Evolution and Vertical Structure of an Undular Bore Observed on 20 June 2015 during PECAN

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ABSTRACT

This study documents the evolution of an impressive, largely undular bore triggered by an MCS-generated density current on 20 June 2015, observed as part of the Plains Elevated Convection at Night (PECAN) experiment. The University of Wyoming King Air with profiling nadir- and zenith-viewing lidars sampled the south-bound bore from the time the first bore wave emerged from the nocturnal convective cold pool and where updrafts over 10 m s\(^{-1}\) and turbulence in the wave’s wake were encountered, through the early dissipative stage in which the leading wave began to lose amplitude and speed. Through most of the bore’s life cycle, its second wave had a higher or equal amplitude relative to the leading wave. Striking roll clouds formed in wave crests and wave energy was detected to about 5 km AGL. The upstream environment indicates a negative Scorer parameter region due to flow reversal at midlevels, providing a wave trapping mechanism. The observed bore strength of 2.4–2.9 and speed of 15–16 m s\(^{-1}\) agree well with values predicted from hydraulic theory. Surface and profiling measurements collected later in the bore’s life cycle, just after sunrise, indicate a transition to a soliton.

1. Introduction

To a stationary observer, the passage of a bore results in a net increase in the fluid mass overhead, unlike the passage of an internal gravity wave [see 19–22 in Simpson (1997)]. A bore essentially is a hydraulic jump. Unlike a density (or gravity) current, it transports little mass and does not advect a reservoir of denser fluid. Atmospheric bores are frequently observed to propagate on a surface-based temperature inversion at night or during the early morning hours, and are generally induced by a density current (e.g., a cold front or a thunderstorm outflow boundary) (Koch et al. 1991). One main characteristic of bores is that they cause the stable boundary layer (SBL) to deepen and to remain deep following bore passage (Koch et al. 2008b). Thus, the hydrostatic pressure increases at the surface due to column cooling (Koch et al. 1991). But unlike density currents, the passage of a bore does not result in any pronounced surface cooling. Surface warming may actually occur because air from above the inversion can be turbulently mixed down (Koch et al. 2005). There is typically cooling aloft due to adiabatic lifting of stratified air. Just like density currents, bores cause the surface wind to shift to the direction of bore propagation, but while density currents are marked by feeder flow (rear-to-front flow, in a frame of reference moving with the density current head) (e.g., Wakimoto 1982; Mueller and Carbone 1987; Geerts et al. 2006), bores usually are marked by bore-relative front-to-rear flow at all levels (Liu and Moncrieff 2000). A bore and its trailing waves are sometimes visually evident from the associated roll/wave clouds (Smith 1988; Coleman et al. 2010).

Bores and related gravity waves are of great interest because they can 1) modify the stability of the nocturnal SBL and thus the dispersal of pollutants; 2) transport energy, momentum, and moisture (e.g., Koch et al. 1991; Martin and Johnson 2008); 3) trigger convection (e.g., Koch and Clark 1999; Locatelli et al. 2002; Coleman and Knupp 2011); 4) may contribute to the maintenance and longevity of nocturnal continental mesoscale convective systems (MCSs) (e.g., French and Parker 2010; Geerts et al. 2017); and 5) may produce turbulence and wind shear, which can be hazardous to aviation (e.g., Christie 1983; Koch et al. 2008a).

The objective of this study is to describe the evolution of the vertical structure of an undular bore triggered by a cold pool associated with a nocturnal MCS. This observational case study is unique in that it explores the evolution of the vertical structure of a bore. While other

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fixed, surface-based profiling observations depict a blend of distance and time evolution, this study uses airborne profiling lidar data, providing a series of quasi-instantaneous vertical transects, at a time resolution corresponding with the frequency of flight traverses across the bore (in this case 10–15 min). Another advantage to using an aircraft is the ability to directly measure thermodynamic and wind variables at flight level to determine phase relationships aloft.

The present study focuses on the interaction between a MCS-generated cold pool outflow boundary and a nocturnal SBL. This interaction may yield bores and/or solitons, as has been documented using observations (Doviak and Ge 1984; Fulton et al. 1990; Mahapatra et al. 1991; Knupp 2006; Koch et al. 2008a,b; Martin and Johnson 2008; Coleman et al. 2009, 2010; Coleman and Knupp 2011) and numerical simulations (e.g., Crook and Miller 1985; Crook 1986; Haase and Smith 1989; Liu and Moncrieff 2000; Martin and Johnson 2008).

This paper is structured as follows. Section 2 provides some relevant background about the dynamics and characteristics of bores. The datasets and research methods are described in section 3. Airborne lidar and in situ observations are described in section 4, and surface-based measurements are described in section 5. A stability and wave-trapping analysis is presented in section 6, and a comparison with theory is presented in section 7. Conclusions follow in section 8.

2. Bores and solitons

The bore strength $S$, defined as

$$S = \frac{h_1}{h_o},$$

(1)

can be used to discriminate between undular and turbulent bores, particularly on the backside of the waves (Rottman and Simpson 1989). In Eq. (1), $h_o$ is the depth of the undisturbed SBL and $h_1$ is the bore height, specifically the height of the midpoint between wave crests and troughs, or alternatively defined as the depth of the lower layer after bore passage. For $1 < S < 2$, bores are undular, with little or no turbulent mixing. The bore head only envelops a small amount of the density current fluid and quickly leaves it behind. The undulations dissipate the excess potential energy created along the bore front. When $2 < S < 4$, the bore is still mostly smooth, but due to shear instability, there is some mixing and turbulence behind the leading wave; the bore ingests more of the density current fluid. When $S > 4$, mixing is intense and there are no noticeable waves behind the leading jump; the bore looks essentially like a density current and distinguishing between the two features is difficult.

Klemp et al. (1997) develop a theory for internal bore propagation, assuming energy is lost not in the expanding lower layer (as in the classical hydraulic theory), but rather in the contracting upper fluid layer. According to Klemp et al. (1997) [their Eq. (14)], the speed of an internal bore $C_b$ for a very large total fluid depth is given by

$$C_b = \sqrt{g' h_o \frac{S}{\sqrt{(1 + S)/2}}} = \sqrt{g \frac{\Delta \theta}{\theta_o} h_o \frac{S}{\sqrt{(1 + S)/2}},}$$

(2)

which becomes the density current speed $\sqrt{g'(\Delta \theta_o/\theta_o)h_o}$ for $S \rightarrow \infty$. Here the vertical difference in virtual potential temperature $\theta_o$ across the inversion $(\Delta \theta_o)$ is used, instead of horizontal density differences [specifically, the reduced gravity term $g' = g(\rho_{\text{high}} - \rho_{\text{low}})/\rho_{\text{high}}$] used in laboratory fluid studies. This equation is simplistic in that it assumes a two-layer Boussinesq fluid and does not take into account mixing between the layers, wind shear, and any nonhydrostatic effects. More complex models for internal bores can be found in the literature, including in Borden and Meiburg (2013), Thorpe and Li (2014), and Baines (2016), but they all assume two-layer flow, rather than a more realistic continuously stratified fluid.

Of importance to atmospheric bore maintenance is a wave duct and some mechanism for wave energy reflectance and wave trapping. Three environmental conditions are favorable for bore longevity (Crook 1988): 1) a layer of smaller Scorer parameter above the SBL, 2) curvature of the low-level wind profile normal to the bore, and 3) evanescence aloft due to winds that oppose the direction of bore propagation or due to low stability. The Scorer parameter $l_s$, which comes from the Taylor–Goldstein equation, is given by the first two terms on the right-hand side of Eq. (3):

$$m^2 = \ell^2 - k^2 = \frac{N_m^2}{(U - C_b)^2} - \frac{\partial^2 U}{\partial z^2} - \frac{1}{(U - C_b)} - k^2,$$

(3)

where $N_m$ is the moist Brunt–Väisälä frequency, $U$ is the environmental wind speed in the direction of bore movement, $(k, m)$ is the (horizontal, vertical) wavenumber, and $z$ is the distance in the vertical (e.g., Koch et al. 2005). Technically, wave trapping requires $m^2 < 0$, but in this study $k^2$ will be omitted due to relatively long horizontal wavelengths. Thus, $m^2$ is considered to be roughly equivalent to $\ell^2$. It follows from Eq. (3) that a decrease in static stability above a SBL and/or
increasing flow curvature \( \partial^2 U / \partial z^2 \) decreases \( I \) with height. The first two conditions listed above are satisfied when \( I \) becomes negative (or decreases rapidly with height), and when a region of maximum curvature falls within a layer of low stability (Crook 1988). Such wave trapping mechanism commonly applies in the Great Plains in cases of south-bound bores encountering the southerly nocturnal low-level jet (LLJ) (Koch et al. 1991). A critical layer, where \( C_b \) equals \( U \) and the Richardson number is less than 0.25, can also prolong bore lifetime due to wave reflectance. The critical layer should not be in the duct, however, because wave energy would then be absorbed in the duct (Lindzen and Tung 1976).

Often associated with atmospheric bores are solitary waves and solitons. A solitary wave is a single gravity wave of elevation (or, less frequently, of depression) which propagates without changing form, at a constant velocity faster than the linear gravity wave phase speed (Christie et al. 1978). In effect, a solitary wave is part of an amplitude-ordered wave train in which trailing waves have a much smaller amplitude. Such a wave train, referred to as a soliton, develops when an internal bore begins to decay and evolves asymptotically to become a soliton (Koch et al. 2005; Christie et al. 1981). The main distinguishing point between solitary waves and bores is that the SBL top may oscillate during the passage of a soliton, but there is no sustained elevation of the SBL. Therefore, surface pressure will oscillate, but not be noticeably higher after the wave train passes (Simpson 1997). Also, solitary waves, unlike bores, do not cause a substantial wind shift at the surface (Knupp 2006). Perturbations in parameters mostly occur within the inversion level or SBL top (Christie et al. 1979).

Atmospheric bores and solitary waves are triggered in two principal ways: (i) interactions between density currents (such as sea breezes, convective cold pool outflows, katabatic flows and cold fronts) and an SBL (Clarke 1972; Clarke et al. 1981; Smith 1988; Clarke 1998; Goler 2009; Hartung and Sitkowski 2010); and (ii) collisions between density currents (Smith 1988; Wakimoto and Kingsmill 1995; Goler and Reeder 2004; Goler 2009).

### 3. Data and methodology

This study focuses on the structure and evolution of a bore near Grand Island in central Nebraska around dawn on 20 June 2015 during the Plains Elevated Convection at Night (PECAN) field campaign (Geerts et al. 2017). The 20 June bore was sampled primarily by University of Wyoming King Air (UWKA), which employed two lidars: the downward-pointing Compact Raman lidar (CRL) and the upward-pointing Wyoming Cloud lidar (WCL).

The CRL is a multichannel rotational Raman lidar that utilizes a flashlamp-pumped Nd:YAG laser at \( \sim 355\text{-nm} \) wavelength (Liu et al. 2014). In its original design, it had \( \text{N}_2 \) and \( \text{H}_2\text{O} \) vibrational–rotational Raman channels for water vapor measurements, and two elastic channels for aerosol and cloud measurements (e.g., Bergmaier et al. 2014). In 2015, low-\( J \) and high-\( J \) (\( J \) is the rotational quantum number) pure rotational Raman channels were added for temperature measurements (Wu et al. 2016). CRL temperature is not used here because, for the purpose of this study, the spatial resolution at relevant ranges is inadequate. This study only shows CRL profiles of water vapor mixing ratio and lidar scattering ratio (LSR). LSR is a measure of aerosol loading. The LSR is the ratio of total backscattering to molecular backscattering, which is determined with the elastic and \( \text{N}_2 \) Raman channel measurements. It is a normalized quantity where a value of 1.0 represents scattering by air molecules only and not aerosols. The CRL can provide useful data starting at 30 m from the aircraft with a vertical resolution of 0.6 m. This was averaged to 1.5 m for this study. For the water vapor and aerosol channels, the CRL’s horizontal resolution is about 90 m, considering the aircraft’s speed of \( \sim 90 \text{ m s}^{-1} \).

The WCL is a 355-nm wavelength compact elastic lidar, which provides attenuated backscattering coefficient and depolarization ratio profiles (Wang et al. 2009). The WCL has an along-beam resolution of 1.5 m, and the cross-beam resolution is often between 5 and 20 m, depending on pulse average number for each saved profile. WCL data were averaged to the same time resolution (1 s) as the CRL. The lidar overlap factor was determined and corrected based on the molecular signal collected above the boundary layer (Wang et al. 2009), so that the first useful data point is approximately 30 m above the UWKA. The WCL LSR is derived with WCL attenuated backscattering coefficients and estimated molecular backscattering coefficients based on flight level temperature and pressure measurements and the standard temperature lapse rate. Those profiles are then combined with CRL LSR profiles, resulting in a full LSR profile centered at flight level. The lidar datasets are corrected for aircraft pitch and roll, so data are a function of altitude instead of range. The transects shown in this study all maintain the same height-to-width aspect ratio.

In situ UWKA variables utilized in this study include vertical velocity, wind speed and direction, potential temperature, and equivalent potential temperature. In addition to UWKA measurements, this study uses data
from the Fixed PECAN Integrated Sounding Array (PISA) 4 (FP4) in Minden, Nebraska (Fig. 1). FP4 data include a radiosonde, a 915-MHz profiler, an Atmospheric Emitted Radiance Interferometer (AERI), and a surface weather station. We also use reflectivity and radial velocity data from the KLNX (North Platte, Nebraska), KOAX (Omaha, Nebraska), and KUEX (Hastings, Nebraska) WSR-88Ds, as well as some operational surface and upper-air data.

4. UKWA-observed structure and evolution

a. Cold pool source and ambient vertical structure

During the second half of the night on 20 June 2015, rather strong westerly winds were present aloft over the northern Great Plains ahead of a Pacific Northwest trough and lee cyclogenesis was induced over the western South Dakota–Nebraska border. A substantial MCS developed over the western part of South Dakota around 0300 UTC. It tracked across South Dakota through the night, producing large hail and winds over 65 kt (33.4 m s$^{-1}$). Farther to the south in Kansas and southern Nebraska, a southwesterly LLJ was present, peaking at $\sim$25 m s$^{-1}$ around 0600 UTC. This nocturnal LLJ is evident in a sounding released at 1000 UTC at FP4 (Fig. 1), to the south of the MCS and outflow boundary (Fig. 2). Also evident in this sounding is a strong SBL, with an 11-K potential temperature ($\theta$) increase in the lowest 377 m. We use this value for the SBL depth $h_o$. This SBL sits below a less stratified stable layer (up to $\sim$800 hPa or 1.3 km AGL), which in turn lies below a deep elevated mixed layer (up to 450 hPa or 6.0 km AGL). While the SBL was fairly moist, a deep, dry layer is found up to about 500 hPa. This dry air is important as it supports cloud-free conditions and thus good lidar measurements of the bore (although, as will be shown below, the bore lifting is deep enough to cause condensation within wave crests). It also results in a potentially unstable layer above the SBL: the highest equivalent potential temperature ($\theta_e$) values (below 450 hPa) are found near the surface, implying the presence of potential instability, especially between 850 and 810 hPa, as evident from the wet-bulb temperature profile in Fig. 2.

b. Wave train as observed by the NEXRAD network

The MCS cold pool produced a south-southeastward-propagating wave train evident in the base reflectivity and radial velocity maps of several WSR-88Ds as soon as the outflow boundary emerged from the MCS precipitation echoes. Cells of deep convection were triggered along this boundary to the east of the first flight leg at 0743 UTC (Figs. 3a,b). Time series of KLNX base reflectivity maps show that such bore-induced convection initiation (CI) was common earlier, but after 0743 UTC it was found only farther east as the entire MCS moved eastward. The precipitation echoes near the first flight
leg (Fig. 3a) were weak and decaying, and the closest lightning strikes were \( \sim 100 \) km away. The base reflectivity maps reveal three wave crests, the leading one being best defined (Fig. 3a). During this time and below 725 m AGL, the wind was northerly (positive bore-normal flow) in wave crests (high reflectivity belts), while the background southwesterly flow (negative bore-normal flow) characterized wave troughs, according to KLNX Doppler velocity data (Fig. 4).\(^2\) KLNX base reflectivity maps indicate that the average ground-relative leading-wave bore speed was \( \sim 15 \) m s\(^{-1}\) toward the south-southeast (from \( \sim 350^\circ \)) at this time. In a frame of reference moving with the bore, the low-level wind speed in the wave crests is near zero, while that in the wave troughs displays strong (\( \sim 28 \) m s\(^{-1}\)) front-to-rear flow. A similar velocity pattern near the surface was present in the undular bore documented by Coleman et al. (2010). Given an ambient bore-normal wind of \( \sim 10 \) m s\(^{-1}\) from 170° (Fig. 2), this bore system (crests + troughs) does transport mass into the ambient flow. Positive wind anomalies in wave crests are consistent with wave polarization relationships for trapped linear gravity waves. Above 725 m AGL, winds in wave crests shift to the prevailing southwesterly direction, while the direction in the trough remains unchanged. This results in a smaller crest-to-trough wave amplitude aloft, up to 2.3 km AGL, the highest level with sufficiently strong echoes (Fig. 4).

The UWKA completed 18 transects normal to this wave train between 0744 and 1034 UTC.

c. Sample UWKA transects

The UWKA only penetrated the leading wave of the bore system in its first inbound leg (“leg 1”). The corresponding outbound transect (“leg 2”) is shown in

\( ^2 \) The range-to-height equation used to assign beam center height values in Fig. 4 is that used in the WSR-88D algorithm, which is a simplification of Eq. (2.28b) in Doviak and Zrnić (1993), which assumes a constant index of refraction. Because of the high stability at relevant heights (Fig. 2), the beam height may be overestimated somewhat.
Figs. 5a–e. At this initial stage of the wave train evolution, the leading wave roll cloud reached above flight level (3.0 km MSL or ~2.3 km AGL). The prebore boundary layer air had a low aerosol content (LSR values close to 1.0, Fig. 5a) as well as pronounced moisture stratification (Fig. 5b). Bore-relative wind speed increased in the wave crest (Fig. 5c, negative values indicating front-to-rear flow). It follows from the phase relation between flight-level vertical velocity (Fig. 5d) and the parcel displacement suggested by the moisture structure (Fig. 5b) that the flow is front to rear. In the flow’s approach to the wave crest, \( \theta \) dropped and \( \theta_c \) increased rapidly (Fig. 5e), indicating that this air originated from well below flight level, in fact from the stable layer top at 890 hPa (377 km AGL) (Fig. 2). Indeed, some of the moist SBL air was carried upward over the wave (Fig. 5b) and at flight level the crest was preceded by a smooth updraft of nearly 9 m s\(^{-1}\), peaking

FIG. 3. (a),(c) 0.5° reflectivity and (b),(d) 0.5° radial velocity from the KLNX radar at (a),(b) 0743 and (c),(d) 0830 UTC 20 Jun 2015. The black/blue line indicates the UWKA flight track; the blue part depicts the 10 min prior to radar image time. Positive Doppler velocities are outbound. Range rings are plotted every 50 km.

FIG. 4. Bore-normal winds derived from WSR-88D and UWKA flight level at ~0800–0900 UTC. The dashed line represents the phase speed of the bore.
at the time of most rapid $\theta_c$ and $\theta$ changes. The negative bore-relative winds (Fig. 5c) suggest that the UWKA did not sample a density current, which requires rear-to-front flow.

The UWKA penetrated the wave train to the second wave on the second inbound leg (leg 3). Here, and on the subsequent outbound leg (leg 4; Figs. 5f–j), and in fact in all subsequent transects, the second wave had an equal or higher amplitude and displaced air parcels higher than the first wave. The reason for such pattern is unclear. The strongest updraft observed on this flight ($18.2 \text{ m s}^{-1}$) was measured in advance of the second wave on leg 4 (Fig. 5i). While the leading wave was quite undular (smooth), wave breaking is evident mainly on the lee side of the second wave, as evident from the depiction of the cloud edge (Fig. 5f) and from flight-level turbulence measurements. These measurements indicate eddy dissipation rate (EDR) values up to $0.45 \text{ m}^{2/3} \text{s}^{-1}$ on leg 4, implying turbulence in the “moderate to severe” category (Strauss et al. 2015). A similar sequence of an undular leading wave and breaking trailing wave is documented in Mahapatra et al. (1991).

By the time of leg 4, the leading wave appeared to have weakened, being more removed from the cold pool core. Compared to leg 2, the leading wave’s updraft decreased by a factor of 2, the flight-level $\theta_c$ and $\theta$ perturbations (cf. the prebore environment) by about a third, and the maximum vertical displacement by about 1 km, according to a comparison of Figs. 5a–e with Figs. 5f–j. The extreme values of $\theta_c$ and $\theta$ in the leading wave crest indicate that the SBL no longer was lofted to flight level—only the less stable layer above the SBL was detected at flight level (Fig. 2).

By 0830 UTC, most of the precipitating cells behind the leading line had dissipated near the flight track and the UWKA was able to sample the full wave train (Figs. 3c,d). Only two full radar fine lines are apparent at this time. An example of this double wave structure is shown in Figs. 5k–o (the ninth flight leg, 0840 UTC). Both waves are quite undular at this time. The second wave extends only to a slightly higher altitude.

By 0900 UTC a mature wave train with four–five waves had developed, according to KLNX base reflectivity imagery (Figs. 6a,b). The UWKA remained focused on the two leading waves in leg 12 (Figs. 5p–t). At this time the leading wave had weakened noticeably in terms of apparent parcel vertical displacement and
magnitude of flight-level perturbations, while the second wave maintained its amplitude. The CRL signal was attenuated in the trough between the leading and second wave clouds, most likely due to large, hygroscopic aerosols (Fig. 5p). This indicates limited subsidence in the wave trough, and increased turbulence and mixing behind the cloud top of the leading wave.

After leg 12, the UWKA moved eastward to “area 2” (Fig. 1) to document bore diversity closer to the propagating MCS. The KOAX radar base elevation scan depicts only two well-defined wave crests near the flight track around 1000 UTC, although there may have been more, too low for the radar to see at this large range (Figs. 6c,d). Again, the wave crests were moving toward the south-southeast at \( \frac{15}{1000} \text{m s}^{-1} \). Around this time, the aircraft sampled a wave train with three crests (Fig. 7). Characteristic of a bore (Simpson 1997), sustained deepening of the SBL occurred, as evident in the depth of the \( 13 \text{ g kg}^{-1} \) isohume (yellow in Fig. 7b). Here the 0.5–1.5 km AGL stable layer (above the more humid SBL, Fig. 2) was more humid than in area 1 (Figs. 5b,g,l,q). Greater boundary layer moisture raises convective available potential energy (CAPE), yet the prebore sounding reveals much convective inhibition for any source level with CAPE. The lidar LSR (Fig. 7a) reveals a layer of likely large (deliquescent) aerosol at \( \frac{1.4}{1.5} \text{ km AGL in the prebore environment. This layer does not become saturated above the leading wave, but it} \) does above the higher second wave, following a vertical displacement of \( \frac{1.6}{1.5} \text{ km. The resulting wave cloud, capping near flight level, largely hides the wave structure below, although the CRL signal only becomes fully attenuated by a second, lower wave cloud, capping near 2 km AGL. The second wave again had the highest crest-to-trough amplitude, with a layer lift of about 1.7 km, followed by the third wave (\( \frac{1.2}{1.5} \text{ km vertical lift}), and finally the weakening leading wave with just 0.9-km vertical lift (Figs. 7a,b). At flight level, now 500 m higher (3.5 km MSL, 3.0 km AGL), the second wave also displays a higher amplitude than the two others in terms of in situ variables. The variations of \( \theta \) and especially \( w \) are rather undular and of rather small magnitude (Figs. 7c–e). Remarkably, these wavelike variations are not matched in the bore-normal wind (Fig. 7c). WCL data reveal a third layer of clouds near 5.0 km AGL, corresponding to a rather humid layer in the prebore sounding (Fig. 2). Onboard visual observations in the early-dawn light indicated several altostratus cloud lines aligned with the lower wave clouds. The high moisture content in the SBL and above the residual mixed layer likely promoted the formation of these stable wave clouds (Crook 1986).

**d. Wave train evolution: Aircraft measurements**

To summarize the vertical structure and evolution of the system, all legs were assembled into outline schematics for areas 1 and 2 (Fig. 8). The horizontal position of each wave outline in this figure was determined by projecting the location of the leading wave crest
(sampled by the CRL) during each leg onto a common bore-normal transect. Then the horizontal displacement in Fig. 8 was calculated as the distance of the leading wave crest between successive flight legs along this common transect. This, together with time difference, then provides a lidar-based estimate of phase speed, which is listed in Table 1. The outlines in Fig. 8 are traces of the mixing ratio isopleth, which corresponds to the top of the SBL (Fig. 2). Assuming steady flow, such isohumes correspond to parcel trajectories since water vapor is conserved in the absence of clouds (i.e., up to the cloud edge) (Koch et al. 1991). In that sense, and to the extent the flow is adiabatic, the outlines can also be interpreted as isentropes. The isohumes across the second wave are somewhat uncertain for legs 15–18 because of concealment by a wave cloud between flight level and the isohume (Figs. 7a,b). Hence, the dashed line section in Fig. 8 for these legs.

The flight legs in area 1 (Fig. 8) reveal a deamplification of the leading wave and a general increase in wavelength between the leading and second wave from ~12 km in the first sample to 15–20 km after several legs, to 25 km by leg 12. The increase in wavelength from the end of the mature stage through decay has been noted in other studies (Hartung and Sitkowski 2010; Koch et al. 2008b). Also, the reader may note the ragged, turbulent nature of the trailing edge of the leading wave in legs 10, 11, and 12 (Fig. 8). Martin and Johnson (2008) made a similar observation in a bore, although they did not document bore evolution. Laboratory studies indicate that turbulent mixing on the trailing, downstream face of the leading wave should emerge as the bore strength increases from 2 to 3 (Rottman and Simpson 1989).

Farther to the east, in area 2, the second wave was more turbulent and the speed of the wave train decreased as it progressed (Fig. 8). The wavelength trend at this stage and location was slightly different: waves in legs 13–16 had quite long wavelengths, between 20 and 30 km, but as the leading wave began to weaken dramatically in leg 17, the wavelength between the leading wave and second wave actually decreased. Crest displacement analysis shows that this was due to a decrease in speed of the leading wave but not the second wave (i.e., the second wave was catching up). The spacing between the second and third waves remained large. Flattening of the leading wave may have been due to the longer wavelength component of the bore beginning to take over the shorter wavelength components (Mahapatra et al. 1991). Koch et al. (2008b) also saw the leading wave flatten with time, which they hypothesized precipitated into the decay of the entire wave train. In our case, the wave train persisted for at least another 2 h, until daytime surface heating destroyed the SBL. At no time in its evolution are the waves amplitude ordered, a characteristic of solitary waves (Christie et al. 1979). The wavelengths observed here are rather large: estimates closer to 5–10 km are reported in other studies (Coleman et al. 2009, 2010; Coleman and Knupp 2011; Koch et al. 2008b).

The evolution of flight-level perturbations is summarized in Table 1. Both the leading wave speed and the wave amplitude (in terms of updraft strength, vertical displacement, and temperature anomalies) generally decreased over the nearly 3-h period of aircraft observations. The maximum updraft recorded occurred during the leg where the UWKA was first able to fly through the second wave. This corroborates the strong, high-amplitude nature of the second wave. All θ minima were recorded when the UWKA penetrated a wave crest cloud. The θ minima indicate that air from the top of the SBL (377 m AGL) was lifted several kilometers to flight level. Koch et al. (1991) also found several degrees of cooling in the lower troposphere upon bore passage at corresponding levels.
e. Wave train evolution: Measurements at FP4

The wave train was sampled by FP4 (Fig. 1) starting at 1150 UTC (i.e., 76 min after the last UWKA flight leg, and 45 min after sunrise). In close proximity to the KUEX radar, four–five wave crests are apparent (Fig. 9). The 915-MHz profiler SNR and vertical velocity time–height transects show even more wave crests passing through over several hours, above the developing convective boundary layer (Fig. 10). The first two waves had the highest amplitude, with vertical displacements of 1.0–1.5 km, at least for air parcels below the original flight level (3.0 km AGL) (Fig. 10a). The vertical velocity couplet, peaking at up to 5 m s\(^{-1}\) up and down, extend from the surface to the highest level with sufficient SNR (~3.5 km AGL), but the peak values occur lower (~1.0 km AGL) (Fig. 10b). The vertical velocity couplet was strongest in the leading wave, similar to Martin and Johnson (2008). The wave train appears more amplitude ordered at this time, although the first two waves are of roughly equal size (Fig. 10). (The second wave has a greater crest-to-trough vertical displacement and a longer wavelength than the leading wave, but the leading wave has the strongest updraft.) The trailing weaker waves also have a shorter wavelength. Since the abscissa (time) combines translation and evolution, it is not clear whether these changes are spatial or temporal. The distance between the first two crests is estimated at 31 km (given a bore speed of ~15 m s\(^{-1}\)), which is only slightly longer compared to the final lidar-based estimates (Fig. 8). The waves become decoupled from the surface about two hours after sunrise (1305 UTC) as a surface mixed layer develops (Fig. 10b). The 915-MHz wind profiler cannot resolve strong wavelike variations in wind, as the averaging period for these data cannot be reduced to less than

<table>
<thead>
<tr>
<th>Time (UTC)</th>
<th>Leg No.</th>
<th>Area</th>
<th>Flight level (km, AGL)</th>
<th>Peak updraft (m s(^{-1}))</th>
<th>Max cooling ((\theta' &lt; 0)) (K)</th>
<th>Source level (km below flight level)</th>
<th>Leading wave phase speed (m s(^{-1}))</th>
</tr>
</thead>
<tbody>
<tr>
<td>0740–0800</td>
<td>1–2</td>
<td>1</td>
<td>2.3</td>
<td>8.3</td>
<td>–7.5</td>
<td>1.7</td>
<td>22.0</td>
</tr>
<tr>
<td>0800–0830</td>
<td>3–8</td>
<td>1</td>
<td>2.3</td>
<td>5.2</td>
<td>9.6</td>
<td>–7.4</td>
<td>1.3</td>
</tr>
<tr>
<td>0830–0900</td>
<td>9–12</td>
<td>1</td>
<td>2.3</td>
<td>4.3</td>
<td>6.2</td>
<td>–7.1</td>
<td>1.0</td>
</tr>
<tr>
<td>0930–1030</td>
<td>13–18</td>
<td>2</td>
<td>3.0</td>
<td>2.3</td>
<td>3.9</td>
<td>–2.1</td>
<td>1.2</td>
</tr>
</tbody>
</table>
10 min (Fig. 10c). (The SNR and vertical velocity data in Figs. 10a and 10b have a resolution of 1.5–2.0 min.)

The leading wave corresponds with a dramatic change in low-level momentum. This change is consistent with the changes in the wind profile at great distances across the bore (Fig. 2), but the concentrated change at the bore front indicates that the bore transported significant momentum. The wind profiles from this 915-MHz profiler are compared to KUEX radial velocity values normal to the bore, both in wave crests and troughs (Fig. 11). The wave pattern derived from scanning WSR-88D is essentially the same wind anomalies as observed as seen three hours earlier (Fig. 4): in a frame of reference moving with the bore, the low-level wind speed is near zero in the wave crests, and highly negative in the troughs. Tremendous wind shear is present in the wave crest at ~0.6 km AGL, again matching the earlier observations. The wind profiler does not have the resolution to fully resolve this wave, but the sign of the low-level wind (and wind shear) anomalies match WSR-88D observations.

Continuous surface data at FP4 (Fig. 12) show oscillations in the bore-relative wind speed consistent with the WSR-88D and wind profiler data near the ground: the surface wind shifted to northerly in crests, in the direction of bore motion, but never exceeded the bore speed (no rear-to-front flow), consistent with Coleman et al. (2010). CRESTS are defined here as periods with positive $p^\prime$ (Fig. 12a), a hydrostatic perturbation due to the cooling aloft as documented in crests at flight level. Here $p^\prime$ is defined as the surface pressure perturbation from the 6-h mean. Unlike at flight level where strong front-to-rear flow occurred in crests, the surface front-to-rear flow in crests is anomalously weak. The leading wave had a $p^\prime$ value of ~2 hPa. While the leading four waves are well defined in the surface bore-relative wind and $p^\prime$ fields, the wave signature is not nearly as obvious in the temperature and moisture fields, as expected. The leading wave caused ~0.5 K of cooling at the surface, which is benign in comparison with $\theta^\prime$ values at flight level. The sharp, but minor rise in temperature after the leading wave could have been due to the turbulent mixing on the downstream wave face.

There was a significant change in the boundary layer temperature profile right above the surface layer though, according to AERI data (Fig. 13a). Significant cooling below 2 km AGL, mainly between 0.5 and 1.0 km, was sustained for ~2 h until the daytime heating took over. Hydrostatically, the AERI-estimated cooling in the lowest 2 km should produce a ~4 hPa pressure rise, about twice the observed value (Fig. 12a). The nearly 10-K SBL inversion vanished in the first crest and never fully recovered in the following wave troughs. Koch et al. (1991) also saw bore passage result in destruction of the SBL. Rottman and Simpson (1989) note that the downstream fluid is dramatically changed after bore mixing. (In the absence of daytime surface heating, the SBL may have been restored of course.) The AERI water vapor mixing ratio profile (Fig. 13b) shows that sustained moistening occurred above the SBL due to ascent in the first wave crest, resulting in a cloud based at 2 km AGL and slight loss of water vapor near the surface, as observed also by a surface humidity sensor (not shown). The 8 g kg$^{-1}$ isopleth was lifted about 1.7 km after the first couple of waves passed. Thus, the cooling of the lowest 2 km was due primarily to (adiabatic) bore lifting, although, since the bore carried some momentum, there may have been some cold-air advection also, explaining the observed surface cooling with the passage of the leading wave (Fig. 12b). Knupp (2006) also noted parcel displacements of about 2 km and moistening above the bore head (instead of near the surface) due to the permanent lofting of parcels.

5. Undular bore vertical structure

The observed flight-level and near-surface phase relationships are summarized in a schematic in Fig. 14,
together with bore-normal wind and potential temperature profiles on opposite sides of the bore. One caveat is that UWKA and FP4 data are merged, whereas in reality these data were collected several hours apart and, in fact, in that period the perturbation changed from a bore with a low-amplitude leading wave to an amplitude-ordered system reminiscent of a soliton, as will be discussed below. This change is ignored in the schematic of Fig. 14.

Near the surface, the pressure perturbations are largely hydrostatic and the wave crests display a wind component in the direction of the bore propagation.
(northerly), roughly matching the bore phase speed itself (Figs. 4 and 11). Thus, characteristic of a bore of high strength ($S$ large) and unlike linear gravity waves, this disturbance transports momentum (opposing the LLJ) as a material fluid in the SBL. The UWKA mostly flies above this material fluid; there wave crests display a southerly (front to rear) wind anomaly (Figs. 4 and 11), on account of the lofting of the LLJ from below (see schematic prebore wind profile on the right in Fig. 14). Flight-level measurements indicate that the momentum and temperature variations ($u'$, $w'$, and $\theta'$) are consistent with a trapped gravity wave. Wave breaking aloft is observed in the lee of mainly the second wave crest in several transects (Figs. 5, 7, and 8).

Koch et al. (2008a) discussed the material fluid encompassed by bores and described the bore in their case causing an “indirect influence” on the waves detected aloft due to lifting and the propagation of wave energy to higher altitudes. They also noted elevated aerosol layers (like in Fig. 7) and commented that the indirect influence could produce cloud bands above the bore crests. This was evident in this case as altostratus bands (e.g., Fig. 7) and was recorded in Koch et al. (1991) as scattered fractocumulus clouds.

6. Stability and wave trapping analysis

Low-level stability provided by the nocturnal inversion allowed the formation of a temperature duct for wave propagation. Strong low-level stability is evident in the FP4 sounding at 1000 UTC, 1.8 h before bore arrival (Figs. 2 and 14). Because FP4 did not launch another radiosonde after the waves passed, the 1110 UTC (nominally 1200 UTC) North Platte sounding 163 km to the northwest was used to describe the postbore environment (Fig. 1). Radar data indicate that the leading wave passed over North Platte at ~1045 UTC, and that the radiosonde was launched just after the passage of the second wave crest. The red sounding in Fig. 2 confirms this passage, showing light north-northwesterly winds near the surface and an increase in the inversion layer to 803 hPa (~1.1 km AGL). The observed increases in the stable layer depth and wind shift are classic signs of bore passage (Koch et al. 2008a). The North Platte sounding reveals little...
vertical mixing, suggesting the passage of an undular bore of low strength $S$ in that area.

The prebore FP4 sounding data were redistributed on a coarse grid ($\delta z = 200$ m) to obtain smoother profiles of bore-normal wind $U$, bore-relative wind $U - C_b$, the square of the Scorer parameter $l^2$, and $\theta$ (Fig. 15). The Scorer parameter (Fig. 15b) further was filtered using a Fourier transform with a low-pass wavelength of 1 km. The wind profile highlights a strong LLJ within the inversion layer and a decrease in wind speed above the LLJ. The bore-normal wind $U$ was positive from about 3.3 to 6.4 km AGL, due to northwest winds (Fig. 2) associated with a passing synoptic-scale trough. While the bore-relative wind speed decreased significantly with height, there was no critical layer ($U - C_b = 0$) to absorb the wave energy. Two layers of strong curvature in the wind profiles correspond to $l^2 < 0$ at 1.7 and 4.1 km AGL. Karyampudi et al. (1995) and Koch et al. (2008b) also found a region of $l^2 < 0$ just above the southerly LLJ, at a similar height. Did either of these two $l^2 < 0$ layers enable the wave trapping needed for this undular bore to be long lived (Crook 1988)? In other words, was the ducting layer below $l^2 < 0$ (whose
depth is denoted as $H$) sufficiently deep to support 1/4 of the vertical wavelength ($\lambda_z = 2\pi/m = 2\pi/l$) (i.e., does $H \approx \lambda_z/4$)? A quarter of the vertical wavelength is 2.7 (2.3) km for the lower (upper) $\lambda_z$. Thus, only the upper layer at $H = 4.1$ km (with an even lower $\lambda_z^2$) is deep enough for wave trapping.

Indeed, wave activity was detected up to just above 4.0 km AGL (Fig. 7). That level corresponds nearly exactly to the second $\lambda_z^2 < 0$ layer. Some wave energy leaked through this layer, as altostratus cloud bands aligned with the bore crests were present between 5 and 6 km AGL. In $\lambda_z^2 < 0$ layers, waves are evanescent and reflection occurs, although not perfect reflection as the layer is finite (Lindzen and Tung 1976). Wave energy reflection was likely also enhanced by the static stability profile (Fig. 15c), with a deep residual mixed layer above the low-level stable duct (Rottman and Einaudi 1993). In fact, all of the wave trapping criteria in Crook (1986) are met in this case: the MCS cold pool boundary evolved into a bore once a strong SBL had developed; there was very weak stability in the significant $\lambda_z^2 < 0$ region, and winds aloft (above 7 km AGL) opposed the bore motion (Crook 1986). The waves also lacked any significant tilt with height, at least up to \( \sim 4 \) km AGL, which supports the theory that waves were trapped (Koch et al. 2008b).

7. Discussion

a. Bore or soliton?

To properly classify the 20 June wave train as a bore, we compute theoretical values based on bore and solitary wave theories and compare them to observed values (Table 2). Data sources include the 1000 UTC FP4 radiosonde, the 1110 UTC North Platte radiosonde, the WSR-88D base velocity, and the CRL on board the UWKA. We focus on the early phase with airborne measurements (0732–1034 UTC), rather than the later phase after 1150 UTC with FP4 data. The redundancy in methods listed in Table 2 provides an estimate of uncertainty. The observed bore strength [Eq. (1)] of 2.39–2.89 implies, based on the theory of Rottman and Simpson (1989), that the wave train was mostly smooth with mixing and turbulence on the trailing edge of waves, particularly the leading wave. This agrees well with CRL observations (e.g., Fig. 8).

Lidar and WSR-88D observations agree on a bore speed of 15–16 m s\(^{-1}\). This is ground relative, facing a strong LLJ
headwind. To compare this against the theoretical bore speed estimate in Eq. (2), we subtract the magnitude of the head wind. This assumption of Galilean invariance was made also by Kingsmill and Crook (2003) and Coleman et al. (2009, 2010). The prebore bore-normal head wind was about 16 m s$^{-1}$ based over the postbore SBL depth $h_1$ (Fig. 2). When considering the ground-relative adjustment, the predicted values of $C_b$ of 14.1–16.3 m s$^{-1}$ (Table 2) are close to the observed values. These results support the hypothesis that the observed feature was a bore.

This bore does not evolve into a soliton during the airborne measurement period, because the second wave retains the highest amplitude, the isohumes (isentropes) retain elevation upon bore passage (Fig. 8), the wave passage causes a substantial wind shift (Figs. 3, 6, 9, and 12a), and the predicted soliton speed is much smaller than the observed speed. The theoretical speed of the leading solitary wave in a dispersive, amplitude-ordered wave train (soliton) can be estimated as follows (Christie et al. 1978):

$$C_s = \sqrt{gh_0 \frac{(\rho_1 - \rho_2)}{\rho_1} \left(1 + \frac{3}{4} \frac{a}{h_0}\right)}, \quad (4)$$

where $\rho_1$ is the density of the density current, $\rho_2$ is the density of the stable layer, and $a$ is the maximum amplitude of the wave. Using ($\rho_1$, $\rho_2$) values obtained from an upwind station that witnessed the passage of the parent density current (see below), KAIA (Fig. 1), and assuming $a = 1.7$ km (Fig. 7) and Galilean invariance, the ground-relative value for $C_s$ is only $\sim$5 m s$^{-1}$, about one-third of the observed speed of the leading bore wave. The speed of propagation did not decrease significantly from the leading wave through the train, unlike the amplitude-ordered solitary wave train described in Koch et al. (2008a). Thus, the undulations behind the bore’s leading edge essentially are vertically stacked trapped waves moving at the same speed as the bore head (Koch et al. 2008b).

Later on, at FP4, the perturbation does appear to transition into a soliton, as the waves are mostly amplitude ordered (Figs. 10 and 12), although even then the surface pressure increase is sustained after the wave train passes (Fig. 12a). The residual waves after $\sim$1320 UTC certainly seem to represent a soliton above a developing convective BL (Christie et al. 1979).

b. Parent density current

As the MCS-generated density current was vital to the formation of the bore [in fact 40% of the energy of a density current can be used to sustain a bore; White and Helfrich (2012)], investigation into its characteristics is important. An SBL topped by a neutral layer should cause the density current depth to decrease, but the density current speed to increase, as compared to a neutral atmosphere (Liu and Moncrieff 2000). The following expression for the density current depth $d_o$ (Koch et al. 1991) is used:

$$d_o = \frac{\theta_{wc} \Delta p}{\rho_w g \left(\frac{\theta_{wc}}{\theta_{wc}}\right)^{\frac{1-k}{1-k}}} - \theta_{wc}. \quad (5)$$

[In contrast with Koch et al. (1991), we use virtual potential temperature $\theta_{wc}$ instead of temperature. The subscripts refer to the warm ($w$) and cold ($c$) side of the density current. Here $k = R_d/C_p$ with $R_d$ being the dry air gas constant, and $C_p$ being the specific heat for a constant pressure process. The difference in surface pressure ($\Delta p = p_c - p_w$) is assumed to be hydrostatic. This equation further assumes the density is constant throughout the depth of the density current.]

The MCS cold pool outflow boundary density current transformed into an undular structure with multiple radar fine lines before the first UWKA transect ab initio (i.e., as soon as the boundary peeled away from precipitation) (whose echoes overwhelmed any clear-air reflectivity) (Fig. 3a). Yet the boundary appeared as a single-fine-line boundary that passed over the Alliance, Nebraska, ASOS station (KAIA) farther west (Fig. 1), around the same time as the first UWKA transect in area 1. Warm sector values were taken at KAIA at 0730 UTC, and cold sector values were taken at 0854 UTC. The surface wind shifted from the southwest (at 2 m s$^{-1}$) to the

---

### Table 2. Observed and theoretical values of bore and density current parameters based on data collected between 0730 and 1030 UTC.

<table>
<thead>
<tr>
<th>Parameters</th>
<th>Frame of reference</th>
<th>Lidar method</th>
<th>Sounding or radar method</th>
</tr>
</thead>
<tbody>
<tr>
<td>Bore strength $S$ (observed)</td>
<td>Intrinsic</td>
<td>2.39</td>
<td>2.89</td>
</tr>
<tr>
<td>Bore speed $C_b$ [theoretical: Eq. (2)] (m s$^{-1}$)</td>
<td>Intrinsic</td>
<td>30.1</td>
<td>32.3</td>
</tr>
<tr>
<td>Bore speed $C_b$ (observed) (m s$^{-1}$)</td>
<td>Ground-relative</td>
<td>14.1</td>
<td>16.3</td>
</tr>
<tr>
<td>Density current speed $C_{dc}$ (observed) (m s$^{-1}$)</td>
<td>Ground-relative</td>
<td>15.7</td>
<td>15.3</td>
</tr>
<tr>
<td>Froude number</td>
<td>—</td>
<td>21.4</td>
<td></td>
</tr>
<tr>
<td>Density current depth from KAIA surface data (m AGL): 1215</td>
<td></td>
<td>1.48</td>
<td>1.72</td>
</tr>
</tbody>
</table>
north with the passage, the temperature dropped about 3.3 K, and the pressure rose 1.6 hPa. The cooling and wind shift suggest that the radar fine line passing KAIA marked a surface-based density current, not an intrusive current that propagated above the SBL. Based on the KAIA values, the density current depth $d_o$ is estimated at 1215 m according to Eq. (5).

Following the methodology of Koch et al. (2008b), the density current speed $C_{dc}$ was taken to be that determined from radar imagery. This was estimated from the speed of the radar fine line that initially developed from the MCS in southwestern South Dakota and propagated toward and across KAIA. This yields $C_{dc} = 21.4 \text{ m s}^{-1}$. There was a light head wind at KAIA at the time, thus the intrinsic speed is estimated to be slightly higher at 22.6 m s$^{-1}$. The intrinsic bore speed $C_b$ was higher than this farther east near area 1 (Table 2), allowing the density current to transform into a relatively faster bore.

c. Bore flow regime

To determine which flow regime was responsible for producing the bore (Fig. 16), we need an estimate of the normalized density current depth, $d_o/h_o$, and the Froude number $Fr$. The latter is given by

$$Fr = \frac{U}{\sqrt{g' h_o}} = \frac{C_{dc}}{C_{gw}}.$$  

Here $g' = g(\rho_c - \rho_o)/\rho_w$ is reduced gravity, $U$ is the speed of the obstacle moving through the quiescent fluid (here equal to $C_{dc}$ given in Table 2, as it is the density current that is moving), and $C_{gw}$ is the long-wave gravity speed. We can estimate $C_{gw} = \sqrt{g' h_o}$ from the SBL height ($h_o = 377$ m AGL) and the potential temperature difference, both inferred from the prebore sounding (Fig. 2). This gives $Fr = 1.72$. Allowance for a deeper SBL ($h_o \approx 500$ m) as suggested by CRL data (Fig. 5) yields $Fr = 1.48$. The average Froude number is then 1.60.

These Froude number estimates combined with normalized density current depth estimates define the colored dots in Fig. 16. Points falling in regime A represent unblocked flow (the preboundary air readily moves over the moving obstacle) and those in regime C represent completely blocked flow, in which density currents are intrusive (decoupled from the surface), or dissipative (Rottman and Simpson 1989). Regime B is partially blocked flow in which bores occur, at various strengths as indicated in Fig. 16. Bores may also result from dissipative density currents (regime C) (Rottman and Simpson 1989). Both the average and lidar method points are within the partially blocked regime, but the sounding method is in regime C. The predicted bore strength $S$ for the lidar method is 2.7, which is within the range of observed estimates mentioned in Table 2. The absissa coordinate is very sensitive to $h_o$, so the error bars could be relatively large. But all estimates within a broad range of uncertainty fall in either bore flow regime B or C during the aircraft measurement phase. This further supports the claim that this feature was a bore.

8. Conclusions

A bore was observed on 20 June 2015 during the PECAN field campaign by an instrumented aircraft, a profiling site, and several scanning radars. This bore was triggered by an MCS-generated cold pool protruding into the LLJ, located near the top of a nocturnal SBL. The aircraft with profiling lidars provided a unique opportunity to capture the evolution of the system along with high-resolution vertical aerosol and moisture structure. Initially, the bore appeared as a high-amplitude multiwave system that triggered surface-based convection in a potentially unstably atmosphere. Aircraft sampling of the wave train only occurred later on, when deep convection no longer was triggered. Initially high vertical velocities were recorded in the leading wave, up to $10 \text{ m s}^{-1}$. As the wave train propagated away from the convective cold pool, the leading wave weakened while the second wave
remained robust. Later measurements of the full wave train with three crests indicate that the second wave retained the highest amplitude. Moisture was lofted from the SBL over a depth of up to 2 km. The wave crests generated long roll clouds. As the wave train started to decay, the bore speed decreased, the horizontal wavelength increased, and the amplitude of mainly the leading wave decreased.

The UWKA, surface, and supplemental observations together all make a strong case that the phenomenon is an undular bore. The observational evidence is summarized below:

1) Airborne Raman lidar observations reveal a sustained elevation of the moist SBL top behind the waves. A comparison between pre- and postbore soundings and wind profiler data indicates a ~700 m deepening of the SBL, and a ~10 m s⁻¹ change in momentum in the lowest 1 km.

2) The vertical displacement depicted by the airborne lidars is wavelike and flight-level observations indicated impressive updrafts that lead cooling by a quadrature phase shift.

3) The airborne lidars reveal significant wave breaking and turbulence, initially on the lee side of the leading wave, and later in the evolution of the bore on the lee side of the second wave. The semiturbulent nature of the waves is consistent with a high bore strength, S, estimated at between 2 < S < 3.

4) Both the lidar- and radar-observed estimates of the speed of the leading wave (~15 m s⁻¹) are consistent with this theory.

5) Low-level forward momentum at about the same speed as the bore is present in the wave crests, and front-to-rear flow in strength matching the prebore LLJ in the wave troughs, indicating a high-strength bore rather than a density current. At flight level (2.3–3.0 km AGL), front-to-rear flow occurs in both wave crests and troughs with stronger bore-relative wind in crests.

6) Analysis of the environmental conditions before wave passage highlight a strong SBL, which acts as a duct for the waves, and curvature in the wind profile between ~3 and 4.5 km AGL. That curvature within the residual mixed layer causes a significant reduction in the Scorer parameter and at least partial reflection of wave energy. The phase relationships between flow, vertical velocity, and temperature, and the lack of phase tilt with height, confirm wave energy trapping.

7) Surface observations a few hours later in the bore’s life depict an amplitude-ordered wave train with forward wind anomalies in wave crests (defined as periods of cooling aloft, resulting in positive pressure perturbations at the surface), no significant temperature drop, suggesting a transition to a soliton. Even so, bore characteristics remain, in particular the sustained surface pressure increase and the sustained change in moist boundary layer depth and in momentum.

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**REFERENCES**


——, ——, and D. Herzmann, 2009: The spectacular undular bore TMGOTG, which is funded by NSF.


