the “mixing zone” (Mahrt 1991; Ziegler and Hane 1993; Hane et al. 1993; Crawford and Bluestein 1997).

d. Convection initiation

The KCYS WSR-88D 0.5° base reflectivity images between 2025 and 0032 UTC highlight developing convection in SE WY and NE (Fig. 7). The onset of CI, defined here as the first radar echo exceeding 40 dBZ, was observed in SE WY near 2025 UTC about 30 km east of the DL (Fig. 7a). The embryo of this storm
(“storm 1”) was evident at 2000 UTC (Fig. 5) closer to the DL. Deep convection typically first appears on radar a few tens of kilometers to the east of the DL in the SGP (Ziegler and Rasmussen 1998). The first heavy rain echo was closest to the DL just northeast of a minor mesoscale bulge that formed between 2000 and 2100 UTC (see Fig. 6a). This bulge may have served as a site for regionally enhanced surface convergence (\(\Delta u\) estimated at \(\sim10\) m s\(^{-1}\) over a few kilometers across the DL) and upward motion, as wind blew from the southwest on the dry side (i.e., more normal to the DL just north of the bulge). Some scattered thin stratocumuli were present in the shallow moist air mass just east of the LR crest around 1945 UTC (the time the research aircraft with
a forward camera first crossed the DL; see Part II), so the lifted condensation level (LCL) must have been low and close to the DL. GOES-13 visible imagery (not shown) indicates that the first appearance of convective clouds was as early as 1915 UTC, and that they were located at or within a few kilometers of the DL. Subsequently, the deepening convective towers were advected toward the northeast by midlevel winds. The deepening was possible by high CAPE values in the BL air just east of the DL (see Part II). The first deep convection clearly was sheared toward the northeast, with a v-shaped echo extending to the northeast (Fig. 7a). Along the southern segment of the DL (KGXY and KFNL), where the airflow was only weakly confluent (Fig. 7a) and the terrain lower, no CI occurred.

GOES-13 visible imagery (not shown) indicates that a band of deepening cumulus formed at about a 45° angle to the DL, starting near the highest part of the Cheyenne Ridge. By 1945 UTC, this band extended about 55 km northeastward from the DL. By 2025 UTC, the cold cloud tops in the GOES infrared imagery (not shown) suggest that a cumulonimbus anvil had developed within this band, extending ~100 km to the northeast. The heavy rain echo observed at 2025 UTC (Fig. 7a) was near the southwest end of the anvil.

During the development stages of convection, storm 1 (consisting of a succession of cells) remained quasi-stationary and intensified until about 2130 UTC. Storm 1 then started moving eastward, became severe by 2130 UTC, and produced tennis-ball-sized hail at 2200 UTC. Storm 1 continued eastward, intensified, produced a mesocyclone, and spawned a tornado by 2245 UTC just across the state line in NE (SPC storm reports). The leading edge of a well-defined outflow boundary trailed storm 1 at 2250 UTC, as shown by the newly formed fineline in the reflectivity data (Fig. 7b). A second storm (“storm 2”) emerged about 25 km northeast of the intersection between the DL and this outflow boundary at 2250 UTC (Fig. 7b). Another storm (“storm 3”) initiated in close proximity to the intersection of the apex of the mesoscale DL bulge and the outflow boundary from storm 1 (Fig. 7c). Enhanced low-level convergence and the necessary lift for air parcels to reach their LFC appear to have been present at this triple point. Storms 2 and 3 did not become severe as they remained over the cold pool in the wake of storm 1.
By 2332 UTC, the outflow boundary from storm 1 had grown and extended farther south (Fig. 7c). As this outflow boundary passed Kimble, NE (KIBM), between 2253 (Fig. 7b) and 2332 UTC (Fig. 7c), the temperature dropped 5 K, and the wind shifted nearly 180°. A saturation point analysis (Betts 1982) confirms evidence from the KCYS radar data that KIBM was within the cold pool from storm 1 after 2332 UTC (not shown).

4. WRF simulations

a. Model configuration

To better understand the finescale horizontal and vertical structures of the DL, its evolution, and CI along the DL, we employ the nonhydrostatic WRF-ARW (V3.3.1) for the 22 June 2010 case. The model is initialized and bounded by the 12-km NAM, where gridded binary (GRIB) formatted 3-hourly NAM files are interpolated to the resolution of the WRF domains. We use a 4-km-resolution outer domain (domain 1) and an interactive (two way), nested, 1-km-resolution inner domain (domain 2) (Fig. 8). Domain 1 is used to provide a broader context of the DL characteristics and its environment, and domain 2 is used to resolve phenomena related to the CI along the DL.

Table 1 provides a basic overview of the WRF configuration and physics choices used in simulating the DL in SE WY. The model consists of a total of 40 terrain-following sigma levels; 3D model output is interpolated to pressure levels for further analysis. The terrain files used have a better resolution than the domain resolution (4 or 1 km). The unified Noah land surface model (Ek et al. 2003) is used, with initial soil moisture at four levels set at spatially resolved June climatological values. Soil moisture is important as it affects the partitioning of surface heat fluxes, and thus convective BL depth, an essential ingredient in DL formation and evolution.

b. Model validation

In an effort to validate the model results, we overlay surface station observations and the quasi-stationary
radar fineline position between 1600 and 2000 UTC (Figs. 5 and 6a) over the modeled (domain 1) 2-m $q_v$ and 10-m wind barbs (Figs. 9a,b). The 5 g kg$^{-1}$ $q_v$ contour is used to approximate the modeled DL position, because it compares well to a developing band of strong convergence with a large humidity gradient, yet it is a continuous line (unlike gradient contours). (Any isohume between 4.0 and 5.5 g kg$^{-1}$ yields essentially the same DL position.) The model generates a distinct boundary between low- (high-) $q_v$ air to the west (east) as early as 1700 UTC (Fig. 9a). The modeled DL orientation and location agree well with the observations. The agreement between the observations and the model is best in northeast CO, whereas the modeled DL bulges over the Cheyenne Ridge, ~25 km east of the observed DL position at 1700 UTC and up to ~65 km at 2000 UTC. This eastward offset is consistent with a model overestimation of the zonal wind $u$ at KLAR by 3 m s$^{-1}$ at 1700 and 2000 UTC. Other possible reasons are that the moist layer is too shallow or that the BL deepens too rapidly, thus giving way to drier westerly flow. Comparisons between WRF simulations and observations in the SGP have shown that the model DL position tends to have an eastward bias (Coniglio et al. 2010; Coffer et al. 2013). In our case, the excessive bulge follows the pattern of the higher terrain (Fig. 9b). There is reasonable agreement with the station winds, especially east of the DL, in regions with moderately strong wind, and at the later time (2000 UTC). In the dry air to the west and in the moist air east of the model DL bulge, there is good agreement between the observed and modeled near-surface model $q_v$ values (discrepancies generally less than 2 g kg$^{-1}$ at 1700 and 2000 UTC). A scatterplot of model against station data (not shown) indicates a general negative $q_v$ bias, mainly on the dry side at 1700 UTC and on the moist side at 2000 UTC, confirming that the BL may deepen too rapidly in the model. Given the excessive eastward bulge in the model, there is a negative $q_v$ bias also just to the east of the observed DL (e.g., at KCYS).

We also compare the modeled (domain 1) 2-m $\theta_v$ and 10-m wind barbs to station observations at 2000 UTC (Fig. 9c). The model produces a distinct buoyancy gradient in the vicinity of the predicted DL location, with generally higher- (lower-) $\theta_v$ air to the west (east) of the 5 g kg$^{-1}$ contour. The buoyancy gradient is strongest north of the DL bulge; convergence is strongest here as well. This is where convection first erupts (section 4d). The highest-$\theta_v$ air is predicted to be just west of the DL boundary, over the highest terrain. The slight overprediction of $\theta_v$ at KCYS is consistent with the model’s eastward-biased DL bulge. The agreement is better to the east and west of this bulge.

FIG. 9. Model (domain 1) 2-m $q_v$ and 10-m wind barbs compared to near-surface observations, and approximate DL position for (a) 1700 and (b) 2000 UTC 22 Jun 2010. The station model plot in white only includes the station ID (top), wind barbs (center), and specific humidity (bottom right). The DL in white is the approximate observed position of the quasi-stationary DL between 1600 and 2000 UTC (from Fig. 6a). The background modeled $q_v$ is shaded according to the color legend (g kg$^{-1}$) on the right side of the figure, with the 5 g kg$^{-1}$ isohume highlighted as a black solid line. This line represents the approximate modeled position of the DL boundary. The modeled wind barbs are shown in black (full barb is 5 m s$^{-1}$). The purple lines represent terrain contours, decreasing from west to east, at a 100-m interval between 2000 and 1700 m. (c) As in (b), but for model (domain 1) 2-m $\theta_v$ (color shaded, K) and 10-m wind barbs compared to the observed DL location (red line) and station observations, with only the station ID (top), wind barbs (center), and $\theta_v$ (bottom right, K) shown in red.
Fig. 9 indicates that the model simulates a distinct low-level airmass boundary near the observed DL in SE WY, but it formed a DL bulge that is too strong over the Cheyenne Ridge too early.

c. **Airmass boundary, flow, and cloud initiation**

We now examine a zonal cross section of $q_y$, $\theta_e$, and vector tangential winds. The cross section extends from SE WY into NE, and intersects the DL. Convection initiated near this transect along the DL around 2000 UTC. In the morning hours (1500 UTC; Fig. 10a) the flow near the surface is weak, and there is a shallow layer of high-$q_y$ (low $\theta_e$) air well east of the LR. This shallow moist layer intersects the sloping topography and is capped by much drier air. Thus, the layer between 850 and 750 hPa is potentially unstable ($\frac{\partial \theta_e}{\partial z} < 0$); this instability remains into the afternoon. In the early afternoon (2000 UTC; Fig. 10b), a convergent DL boundary has formed, and the shallow moist layer has deepened, but the DL remains east of the LR. The weak easterly flow in this moist layer advects water vapor toward the DL.
The dry convective BL reaches up to 500 hPa west of the DL at 2000 and 0100 UTC (Fig. 10c). Air rises on the warm dry side of the DL over a great depth (up to at least 500 hPa) and subsides to the east; that is, the DL is a material boundary (like a mountain obstacle) to the westerly flow. Some moist air is lifted and entrained into the dry westerly air mass at the DL. The leading moist-air $\theta_v$ line is steeply sloped from the surface DL (blue arrow), reaching well above the LR, with trailing waves suggestive of trapped lee waves over a density current (Figs. 10b,c). The $q_v$ and vertical velocity fields are consistent with this gravity wave interpretation. The deepening moist layer is essential to CI, which occurs in the model just north of this transect near 2000 UTC. At 0100 UTC (Fig. 10c), the low-level westerly flow on the dry side has strengthened, leading to stronger convergence and larger zonal $q_v$ and $\theta_v$ gradients at the DL boundary, which has propagated eastward. Farther east in NE, active deep convection is intersected.

Analyses in Figs. 9 and 10 suggest that the $5 \text{ g kg}^{-1} q_v$ contour is a good surrogate for the position of the DL boundary, while also useful for identifying moisture plumes as the contours become upright. Thus, Fig. 11 presents a 3D analysis of the $5 \text{ g kg}^{-1} q_v$ isosurface, shaded by the height above mean sea level (MSL) of the isosurface, as well as the 0.01 g m$^{-3}$ total (ice + liquid) water mixing ratio isosurface, used to represent the predicted boundaries of clouds (e.g., Fig. 5e in Wang et al. 2009). The purpose of this analysis is to visualize the shallow moist layer trapped by the terrain and to relate the model-predicted convective clouds to water vapor plumes emerging from this moist layer. At 1500 UTC (Fig. 11a), the $5 \text{ g kg}^{-1}$ isosurface is relatively flat and low in height ($\approx 2 \text{ km MSL}$), confined to areas east of the LR. At 2000 UTC (Fig. 11b), the $5 \text{ g kg}^{-1}$ isosurface has clear topography: a greater depth of moisture (and thus lower CIN and greater CAPE) is found just east of the DL. The $5 \text{ g kg}^{-1}$ isosurface along the edge of the DL boundary approaches 4 km MSL, or $\approx 2$ km above ground level (AGL), with a maximum just east of the DL bulge ($\approx 2.5 \text{ km AGL}$). The first model cumulus clouds appear between 1900 and 2000 UTC, just NE of this bulge and LR, above relatively high peaks ($\approx 2 \text{ km AGL}$) in the $5 \text{ g kg}^{-1}$ isohume (Fig. 11b). This result is in qualitative agreement with the GOES-13 imagery (not shown), which reveals the first convective clouds near the DL at 1915 UTC. At 0100 UTC (Fig. 11c), the DL boundary has moved to the east, and the moist plumes have deepened significantly in some areas. Some points along the DL boundary, masked by the base of deep convective clouds in Fig. 11c, have $5 \text{ g kg}^{-1}$ isohume peaks at 5–6 km MSL (3–4 km AGL). There are also new convective towers that form just to the east-northeast of the LR at 0100 UTC. These deeper clouds are in qualitative agreement with Fig. 7, where radar observations show the initiation of new storms in this region (storms 2 and 3).

Overall, Fig. 11 indicates that shallow clouds at the DL boundary occur when moisture plumes, emerging from the shallow moist layer, reach a height of $\approx 2 \text{ km AGL}$. Deep moist convection farther to the east of the DL boundary appears to be associated with moisture plumes reaching heights near 3 km AGL.

d. Eastward progression in southeast Wyoming

Here, we analyze hourly changes in wind, stability, and humidity profiles over KCYS in Fig. 12, using our domain 1 WRF output. In section 4b, we discussed the early DL passage over KCYS (1700–1800 UTC). The time–height transect in Fig. 12 further indicates that the westerly surface wind increase in the afternoon is at least partly due to a deepening of the convective BL and, thus, the mixing of westerly momentum from increasing depth toward the surface (Fig. 12). Figure 12 also indicates some veering of the wind (warm-air advection) from the surface to 500 hPa around 1700–1800 UTC, followed by backing of the wind (cold-air advection) over the same depth after 0000 UTC, accompanied by slight cooling and drying at midlevels (isentropes slope up, and relative humidity decreases), consistent with subsidence. These changes are consistent with the passage of a weak upper-level trough, consistent with Figs. 3 and 6b, respectively. The diurnal BL deepening (and ingestion of westerly momentum) and the synoptic evolution conspire to drive the DL boundary eastward, overwhelming any westward acceleration imposed by the afternoon buoyancy gradient across the DL.

e. Initiation of convection in southeast Wyoming

We now use domain 2 data to investigate the CI along the 22 June 2010 DL in SE WY. Specifically, we aim to compare the observed CI to that predicted by the model, and to examine how the along-DL variability relates to the locations of CI.

1) PREDICTED AND OBSERVED CONVECTION INITIATION

The first convective precipitation in the model (reflectivity $>20 \text{ dBZ}$) emerged by 1900 UTC, about 15–20 km northeast of the DL bulge (Fig. 13a). In reality, moist convection was not observed until about an hour later, at 2000 UTC, just southwest of the predicted convection (Fig. 13b). (The actual radar echoes are shown using a different color table in Fig. 13, the one with the
Fig. 11. Domain 1 isosurface of $q_y$ (5 g kg$^{-1}$) colored by geopotential height above sea level (top color scale, m), 0.01 g m$^{-3}$ condensed water mixing ratio isosurface (white-gray shading, outlining cloud boundaries), and surface topography (left color scale, m) at (a) 1500, (b) 2000, and (c) 0100 UTC. The wireframe box has a vertical range of 0–16 km MSL.
ground clutter.) The prematurity of the model convection, and the eastward offset of model CI, may partly be due to the development of the DL bulge in WRF, too early and too far to the east (Fig. 9). The distance between the first incidence of convection and the DL location agrees well, however. For the convection just northeast of the DL bulge, the distances between the center of the initial convective cell and the DL location were about 15 km for the model (1900 UTC; Fig. 13a) and 25 km for the observation (2000 UTC; Fig. 13b). Farther to the north, there is even better agreement between the predicted and observed convective cells, where both indicate convection to occur at the DL boundary near a local terrain peak. As Fig. 9 suggests, this may be due to the better agreement between the observed and predicted DL position north of the DL bulge. Overall, these comparisons show promise, and allow us to further use the model results to investigate CI along the DL.

The DL boundary in domain 2 is rather distorted (Fig. 13), which may induce localized areas of enhanced convergence and vorticity. In fact, inside two 35 × 35 km² boxes (A and B), the model produces a cyclonic rotation at the DL boundary (Fig. 13a). In both cases CI occurs, at 1900 UTC in box A and at 2000 UTC in box B, to the northeast of the cyclonic vortex at the DL. The relation between CI and these vortices is investigated in the next section.

2) ALONG-LINE VARIABILITY, CONVECTIVE ROLLS, AND VORTICES

Along-DL variability and its role in CI along DLs has been studied using both observations (e.g., Hane et al. 1997; 2002; Murphey et al. 2006; Wakimoto and Murphey 2009) and models (e.g., Ziegler et al. 1997; Xue and Martin 2006a,b). Although terrain and large-scale variability invariably play a role in determining where convection may initiate along a DL, smaller-scale perturbations along the DL may also play a role. Such perturbations may be triggered by intersecting horizontal convective rolls (HCRs). HCRs are counterrotating horizontal vortex tubes that appear in convective BLs with moderate flow and vertical wind shear (Etling and Brown 1993). They appear as regularly spaced bands of BL convergence, alternating with divergence bands. Atkins et al. (1998) observed HCRs in the DL environment using aircraft and radar data, where Hane et al. (2001) had also observed two finelines and suggested that they were due to HCRs. The interaction of HCR convergence bands with the DL boundary and resulting CI has been examined in several modeling studies (Peckham et al. 2004; Xue and Martin 2006a,b). In particular, the strongest low-level convergence, upward motion, cloud formation, and CI have been shown to occur where HCR convergence bands intersect the
eastern side of the strongest moisture gradient along the DL (Xue and Martin 2006b).

Intersecting HCRs or other sources may trigger undulations along the DL, especially when strong horizontal wind shear exists along the DL (Markowski and Richardson 2010, chapter 5.2). These undulations yield alternating regions of convergence–divergence and cyclonic–anticyclonic vorticity. The cyclonic vortices dominate because of a background cyclonic shear vorticity along the DL. Such vortices have been observed to intensify in the convergent flow along DLS (e.g., Murphey et al. 2006; Buban et al. 2007; Marquis et al. 2007). They tend to be less than 4 km in diameter and, thus, are referred to as misocyclones (Fujita 1981).

We use domain 2 model output to investigate HCRs, vortices along the DL, and CI. There is strong moisture convergence along the DL (Fig. 14a), yet there is no evidence of a thermally forced mountain–plain circulation pattern with convergence and rising motion along the majority of the LR crest (boldface black lines), which is along 105.43°W. The only exception is north of about 42°N, where three local terrain peaks
are associated with areas of convergence and rising motion extending eastward from the DL boundary. The maximum vertical velocity along the DL drastically increases north of about 41.5°N (i.e., north of the DL bulge; Fig. 14b). There is also significant variation in peak updraft strength in this region, which indicates a high level of along-DL variability. South of 41.5°N, the maximum updrafts along the DL boundary are relatively constant and much weaker (<1 m s⁻¹), and there is no CI predicted or observed (Figs. 7 and 13). The good correlation between peak updraft and surface convergence along the DL (Fig. 14b) confirms that the convergence shown in Fig. 14a is rather deep, and that the 5 g kg⁻¹ isohume adequately represents the DL in this case.

Figure 14a also indicates a clear signature of HCR convergence bands, mostly oriented normal to the DL both east and west of the DL boundary. The general
orientation of the HCRs is aligned with the low-level wind shear vectors (Fig. 14a), especially east of the DL, consistent with HCR theory (Etling and Brown 1993) and simulations (Peckham et al. 2004; Xue and Martin 2006a). The HCR bands are more intense on the east side of the DL, becoming cellular farther east, whereas they are more elongated and weaker on the west side. This is somewhat in contrast to Xue and Martin (2006b), in which the HCRs on the west side of the DL were found to be more intense and deeper. It does appear, however, that the western HCRs interact similarly with the DL boundary as do those in Xue and Martin (2006b), but possibly with a different magnitude. The HCRs have an approximate wavelength of 5 km east of the DL boundary. The aspect ratio [defined as the ratio of the distance between HCR-generated updrafts and the DL depth; Xue and Martin (2006a)] ranges between 2 and 6 in this region. The more elongated and organized HCRs on the west side have a wavelength of 6 km and an aspect ratio between 1 and 2.

The main HCR convergence bands in Fig. 14a are associated with enhanced convergence and rising motion along the DL; that is, there are pockets of positive interference between the DL solenoidal convergence and the HCR convergence, especially in the northern region of the DL bulge, leading to stronger updrafts. In fact, the three highest maximum vertical velocities (>5 m s\(^{-1}\)) (highlighted in Figs. 14a,b) occur near the intersection of HCRs from the west side of the DL. Figure 14b indicates that the maximum vertical velocities in this region of the DL boundary were strong. The updraft peak in the western region of box A at 1900 UTC (Fig. 14a) was just west of the location of the first precipitating cumulus cloud in the model at 1900 UTC (Fig. 13a). That cumulus cloud emerged from a similar updraft peak along the DL and was subsequently advected toward the east-northeast by midlevel winds (Fig. 7; section 3d). The fact that the first convective cell appeared downwind of an HCR–DL intersection is in agreement with Peckham et al. (2004) and Xue and Martin (2006b).

HCRs intersect the DL in box B as well, with even stronger model vertical velocities at the DL boundary, some exceeding 5.5 m s\(^{-1}\) (Figs. 14a,b). As shown in Fig. 13b, convection was also predicted an hour later at the DL boundary in box B. Model output suggests the same CI mechanism (i.e., the superposition of DL convergence with HCR convergence), but in this case the first convective cell occurred at the DL, rather than downwind (Fig. 13b). In this example, the CI locations were also just downwind of the highest terrain, although the convergence peaks along the DL, not at the terrain crest (Fig. 14a). CI also occurred at the DL to the north of box B (>42°N), in agreement with the observations (Fig. 13b). Here, the convergence and localized terrain peaks are more correlated, and thus the CI may be induced by thermal mountain–plain circulations, not HCRs.

As previously mentioned, the model DL boundary just north of the bulge is rather wavelike at 1900 UTC (Figs. 13a and 14a). This wavelike appearance of the DL boundary and the development of vortices can be due to the interaction of HCRs with the DL boundary (Xue and Martin 2006b). A 1-km grid spacing only marginally captures features 4 km in diameter. The main vortices in our domain (relative vorticity > 2 × 10\(^{-3}\) s\(^{-1}\)) are 4–10 km in diameter (Fig. 15). Thus, we shall refer to them as mesocyclones, rather than misocyclones (<4 km), cognizant that misocyclones would likely be present in higher-resolution simulations (e.g., Buban et al. 2012). The vortices deform the DL boundary, yielding a line-echo wave pattern, while also influencing the location of the updrafts (Murphey et al. 2006; Wakimoto and Murphey 2009). In an attempt to further address the influences of these HCRs and mesocyclones on CI, Fig. 15 shows a smaller domain that focuses on the northwest corner of box A and southeast corner of box B. This is the region of most HCR interaction with the DL boundary, strongest buoyancy gradient, strongest moisture convergence, upward vertical motion, and wave-like appearance of the SE WY DL. A few strong surface mesocyclones are evident in Fig. 15. An hour prior to CI along the DL in SE WY (1800 UTC; Fig. 15a), there are three main mesocyclones (labeled M1–M3) stretching along the DL boundary. The cyclonic circulation transports warm, dry air to the north, and cool, moist air to the south, yielding a wavelike DL deformation. The associated three main updraft cores (labeled U1–U3) are generally just east of the mesocyclones, which makes intuitive sense since that is where warm dry air is lifted over the cooler moist air. This spatial arrangement is consistent with previous observations (e.g., Murphey et al. 2006; Buban et al. 2012). Thus, mesocyclones can focus and locally enhance the low-level convergence and upward motion, similar to results from Murphey et al. (2006). Judging by their location, the main mesocyclones may have been triggered at the intersection of the dominant HCR convergence lines with the DL boundary. Such triggering may not occur at every intersection, as the HCR spacing is smaller than the typical mesocyclone spacing. This is especially true with mesocyclone M3, where several moisture convergence bands appear to coalesce (Fig. 15a). The first deep convection is predicted to occur at CI (Fig. 15b) at 1900 UTC, some 15 km east of the updraft associated with M3; this convection eventually leads to a supercell storm, as observed (Fig. 7).
Different mesocyclones (Ma–Mc) are present an hour later at 1900 UTC (Fig. 15b). Misocyclones along DLs have been shown to form, amplify, and merge or decay on time scales of less than an hour (Lee and Wilhelmson 1997; Murphey et al. 2006; Marquis et al. 2007), so mesocyclones M1–M3 in Fig. 15a are probably not the same as Ma–Mc in Fig. 15b. New updraft cores (Ua–Uc) can be seen to the east of the mesocyclones. Clearly, the mesocyclone’s core has the ability to suppress upward motion, as it is generally positioned between locations of updraft cores, most clearly for Mb (Fig. 15b). This is due to a downward perturbation pressure gradient (Murphey et al. 2006). Farther to the north along the DL, the convection that occurs an hour later (Fig. 15b; C2*) seems to be a result of a strong HCR intersecting the DL. This interaction allowed for sufficient lift, as well as CI at the DL rather than ahead of the DL at this location.

In short, the northern edge of the DL bulge appears to be the preferential region along the DL for CI (C1 and C2* in Fig. 15b), because the DL there is normal to the wind on opposite sides (southwest wind on the dry side and northeast wind on the moist side). On finer scales, CI is concentrated near intersections of major HCR.
5. Discussion

This study provides the first detailed analysis of a DL in SE WY. This DL was critical in the formation of a severe thunderstorm, which spawned large hail and tornadoes. Although NWS regional office forecasters at KCYS do recognize the occurrence of DLs in SE WY, they are not analyzed as DLs on HPC surface charts for any of the ~40 DL events analyzed by the third author (PB). One can argue that the NWS HPC surface analysis of the DL boundary on 2100 UTC 22 June 2010, as seen in Fig. 2, is inaccurate in that it fails to show a DL in SE WY. Our study remains rather myopic though, as it has not shown the larger-scale extent of this DL for lack of station and clear-air radar data, especially the DL’s evolution farther north (where the terrain is steeper again), and its linkage with the DL south of the Colorado low (Fig. 2). In the absence of evidence for a larger-scale DL boundary, we do not argue that the HPC analyses are wrong. Outflow boundaries from mesoscale convective systems may be longer than the SE WY dryline. Nevertheless, this humidity boundary in SE WY has all the characteristics of a DL as observed in the SGP, over similarly sloping, although higher, terrain.

This paper serves to improve awareness of DLs and their relation to deep moist convection in this region. Given the low density of surface stations in WY, it remains difficult to pinpoint DL boundaries there, especially at a range >70 km from KCYS. Thus, DLs in SE WY remain poorly documented and rarely recognized.

For that reason we have embarked on a climatological study of DLs in SE WY, to characterize their meso-α-scale environment, their vertical structure, and their potential for CI (PB). Large variations in mixing ratio are observed along the DL over horizontal distances of less than 1 km, which can have a profound effect on CAPE and thus CI, as will be shown in Part II, using airborne Raman lidar data.

6. Conclusions

A detailed analysis of a DL observed on 22 June 2010 in SE WY is presented. Observations indicate that this DL has characteristics of DLs that form in the SGP. The DL exhibited a strongly convergent moisture boundary, along with a modest buoyancy gradient, with denser air to the east. The DL formed around noon along the gently ascending terrain east of the LR, as the shallow moist layer intersected the terrain at ~2200 m MSL. The moisture convergence along the DL led to a deepening of the moist layer. The DL was quasi stationary at first, and then started to move toward the east in the afternoon, due to an increase of westerly flow at the surface associated with the diurnally deepening convective BL, and by the eastward progression of a weak upper-level trough. A bulge formed in the progressing DL over the relatively higher terrain of the Cheyenne Ridge, as the moist air wedge mixed out earlier there. Numerical simulations show that DL convergence and the buoyancy gradient were strongest along the north side of this bulge, and that this part of the DL was characterized by HCR convergence bands intersecting the DL, and mesocyclones along the DL, leading to strong updrafts and moisture plumes reaching the LCL and LFC, and leading further to CI at or just east of the DL. Both in the model and in reality the primary CI led to a supercell storm.

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