A Possible Mechanism of Tornadogenesis Associated with the Interaction between a Supercell and an Outflow Boundary without Horizontal Shear

TAKUMI HONDA*

Department of Earth and Planetary Sciences, Graduate School of Sciences, Kyushu University, Fukuoka, Japan

TETSUYA KAWANO

Department of Earth and Planetary Sciences, Faculty of Sciences, Kyushu University, Fukuoka, Japan

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ABSTRACT

Using the insert–restart method developed in this study, tornadogenesis processes associated with the interaction between a supercell and an outflow boundary were investigated. A highly idealized outflow boundary promoted surface vortex intensification in a preconditioned supercell having a strong surface vortex. In particular, a tornado-like vortex was observed in an experiment with a moderate outflow boundary. This intensification was associated with the enhancement of the near-surface horizontal convergence by the outflow boundary. The optimal coldness of the outflow boundary was explained by the balance between the enhancement of the near-surface horizontal convergence and the buoyancy reduction by the outflow boundary. Sensitivity experiments demonstrated that, despite a less favorable environment, a storm with a boundary interaction might reach a similar maximum near-surface vorticity as a storm without a boundary in a more favorable environment.

1. Introduction

Most strong tornadoes are associated with supercells (Trapp et al. 2005). A large number of numerical (Klemp and Wilhelmson 1978a,b; Rotunno and Klemp 1985; Wicker and Wilhelmson 1995; Mashiko et al. 2009; Noda and Niino 2010; Dahl et al. 2012, 2014) and observational (Rasmussen et al. 1994; Straka et al. 2007; Markowski et al. 2008, 2012a,b; Wurman et al. 2012; Koshiba et al. 2013) studies have focused on the dynamics of supercells and related tornadogenesis. For example, Wicker and Wilhelmson (1995) conducted an idealized numerical simulation and showed that a rear-flank downdraft (RFD) plays a role in tornadogenesis. The importance of such downdraft and related baroclinic vorticity generation has been indicated by theoretical (Davies-Jones and Brooks 1993), observational (Straka et al. 2007; Markowski et al. 2008, 2012a,b), and recent numerical (Dahl et al. 2014) investigations. The baroclinically generated horizontal vorticity and its subsequent tilting within downdrafts contribute to the formation of near-surface vertical vorticity (Davies-Jones and Brooks 1993). In addition, Brooks et al. (2003) showed that tornadogenesis is favored in environments of strong low-level shear and high near-surface moisture. Recently, Markowski and Richardson (2014, hereafter MR14) conducted idealized toy model simulations of pseudostorms varying the intensity of the cold pool and low-level shear. Their results revealed that a moderate cold pool with strong low-level shear can spawn a strong cyclonic rotation near the surface, while a strong cold pool cannot induce such strong rotation because of the difficulty in lifting less-buoyant parcels from near the surface. In addition, MR14 demonstrated that strong low-level shear lowers the

* Current affiliation: RIKEN, Kobe, Japan.

**Corresponding author address:** Takumi Honda, RIKEN Advanced Institute for Computational Science, 7-1-26 Minatojima-minami-machi, Chuo-ku, Kobe, Hyogo 650-0047, Japan.

E-mail: takumi.honda@riken.jp

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base of a low-level mesocyclone (MC), enhancing the upward dynamical perturbation pressure gradient force (DPGF) near the surface.

Despite many studies, some aspects regarding supercellular tornadogenesis are not yet well understood. For example, the effect of surface friction (Schenkman et al. 2014), radiation processes (Frame and Markowski 2013), and storm–thermal boundary interactions (Schultz et al. 2014) remain unknown. In particular, storm–thermal boundary interactions play an important role in some cases of tornadogenesis, since many tornadoes are observed near mesoscale thermal boundaries (Maddox et al. 1980; Markowski et al. 1998; Rasmussen et al. 2000; Fierro et al. 2006; Schultz et al. 2014).

Maddox et al. (1980) proposed a physical model that explains why meso-β-scale moisture convergence and vertical shear are enhanced near thermal boundaries and showed that storms moving across (parallel or along) a thermal boundary tended to spawn short-lived (long lived) tornadoes. Doswell et al. (2002) reported that an observed supercell was only tornadic early in its life when it first encountered a mesoscale cold pool. As the supercell progressed farther into the cold pool, it became elevated and ceased producing tornadoes.

A storm–storm interaction can be regarded as an example of a storm–thermal boundary interaction, especially just before mergers, although additional physical processes (e.g., updraft interactions) may appear after mergers. Indeed, Lee et al. (2006a,b) reported that, during the Illinois tornado outbreak in 1996, a number of supercells and other convective cells formed and repeatedly merged or separated from each other. Lee et al. (2006b) stated that, during this outbreak, the tornadoes were frequently observed just before or after storm mergers. Recently, Hastings et al. (2012, 2014) conducted idealized simulations regarding mergers in supercells and showed that some types of mergers enhance near-surface mesocyclogenesis. They suggested that this result is caused by the difference in the rear-flank baroclinicity between post- and premerger storms. Regarding another observed tornado, Wurman et al. (2007) analyzed dual-Doppler radar observations and showed that storm mergers may favor successive tornadogenesis by enhancing convergence. In particular, the tornadoes occurring just before storm mergers were possibly influenced by the interaction between a supercell and a cold pool induced by other convective cells (i.e., outflow boundary). In this study, we attempt to better understand the storm–outflow boundary interaction and related tornadogenesis.

Previous studies have addressed the influence of outflow boundaries on the evolution of supercells and tornadogenesis. For example, Atkins et al. (1999, hereafter A99) conducted idealized numerical experiments using an idealized heterogeneous base state derived from a composite of proximity soundings. In one of their experiments, supercellular convection formed near an airmass boundary and crossed it. Their results revealed that the horizontal vorticity generation at the forward flank of a simulated supercell is enhanced by the baroclinicity across the airmass boundary, promoting the development of a near-surface MC2 through tilting. It is important to note that A99 adopted the Kessler warm-rain microphysics parameterization (Kessler 1969), which overestimates the evaporative cooling and subsequent baroclinic generation of horizontal vorticity (Gilmore et al. 2004). In addition, since they used coarse horizontal grid spacing, their simulations could not resolve the tornadogenesis processes associated with the interaction between a supercell and an outflow boundary. On the other hand, Ziegler et al. (2010) simulated the Binger tornadic supercell interacting with a slowly varying, idealized horizontally heterogeneous base state based on observations. They revealed that horizontal variations in the thermodynamic properties contributed to the long-term maintenance of the Binger supercell.

To make it possible to conduct high-resolution numerical experiments on tornadogenesis associated with the interaction between a supercell and an outflow boundary, the following approaches are useful. Richardson et al. (2007, hereafter R07) introduced large modifications in the lateral boundary conditions and advection schemes to produce the correct environmental variation as a small domain moved through a prescribed steady-state environment. These techniques and the use of analytic wind profiles that satisfy steady-state conditions allowed R07 to investigate the influence of horizontally heterogeneous steady vertical shear on convective cells. Their experiments with a horizontally varying vertical shear revealed that the development of new convective cells on the stronger shear flank is promoted, provided that convection is initiated in a relatively weak shear regime.

Another approach was recently developed by Nowotarski et al. (2014, 2015), who investigated the influence of a quasi-two-dimensional convective boundary layer (CBL) on a simulated supercell. They first initiated the CBL with random thermal perturbations and surface heat fluxes. Then a thermal bubble was inserted within the horizontally inhomogeneous environment including the CBL, after which the calculation was restarted. As a

2 The definition of low-level MCs in A99 differs from that used in this manuscript. A99 originally referred to circulations at 500 m AGL as low-level MCs.
result, both the simulated supercell and the CBL were observed within the calculation domain. Using this method, Nowotarski et al. (2015) found that near-surface circulation in a supercell is enhanced (disrupted), provided that the environmental wind profile causes the storm to move parallel (perpendicular) to the roll convection. Moreover, they demonstrated that the cloud shading effect suppresses such roll convection.

In this study, we first initiate a supercell by a thermal bubble embedded in a horizontally homogeneous base state. After that, a potential temperature perturbation in a near-surface layer, corresponding to a highly idealized outflow boundary, is inserted. Using this modified field, the calculation is restarted. This method allows the supercell to fully develop before interacting with an outflow boundary, even for a relatively small calculation domain. Moreover, the time dependence of such interactions can be examined using different restart times. Another advantage of this method is that it does not require any modifications of the model boundary conditions or advection schemes. This method is referred to as the insert–restart method in this paper.

The main goal of this study is to clarify the tornado-genesis mechanisms associated with the interaction between a supercell and an outflow boundary. Although outflow boundaries are often accompanied by horizontal shear on their leading edge (Lee and Wilhelmson 1997), we perform high-resolution numerical experiments regarding outflow boundaries without horizontal shear using a better microphysics parameterization. This is because the inclusion of horizontal shear could have complicated the analysis. In a separate, forthcoming study, we consider the interaction between a supercell and an outflow boundary with horizontal shear, which possibly modulates tornadogenesis. Although both the radiation processes (Doswell et al. 2002; Frame and Markowski 2013; Nowotarski et al. 2015) and surface friction (Shimose 2009; Schenkman et al. 2014) may affect tornadogenesis and the outflow boundaries, their effect is beyond the scope of this study. This paper is organized as follows. A brief description of the numerical model and methodology are given in section 2, and the experimental results are described in section 3. A discussion is given in section 4, and the findings are summarized in section 5. The formulations used in this study are described in the appendix.

2. Methodology

a. Numerical model

The Advanced Research Weather Research and Forecasting (WRF) Model version 3.4 (Skamarock et al. 2008) was used in this study. The governing equations of the WRF Model are fully compressible and non-hydrostatic. The calculation domain was \(85 \times 70 \times 20\) km\(^3\), with a horizontal grid spacing of 100 m. The vertical grid spacing varied between 23 m at the bottom and 2095 m at the top of the model (48 levels). The lateral boundary conditions for velocity and scalars were open and a zero-gradient condition in the inflow boundaries, respectively. The phase speed of gravity waves at the lateral boundaries was fixed at a value of 45 m s\(^{-1}\), as in R07. A rigid wall with a sponge layer located between the top of the model and 14 km above ground level (AGL) was used as the top boundary condition. The bottom and top boundary conditions were a free slip. The subgrid-scale turbulence was represented by the 1.5-order turbulent kinetic energy closure. For the microphysics parameterization, the Morrison et al. (2009) scheme, which predicts the mixing ratios of water vapor, cloud water, rain, cloud ice, snow, and graupel, and the number concentrations of rain, cloud ice, snow, and graupel, was adopted. For simplicity, the Coriolis force and radiation processes were excluded.

b. Base state

1) THERMODYNAMIC PROFILE

We generated base-state thermodynamic profiles based on McCaul and Weisman (2001), James et al. (2006), and James and Markowski (2010). The primary objective of their methods was to change the amount of the convective available potential energy (CAPE) with a fixed shape of the buoyancy profile, since the shape of the buoyancy profile possibly influences the behavior of supercells (McCaul and Weisman 2001). First, the surface pressure \(p_s = 1000\) hPa, and the height \(H_{\text{bl}} = 1.0\) km and pressure \(p_{\text{bl}} = 885\) hPa at the top of the boundary layer were determined. These values gave a scale height \(H = H_{\text{bl}}/\ln(p_s/p_{\text{bl}})\) of 8.1 km, which was used to produce the vertical profile of pressure. In the boundary layer, the lapse rates of the potential temperature and water vapor were held constant at \(-1\) K km\(^{-1}\) and 3 g kg\(^{-1}\) km\(^{-1}\), respectively. At the top of the boundary layer, the relative humidity (RH) and equivalent potential temperature were set at 95% and 340 K, respectively. According to James and Markowski (2010), the RH profile between the top of the boundary layer and the tropopause \(H_{\text{trop}} = 12\) km) was derived from the following function:

\[
\text{RH} = 0.95 - 0.45 \left( \frac{z - H_{\text{bl}}}{H_{\text{trop}} - H_{\text{bl}}} \right)^{0.7},
\]
where $z$ represents the height above the ground. Next, the buoyancy profile for a parcel ascending pseudoadiabatically from the top of the boundary layer was specified by the following equation (James et al. 2006):

$$b(z') = A_b \frac{m^2}{H_b} z' \exp\left(-\frac{m z'}{H_b}\right) - B_b z'^3,$$

(2)

where $z'$, $H_b$, and $m$ are the height above the boundary layer top, the scale height for the buoyancy profile ($H_b = 12.75$ km), and the parameter affecting the shape of the buoyancy profile, respectively. Parameters $A_b$ and $B_b$ were defined as

$$A_b = \frac{E}{1 - \left(1 + \frac{m z_{el}^2}{H_b} + \frac{m^2 z_{el}^4}{4 H_b^2}\right) \exp\left(-\frac{m z_{el}}{H_b}\right)},$$

(3)

and

$$B_b = A_b \frac{m^2}{H_b z_{el}} \exp\left(-\frac{m z_{el}}{H_b}\right),$$

(4)

where $z_{el}$ and $E$ are the equilibrium level and the parameter that controls the magnitude of CAPE, respectively. The base-state potential temperature profile was determined to satisfy the buoyancy profile derived from Eq. (2), the pressure profile, and the RH profile based on Eq. (1). The RH above the tropopause was held constant at 25%, and the stratosphere was assumed to be isothermal. Two base-state soundings were produced with a fixed buoyancy profile in which the maximum buoyancy was placed at the midtroposphere as in James et al. (2006) and James and Markowski (2010). Figure 1 shows the skew $T$–log $p$ diagrams for each base state. One profile was characterized as a large CAPE environment, in which the mixed-layer CAPE (mlCAPE) was 3156 J kg$^{-1}$ (Fig. 1a). This sounding was referred to as CAPE3200. The other sounding, referred to as CAPE700, had a small CAPE (mlCAPE = 704 J kg$^{-1}$) (Fig. 1b). The sounding parameters $E$ and $m$ for CAPE3200 (CAPE700), determined by trial and error, were 3610 and 0.8 (1000 and 3.2), respectively.

Some environmental parameters for each sounding are shown in Fig. 1. For example, the mixed-layer lifting
condensation level (mlLCL) for CAPE3200 was 1193 m (Fig. 1a). The environmental parameters for the wind profile and state within the boundary layer were almost the same for CAPE3200 and CAPE700, since both parameters were fixed, as shown below. Thus, CAPE3200 (CAPE700) was characterized by a larger (smaller) significant tornado parameter (STP; Thompson et al. 2003). Indeed, STP for CAPE3200 (CAPE700) was 1.40 (0.31) (Fig. 1). Since the likelihood of significant tornadoes strongly depends on STP (Thompson et al. 2003), the CAPE700 sounding was regarded as a less favorable environment for tornadogenesis compared to CAPE3200.

2) WIND PROFILE

The base-state wind hodograph was produced as follows (Fig. 2). First, the angle of the hodograph rotation within 0–2 km AGL (α2km = 90°), the magnitude of the 0–1 km AGL horizontal wind shear vector $S_{1km}$, and the magnitude of the 0–6 km AGL horizontal wind shear vector $U_{6km}$ were determined. Then the radius of the hodograph $R$ below 2 km AGL was derived as

$$R = S_{1km} \times \sin \left( \frac{\pi}{2} - \frac{\alpha_{2km}}{4} \right) / \cos \left( \frac{\pi}{2} - \frac{\alpha_{2km}}{2} \right).$$

For the 2–6 km AGL layer, the meridional wind was fixed, and only the zonal wind increased linearly with the height, satisfying the given $U_{6km}$ magnitude.

magnitudes of $S_{1km}$ and $U_{6km}$ were 7 and 20 m s$^{-1}$, respectively, based on the relatively weak magnitudes for tornadic events given by Thompson et al. (2003). This is because the CAPE3200 sounding with stronger vertical shear produces an environment so favorable for strong tornadoes that it is difficult to see the effects of the outflow boundary interaction. Finally, the storm motion identified from a pilot simulation was subtracted from the hodograph. The resulting wind profile (Fig. 1) was used for all experiments.

c. Overview of the control experiments

The control experiments were performed using a horizontally homogeneous base state (CAPE3200 or CAPE700) with a 4-K ellipsoidal thermal bubble. The height of the bubble center was 1.5 km AGL. The horizontal and vertical radii of this bubble were 10 and 1.5 km, respectively. An overview of the control experiments (i.e., no outflow boundary) with the CAPE3200 (CNTL) and CAPE700 sounding (CNTL CA7) is given in the following paragraphs.

Figure 3 presents the horizontal distributions of the vertical velocity, vertical vorticity, and surface cold pool for both experiments. A typical developed supercell was reproduced in CNTL [i.e., an MC, updraft region, and surface cold pool were found (Figs. 3a,b)]. Although a typical supercell was also simulated in CNTL CA7, the convection evolution was somewhat different from that in CNTL; that is, the MC and updraft of this storm at 140 min were weak (Figs. 3b,d), which is consistent with Weisman and Klemp (1982), who demonstrated that low-CAPE environments are not favorable for the development of strong supercells.

d. Insert–restart method

The insert–restart method adopted in this study is explained in this subsection. Figure 4 shows the horizontal and zonal–vertical cross sections from a CAPE3200 experiment using the insert–restart method. In this method, the integration of CNTL was stopped at a certain time for a restart (e.g., 90 min; Fig. 3a). Then a near-surface negative potential temperature perturbation $\theta_B$, corresponding to a highly idealized

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Footnotes:

3. The amount of the STP value depends not only on mlCAPE but also on the vertical shear and near-surface moisture. The influence of the latter two factors is beyond the scope of this study.

4. Thompson et al. (2003) showed the magnitude of the 0–6 km shear vector, which somewhat differs from the $U_{6km}$ used in this study. The 0–6 km shear vector in this study was 21.4 m s$^{-1}$, which was comparable to the median value of the weakly tornadic supercells in Thompson et al. (2003) (22.8 m s$^{-1}$; see their Fig. 8).

5. Throughout the manuscript, $\theta_B$ represents the inserted potential temperature perturbation associated with a highly idealized outflow boundary. On the other hand, $\theta$ simply indicates the potential temperature deviation from its base-state component.
outflow boundary, was inserted in an eastern region of
the domain (Figs. 4a,c). Since the boundary-parallel
wind perturbation was not inserted, the outflow
boundary had initially no horizontal shear. The \( u_0 \) value
was 2 K, which is comparable to the intensity of the
cold pool in a developed supercell (Figs. 3, 4a,c) and was
held constant in the eastern region of the domain
(Figs. 4a,c). At the western edge of this region, a 10-km-
wide transition region, with the amplitude of the
\( u_0 \) linearly decreasing westward, was introduced (Figs. 4a,c).
The perturbation was inserted only below the top of
the boundary layer (1 km AGL) with a transition
layer between 0.96 and 1.1 km AGL. After that, the
calculation was restarted. The perturbation moved
westward as a density current affected by the near-surface
base-state easterly wind (Fig. 1), approaching a de-
veloped supercell located in the western region of the
domain (Figs. 4b,d).

A set of experiments was conducted using the insert-
restart method. Cold-air perturbation (\( \theta_B = -2 \) K) ex-
periments using CAPE3200 with a restart time of 10, 60,
and 90 min were referred to as CP2T10, CP2T60, and
CP2T90, respectively. The timing dependence of the
supercell–outflow boundary interaction was examined
by comparing these experiments in section 4. To exam-
ine the dependence of the results on the
\( u_0 \) amplitude,
we also conducted CP1T90, CP3T90, CP4T90, and
CP5T90 experiments for different \( u_0 \) amplitudes. The
CAPE3200 experiments are summarized in Table 1. In
all experiments, \( \theta_B \) approached the developed supercell
about 40–50 min after the restart time, since the \( \theta_B \) loca-
tion was adjusted in each experiment (not shown).
The integration time after the restart was 60 min for all
experiments (e.g., the CP2T60 integration restarted at
60 min and finished at 120 min). To discuss the influence
of the base-state mlCAPE (STP), we also performed
similar experiments using CAPE700 (CP2T10CA7,

\[ \theta'_{B} \text{ at 0.1 km AGL (color shading; K), vertical velocity at 0.8 km AGL (red lines; 1 and 2 m s}^{-1}) \text{, and vertical vorticity at 0.8 km AGL (black lines; 0.01 s}^{-1}) \text{ in (a),(b) CNTL and (c),(d) CNTLCA7 at (a),(c) 90 and (b),(d) 140 min. Blue lines in (a) and (c) are the initial potential temperature perturbations at 1.5 km AGL (1-K interval).} \]
CP4T10CA7, CP6T10CA7, CP8T10CA7, CP10T10CA7, and CP2T60CA7; see Table 1). The results for the CAPE700 sounding are compared in section 4.

3. Results

In CP4T90, a surface vortex below the supercell was intensified by the supercell–outflow boundary interaction. Figure 5 shows the three-dimensional distributions of the vertical vorticity and $\theta'$ in CP4T90 and CNTL. At 122 min, a supercell with a strong updraft and an MC was found. This storm had arching vortex lines passing near a surface vortex, as reported by several observational and numerical studies (Markowski et al. 2012a; MR14). At 135 min, the base of the $w = 10\text{ m s}^{-1}$ isosurface lowered, reflecting the enhancement of the near-surface updraft. The outflow boundary was located around $X = 38\text{ km}$ at 135 min and reached the supercell at 140.5 min. Clearly, both the surface vortex and the near-surface updraft were stronger in CP4T90 than in CNTL (Figs. 5c,d).

The surface vortex intensification in CP4T90 was associated with the enhancement of the near-surface horizontal convergence. Figures 6a–h show the horizontal cross sections of the near-surface convergence below the supercell in CNTL and CP4T90. At 122.5 min, a surface vortex was located near the convergence region both in CNTL and CP4T90 (Figs. 6a,e). After that, the vortex and the surface convergence began to intensify in CP4T90 compared to CNTL (Figs. 6b,f). The main convergence zone of the outflow boundary was located around $X = 42\text{ km}$ at 130.5 min, and the

<table>
<thead>
<tr>
<th>Name</th>
<th>$\theta'$ (K)</th>
<th>Restart time (min)</th>
<th>mlCAPE (J kg$^{-1}$)</th>
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<tbody>
<tr>
<td>CNTL</td>
<td>—</td>
<td>—</td>
<td>3156</td>
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<tr>
<td>CP2T90</td>
<td>-2</td>
<td>90</td>
<td>3156</td>
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<td>-2</td>
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near-surface easterly wind was enhanced in front of the leading edge of the outflow boundary in CP4T90 (Figs. 6b,f). As shown below, this wind perturbation, which resulted in the enhancement of the surface convergence before the arrival of the outflow boundary at the surface vortex, was associated with a high-pressure perturbation on the leading edge of the outflow boundary. After the arrival of the outflow boundary, the surface convergence and vertical vorticity markedly increased in CP4T90, which had a vertical vorticity peak of 0.63 s⁻¹, twice the CNTL peak (Figs. 6c,d,g,h).

The time evolutions of the maximum vertical vorticity (ζmax) and maximum horizontal wind speed (UVmax) at 100 m AGL are presented in Figs. 6i and 6j. Both the ζmax and UVmax were larger in CP4T90 than in CNTL. In particular, the UVmax in CP4T90 exceeded 30 m s⁻¹. This tornado-like vortex⁷ in CP4T90 immediately dissipated (Figs. 6i,j), in agreement with some observational studies regarding tornadoes associated with storm–thermal boundary interactions (Maddox et al. 1980; Doswell et al. 2002; Wurman et al. 2007).

The surface vortex intensification by the supercell–outflow boundary interaction was observed in other experiments with different amplitudes of u₀B within an outflow boundary. Figure 7 shows the surface convergence and times series of ζmax and UVmax in CP1T90, CP2T90, CP3T90, CP4T90, and CP5T90. In all of these experiments, the surface convergence, vertical vorticity, and UVmax were enhanced by the interaction with an outflow boundary. Although the easterly wind perturbation associated with the outflow boundary increased with the amplitude of u₀B, the ζmax and UVmax enhancements were most prominent in the experiments with a moderate u₀B amplitude (i.e., CP2T90, CP3T90, and CP4T90). In these experiments, ζmax and UVmax began to increase around 123 min compared to CNTL, although the outflow boundary was still about 15 km distant from the supercell (Figs. 7e,f, 4b,d). This remote effect of the outflow boundary, which might be caused by model adjustments, will be discussed later.

The enhancements of the horizontal convergence and easterly wind by the outflow boundary are explained by density current dynamics. Figures 8a and 8b show the

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⁷ Actually, the horizontal grid interval of this study (100 m) was not sufficient to resolve the tornado structure. Thus, the observed strong surface vortices are “tornado-like” vortices. For simplicity, the formation of tornado-like vortices, corresponding to the prominent surface vortex intensification, is referred to as tornadogenesis.
horizontal distributions of the theoretical phase speed of the density currents derived from Eq. (A1) (see appendix) and the simulated zonal wind perturbation in CP4T90 and CP5T90. The intensity of the zonal wind perturbation (contours in Figs. 8a,b) was comparable to the estimated phase speed of the density current (shading in Figs. 8a,b). Figures 8c–f show the zonal–vertical cross sections of the outflow boundaries in CP4T90 and CP5T90. Since the outflow boundary moved westward as a density current, the horizontal convergence was enhanced at the leading edge of the outflow boundaries (Figs. 8c,d). Both the estimated phase speed of the density current and the associated horizontal convergence were stronger in CP5T90 than in CP4T90 (Figs. 8a–d), since a density current consisting of cooler (negatively buoyant) parcels propagates more rapidly than one consisting of relatively warmer (buoyant) parcels, provided that the depth of the density currents is nearly identical [Eq. (A1)]. Indeed, as shown in the zonal–vertical cross sections of the density potential temperature perturbation $\theta_p$ [Eq. (A4) in the appendix], the outflow boundary in CP5T90 had more negatively buoyant parcels than in CP4T90 (Figs. 8e,f). Moreover, the easterly wind was also enhanced in the west (front) of the outflow boundary ($X = 39$ km in Figs. 8c,d). A high perturbation pressure $p'$ on the leading edge of the outflow boundary and associated westward pressure gradient force, as noted by Bryan and Rotunno (2014), were partly responsible for these easterly wind perturbations (Figs. 8c–f). However, since no pressure perturbations with the inserted cold air were applied in this study, model adjustments might result in the easterly wind perturbations, especially far from the main convergence zone of the outflow boundary. We emphasize that the largest differences occurred after 135.5 min, when the outflow boundary reached the

![Horizontal Convergence, Vorticity, and Winds at 97 m AGL](image)

**Fig. 6.** (a)–(h) Horizontal cross sections of the horizontal convergence (color shading; $3 \times 10^{-2}$ s$^{-1}$) at 97 m AGL in (a)–(d) CNTL and (e)–(h) CP4T90 at (a),(e) 122.5; (b),(f) 130.5; (c),(g) 135.5; and (d),(h) 140.5 min. Black contours are the vertical vorticity (0.01, 0.1, 0.2, and 0.3 s$^{-1}$) at 97 m AGL. Magenta dots show the regions where the potential temperature perturbation is less than $1.5$ K. Vectors are (a)–(d) the horizontal wind at 97 m AGL in CNTL and (e)–(h) the differential horizontal wind between CP4T90 and CNTL at each time. Vectors in (e)–(h) are drawn on the grid points where the differential wind is greater than 0.5 m s$^{-1}$. The maximum vertical vorticity ($\zeta_{\text{max}}$; s$^{-1}$) and horizontal wind speed (UV$\text{max}$; m s$^{-1}$) in each frame are shown at the top of each panel. Also shown are time series of (i) $\zeta_{\text{max}}$ (s$^{-1}$) and (j) UV$\text{max}$ (m s$^{-1}$) at 100 m AGL for CNTL (black lines) and CP4T90 (blue lines), respectively. The values of $\zeta_{\text{max}}$ and UV$\text{max}$ are sought within the region that corresponds to (a)–(h).
supercell (Figs. 6a–h). Thus, it is clear that the outflow boundary significantly contributed to the surface vortex intensification.

Negatively buoyant parcels within the outflow boundary influence the acceleration of the upward velocity and surface vortex stretching, provided that they penetrate around and flow through the surface vortex. To clarify the penetration of these negatively buoyant parcels, we calculated 6-min forward trajectories using a fourth-order Runge–Kutta method with a time step of 0.1 s. The wind fields stored at 6-s intervals were linearly spatiotemporally interpolated. To avoid extrapolation, the parcels that went below the lowest model level (12 m AGL) were not analyzed. Figure 9 depicts the zonal–vertical and horizontal cross sections and projections of the forward trajectories in CNTL, CP4T90, and CP5T90. We seeded 20 parcels with 200-m meridional intervals at both 50 and 150 m AGL (i.e., total number of 40 parcels) at 135 min (black squares in Fig. 9).

A forward-trajectory analysis revealed that some of the negatively buoyant parcels behind the outflow boundary penetrated inward just above the surface vortex in both CP4T90 and CP5T90 (Fig. 9). In contrast, such penetration of the seeded (forward flank) parcels was not observed in CNTL, in agreement with Dahl et al. (2012). This difference indicates that the interaction with the outflow boundary affected the origin of parcels around the surface vortex. The penetrating parcels in CP4T90 and CP5T90 were initially more negatively buoyant than the seeded parcels in CNTL, so that their ascending motion during the 6-min time integration was weaker than in CNTL (Figs. 9a–c). However, as mentioned earlier, the interaction with the outflow boundary enhanced the near-surface horizontal convergence (equivalent to the vertical gradient of the vertical velocity under the Boussinesq approximation; color shading in Figs. 9a–c). As a result, the surface vortex was strengthened, although some negatively buoyant parcels penetrated around the surface vortex. The buoyancy reduction around the surface vortex by the outflow boundary is more clearly demonstrated in section 4c.

Unfortunately, we could not conduct a sufficiently accurate Lagrangian vorticity budget analysis for the surface vortex because of the lower boundary condition for the trajectory calculation (Dahl et al. 2014). The interaction with the outflow boundary resulted in the modulation of the near-surface baroclinicity, strongly influencing the vorticity budget. Such modulation is clearly depicted in Figs. 10a and 10b, which show the horizontal distributions of $\theta^\prime$ (shading) and approximate intensity of the baroclinic horizontal vorticity.
generation (red vectors in Figs. 10a,b) in CNTL and CP4T90, respectively. In particular, the rear-flank baroclinicity largely differed between CNTL and CP4T90 (Figs. 10a,b). Actually, similar modification of the rear-flank baroclinicity was found in a storm merger (Hastings et al. 2012, 2014). The influence of the difference in the rear-flank baroclinicity on the surface vorticity budget will be examined in the future.

In summary, an outflow boundary promoted tornado genesis in a supercell through the enhancement of the horizontal convergence near the surface and the subsequent vortex stretching. This process is similar to that indicated by Wurman et al. (2007). Furthermore, the outflow boundary played a role in increasing the surface baroclinicity, especially in the rear-flank region. In addition to the enhancements of the near-surface horizontal convergence and associated vortex stretching, this might also contribute to the near-surface vorticity budget. Although a surface vortex significantly intensified through the supercell–outflow boundary interaction, the resulting tornado-like vortex dissipated immediately. This is consistent with observational studies regarding this type of the storm–thermal boundary interaction (Maddox et al. 1980; Doswell et al. 2002; Wurman et al. 2007).

4. Discussion

a. Dependence on the interaction timing: Preconditioning

The surface vortex intensification depended on the timing of the interaction. Figure 11 shows the horizontal cross sections of the near-surface convergence and the time evolution of \( \zeta_{\text{max}} \) and \( U/V_{\text{max}} \) in CP2T60 and CNTL. The value of \( \zeta_{\text{max}} \) in CP2T60 was much smaller than in CP2T90 (Figs. 7e, 11e): that is, the surface vortex intensification observed in CP2T90 was not found in
The surface vortex in CNTL was much weaker at 98 min than at 122.5 min (Figs. 6a, 11c), and this was the only difference that could plausibly account for the lack of surface vortex intensification in CP2T60. In other words, the surface vortex intensification through storm–outflow boundary interaction required a preconditioned supercell possessing a strong surface vortex.

The preconditioning of the supercell may be cyclically satisfied, according to the cyclic time evolution of supercells and MCs (Adlerman et al. 1999). Figure 12 shows the time evolution of the maximum low-level updraft helicity (UH) in CNTL. Two UH peaks were found at 130–140 and 50–60 min, respectively. Thus, in addition to the 130–140-min interaction, we simulated another timing interaction at 50–60 min using the CAPE3200 sounding. Additional experiments restarting at 10 min are also listed in Table 1 (i.e., CP2T10, CP4T10, CP6T10, and CP7T10). Figure 13 shows the horizontal cross sections of the near-surface horizontal convergence and the time evolution of the $\zeta_{\text{max}}$ and UVmax. The interaction with the outflow boundary at 50–60 min also resulted in a surface vortex intensification (Fig. 13). However, the $\zeta_{\text{max}}$ observed in these additional experiments was weaker than at 130–140 min (Figs. 7, 13), since both the near-surface circulation and low-level UH in CNTL at 50–60 min were weaker than at 130–140 min (Figs. 7, 12, 13), meaning that the preconditioning of the supercell was not fully satisfied at 50–60 min.

b. Influence of environmental mCAPE (STP)

In this subsection, the dependence of the surface vortex intensification on the environmental mCAPE (STP) is examined, since the significant tornado likelihood strongly depends on STP (Thompson et al. 2003), which is associated with mCAPE. Using the output from CNTLCA7, some additional experiments were performed, as shown in Table 1 (i.e., CP2T10CA7, CP4T10CA7, CP6T10CA7, CP8T10CA7, CP10T10CA7, and CP2T10CA7). Since the lifetime of a supercell simulated in CNTLCA7 was much shorter than in CNTL (Fig. 3), the additional experiments were restarted earlier than for the CAPE3200 experiments. Even for the experiments with an earlier restart time (10 min), the simulated storm was fully developed before the storm–outflow boundary interaction (not shown). As in the CAPE3200 experiments, the $\theta''_p$ location was adjusted so that the inserted outflow boundary approached the supercell 40–50 min after the restart.
Figure 14 shows the horizontal cross sections of the near-surface horizontal convergence and the time series of the $z_{\text{max}}$ and UVmax for the CAPE700 experiments. The value of $z_{\text{max}}$ was smaller in CNTLCA7 than in CNTL (Figs. 14a,e, 6a–d,i), reflecting a relatively low mlCAPE (STP; Figs. 1a,b). As the inserted outflow boundary arrived around the supercell, the surface vortex intensified (Fig. 14). As in the CAPE3200 experiments, such vortex intensification was not observed in an experiment with a different restart time (CP2T60CA7; not shown), meaning that the surface vortex intensification depended on the timing of the supercell–outflow boundary interaction even for a relatively low mlCAPE (STP). In addition, the surface vortex intensification depended on the amplitude of $u_0$, as observed in the CAPE3200 (Figs. 14, 7). The optimal $u_0$ value for the CAPE700 experiments was $28$ K, being much cooler than for the CAPE3200 experiments (from $22$ to $-4$ K; Figs. 13e, 7e). This difference is discussed in section 4c.

The optimal timing of the interaction was earlier in the CAPE700 experiments (45–55 min; Fig. 13e) than in the CAPE3200 experiments (130–140 min; Figs. 7e, 11e), which is attributed to the time evolution of the control supercell. In CNTLCA7, the amplitude of the low-level UH reached a maximum at 45–55 min, gradually dissipating afterward (Fig. 12). Thus, the preconditioning of the supercell was satisfied earlier in CAPE700 than in CAPE3200.

In summary, the surface vortex intensification associated with the supercell–outflow boundary interaction occurred even for a low-mlCAPE (STP) environment, provided that a supercell was preconditioned. Note that the $z_{\text{max}}$ in CP8T10CA7 was $0.3$ s$^{-1}$ (Fig. 14e). Although STP for the CAPE700 sounding (0.31) was much smaller than for the CAPE3200 sounding (1.40; Fig. 1), the vorticity maximum was comparable to the CAPE3200 experiment without an outflow boundary (CNTL; see the black line in Fig. 6i). Despite a less favorable environment, a storm with a boundary interaction might reach a similar maximum near-surface vorticity as a storm without a boundary in a more favorable environment.

c. Optimal intensity of the outflow boundary

The optimal outflow boundary intensities observed in this study were $-2$ to $-4$ and $-8$ K for the CAPE3200 experiment

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**Figure 10.** Horizontal distributions of the potential temperature perturbation (color shading; K) at 0.1 km AGL for (a) CNTL at 138 min, (b) CP4T90 at 138 min, and (c) CP4T10CA7 at 48 min. Using the density potential temperature perturbation $\theta'_d$ [see Eq. (A4)], red arrows show approximate baroclinic vorticity generation ($\partial \theta'_d/\partial y$, $-\partial \theta'_d/\partial x$). Black circles denote the location of the strongest surface vortex in each panel. The vertical vorticity of the strongest vortex is shown in the lower-right corner of each panel. Also shown are time evolutions of $\theta'_d$ within (d) box A, (e) box B, and (f) box C in (a)–(c), respectively. The depths of the boxes are 0.1 km (0.1–0.2 km AGL). Times $t$ are (d) 138, (e) 138, and (f) 48 min. The light purple and red regions in (e) and (f) indicate the reduction in $\theta'_d$ between $t=6$ and $t+8$ min in CP4T90 and CP4T10CA7, respectively.
and CAPE700 experiments, respectively (Figs. 7e, 14e). This difference was related to the intensity of the cold pool in the simulated supercells. Figure 10c depicts the horizontal distributions of $u_0$ near the surface in CP4T10CA7. The cold pool of the supercell was warmer in CP4T90 than in CP4T10CA7 (Figs. 10b,c). Since their $u_B^0$ was identical ($\sim 4$ K), the instantaneous baroclinic vorticity generation in the rear-flank region was more strongly enhanced in CP4T90 compared to CP4T10CA7 (red vectors in Figs. 10b,c). This difference might partly contribute the vorticity difference between CP4T90 and CP4T10CA7, although we failed to justify the contribution by the Lagrangian vorticity budget analysis.

On the other hand, the reduction of the surface buoyancy due to the interaction was larger in a supercell with a weak cold pool, compared to a supercell with a strong cold pool. Figures 10d–f show the time evolutions of the area-mean $\theta_p^0$ within rectangles (black rectangles in Figs. 10a–c) surrounding the surface vortex. The interaction with the outflow boundary reduced the area-mean $\theta_p^0$ in CP4T90 compared to CNTL (Figs. 10d,e). This reduction was of about 3 K in CP4T90 (shaded purple in Fig. 10e), being larger than in CP4T10CA7 (about 2 K; shaded red in Fig. 10f). It is important to note that too-strong cold pools suppress the intensification of surface vortices since they consist of significantly negatively buoyant parcels (e.g., Markowski and Richardson 2010; MR14). Thus, the optimal outflow boundary intensity reflects the buoyancy balance between the cold pool and outflow boundary. A negatively buoyant outflow boundary results in not only an increase of near-surface convergence but also a reduction of near-surface buoyancy around a surface vortex within a supercell. Amplitude of the reduction depends on the cold pool buoyancy, so that it becomes small if the supercell has a strong cold pool. However, such a strong cold pool is not suitable for tornadogenesis. Consequently, a moderate buoyancy difference between a cold pool in a supercell and an outflow boundary results in the most prominent surface vortex intensification.

Our results do not suggest that the optimal intensity of the outflow boundary in CAPE700 was always weaker than for the CAPE3200 experiments. The cold pool of the simulated supercell in CAPE3200 was often stronger than in CAPE700 (Fig. 3). Thus, the surface vortex intensity
intensification depended not on the base-state mCAPE (STP) but on the intensities of the outflow boundary and cold pool in a supercell just before the supercell–outflow boundary interaction.

d. Role of the low-level shear

In section 3, we showed that the enhanced near-surface easterly wind associated with the outflow boundary promoted vortex stretching and subsequent tornadogenesis. In contrast, several studies (Maddox et al. 1980; A99) suggested that the horizontal variation in the low-level vertical shear across a thermal boundary strongly influenced supercells and MCs. In particular, A99 revealed that the near-surface MC was intensified by the tilting of the streamwise vorticity associated with a low-level thermal boundary. Recently, MR14 indicated that strong low-level shear with a high streamwise vorticity lowers the base of the low-level MC, resulting in strong dynamical lifting near the surface. A rotation increase aloft induces dynamical suction (Lilly 1986), promoting vortex stretching near the surface (Wicker and Wilhelmson 1995; Noda and Niino 2010). In this subsection, we discuss the role of the low-level vertical shear in the surface vortex intensification.

The enhanced easterly wind near the surface corresponds to an increase of the low-level vertical shear. This is because the base-state westerly wind increases with height (Figs. 1, 2). Figure 15 shows the horizontal cross sections of the low-level UH, 0–2-km mean horizontal vorticity, and 0–1-km mean DPGF for CNTL and CP4T90. As the outflow boundary approached the supercell, the northward component of the horizontal vorticity, associated with the enhanced easterly wind near the surface, and the baroclinic vorticity generation on the leading edge of the outflow boundary increased in CP4T90 in comparison with CNTL (Figs. 10a,b, 15). Part of the increased horizontal vorticity corresponded to the streamwise vorticity since the low-level (0–2 km AGL) mean wind was northwestward. Such streamwise vorticity may intensify the rotation aloft and associated dynamical lifting through tilting, as shown by MR14. Indeed, both low-level UH and DPGF intensified at 140 min in CP4T90 compared to CNTL (Figs. 15c,f). These seem to be associated with a local increase of the streamwise component of the 0–2-km mean horizontal vorticity in CP4T90 (blue contours in Fig. 15f). However, it is difficult to quantify the increase of UH by this effect because low-level horizontal vorticity was not identical between CNTL and CP4T90 even before the boundary arrived (Fig. 15). The low-level UH began to increase before the arrival of the enhanced horizontal vorticity near a large UH region (\(X = 36 \text{ km at 134 min; Fig. 15c,e}\)). The preceding increase of the low-level UH resulted from the enhancement of the near-surface horizontal convergence and subsequent updraft acceleration (Figs. 5, 6).
In summary, an outflow boundary enhanced not only the horizontal convergence near the surface but also the low-level vertical shear and associated near-surface dynamical lifting, as in A99 and MR14. This process was also responsible for the surface vortex intensification associated with the supercell–outflow boundary interaction.

5. Summary and concluding remarks

A mechanism of tornadogenesis associated with the supercell–outflow boundary interaction was examined. We adopted the insert–restart method, in which a near-surface cold-air perturbation corresponding to a highly idealized outflow boundary was inserted in the calculation domain containing a developed supercell. After that, the calculation was restarted and the inserted outflow boundary interacted with the supercell. We also conducted several experiments with different perturbation amplitudes and restart times.

An outflow boundary promoted surface vortex intensification in a preconditioned supercell having a strong surface vortex. In particular, a tornado-like vortex was observed in an experiment with a moderate outflow boundary, associated with the enhancement of the near-surface horizontal convergence and subsequent vortex stretching by the outflow boundary. In addition, the outflow boundary modulated the low-level shear and near-surface baroclinicity around the supercell, with both potentially influencing the surface vortex intensification.

The surface vortex intensification was most prominent when a moderately cold outflow boundary arrived at a preconditioned supercell having a strong surface vortex. This optimal intensity of the outflow boundary was associated with the balance between the horizontal convergence intensification and the parcel buoyancy reduction by the outflow boundary. Furthermore, the surface vortex intensification associated with the supercell–outflow boundary interaction was independent of the base-state mCAPE (STP). Despite less favorable environment, a storm with a boundary interaction might reach a similar maximum near-surface vorticity as a storm without a boundary in a more favorable environment. The mCAPE variation affected the timing of the preconditioning and the intensity of the cold pool in the supercell, modulating the surface vortex intensification.

In this study, we did not consider some factors that possibly affect tornadogenesis associated with storm–outflow boundary interaction. In particular, our discussion regarding the low-level shear and associated dynamical lifting indicated that the environmental hodograph shape strongly affects tornadogenesis. Indeed, Esterheld and Giuliano (2008) indicated that the angle between the near-surface storm-relative inflow and the near-surface shear vector allows a distinction.
between significant tornadic storms and nontornadic storms. A similar discussion was also conducted by Nowotarski and Jensen (2013) using a large sample of observed storms. Moreover, Maddox et al. (1980) and A99 showed the importance of the angle between a supercell and an outflow boundary, although our simulations adopted only one direction. Our simulations also excluded the surface friction and radiation processes, both of which may modulate the supercell, the outflow boundary, and their interaction (Lee and Wilhelmson 1997; Doswell et al. 2002; Frame and Markowski 2013; Schenkman et al. 2014; Nowotarski et al. 2015). Furthermore, although we fixed the shape of the inserted cold pool as a long wall, a circular cold pool is probably more realistic. Therefore, more comprehensive simulations in a large parameter space are required in the future.

One of the largest assumptions in this study is that outflow boundaries initially have no horizontal shear on their leading edge, although they often accompany such shear (Lee and Wilhelmson 1997). The interaction between a supercell and an outflow boundary with horizontal shear is a current subject of our research.

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APPENDIX

Formulations Used in This Study

The phase speed $c$ of a density current in shear is approximated as (e.g., Markowski and Richardson 2010)
where $H$ and $B_f$ represent the depth of the density current and buoyancy forcing, respectively. According to Doswell and Markowski (2004) and Markowski (2002), $B_f$ in the Boussinesq approximation is defined as

$$B_f = B - \frac{1}{\rho_0} \frac{\partial p_h'}{\partial z}$$  \hspace{1cm} (A2)

where $B$, $\rho_0$, $p_h'$, $\theta$, $q_v$, $r_h$, and $g$ represent thermal buoyancy, constant air density, buoyancy pressure perturbation, potential temperature, the mixing ratios of water vapor and hydrometeors, and gravitational acceleration, respectively. The overbars indicate the base-state component, and the prime stands for the deviation from the base state. In addition, the density potential temperature $\theta_\rho$ can be used as a measure of parcel buoyancy; $\theta_\rho$ is defined as

$$\theta_\rho = \theta(1 + 0.61q_v - r_h).$$  \hspace{1cm} (A4)

Under the Boussinesq approximation, dynamical pressure perturbation $p'_d$ and $p'_b$ satisfy the following relationships (e.g., Markowski and Richardson 2010):

$$\alpha_0 \nabla^2 p'_d = -c^2_{ij} + \frac{1}{2} | \omega^2 |$$  \hspace{1cm} (A5)

and

$$\alpha_0 \nabla^2 p'_b = \frac{\partial B}{\partial z}.$$  \hspace{1cm} (A6)

where

$$c^2_{ij} = \frac{1}{4} \sum_{i=1}^{3} \sum_{j=1}^{3} \left( \frac{\partial u_i}{\partial x_j} + \frac{\partial u_j}{\partial x_i} \right)^2,$$  \hspace{1cm} (A7)

and $u_1 = u, u_2 = v, u_3 = w, x_1 = x, x_2 = y,$ and $x_3 = z$. The variables $\alpha_0$ and $\omega$ represent the specific volume (constant) and three-dimensional vorticity vector, respectively.

To solve a Poisson equation [Eq. (A6)], three-dimensional boundary conditions are required. Since the calculation domain is relatively small, cyclic lateral boundary conditions are not appropriate in this study. Thus, for the lateral boundary conditions, the Neumann condition $\left( \frac{\partial p'_d}{\partial x} = \frac{\partial p'_b}{\partial y} = 0 \right)$ is used. According to Rotunno and Klemp (1982),

$$\frac{\partial p'_d}{\partial z} = B$$  \hspace{1cm} (A8)

is satisfied at the bottom and top boundaries because vertical velocity is zero there. Although $\frac{\partial p'_d}{\partial z} = 0$ at the bottom and top boundaries is assumed in this study, individual conditions for $\frac{\partial p'_d}{\partial z}$ and $\frac{\partial p'_b}{\partial z}$ are somewhat arbitrary (Markowski and Richardson 2010).

To derive $p'_d$ and $p'_b$, we solve the Poisson equation [Eq. (A6)] and then obtain $p'_b$. By subtracting $p'_b$ from the total perturbation pressure, $p' (= p'_d + p'_b)$. $p'_d$ can be calculated as the residual component. We then use $p'_d$ for the calculation of the dynamical perturbation pressure gradient force.

To examine the time evolution of the low-level MC, we used updraft helicity (UH; Kain et al. 2008). In this study, the low-level UH is calculated as

$$UH = \int_{0}^{2 \text{km}} \xi w \, dz.$$  \hspace{1cm} (A9)

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