Convection Initiation and Bore Formation Following the Collision of Mesoscale Boundaries over a Developing Stable Boundary Layer: A Case Study from PECAN

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ABSTRACT: This observational study documents the consequences of a collision between two converging shallow atmospheric boundaries over the central Great Plains on the evening of 7 June 2015. This study uses data from a profiling airborne Raman lidar [the compact Raman lidar (CRL)] and other airborne and ground-based data collected during the Plains Elevated Convection at Night (PECAN) field campaign to investigate the collision between a weak cold front and the outflow from an MCS. The collision between these boundaries led to the lofting of high-CAPE, low-CIN air, resulting in deep convection, as well as an undular bore. Both boundaries behaved as density currents prior to collision. Because the MCS outflow boundary was denser and less deep than the cold-frontal air mass, the bore propagated over the latter. This bore was tracked by the CRL for about 3 h as it traveled north over the shallow cold-frontal surface and evolved into a soliton. This case study is unique by using the high temporal and spatial resolution of airborne Raman lidar measurements to describe the thermodynamic structure of interacting boundaries and a resulting bore.

KEYWORDS: Small scale processes; Boundary layer; Storm environments; Aircraft observations; Lidars/Lidar observations

1. Introduction

Convergent boundaries may deepen boundary layer (BL) humidity to create conditions favorable for initiating deep convection in areas where deep convection otherwise is inhibited (Wilson and Schreiber 1986; Geerts et al. 2017). Over land in the warm season, these boundaries are visible in low-elevation radar reflectivity scans as “fine lines” because of small, weakly flying insects lofted into the BL (Russell and Wilson 1997; Geerts and Miao 2005). Since Purdom (1982) noted the potential for using these radar fine lines for identifying possible locations for convection initiation (CI), forecasters have been monitoring these fine lines as harbingers of possibly hazardous weather. In addition to enhanced scattering, a wind shift is usually present across the boundary, which provides another method for boundary detection via Doppler radar (Wilson and Schreiber 1986) or Doppler lidar (Intieri et al. 1990).

In the daytime convective BL, over flat terrain, these boundaries are solenoidally forced, i.e., driven by a density difference (Miao and Geerts 2007), created by thunderstorm outflows (Wakimoto 1982), synoptic fronts (Geerts et al. 2006), or other mechanisms. When the land surface starts to cool in the evening and a stable boundary layer (SBL) forms, these density currents may transition to bores and solitary waves (e.g., Clarke 1972). A bore is a gravity wave response generated when a density current intrudes into a stable layer (Crook 1988; Rottman and Simpson 1989; Haghi et al. 2019). The importance of low-level shear on the behavior of undular bores is less understood (Haghi et al. 2019). One of the key objectives of the 2015 Plains Elevated Convection at Night (PECAN) field campaign regards the role of bores in CI and the maintenance of MCSs across the Great Plains (Geerts et al. 2017).

The dynamics of CI along a convergent boundary are generally well understood (e.g., Harrison et al. 2009). CI occurs when a sufficiently large air parcel (i.e., a parcel surviving entrainment/detrainment) is lifted above its level of free convection (LFC). In the presence of a SBL and forced vertical motion, profiles of parcel LFC and CAPE can be examined to estimate whether a parcel emerging from any layer can realize elevated CAPE (e.g., Grasmick et al. 2018). In a well-mixed BL, ambient wind shear matters as well: new convective cells are more likely to form (and MCSs are more likely maintained) if the solenoidally driven horizontal vorticity of a boundary matches the vorticity of the ambient vertical wind shear in magnitude, but is opposite in sign, according to the Rotunno–Klemp–Weisman (RKW) theory (Rotunno et al. 1988).

CI is often difficult along a single propagating boundary. Collisions between boundaries propagating in different directions and/or at different speeds increase the likelihood for CI because they lead to transient deeper lifting from enhanced low-level convergence (Kingsmill 1995). Different boundary collisions have been described, for example, between two thunderstorm outflows (Karan and Knupp 2009), between a sea-breeze front and an outflow boundary (Kingsmill 1995), and between a dryline and a cold front (e.g., Wakimoto et al. 2006). CI along boundary collisions is difficult to predict: the

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boundary characteristics that determine upward displacement (such as depth, temperature, boundary-relative wind, and orientation) are usually inadequately known and any resulting CI is often displaced horizontally (Weckwerth and Parsons 2006).

Density current collisions frequently generate atmospheric bores (Kingsmill and Crook 2003; Karan and Knupp 2009). Simpson (1987) conceptualized bore generation in density current collisions based on laboratory experiments. Kingsmill and Crook (2003) applied the Simpson model to cases of colliding gust fronts and sea-breeze fronts. Theoretically, if two colliding density currents are similar in depth and density, the structural evolution is symmetrical: the region of colliding density current mass bulges upward and this bulge subsequently evolves into two bores, propagating away from the collision line in opposite directions. However, most colliding currents are not identical, so collisions commonly result in the less dense current overriding the denser current (Fig. 1 in Kingsmill and Crook 2003).

This paper documents a collision between thunderstorm outflow (i.e., “gust front”) and a cold front, which results in both CI and a bore. As will be shown below, the gust front is denser and shallower than the encroaching cold-frontal air mass, producing a highly nonsymmetrical evolution (see Fig. 1). Furthermore, the bore in this study was observed propagating in an uncommon direction, toward the north, which is rare among published case studies that have documented eastward and southward propagating bores in the Great Plains (Grasmick et al. 2018; Haghi et al. 2019; Johnson and Wang 2019; Loveless et al. 2019; Parsons et al. 2019).

The Raman lidar and ground-based network provides a thermodynamic perspective of the environment before, during, and after the collision and bore formation. The primary objective of this study is to document the evolution of the colliding density currents, the resulting CI, and bore, by means of high-resolution vertical transects of humidity and temperature obtained from an airborne Raman lidar. The current, sparsely spaced 12-hourly sounding network offers low-resolution horizontal and temporal profiles. Surface stations are also sparse. Geostationary satellites have high temporal resolution but do not provide adequate vertical resolution in the BL. As a result, a small-scale, rapidly evolving convective plume from a convergent boundary is difficult to monitor using the current operational network. The rare airborne measurements presented here have sufficient resolution to describe the density current collision and ensuing deep convection.

Our analysis of CI uses data from the airborne Raman lidar to derive variables such as CAPE and CIN and includes an array of ground-based measurements to reveal the evolving vertical structure during the boundary collision. Such thermodynamic analysis is a powerful tool to assess convective potential and improve CI forecast (Weckwerth and Parsons 2006).

Section 2 describes the datasets and instruments for this case study. An overview of the mesoscale environment and Raman lidar transects are presented in section 3. The vertical structure of the merged boundaries is detailed in section 4. The resulting bore, and environmental conditions that support bore formation and propagation, are analyzed in section 5. Conclusions are presented in section 6.

2. Data and methodology

a. The 2015 PECAN campaign

The 2015 PECAN field campaign aimed to understand the driving mechanisms of nocturnal MCSs, bores, and CI in the presence of a SBL and a low-level jet (Geerts et al. 2017). Cold-pool-generated bores and related solitary waves were found to be relatively common during PECAN. Among other objectives, the PECAN campaign aimed to advance the understanding of the generation and evolution of bores, and their role in CI. This study contributes to that objective.

b. PECAN airborne data

Three aircraft were deployed in PECAN, but the University of Wyoming King Air (UWKA) was the only aircraft to focus on the lower-tropospheric environment where CI and bores/solitary waves originate. The UWKA carried an array of in situ probes and the compact Raman lidar (CRL) (Wu et al. 2016). The CRL can provide accurate retrievals of water vapor mixing
ratio (WVMR), lidar scattering ratio (LSR), and temperature in the lower troposphere at a resolution of ~300 m horizontally and ~100 m vertically (Liu et al. 2014; Wu et al. 2016). The CRL provides WVMR with a mean difference of 0.2 kg m$^{-1}$ compared to in situ measurements (Wang 2020). Due to the photomultiplier tube ring noise in CRL’s temperature measurements in PECAN (at least until this was fixed on 15 June) (Wang et al. 2016), CRL temperature measurements only can be used to provide horizontal temperature variations in the case analyzed here. Therefore, we calculate the temperature in cross sections below the flight track by adding the CRL-derived horizontal temperature variation at each level to a mean low-level temperature profile from proximity radiosonde data.

Random errors within CRL temperature measurements increase with range. As a result, the near-surface random errors can be up to 1.5 K for the flight level used in this case (Wu et al. 2016). Aircraft banking also introduces surface signal contaminations in temperature retrievals near the surface. Thus, the lowest 200-m temperature layer is removed. Flight-level in situ WVMR and virtual potential temperature measurements are added to the top of CRL cross sections in the figures shown below, as a means of CRL validation.

The LSR is a normalized parameter of total backscattering to molecular backscattering. The minimum value (1.0) represents scattering by air molecules only, and larger values (>1.0) represent additional aerosol backscattering. The CRL, which was pointed at nadir, was complemented by the zenith-pointing Wyoming Cloud Lidar (WCL; University of Wyoming-Flight Center 2007; Wang et al. 2009), to provide lower-tropospheric LSR profiles across flight level. The synergy of flight-level measurement (1 and 10 Hz) with a full lower-tropospheric LSR profile (above and below flight level) is uniquely powerful for studying the dynamic and thermodynamic environment across the convergent boundaries. In this study, we use mean sea level (MSL) to reference height because of varying ground elevation along flight legs, and the ease with which airborne, surface, and profile observations can be compared. The mean height of the ground surface over the domain of interest is 590 m MSL.

CAPE, CIN, and the distance to each parcel’s level of free convection (LFC) are computed following Grasmick et al. (2018) and Lin et al. (2019). In essence, CRL temperature and moisture at each time step (interval of 3 s) at any height between the ground and flight level are combined with proximity radiosonde data above flight level (in this case, the 0300 UTC radiosonde released from MP1) to reconstruct a sounding for each UWKA profile. Then, each thermodynamic profile is used to calculate the variation of lifting condensation level (LCL), LFC, CAPE, and CIN along the flight track, for each parcel between the surface and flight level.

c. Ground-based data in PECAN

During PECAN, a series of surface mesonet vehicles were deployed, equipped with roof-mounted instruments to measure temperature, pressure, humidity, and wind at 1 Hz (Roberts et al. 2008). Data from one of these vehicles, called the “mobile mesonet 1” (MM1), are used to characterize the surface characteristics of the different air masses as it crossed the convergent boundary (Waugh and Ziegler 2017).

PECAN Integrated Sounding Arrays (PISAs) with four mobile PISA units (MP) and six fixed units (FP) were deployed during intensive observation periods (IOPs). The FPs were continuously operating. MPs were entirely contained within a vehicle and were deployed to different locations in different IOPs, but remained stationary for individual IOPs. Specifically, surface-based and radiosonde data from MP1 and radiosonde data from FP2 are used in this study (Klein et al. 2016; Turner 2016; Vermuesch 2015). MP1 is equipped with instruments to measure surface-based temperature, pressure, humidity, and wind at 1 Hz. Two MP1 radiosondes, spaced 3 h apart, supplement the CRL temperature and mixing ratio retrieved below the UWKA flight level.

Finally, Atmospheric Emitted Radiance Interferometer (AERI) data from MP1 were used in this study (Turner 2018). An AERI is a ground-based remote sensor measuring spectrally resolved downwelling infrared radiation to retrieve vertical profiles of water vapor and temperature, using an optimal estimation-based physical retrieval algorithm (Knuteson et al. 2004a, b). In clear-sky situations (or cloud bases above 2 km), the mean bias errors with respect to radiosonde profiles are less than 0.2 K and 0.3 g kg$^{-1}$ for temperature and water vapor mixing ratio, respectively, but its random errors could be significantly higher than the mean biases (Turner and Löhner 2014).

d. Operational data

This study employs the operational network of radar and surface observations. The radar reflectivity factor and mean radial velocity from the Next Generation Weather Radar (NEXRAD) system (Crum et al. 1993) are used to show precipitation and wind fields in the proximity of the colliding boundaries and to place the UWKA flight track in the context of the MCS, cold front, outflow boundaries, and the bore.

The Meteorological Assimilation Data Ingest System (MADIS) is a fine-scale grid of interpolated surface meteorological observations (UCAR/NCAR–Earth Observing Laboratory 2011), used to initialize and evaluate weather and climate models, provided by the National Oceanic and Atmospheric Administration (NOAA). MADIS combines NOAA data sources with non-NOAA data into a common format with multiple quality control procedures (details at https://madis.ncep.noaa.gov/madis_qc.shtml) and provides a finer-resolution depiction of environmental conditions than synoptic-scale data (Miller et al. 2005).

3. Overview of the mesoscale environment and airmass characteristics

At 0200 UTC 8 June [2100 central daylight time (CDT) 7 June, 3 min after local sunset], a southwest–northeast-oriented MCS extended from the central Texas panhandle to northern Oklahoma (Fig. 2). This was the target of PECAN IOP 6, which was designed to document Cl, predicted to develop north of the MCS, in the vicinity of a weak cold front. The cold front propagated southeastward in the area and period of interest, and was followed by cold-air advection. This cold front is apparent in the MADIS data (Fig. 3d) at the boundary between northerly and southerly flow in south-central Kansas. The UWKA flight track is...
shown within the area of interest (small red box in Fig. 2), where it sampled the cold front, MCS outflow, and post-collision boundary. The MCS was located on the south side of the cold front, just south of the red box in Fig. 2. Its spreading cold pool is observable as relatively cool, moist, diffluent flow (Figs. 3a,b). The MCS outflow boundary is evident in the MADIS surface data as the northern boundary of this cold pool.

Both boundaries are evident in the Wichita, Kansas (KICT), Doppler radar base reflectivity map as fine lines. The outflow boundary can be seen moving northwesterly, and colliding with the southeast-moving cold front, with the collision point moving eastward (Figs. 4a–d). Weak westerly flow is observed between the two boundaries, weaker than the flow behind the outflow boundary (Figs. 4f and 4g). Stronger northwesterly flow is also observed behind the cold front, evident in Doppler radial velocity data (Figs. 4e–h) and in MADIS data (Figs. 3c and 3d). The speed of the surface cold front was \(8 \text{ ms}^{-1}\) as it passed the MP1 site (estimated from radar loops in the PECAN Catalog Map: http://catalog.eol.ucar.edu/maps/pecan). Behind the cold front, the front-normal wind speed exceeded \(10 \text{ m s}^{-1}\) at all levels below 1.25 km MSL (Fig. 5a). Since the wind speed behind the cold front was larger than the front’s propagation speed, there was front-relative forward mass transport, as is common for cold fronts propagating as density currents.

The MP1 data document the precollision cold front (Fig. 6). The air immediately behind the cold front was only slightly cooler than the prefrontal air (Fig. 6a), and changes in wind speed and direction were benign (Fig. 6c). After passage, the surface air temperature continued to decrease with time while the air density and pressure increased (Fig. 6b). The water

![Image](https://example.com/image.png)
vapor mixing ratio was the most prominent discriminator of these air masses; frontal passage at MP1 was most notable by an increase in WVMR, from \( \sim 10 \) to \( \sim 15 \, \text{g kg}^{-1} \) (Fig. 6a). Northeast of MP1, frontal convergence and ascent was sufficient to trigger deep convection (Figs. 4a and 4b), in contrast with our region of interest, where a boundary collision was required to initiate convection. About 3 h after cold-frontal passage, MP1 witnessed the postcollision outflow boundary passage (Fig. 6). These observations will be discussed later.

The FP2 site, 55 km southwest of MP1 (Fig. 4), sampled the precollision MCS outflow boundary. This boundary moved to the northeast at 5.6 m s\(^{-1}\) when it passed FP2 around 0200 UTC, while the normal wind speed behind the outflow, averaged over the lowest 1 km above the surface, was \( \sim 7.0 \, \text{m s}^{-1} \) at 0306 UTC (Fig. 5b). These observations show boundary-normal forward mass transport (the wind speed behind the outflow is larger than outflow’s propagation speed), confirming that the MCS outflow also was propagating as a density current.

4. CI and convective potential
   a. Thermodynamic vertical structure of a boundary collision

The UWKA flew inbound and outbound legs (relative to the MCS) over the region of CI, profiling the thermodynamic...
structure before and during the initiation of isolated deep convection resulting from the collision of the cold front and outflow boundary. The first flight leg (the red line in Fig. 7f and in Fig. 4b) sampled the collision between the cold front and outflow boundary, around 0220 UTC. The airborne lidars provide information about the vertical structure of WVMR, LSR, and virtual potential temperature $\theta_v$ (Figs. 7a–c). The profiling observations show 850-m-deep warmer, moister air ($\sim 15 \text{ g kg}^{-1}$ of WVMR below 1.4 km MSL, Fig. 7a) moving southward behind the cold front approaching the MCS. At the same time, cooler, drier ($\sim 8 \text{ g kg}^{-1}$) outflow from the parent MCS flowed northward. The MP1 radiosonde profile at 0300 UTC also shows the mixed, albeit stabilizing (occurring later after sunset), moist layer from 0.75 to 1.25 km with $\sim 15 \text{ g kg}^{-1}$ of WVMR behind the cold front (Fig. 5a).

The $\sim 7 \text{ g kg}^{-1}$ WVMR gradient across the colliding boundaries, over just a few kilometers, is quite impressive, and, as will be shown below, has significant implications about convective potential. This gradient is confirmed by the MM1 vehicle driving southward across these colliding boundaries, roughly simultaneous with the UWKA flight track (Fig. 8): the WVMR decreased sharply from 15 to 10 g kg$^{-1}$, the temperature dropped from 27$^\circ$ to 21$^\circ$C, $\theta_v$ decreased by 7 K (indicating that the MCS cold pool was denser at this location), sea level pressure (SLP) increased (indicating, hydrostatically, a lower $\theta_v$ over some depth), and wind shifted from northeast to southwest (Fig. 8). These surface-based measurements are consistent with large near surface water vapor and temperature variations observed by the CRL (Fig. 7).

The transect of $\theta_v$ shows the MCS outflow as a nearly 600-m-deep surface-based cold pool, with slightly ($\sim 2$ K) lower $\theta_v$ than the postfrontal air: the average CRL-retrieved $\theta_v$, below 1.10 km MSL was 307 K in the cold-frontal density current (0215–0219 UTC, Fig. 7c), compared to 305 K in the MCS outflow (0221–0225 UTC). This gradient provides sufficient solenoidal forcing for the lofting of the less dense air mass (e.g., Miao and Geerts 2007). The preexisting BL air between the cold front and MCS outflow, lofted above the MCS outflow due to the boundary collision, contained more water vapor than the MCS cold pool.

As a result of the collision, a plume of high-WVMR (about $11 \text{ g kg}^{-1}$), cold-frontal air was lofted to flight-level, 2.5 km MSL (Figs. 7a and 7b), with a peak flight-level vertical velocity of 3.0 m s$^{-1}$, northwesterly wind (Fig. 7d), and higher humidity (Fig. 7e). This lofting is due to the slightly higher density (lower $\theta_v$) of the “outflow boundary” density current, and is shown schematically in Fig. 1. The upward-looking lidar captured this plume as a nonprecipitating cloud base at a height of around 3.0 km MSL (Fig. 7b). The top of the cold frontal surface lifted from its original height of $\sim 1.4 \text{ km MSL}$ ($\sim 0.8 \text{ km above}$
ground level) to at least 3.0 km, a ~1.6-km vertical displacement (Fig. 7b).

The two sides of the merged boundary were also differentiable by aerosol content. The postfrontal air has a large LSR values (>1.15) up to 1.5 km MSL, while the MCS outflow appears nearly devoid of aerosols at low levels (LSR ~ 1) (Fig. 7b), likely due to moist scavenging within/under the MCS, or entrainment of air from above the BL.

The MCS cold pool properties remained fairly homogeneous as the UWKA paralleled the storm toward the northeast (Fig. 7f). Next, the UWKA made two more transects across the boundary collision point (green and orange solid lines below Fig. 7c), revealing spatial and temporal changes. The UWKA flew just above a cloud on the third transect, as evident from the attenuation of the CRL signal near 2.4 km MSL (black boxes in Figs. 7a and 7b at 0226 UTC). This cloud was coincident with the collision point and was convective, with flight-level updrafts up to 7 m s⁻¹ (Fig. 7d).

During the final transect in this region (orange solid line below Fig. 7c, southbound leg), the UWKA had descended to a lower flight level and passed through this convective cloud (Fig. 7b, 0229–0231 UTC); the in situ instruments detected liquid water up to 0.08 g m⁻³ (Fig. 7d), 5 m s⁻¹ updrafts, and enhanced aerosol concentration and humidity around cloud (black boxes in Figs. 7a and 7b, 0233 UTC). A brief, intense (~10 m s⁻¹) downdraft was present on the south side of this cloud, just behind the MCS outflow boundary. As soon as the UWKA completed this final transect, the convective cloud produced heavy precipitation (reflectivity values exceeding 50 dBZ) (black dotted boxes in Figs. 4c and 4d), and in-cloud lightning was observed (not shown), which forced the UWKA to leave the area.

b. Why did boundary collision lead to deep convection along the flight transects?

There is no doubt that the boundary collision was instrumental to CI in this case. The first argument is purely kinematic: the BL flow was highly convergent (Fig. 4g), and this BL convergence led to a strong but transient updraft in the region of the UWKA transects (Fig. 7). This was not the case farther west, where the cold front and MCS outflow had already collided by 0200 UTC (Figs. 4a and 4e). This may be due to weaker outflow farther away from the parent MCS (Figs. 4b,c,f,g). Another possibility could be the orientation of the two fine lines, which allowed intermediate air to escape to the east without being forced vertically (Wilson and Schreiber 1986; Frank and Kucera 2003). This westerly acceleration ahead of the collision point (which moved eastward like a zipper; Figs. 4a–d) may have enhanced convergence, eventually leading to the observed CI. Doppler velocity at KICT does show an increased westerly wind between the two boundaries, which was also resolved as an increasing westerly wind by the MADIS data (Fig. 3c).

The second argument is thermodynamic. Transects of CAPE, CIN, and minimum vertical displacement for CI (derived from CRL and sounding data, see section 2c) reveal that air parcels on different sides of the colliding boundaries have different convective potential (Fig. 9). The relatively warm and moist postfrontal air mass had large values of CAPE (with pockets > 1000 J kg⁻¹) and modest CIN (<300 J kg⁻¹). The MCS outflow contained less CAPE (generally <200 J kg⁻¹)
and more CIN (generally >400 J kg\(^{-1}\)) below 1.25 km. Therefore, air parcels in the MCS outflow (Figs. 9d and 9e) needed more lift to reach their LCL (1.5 km) and LFC (4 km or no LFC) than the postfrontal air (LCL 1.5 km, and LFC 1.5–2.5 km uplift). Both air masses were colder and more stable than the prefrontal BL air (arising from between the two boundaries), which logically is the most likely source of the observed deep convection.

It is not easy to obtain airborne transects of colliding boundaries at the right time and place and, in fact, the boundary collision observed on 8 June was a stroke of good luck. Unfortunately, there was no precollision flight leg, so this air mass was not sampled by the CRL. The convergence that results from these two air masses colliding lifted a cold-frontal plume about 1.6 km, as seen in the WVMR and LSR plumes (at 0218 UTC in Fig. 7). This lifting is sufficient for air within the lofted cold-frontal plume to overcome environmental CIN, reach its LFC and LCL. The horizontal scale of this lofted moist plume is about 5 km, which may be wide enough for deep convection in a high-CAPE, weakly sheared environment (Peters et al. 2019). In summary, while the observed CI may have been fed mainly by the precollision prefrontal air, the observations presented here demonstrate that the cold-frontal air was lofted sufficiently over the MCS outflow for CI, and may have fed the ensuing deep convection as well.

Clearly, these processes cannot be captured by an operational network. If the lofted moist plume in Fig. 7a were sampled by chance by an operational radiosonde, it would be assigned a much larger footprint, and thus err data assimilation and forecasts.

5. Bore and solitary waves

a. Bore formation and propagation: Observations

In addition to CI, the collision between the gust front and cold front also produced a bore-like wave structure with multiple radar fine lines traveling northward on top of the cold-frontal air mass, in the opposite direction of the cold front’s movement. As the bore passed MP1 around 0345 UTC, surface observations showed periodic variations in wind speed and direction that are more similar to gravity waves rather than a single density current gust front (Fig. 6). The bore gravity waves could be seen propagating toward the north-northwest as multiple, parallel fine lines in the base reflectivity and Doppler velocity of the Dodge City, Kansas (KDDC), radar (Fig. 10). The UWKA executed two sets of racetrack flight patterns that were nearly
perpendicular to these bore waves: the first racetrack consisted of four legs (Figs. 10a and 10b), and the second, farther west, had six (Figs. 10c and 10d).

Nocturnal bores are common during the summer months in the Great Plains and most frequently occur when a density current (usually an MCS outflow boundary) intrudes into a SBL (Geerts et al. 2017; Haghi et al. 2017, 2019; Mueller et al. 2017). The few case studies that have documented bore development from two colliding boundaries, i.e., Wakimoto and Kingsmill (1995), Kingsmill (1995), Kingsmill and Crook (2003) and Karan and Knupp (2009), all relied on continuous radar and surface station or tower measurements, and they all emphasized bore kinematics, obtained from Doppler radar. For instance, Karan and Knupp (2009) display cross sections of winds during the collision of two gust-fronts using a WSR-88D Doppler radar, supplemented with measurements from a 915-MHz wind profiler. Their wind analysis reveals two separate updrafts that combined to form a wider, more intense, short-lived updraft. In their case, isolated CI appeared during the collision 30–40 km away from the collision axis. After the collision, bore wave characteristics were observed over the less dense current, as in our case.

Unlike these previous case studies, this study emphasizes bore thermodynamics in high spatiotemporal resolution inferred from the airborne Raman lidar. Because the collision occurred only 15 min after sunset, a significant SBL had not yet formed. As mentioned in section 4a, the cold-frontal density current was slightly deeper but less dense (~2 K higher $\theta_v$, Fig. 7c) than the MCS outflow boundary. Figure 7a suggests that the cold-frontal and MCS density currents were 850 and 600 m deep, respectively. Both the depth and the density differences have important implications for bore formation following the collision: in this case, the cold-frontal air mass was partially blocked by the shallower but denser MCS outflow, triggering a northward moving bore within the cold-frontal air mass, as illustrated in Fig. 1. We now demonstrate this, using scanning radar and profiling CRL data.

Initial development of a bore occurs when a layer of air is blocked and deepens locally, producing an interface known as a hydraulic jump (typically in liquid), where supercritical flow slows and becomes subcritical relative to a gravity wave. In this case of colliding boundaries, partial blocking by the MCS outflow deepened the cold-frontal air mass on the north side of the collision point (Figs. 10a and 10b). The top of the cold-frontal density current (initially about 1.4 km MSL) was lifted to about 3.0 km MSL (Fig. 7b). The blocking region extended close to 20 km ahead of the MCS gust front (Fig. 11c). Furthermore, the collision does not appear to impede the momentum of the relatively deeper cold-frontal air mass, at least not for a while, because this air mass continues moving southward over the denser thunderstorm outflow (Figs. 4e–h). The elevated cold front is shown schematically at the left (south) side of the transect of Fig. 11c, but it could be farther south, in the area not sampled by the CRL. By 0323 UTC, a singular wave with small undulations has developed ahead of the intruding MCS outflow, on top of the moist cold-frontal air (Figs. 11f and 11g). This is preceded (in space, to the north) by a peak in
vertical motion at flight level (Fig. 11e). By the time of the next UWKA transect (not shown), this first gravity wave has advanced northward, and the maximum vertical velocity exceeded 2 m s\(^{-1}\).

b. Bore formation and propagation: Application of hydraulic theory

The partial blocking of the cold-frontal air mass can be further investigated using the two-layer hydraulic theory initially applied by Rottman and Simpson (1989). They used two nondimensional parameters, Froude number (Fr) and normalized height ratio (D), to describe the resulting flow regime and estimate bore strength. The Froude number is the ratio of the speed of a density current (\(C_{dc}\)) to the speed of a gravity wave (\(C_{gw}\)). The height ratio \(D\) compares the height of the obstacle (\(d_o\)) to the initial depth of the blocked fluid (\(h_o\)). The height ratio \(D\) compares the height of the obstacle (\(d_o\)) to the initial depth of the blocked fluid (\(h_o\)):

\[
Fr = \frac{U}{\sqrt{g h_o}} = \frac{C_{dc}}{C_{gw}} \tag{1}
\]

\[
D = \frac{d_o}{h_o} \tag{2}
\]

The relation between Fr and \(D\) determines the flow regime, i.e., whether a flow is blocked and forms a bore. For this application, the MCS outflow is acting as the blocking mechanism (lower-layer obstacle). However, it is also moving in opposition to the cold-frontal air mass that is lifted over it. The speed of the cold-frontal density current (\(C_{dc}\)) relative to the obstacle is therefore the sum of their two speeds, 13.5 m s\(^{-1}\) in this case (the cold front speed: 5.7 m s\(^{-1}\); outflow speed: 8.8 m s\(^{-1}\)). The speed of a gravity wave in the cold frontal air mass can be calculated using the following equation (Klemp et al. 1997):

\[
C_{gw} = \sqrt{g h_o} = \sqrt{g \frac{\Delta \theta}{\theta_o} h_o}. \tag{3}
\]

The theoretical gravity wave speed \(C_{gw}\) is 6.4 m s\(^{-1}\) calculated from the CRL \(\theta_e\) transect (Fig. 7c; \(\Delta \theta = 307–305\) K) with a cold-frontal airmass height \(h_o\) of 850 m. This yields a Froude number of 1.83. The second parameter, the ratio of the height of the obstacle, i.e., the MCS outflow boundary (\(d_o = 600\) m), to \(h_o\) is only \(D = 0.71\) (Fig. 5c). The Froude number and the nondimensional obstacle height indicate that the cold-frontal
airflow is partially blocked, notwithstanding the very shallow MCS outflow. The bore strength $S$ is defined as $S = h_1/h_0$, where $h_1$ is the after-blocking depth. In this case, $2 < S < 3$, which indicates that the bore is mostly undular with some turbulent mixing on the downstream faces (Simpson 1987).

Bore maintenance requires the presence of a wave duct, or at least some mechanism for wave energy reflectance and wave trapping (Crook 1988). Such wave trapping depends on the profile of the Scorer parameter $l$, defined as

$$l^2 = \frac{N^2}{(U - C)^2} \frac{\partial^2 U}{\partial z^2},$$

where $N^2$ is Brunt–Väisälä frequency, $U$ is the environmental wind speed normal to the direction of bore movement, $C$ is the speed of a bore, and $z$ is the vertical distance. The Scorer parameter was calculated using vertically interpolated MP1 radiosonde data (50-m vertical interval, Fig. 5a). A Scorer parameter decreasing rapidly with height to near-zero values indicates a favorable environment for wave trapping and bore maintenance (Crook 1988). In this case, the bore’s wave energy emanates from about 2.2 km, located between layers of negative Scorer parameter near 2.0 and 2.5 km MSL (Fig. 5a). Bores typically propagate southward in the Great Plains, into the low-level jet, which acts to reflect wave energy (Haghi et al. 2017). The postfrontal environment in this case lacked a

![Fig. 10. As in Fig. 4, but using the KDDC radar data at (a),(e) 0312; (b),(f) 0345; (c),(g) 0418; and (d),(h) 0444 UTC. The red dashed lines represent the earlier flight track. The solid red flight tracks in (a) and (e) are the time period between 0247 (the radar time in Fig. 3c) and 0345 UTC.](image-url)
low-level jet. The southerly shear peaking at the top of the cold air mass (near 1.3 km MSL, Fig. 5a) resulted in a minimum in $\ell^2$ through the curvature term in $\ell^2$ [second term on the right of Eq. (4)]. That level likely served as the base of the duct for a northward moving bore.

c. Solitary wave train

After the UWKA sampled the initial collision (Fig. 7), blocking, and a solitary wave (Fig. 11), the aircraft followed the wave train north and conducted a second racetrack pattern to monitor the evolving wave structure. The UWKA transects were oriented from north-northwest to south-southeast to be perpendicular to the waves as seen on radar (Fig. 10). On leg 1 of this second racetrack (0357 UTC, Fig. 12a), as many as four to five waves had developed and were most apparent in LSR and flight-level vertical velocity and pressure perturbations. The first two waves appear to have similar amplitudes of around 500 m (determined by following the variation the cold frontal surface in the LSR transect) and similar perturbations in vertical velocity and pressure (Figs. 12a–d). The trailing waves decreased in amplitude in a manner similar to a solitary wave train or soliton (Knupp 2006). Wavelength also decreased steadily: it is about 9 km for the larger, leading waves, but only about 3 km for the smallest waves. The LSR cross sections plainly show soliton devolvement occurring behind the bore (in reference to the northward movement of the bore) above the partially blocked region. Note the quadrature phase shift between vertical velocity $w$ and pressure perturbation $p'$ in Figs. 12a and 12e. Interestingly, when the UWKA flew below the wave system in the cold-frontal air mass (Figs. 12a–d), the positive pressure perturbations trailed the updrafts and aligned with the wave ridges, indicating that the $p'$ variations were primarily hydrostatic; however, when the UWKA flew above the wave system in the residual layer (Figs. 12e–h, showing the next leg some 10 min later), the positive pressure perturbations led the updrafts and aligned with the wave troughs. This relationship does not match hydrostatic theory and likely resulted from dynamic forces (similar to the case in Grasmick et al. 2018). Lidar attenuation is occurring at the crest of each wave, indicating the presence of clouds. Behind the last wave, a nearly continuous cloud layer existed (bottom panels in Fig. 12). In the next two flight transects, clouds thickened further in the upper reaches of the cold-frontal air mass, especially toward the rear of the wave train, and the leading wave began to dissipate (not shown). Even so, flight-level vertical velocity and pressure perturbations retained four to five waves.

As the wave train continued northward, it moved over MP1 (Figs. 10c and 10d). The wave train passage at MP1 was most
notable in wind speed and direction (Fig. 13). The observations depict a regular interval of wind speed peaks occurring with a period that decreases from about 20 min between the first two waves to 12–15 min between the last apparent waves. Like the UWKA observations, these also display a soliton-like structure in that the amplitude of the wind maximum decreased with each successive wave. The wind direction shows some pattern of regularity as well, at least across the earlier, larger waves. Upon the outflow arrival (at about 0345 UTC), the wind direction shifted from northerly to southerly. At about the same time, the pressure rose and temperature increased, followed by a quasi-static air density as expected from a bore (Koch et al. 1991), since the bore deepened and mixed the stratified air within the cold-frontal air. Following the passage of the first wave, the wind direction briefly returned to northerly by completing a 360° rotation counterclockwise. A similar wind turning pattern was repeated for the second wave, but the transition was less wavelike. After this, subsequent lower-amplitude waves did not cause such dramatic wind shifts.

First, what appears to be a large-amplitude wave (just after 0230 UTC) was most likely a short-lived updraft that coincidentally passes overhead. A radar reflectivity loop (not shown) reveals a small region of weak echoes drifting over the site during this time. Although likely nonprecipitating, the convergence of BL scatterers beneath the updraft was sufficient for radar detection. This updraft does not appear to be related to any convergence boundary but may have been initiated by the cold front passage about an hour earlier.

Second, the deepening of the moister, slightly cooler cold-frontal air at 0345 UTC marked the arrival of the bore, whose depth (~2.0 km) generally matched LSR estimates (Fig. 11). Numerous waves follow the deepening; they appeared on scanning radar as multiple, parallel fine lines. Behind the bore, cooling occurred above the cold-frontal air mass, where the air was mixed upward and cooled adiabatically (dry or moist depending on the presence of clouds). Although the lifted parcels only reached their LCLs and not their LFCs (Figs. 10 and 12h show the formation of shallow clouds, but no deep convection), the cooling and moistening of upper levels reduces CIN, increasing the likelihood of later convection. Near the surface (MP1 surface observations are included in Fig. 14c), cooling occurred on account of the cold front and MCS outflow arriving (e.g., the 306-K θ_e contour).

6. Summary and discussion

The key findings of this case study, based on the 8 June 2015 IOP during PECAN, are as follows:

FIG. 12. As in Fig. 11, but for racetrack 2.
This study demonstrates how the airborne CRL, in combination with proximity operational radar data and radiosonde data, can be used to characterize the fine-scale thermodynamic vertical structure and convective potential of colliding boundaries and resulting bore formation and evolution. Critical in this analysis are the CRL-based 2D cross sections (time–height) of temperature, humidity, and derived convective parameters (CAPE, CIN, parcel-specific distance to LCL/LFC).

The case illustrates fine-scale storm and environment interactions that trigger deep convection. The “zipper-like” coalescence of an MCS outflow and a cold front led to enhanced convergence ahead of the collision point, initiating CI by lofting near-surface air with large water vapor mixing ratio to greater altitude. In this case, a ~5-km-wide plume of cold-frontal, high-θ_e air was lofted significantly in the collision. The CRL thermodynamic data, as well as the lidar-derived aerosol scattering ratio available above and below flight level, show that the collision pushed the less dense, prefrontal air, possibly mixed with postfrontal air, upward, generated a rising plume of high water vapor and aerosol that reached its LFC and initiated convection.

The collision between the cold front and outflow produced a bore on top of the cold-frontal air mass, in the opposite direction of the cold front. The depth, propagation speed, and density differences of the two boundary layer fluids (the cold-frontal air mass and the denser MCS outflow) suggest that the former was partially blocked, according to the two-layer hydraulic theory initially applied by Rottman and Simpson (1989). The resulting bore initially appeared as a singular wave (triggering deep convection), and then evolved into a soliton, evident on the radar as multiple, parallel fine lines. This soliton propagated northward over the cold-frontal air mass, persisted for at least 3 h after the collision, and modified the environment by cooling and moistening air above the cold-frontal air mass.

This observational study describes the fine-scale thermodynamic vertical structure of the collision between two relatively weak boundaries, one associated with a cold front, the other with a MCS outflow. To our knowledge, this study is unprecedented in that it describes the high spatiotemporal resolution thermodynamic structure of convergent boundaries, resulting in CI, and of an undular bore propagating over the less dense

![Surface measurements of (a) temperature and dewpoint, (b) pressure, and (c) wind speed and wind direction at site MP1 during the bore and soliton passage. The vertical dashed line marks the estimated arrival time of the MCS outflow followed by the periodic behavior of pressure, wind, and wind speed beneath the soliton.](image)
air mass (the cold frontal surface). This analysis complements published density current collision case studies, which primarily present radar kinematic analyses and lack detailed thermodynamic information. This novel thermodynamic analysis is a powerful tool to understand the vertical structure of colliding boundaries and their potential to generate bores or trigger deep convection.

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Fig. 14. (a) AERI temperature and (b) WVMR at MP1 between 0200 and 0500 UTC. (c) The temperature and WVMR measured at the surface. Black contours are virtual potential temperature. A vertical dashed line marks the estimated arrival time of the bore.

Data availability statement. The CRL, in situ, and surface-based data are available from PECAN EOL website (https://data.eol.ucar.edu/master_lists/generated/pecan/).

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