1-2012

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The Impact of Airmass Boundaries on the Propagation of Deep Convection: A Modeling-Based Study in a High-CAPE, Low-Shear Environment

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(Manuscript received 10 November 2010, in final form 14 June 2011)

ABSTRACT

A suite of experiments conducted using a cloud-resolving model is examined to assess the role that preexisting airmass boundaries can play in regulating storm propagation. The 27 May 1997 central Texas tornadic event is used to guide these experiments. The environment of this event was characterized by multiple preexisting airmass boundaries, large CAPE, and weak vertical shear.

Only the experiments with preexisting airmass boundaries produce back-building storm propagation (storm motion in opposition to the mean wind). When both the cold front and dryline are present, storm maintenance occurs through the quasi-continuous maintenance of a set of long-lived updrafts and not through discrete updraft redevelopment. Since the cold front is not required for back building, it is clear that back building in this environment does not require quasi-continuous updraft maintenance. The back-building storm simulated with both the cold front and dryline is found to be anchored to the boundary zipper (the intersection of the cold front and dryline). However, *multiple* preexisting airmass boundaries are not required for back building since experiments with only a dryline also support back building. A conceptual model of back building and boundary zippering is developed that highlights the important role that preexisting boundaries can play in back-building propagation.

1. Introduction

Airmass boundaries are ubiquitous in the atmosphere. More often than not they simply represent benign transitions between air masses. However, airmass boundaries can also have significant impacts on deep convection (moist convection extending through a considerable fraction of the troposphere), affecting its initiation, strength (updraft/downdraft magnitudes), rotation, propagation, and longevity. Of the above, the initiation of deep convection has served as the most common focus of previous research on the impact of airmass boundaries on deep convection and has been reviewed by Weckwerth and Parsons (2006). Airmass boundaries have also been examined for their relationship to storm propagation

(e.g., Newton 1963; Weaver 1979; Wade and Foote 1982; Weaver and Nelson 1982; Wilhelmson and Chen 1982; Bluestein and Jain 1985; Weaver et al. 1994; Atkins et al. 1999; Houston and Wilhelmson 2007a, hereafter HW07a), storm rotation (e.g., Weaver et al. 1994; Wakimoto et al. 1998; Atkins et al. 1999; Rasmussen et al. 2000; Fierro et al. 2006; Houston and Wilhelmson 2007b), tornadogenesis (e.g., Purdom 1976; Maddox et al. 1980; Weaver and Nelson 1982; Simpson et al. 1986; Wilson and Schreiber 1986; Wakimoto and Wilson 1989; Wilczak et al. 1992; Purdom 1993; Weaver et al. 1994; Roberts and Wilson 1995; Lee and Wilhelmson 1997a,b; Markowski et al. 1998; Lee and Wilhelmson 2000; Rasmussen et al. 2000; Caruso and Davies 2005; Houston and Wilhelmson 2007b), storm updraft strength (e.g., Maddox et al. 1980; Weaver et al. 1994; Atkins et al. 1999; Gilmore and Wicker 2002; Fierro et al. 2006), and supercell morphology (e.g., Moller 1982; Doswell et al. 1990; Moller et al. 1990). Although research addressing the relationship between airmass boundaries and deep convection is

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clearly extensive, most of this work has relied on observational data/analysis. Identifying the fundamental mechanisms that regulate the impact of airmass boundaries on deep convection requires augmenting observational analysis with numerical modeling.

In the research documented in this article, experiments are conducted with a cloud-resolving model to complement the observational analysis of the 27 May 1997 central Texas tornadic event conducted by HW07a. In the 27 May 1997 event, at least 12 tornadoes, some violent (including the F5 Jarrell, Texas, tornado), formed from back-building supercells that developed and traveled along several preexisting airmass boundaries that were in place within a high-CAPE, low-shear environment that should have otherwise been unfavorable for supercells and tornadoes (HW07a).

HW07a found that the back-building propagation (storm motion in opposition to the mean flow; Bluestein and Jain 1985) of the 27 May 1997 storms analyzed [referred to as the Lake Belton, Jarrell, and Pedernales Valley storms following the nomenclature of Magsig et al. (1998)] was ultimately a consequence of the “zippering” together of the cold front and dryline during the event. The term zippering was used to characterize the process because the merging of the west-southwest–east-northeast-oriented cold front and southwest–northeast-oriented dryline resembled a zipper closing toward the southwest (Fig. 1). Similar terminology has been used to describe cold front–dryline mergers observed by Koch and Clark (1999) and Wakimoto et al. (2006). Other examples of zippering have been noted by Koch and McCarthy (1982), Parsons et al. (1991), Ross (1987), Hemler et al. (1991), Neiman and Wakimoto (1999), and Parsons et al. (2000). Back-building motion requires that propagation and not translation drives storm motion (Bluestein and Jain 1985). Thus, accounting for the processes associated with boundary zippering that lead to back-building requires accounting for the processes that lead to discrete convective updraft redevelopment or continuous updraft maintenance on the storm’s upstream flank.¹

The objective of the research presented here is to use the 27 May 1997 event to guide an exploration of the impact of preexisting airmass boundaries on the propagation of

FIG. 1. Illustration of zippering from HW07a. The cold front appears as the curve with filled triangles, the dryline is the scalloped curve, and the storm-generated gust front appears as the curve with open triangles.

deep convection in a high-CAPE, low-shear environment. We particularly focus on the role that boundary zippering can play in the upstream updraft redevelopment/maintenance that yields back building. The design of the suite of experiments used, including a brief description of the numerical model, description of the reference state soundings, and explanation of the method used to initialize the airmass boundaries appears in section 2. Results from the analysis of these experiments are presented in section 3 followed by a discussion of a proposed conceptual model of back building and boundary zippering presented in section 4. A summary of the principal conclusions is included in section 5.

2. Experiment design

a. Model description

The numerical model used to conduct the experiments for this work is the Illinois Collaborative Model for Multi-scale Atmospheric Simulations (ICOMMAS). ICOMMAS is a progeny of COMMAS (Wicker and Wilhelmson 1995) and was developed in parallel with L. J. Wicker’s development of the National Severe Storms Laboratory version of COMMAS (NCOMMAS; Coniglio et al. 2006). ICOMMAS has been used to examine cloud-scale processes by Houston and Niyogi (2007) and a complete description of ICOMMAS is documented by Houston (2004). Key elements of the model are included in the appendix and a summary appears below.

ICOMMAS is configured to represent cloud-scale and mesoscale processes. The principal dynamic equations

¹ In contrast to the definition of back building adopted by Bluestein and Jain (1985) in which upstream motion is a consequence of “the periodic appearance of a new cell upstream, relative to cell motion, from an old cell, and the resulting merger of the new cell with the old cell” (i.e., discrete updraft redevelopment), we do not presume a mechanism for upstream motion. Instead, our definition of back building simply requires that motion is in opposition to the mean flow. By this definition, even supercells with significantly deviate motions could be classified as back-building storms.

FIG. 2. (a) Thermodynamic and (b) wind profiles of the reference state air mass (i.e., air mass west of the dryline; thick gray curves) and the air mass to the east of the dryline (thick black curves). The thick dashed curve in (a) represents the temperature of a parcel using a 1000-m-deep mixed layer. The thick dashed (solid) curve in (b) represents the u (v) wind. (The wind profile is visualized with a speed–height diagram instead of a hodograph because the low shear that characterizes this environment renders the hodograph difficult to interpret.)

are nonhydrostatic and supercompressible. Microphysical processes are simulated using the Straka three-class ice, two-class liquid, single-moment parameterization described by Gilmore et al. (2004). [The slope-intercept values and densities for each of the hydrometeor species match those used by Gilmore et al. (2004) and are listed in Table A1.] Subgrid-scale turbulence is represented using a version of the 1.5-order closure parameterization of Klemp and Wilhelmson (1978). ICOMMAS excludes surface fluxes of heat and moisture, topography, and surface drag. Finite differencing of time derivatives utilizes the split-explicit implementation of the third-order Runge–Kutta (RK3) presented by Wicker and Skamarock (2002). Finite differencing of the horizontal and vertical derivatives in the advection equation relies on the fifth- and third-order spatial discretizations, respectively (e.g., Wicker and Skamarock 2002). The Thuburn (1995) multidimensional flux limiter is included to ensure monotonicity (shape preservation) for advection.

The numerical domain is 100 km $\times$ 100 km $\times$ 20 km in size with a horizontal gridpoint spacing of 500 m and a vertical gridpoint spacing of 50 m in the lowest 1 km, geometrically stretched to 450 m at the top of the domain. Lateral domain boundaries are open and vertical domain boundaries are rigid and free slip.

b. Model initialization

The horizontally homogeneous reference state used for all numerical experiments described herein represents

the “warm,” “dry” air mass to the west of the dryline. This reference state is imposed by altering the temperature, moisture, and wind in the lowest portion of the “modified Calvert sounding” from HW07a, which represents conditions to the east of the dryline based on a sounding released near Calvert, Texas, as part of the Texas A&M Convection and Lightning experiment (TEXACAL; Biggerstaff et al. 1997). The surface temperature of the modified Calvert sounding is raised from 32.4° to 33.1°C and surface dewpoint temperature is lowered from 23.9° to 22.0°C. Well-mixed conditions are assumed up to 1151 m AGL, the LCL of the modified Calvert sounding. The thermodynamic profile of the reference state sounding is illustrated in Fig. 2a. (Note that the “east of the dryline” sounding illustrated in Fig. 2 is similar to, but slightly different than, the modified Calvert sounding. A description of the generation of the sounding east of the dryline is described below.) The low-level winds of the modified Calvert sounding are also modified to reflect the air mass west of the dryline. The south-southeast winds (from 170°) of the modified Calvert sounding are changed to west-northwest (from 290°). The reference state winds are also rotated 30° counterclockwise so that the simulated dryline is oriented north–south instead of north-northeast–south-southwest, as shown in Fig. 3. The *rotated* reference state winds are illustrated in Fig. 2b.

Both the dryline and cold front observed during the 27 May 1997 event were found to be essential to storm

FIG. 3. Initial simulated boundaries and observed boundaries. (left) Shading represents the near-surface equivalent potential temperature (lighter shading indicates higher values) and arrows represent the near-surface winds (length indicates speed). (right) The boundaries observed during the 27 May 1997 event (standard surface observation notation is utilized). The black square provides a reference for the approximate size and orientation of the model domain.

propagation, organization, and rotation (HW07a; Houston and Wilhelmson 2007b). The observed boundaries along with the simulated boundaries are illustrated in Fig. 3. The characteristics of the prescribed dryline are set so that the sounding east of the dryline closely resembles the modified Calvert sounding. The dryline is initialized through a slab-symmetric, hydrostatically balanced, negative temperature perturbation and positive water vapor perturbation in an 1100-m-deep block across the eastern half of the domain (recall that the reference state actually represents the air to the west of the dryline). Winds east of the dryline are prescribed using the prestorm surface observations and a linear interpolation to the reference state sounding values at 1100 m. See Fig. 2 for an illustration of the sounding east of the dryline.

Despite numerous precautions taken to balance the initial conditions in the presence of the dryline, the prescribed initial state is not steady. Therefore, for all numerical experiments that include the dryline, the dryline is allowed to adjust to the environment (and vice versa) for 3600 s of integration before any additional boundaries or thermals are included. Moreover, to mitigate the east-west dynamic pressure gradient associated with the low-level convergence along the dryline, a small east-west temperature gradient is initialized across the air mass east of the dryline. The inclusion of this temperature gradient minimally impacts the east-west distribution of CAPE and convective inhibition (CIN) east of the dryline.

It is important to note that this prescription of the dryline does not allow for the initiation of deep convection without

an additional forcing mechanism. As described by HW07a, the observed dryline on 27 May 1997 was responsible for convection initiation only near its intersection with a preexisting nonclassic mesoscale circulation. Since the nonclassic mesoscale circulation is not simulated in these experiments, the dryline is prescribed such that it does not independently initiate deep convection.

The cold front is imposed west of the dryline (after the 3600 s required for the dryline initialization) as an arc extending from the northern domain boundary to the southwest corner (Fig. 3). The temperature within the postfrontal air mass northwest of the cold front is prescribed through a -1 -K temperature perturbation imposed at all points within a 1000-m-deep block. The wind field to the northwest of the front is prescribed through a linear interpolation between the observed winds and the reference state sounding at 1000 m. To allow for adjustment of the front/environment, the advection and microphysics are turned off for 300 s while the momentum and pressure in the vicinity of the cold front respond to the perturbed fields. This procedure is similar to the one utilized by Lee and Wilhelmson (1997b).

One or more sustained thermals are used to initiate isolated convection in all simulations. This approach differs from the more common thermal bubble initialization frequently used in idealized cloud-scale simulations of deep convection. In the approach adopted here, a modest maximum thermal perturbation of 1 K is imposed in an ellipsoid region centered at a height of 500 m with a lateral diameter of 10 km and a vertical diameter of 1 km.

The temperature perturbation in the lower half of the ellipsoid is then held fixed (refreshed) for 1200 s. This approach was found to produce a more gradual and more realistic convection initialization than the conventional thermal bubble initialization.

c. Experiment suite

Five experiments are used in the analysis reported in this article. Each experiment (summarized in Table 1) varies according to the combination of thermals and boundaries used. In the first two experiments, the environment to the east of the dryline (black profiles in Fig. 2) is used as the initial environment. In the first experiment, no airmass boundary is imposed and a single sustained thermal is used to initiate deep convection. Hereafter, this experiment will be referred to as *SingleThrm*. In the second experiment, no airmass boundary is imposed but a line of 6 sustained thermals, each spaced 13.3 km apart, is used to initiate deep convection. This line of thermals has the same general orientation as the cold front and extends from 16 km north of the southern domain boundary to 24 km south of the northern domain boundary. Hereafter, this experiment will be referred to as *MultipleThrm*.

In the third (*BndDry*) and fourth (*BndDryMod*) experiments, the dryline and a single sustained thermal are

TABLE 1. Summary of experiments.

Expt	Thermal forcing	Cold front	Dryline
SingleThrm	One perturbation	No	No
MultipleThrm	Six perturbations	No	No
BndDry	One perturbation	No	Yes
BndDryMod	One perturbation	No	Yes (modified winds west of dryline)
BndColdDry	One perturbation	Yes	Yes

included in the initialization. In the *BndDryMod* experiment, the low-level wind field in the dry air mass is modified to be identical to the flow northwest of the cold front (the temperature and moisture characteristics of the dry air mass are the same as the characteristics west of the dryline in the *BndDry* experiment).

In the fifth experiment, both the dryline and cold front are included (refer to Fig. 3 for the spatial relationship between these boundaries). This experiment (hereafter referred to as *BndColdDry*) is designed to closely resemble the environment associated with the central Texas storm complex of 27 May 1997. Because the observed event began as an isolated storm along the dryline near its intersection with the cold front, a single sustained thermal is used to locally initiate convection. This thermal is started

FIG. 4. Comparison of the *BndDryCold* simulation to the observations. (a) Results from the *BndColdDry* simulation at 8040 s after convection initiation: near-surface equivalent potential temperature is shaded in grayscale at an interval of 4 K and simulated radar reflectivity is shaded in color every 5 dBZ. (b) The radar reflectivity from the 0.5° sweep of the central Texas (GRK) radar at 1908 UTC along with the boundaries. To match the rotation of the reference state used for all simulations, (b) has been rotated. Both (a) and (b) have the same spatial scale.

FIG. 5. Summary of the experiments conducted: (a) SingleThrm, (b) MultipleThrm, (c) BndDry, (d) BndDryMod, and (e) BndColdDry. Near-surface equivalent potential temperature is shaded in grayscale at an interval of 4 K (higher values have lighter shading) and simulated radar reflectivity is shaded in color every 5 dBZ. Simulated reflectivity is computed following Smith et al. (1975). The black arrows denote the direction and relative speed of simulated storm motion and the white arrows denote the direction and relative speed of the density-weighted mean wind. The gray arrows denote the direction of the near-surface winds. The distance between “large” tick marks is 10 km.

shortly before the collision of the dryline and front at a location along the dryline.

3. Results

a. Comparison of the BndColdDry simulation to observations

Since the BndColdDry experiment is designed to most closely resemble the 27 May 1997 event, a comparison of the simulated and observed storm and boundaries is warranted. As illustrated in Fig. 4, the storm and boundaries simulated in the BndColdDry experiment broadly resemble the structure of the observed storm and boundaries

(the observational data have been rotated 30° counterclockwise in Fig. 4 to match the orientation of the model initial conditions and scaled to match the spatial scale of the model data). Note that the relative orientations of the boundaries are very similar as are the relationships between the near-surface precipitation fields and boundaries. Furthermore, temperatures within the simulated near-surface, storm-generated cold pool are also similar to the observed cold pool temperatures. The time-averaged minimum temperature simulated within the cold pool between 7200 and 9720 s after convection initiation (9720 s represents the termination of integration) was 21.0°C and the minimum near-surface temperature during the entire simulation was 20.7°C.

FIG. 5. (Continued)

TABLE 2. Metrics for quantifying storm motion. The deviation is defined as $\Delta \mathbf{V} \equiv \bar{\mathbf{V}}_s - \bar{\mathbf{V}}_0$. The mean wind ($\bar{\mathbf{V}}_0$; the density-weighted wind calculated over the full depth of the sounding) used for these calculations is $(5.6, 5.8) \text{ m s}^{-1}$. The storm motion ($\bar{\mathbf{V}}_s$) (mean wind) vector is illustrated with black (white) arrows in Fig. 5.

Expt	Storm motion ($[\bar{u}_s, \bar{v}_s] \bar{\mathbf{V}}_s $)	Deviation ($[\Delta u, \Delta v] \Delta \mathbf{V} $)	Percent deviation ($100\% \times \Delta \mathbf{V} \cdot \bar{\mathbf{V}}_0 / \bar{\mathbf{V}}_0 ^2$)
SingleThrm	[3.2, 1.3] 3.4	[−2.4, −4.5] 5.1	48.9%
MultipleThrm	[5.0, 0.3] 5.0	[−0.6, −5.5] 5.5	54.1%
BndDry	[0.9, −4.1] 4.2	[−4.7, −9.9] 11.0	104.3%
BndDryMod	[0.9, −8.0] 8.1	[−4.7, −13.8] 14.6	136.4%
BndColdDry	[1.0, −9.5] 9.6	[4.6, −15.3] 16.0	147.6%

Similar values were observed within the storm-generated cold pool of the 27 May 1997 event: at 1914 UTC ($\sim 7200 \text{ s}$ after storm initiation) Temple, Texas (TPL), reported a temperature of 24°C ; at 2153 UTC ($\sim 16\,200 \text{ s}$ after storm initiation) Georgetown municipal airport (T04) reported a temperature of 19°C ; and at 2253 UTC ($\sim 19\,800 \text{ s}$ after storm initiation) Austin, Texas, reported a temperature of 17°C . A notable difference between the simulated and observed fields is the horizontal distribution of precipitation. The observed storm assumes the classic V-notch configuration in the precipitation exhausted downstream by the upper-level flow and a relaxed horizontal gradient in the reflectivity in the downstream precipitation suggesting either precipitation sorting or lateral differences in hydrometeor evaporation/sublimation. In contrast, the simulated storm has a more linear configuration to the downstream precipitation and no V-notch. These differences are most likely related to the rather simplistic treatment of microphysics, but a precise explanation is beyond the scope of this work.

b. Storm propagation

A qualitative summary of storm motion is reflected in the plan view sequences of Fig. 5 with a quantitative assessment of the difference between the mean wind and simulated storm motion illustrated in Table 2. Both reveal that only experiments with preexisting airmass boundaries produce back-building storm propagation. The differences between the mean wind and simulated storm motions (see graphical illustration of vectors in Fig. 5) also illustrate that all experiments produce a storm motion that is to the right of the mean wind. This is consistent with the motion of an archetypical multicell (Chisholm and Renick 1972). However, only the BndColdDry, BndDry, and BndDryMod experiments support a storm motion vector with a component that is markedly in opposition to the mean wind. This relationship between the mean wind and storm motion is quantified via two metrics in Table 2. The first metric, “deviation”, is simply the vector difference between the mean storm motion and the mean wind (this is the density-weighted wind calculated over the full depth of the sounding). The

magnitude of the deviation is nearly the same for SingleThrm and MultipleThrm (without a preexisting boundary) and as much as 3 times larger for the experiments with boundaries. The second metric, “percent deviation,” is computed by taking the projection of the deviation onto the mean wind and normalizing it with the magnitude of the mean wind. The percent deviation will be 100% if the mean wind and storm motion are orthogonal (i.e., there is no component of motion in the direction of the mean wind). Any value greater than 100% indicates storm motion that has a component in the opposite direction of the mean wind. Only the experiments with preexisting boundaries have percent deviation values $>100\%$, thus, the percent deviation metric clearly discriminates between the no-boundary experiments, which produce modest storm motion deviations from the boundary experiments, which produce large deviations. These results also indicate that the zippering of preexisting boundaries is not required for back-building propagation. Both the BndDry and BndDryMod simulations are initialized with only a single boundary yet both produce back-building propagation.

Among the five core experiments, only the BndColdDry experiment with multiple boundaries is found to support storms that back build through quasi-continuous updraft maintenance and not discrete updraft redevelopment. The simulated storm consists of three primary convective updrafts (Fig. 6). The first updraft (“northern updraft 1”) develops around 3200 s and lasts for approximately 1800 s . The second updraft (“southern updraft”) develops around 3360 s $5\text{--}10 \text{ km}$ south of northern updraft 1 and persists through the remainder of the simulation (for 6240 s). The third updraft (“northern updraft 2”) develops around 4800 s on the distorted gust front between the decaying northern updraft 1 and the ongoing southern updraft and persists through the remainder of the integration (for 4800 s). Since only the BndColdDry experiment supports quasi-continuous updraft maintenance, quasi-continuous updraft maintenance is clearly not required for back building; a conclusion that is consistent with the results from Bluestein and Jain (1985) and Parker (2007).

FIG. 6. (left) Isosurface rendering and (right) corresponding plan views for the BndColdDry experiment at (a) 3720 and (b) 8520 s. (left) The isosurface of total cloud + snow (cloud ice, cloud water, and snow) mixing ratio at a value of $5 \times 10^{-8} \text{ kg kg}^{-1}$ along with the surface θ_e shaded at an interval of 4 K (cooler colors correspond to lower values). The “camera” is looking to the northwest. (right) The matching near-surface θ_e field along with the vertical velocity at a height of 5 km (contoured in black every 5 m s^{-1}).

To determine some of the ways in which preexisting boundaries can yield back-building propagation, the mechanisms responsible for back building in the BndColdDry and BndDryMod experiments will be examined. (Unlike the BndDry experiment, both the BndColdDry and BndDryMod experiments produce prominent gust front distortions yielding a prominent “S” shape to the gust front on the south side of the cold pool.) For the BndColdDry experiment, a 3D block of 777 tracers occupying a $1500 \text{ m} \times 9000 \text{ m} \times 200 \text{ m}$ (x, y, z) volume is released into the inflow of the southern updraft at 6000 s. This location was chosen for tracer release because it fell along the trajectory of tracers backward integrated from the mid-levels of the southern updraft. An elongated (north–south) box of tracers is used here and for the BndDryMod

simulation (discussed below) so that the continuity of the airstream flowing toward and successively passing through updrafts can be examined. Tracers are forwards integrated with a fourth-order Runge–Kutta scheme using model output every 120 s and a tracer time step of 15 s. Within this block, a downstream tracer (“D” in Fig. 7) and upstream tracer (“U” in Fig. 7) are identified and followed.² As illustrated in Fig. 7, the upstream tracer reaches the altitude of its downstream counterpart at positions farther south. For example, at 7200 s (all times

² Here and elsewhere in this article, “upstream” and “downstream” are based on the low-level inflow, which is generally south–north.

FIG. 7. Tracer positions at 6720, 7200, 7680, and 8160 s after convection initiation released into the inflow of the BndColdDry simulation. Near-surface values of θ_e are shaded at an interval of 4 K and vertical velocity at 5 km is contoured in black. The initial position of the block of tracers is indicated with a gray rectangle.

will be listed as the time since convection initiation) the upstream tracer reaches an altitude of 1 km approximately 1 km south of the position where the downstream tracer reaches an altitude of 1 km (480 s earlier). This pattern of parcel ascension is a consequence of the zippering of the cold front and dryline that creates the frontal segment (the portion of the cold front in contact with the air mass east of the dryline). Inflow parcels traveling northward along the dryline begin to ascend within the deep forced ascent along the frontal segment once they encounter the intersection of the cold front and dryline (the zipper). As the zipper moves southward, the location where parcels first begin to ascend moves southward. Tracer analysis further reveals that both the downstream and upstream tracers spend approximately the

same amount of time (~ 960 s) along the frontal segment before ascending to a height of 2 km. Thus, the time required for parcels to reach the LFC after passing over the zipper is nearly constant. Therefore, parcels reach their LFC at positions progressively farther south.

The BndDryMod simulation is initialized with only a single boundary so boundary zippering is not possible. Nevertheless, simulated convection exhibits back-building propagation. In contrast to the BndColdDry experiment, back building in the BndDryMod experiment is controlled by discrete convective updraft redevelopment and not quasi-continuous updraft maintenance. Tracers released into the inflow of the BndDryMod experiment at 11 760 s travel along the dryline where they ascended only gradually, then travel along the distorted gust front where parcel

FIG. 8. As in Fig. 7, but for the BndDryMod simulation at 5760 and 6960 s.

ascension is more significant, and finally pass over the undistorted gust front subsequently becoming positively buoyant (Fig. 8). Supplemental tracers integrated backward from the midlevels of convective updrafts that formed to the south of existing convection reveal that the vast majority of parcels composing these convective updrafts immediately following their initiation pass through the forced ascent along either the dryline or the distorted gust front. Figure 9 illustrates an example of these supplemental tracers. The circular and square tracer positions denote tracers that pass through and rise within the forced ascent of the distorted gust front and dryline, respectively. These tracers make up 72% of the tracers for this updraft. Furthermore, the positions at 5640 s reveal that these tracers are the first to ascend above the LFC (~ 1100 m) and are therefore ultimately responsible for convection initiation. These results implicate the forced ascent along the distorted gust front and dryline in the initiation of updrafts south of existing convection. Since the distorted gust front travels southward, so too does the location of updraft formation. Thus, the distorted gust front and dryline are responsible for the back-building propagation simulated in the BndDryMod experiment.

Although the gust front is clearly not a preexisting boundary, a preexisting boundary is necessary for the simulated gust front distortion. The modified dryline used in the BndDryMod experiment is characterized by winds on the west side of the dryline with a northerly component and winds on the east side with a southerly component. This horizontal shear across the dryline yields differential advection of the gust front (the motion of the southward moving gust front is slower on the east

side of the dryline than on the west side) and thus a distortion. Because the distorted gust front depends on the preexisting boundary of this simulation, it is the preexisting boundary that is ultimately responsible for the simulated back-building propagation.

A detailed comparison of the BndDry and BndDryMod simulations is not presented for the sake of brevity. However, it is important to note that, even though the BndDry simulation lacks a prominent gust front distortion, the back building in both the BndModDry and BndDry experiments is ultimately caused by the dryline. Therefore, the BndDry experiment affirms the principal conclusion regarding back building: back building in these experiments requires a preexisting boundary.

4. Discussion

The mechanisms responsible for back building in these simulations are notably different from the mechanism identified by Parker (2007). His experiments did not include preexisting airmass boundaries so clearly the results presented here are not applicable to all cases of back building. He found that new convective updrafts developed on upstream- (storm relative) advancing gust front surges within the larger storm-generated cold pool. Since each new upstream convective cell produced a new gust front surge that spread farther upstream, new updrafts formed at progressively more upstream locations. In contrast, new updrafts simulated in the BndColdDry and BndDryMod experiments are not tied to gust front surges, but are instead tied to preexisting airmass boundaries and gust fronts distorted by these preexisting boundaries.

FIG. 9. Positions of tracers backward integrated from an updraft in the BndDryMod simulation at (a) 5160 and (b) 5640 s. The circular and square tracer positions denote tracers that passed through and rose within the forced ascent of the distorted gust front and dryline, respectively. In the vertical cross sections, vertical velocity is contoured at an interval of 0.2 m s^{-1} , θ_e is shaded at an interval of 4 K, and $5 \times 10^{-8} \text{ kg kg}^{-1}$ mixing ratio isopleths of the total cloud (the combined mixing ratios of cloud water and cloud ice) are contoured in white. In the plan view images, near-surface values of θ_e are shaded at an interval of 4 K. The initial position of the tracers is indicated with a gray polygon.

With the preceding analysis of storm propagation, a conceptual model of boundary zippering and back building can be constructed (Fig. 10). This model is most applicable to the southern updraft of the BndColdDry storm, but a similar model could be constructed for the northern updrafts as well. In this conceptual model, parcels (numbered cubes in Fig. 10) originate in the low levels, travel along the dryline in the lowest 1 km of the atmosphere, reach the intersection of the cold front and dryline (the zipper), and rise steadily as they travel along the frontal segment. After an amount of time $\Delta\tau$, these parcels reach the midtroposphere within the updraft. Since each successive parcel traveling along the dryline reaches the zipper at a point farther south than the previous parcel and the amount of time required to ascend to the midtroposphere is roughly the same, the upstream parcels reach their LFC at locations progressively farther south and the storm back builds to the south quasi-continuously. As noted above, neither boundary zippering nor quasi-continuous

maintenance (which controls the propagation of the BndColdDry storm that the conceptual model is patterned after) is required for back building. Therefore, this conceptual model does not represent the only form of back building. Nevertheless, it does highlight the important role that preexisting boundaries can play in back-building propagation.

5. Summary

A suite of numerical experiments has been conducted with a cloud-resolving model in an effort to examine the impact of preexisting airmass boundaries on storm propagation. The 27 May 1997 central Texas tornadic event was used as the initial condition for the simulations performed. In this event, a cold front and dryline were found to play important roles in storm propagation. When both of these boundaries were included in a simulation, the overall structure and relative orientations of

FIG. 10. Conceptual model of back-building propagation and boundary zipping. The gray arrow in the northeast corner at each time represents the direction of the mean wind.

the storm and airmass boundaries were very similar to the storm–boundary structure and orientations observed in the 27 May 1997 event. The following conclusions emerged from the analysis of the experiments conducted.

Only the experiments with preexisting airmass boundaries produced back-building storm propagation. When both the cold front and dryline were present, storm maintenance occurred through the quasi-continuous maintenance of a set of long-lived updrafts and not through discrete updraft redevelopment. Since the cold front was not required for back building, it is clear that back building does not require quasi-continuous updraft maintenance; a conclusion that is consistent with prior research.

The back-building storm simulated with both the cold front and dryline was found anchored to the boundary zipper (the intersection of the cold front and dryline).

Air parcels entering the storm of this experiment traveled northward along the dryline and began to ascend within the deep forced ascent along the frontal segment (the portion of the cold front in contact with the air mass east of the dryline) once these parcels encountered the zipper. As the zipper moved southward, the location where parcels first began to ascend and (more importantly) the location where these parcels reached the LFC moved southward, leading to back building. In experiments with only one preexisting boundary (the dryline), and therefore no zipper, back building still occurred. With a “modified” dryline (a dryline that supported gust front distortion) back building was found to rely on the forced ascent in place along both the dryline and distorted gust front. Since the distorted gust front traveled southward, so too did the location of updraft formation. Furthermore, because the distorted gust front depended on the preexisting boundary of this simulation, it was the preexisting boundary that was ultimately responsible for the simulated back-building propagation.

Acknowledgments. The authors wish to thank G. Bryan and two anonymous reviewers for their comments, which greatly improved this article. We are particularly grateful to L. Wicker for many beneficial discussions regarding this work. We would also like to thank M. Gilmore, B. Jewett, M. Ramamurthy, and R. Rauber for their comments and suggestions during the early stages of this work. Funding for this work was provided through NSF Grants ATM-9986672 and ATM-0449753. Simulations were performed on the IBM POWER4 p690 at the National Center for Supercomputing Applications.

APPENDIX

Numerical Model Description

a. Dynamics

The principal dynamic equations used in ICOMMAS are identical to the set of equations used in COMMAS (Wicker and Wilhelmson 1995) and very similar to the equations used by Klemp and Wilhelmson (1978, hereafter KW78). The equation set is composed of the nonhydrostatic, supercompressible, Reynolds-averaged Navier–Stokes equations. Predicted quantities are the three components of momentum (u , v , and w), perturbation nondimensional pressure (perturbation Exner function, π), potential temperature θ , mixing coefficient K_m , and the mixing ratios of six water categories: water vapor q_v , cloud water q_c , rainwater q_r , cloud ice q_i , snow q_s , and hail/graupel q_h .

TABLE A1. Summary of tunable model parameters.

Parameter	Description	Value
c_m	Scaling coefficient for mixing length closure	0.2*
c_e	Dissipation coefficient in prognostic TKE equation	0.2*
Pr	Prandtl number (used in turbulence parameterization)	$1/3^{**}$
$(K_m)_{\max}$	Maximum allowable value of K_m (ensures numerical stability)	$1000 \text{ m}^2 \text{ s}^{-1}$
$\gamma_{\max}$	Maximum smoothing coefficient used to account for the anisotropic grid	0.02
c_s	Speed of sound	200 m s^{-1}
ρ_r	Density of a rain hydrometeor	1000 kg m^{-3}
ρ_s	Density of a snow hydrometeor	100 kg m^{-3}
ρ_h	Density of a hail/graupel hydrometeor	900 kg m^{-3}
n_{0r}	Slope intercept for rain	8.0×10^{-6}
n_{0s}	Slope intercept for snow	3.0×10^{-6}
n_{0h}	Slope intercept for hail/graupel	4.0×10^{-4}
α_s	Nondimensional divergence damper	0.01
K_{0R}	Maximum Rayleigh sponge coefficient	0.01 s^{-1}
N_R	Percentage of the domain covered by the Rayleigh sponge	30%

* The values of c_m and c_e used for this work are the same as those used by Klemp and Wilhelmson (1978).

** Follows Deardorff (1972).

b. Prognostic turbulent kinetic energy parameterization

The grid spacing used for this work is incapable of resolving turbulence; thus, a version of the 1.5-order closure turbulence parameterization of KW78 is used to approximate the subgrid-scale mixing due to turbulence. This method predicts the eddy mixing coefficient K_m based on a prognostic equation for turbulent kinetic energy (TKE) and the flux-gradient theory and applies it to the state variables using mixing length theory. The equation for the parcel rate of change of K_m is as follows:

$$\frac{DK_m}{Dt} = \frac{L^2}{2}(S - \bar{S}) - \frac{L^2}{2\text{Pr}}(B - \bar{B}) + \frac{1}{2}\nabla^2 K_m^2 - \frac{c_m c_e}{2L^2} K_m^2. \quad (\text{A1})$$

The four terms on the right-hand side are (from left to right), shear, buoyancy, second-order diffusion, and Rayleigh-type dissipation, where $L = c_m l$; l is the mixing length; c_m is a scaling coefficient; $S(\bar{S})$ and $B(\bar{B})$ are the full (reference state) shear and buoyancy forcings, respectively; Pr is the Prandtl number; and c_e is a dissipation coefficient. It can be shown that if buoyancy, shear, and dissipation are the dominant terms in (A1) then $K_m > 0$ if the Richardson number is less than the Prandtl number. The removal of the reference state shear and buoyancy is done to ensure that the reference state is not modified by turbulence. The values of K_m predicted using this method are capped at a value of $(K_m)_{\max}$ to ensure numerical stability. Values of c_m , c_e , Pr, and $(K_m)_{\max}$ used for this work are listed in Table A1.

Unlike the method of KW78, in this implementation the virtual potential temperature is used instead of potential temperature in the buoyancy term of the TKE tendency equation and the mixing length is given by $l = \Delta z$ not $l = (\Delta x \Delta y \Delta z)^{1/3}$. This latter difference is a consequence of the significantly anisotropic grid in the lowest several kilometers of the domain (50 m vertical, 500 m horizontal). Using a mixing length defined based on the grid point spacing in all three dimensions would yield a mixing length that is biased toward the larger horizontal grid spacing and thus a K_m field that overmixes the vertical distribution of the mixed variable where the vertical grid spacing is substantially less than the horizontal grid spacing. Unfortunately, without taking additional measures, defining the mixing length based solely on the vertical grid spacing would tend to *undermix* the horizontal distribution of the mixed variable where the vertical grid spacing is less than the horizontal grid spacing. Thus, for this work we have chosen to combine a mixing length defined based on the vertical grid spacing with a fourth-order horizontal smoother that compensates for the decreased horizontal mixing. For this smoother, the smoothing coefficient γ is weighted based on the degree to which a grid box is isotropic so that the horizontal smoothing is largest where the undermixing is most significant. Based on this definition, $\gamma = \gamma_{\max}$ at the bottom of the domain and $\gamma = 0$ where the horizontal grid spacing is less than the vertical grid spacing. The value of $\gamma_{\max}$ is listed in Table A1.

The application of the predicted mixing coefficient to the mixing of the three momentum components, potential temperature, and all water species once again follows the methodology of KW78. In this approach the mixing coefficient used for scalar mixing is K_m/Pr

whereas the full eddy mixing coefficient is used for momentum.

c. Numerics

1) TEMPORAL DISCRETIZATION

In this model the temporal discretization is computed using the split-explicit implementation of third-order Runge–Kutta (RK3) developed by Wicker and Skamarock (2002) instead of the forward-time (Wicker and Wilhelmson 1995) or second-order Runge–Kutta (RK2; Wicker and Skamarock 1998) schemes used in COMMAS. In the approach implemented here, the low-frequency (low mach number) modes are advanced using a larger time step interval than that used for the high-frequency (acoustic) modes. High-frequency modes are advanced using first-order forward time. Vertically propagating acoustic modes are solved implicitly.

2) SPATIAL DISCRETIZATION

Fifth- and third-order spatial discretizations are used for the horizontal and vertical derivatives, respectively, in advection. This is in contrast to the second-order approximation used in the original version of COMMAS. These discretizations are implemented following Wicker and Skamarock (2002) through the application of a fifth- (third) order Taylor series expansion to the interfacial values involved in the finite-volume formulation of advection.

3) FLUX LIMITER

Finite-volume approximations that are at least second-order accurate suffer from the tendency to produce non-monotonic (nonshape preserving) behavior (i.e., new extrema in the advected quantity can be created and/or existing extrema can be amplified). A flux limiter (e.g., van Leer 1974; Leonard 1991; Thuburn 1995) is one method for preventing these errors. The underpinning philosophy for the flux limiter is that true monotone schemes are at best first-order accurate (Godunov 1959). Since first-order accurate spatial approximations exhibit significant dissipation, it is impractical to apply a first-order accurate scheme to all portions of the advected field. Thus, the flux limiter is designed to be *locally* first-order accurate in regions of large curvature (especially at shocks), where the field is most susceptible to the generation of spurious oscillations, while remaining higher order in smooth regions of the field.

The Thuburn (1995) multidimensional flux limiter has been implemented in ICOMMAS to correct non-monotonic oscillations *in the scalar quantities* associated with the high-order approximations used for advection. It achieves shape preservation by bounding the “first guess” fluxes out of a given grid box by the fluxes into the grid box, while allowing for the generation and amplification

of physically realistic extrema through divergent flow (Thuburn 1995).

4) DIVERGENCE DAMPER

A divergence damper is included in ICOMMAS to selectively filter acoustic modes without significantly damping more dynamically significant modes such as gravity waves (Skamarock and Klemp 1992). A divergence damper also serves to further stabilize split-explicit scheme (Skamarock and Klemp 1992). It is implemented through the application of the term $\mathbf{A}_\delta \cdot \nabla \delta$ to the 3D equation of motion, where $\mathbf{A}_\delta = \alpha_\delta / \Delta t (\Delta x^2 \hat{\mathbf{i}} + \Delta y^2 \hat{\mathbf{j}} + \Delta z^2 \hat{\mathbf{k}})$ is the dimensional ($\text{m}^2 \text{s}^{-1}$) divergence damper coefficient vector, δ is the divergence, and α_δ is the (constant) non-dimensional divergence damper coefficient. In contrast to COMMAS, the tunable coefficient in this formulation (α_δ) is nondimensional. The value of α_δ used in this work is listed in Table A1.

5) BOUNDARY CONDITIONS

Lateral domain boundaries are treated as open following a formulation similar to that of KW78. The KW78 implementation allows the exhaustion of low Mach number flows (including gravity waves with a phase speed of approximately 30 m s^{-1}) out of the domain and imposes zero advection across the boundary for inflow into the domain. Unfortunately, the absence of boundary-tangential advection of boundary-normal momentum (e.g., the north–south advection of the u component of the flow along the eastern and western boundaries) makes the original KW78 implementation incompatible with the slab-symmetric density current initialization used for this work (described in section 2c): when the original KW78 formulation was used, asymmetries in the slab-symmetric direction were produced at the domain boundaries. Therefore, in ICOMMAS we have implemented lateral boundary conditions in which the boundary-tangential advection of the boundary-normal momentum is included.

Both the upper and lower domain boundaries are set to be rigid and free slip. A Rayleigh “sponge” is imposed on the upper boundary to prevent the reflection of vertically propagating gravity waves. The sponge occupies the upper $N_R\%$ of the domain and is characterized by a damping coefficient that ranges from zero at the bottom of the sponge to a value of K_{0R} at the top. The damping is applied to the perturbations of θ , q , u , v , and w . The values of K_{0R} and N_R used for this work are listed in Table A1.

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