Tornado genesis in a Quasi-Linear Convective System over Kanto Plain in Japan: A Numerical Case Study

EIGO TOCHIMOTO^{a,b} AND HIROSHI NIINO^b

^a National Research Institute for Earth Science and Disaster Resilience, Tsukuba, Japan

^b Atmosphere and Ocean Research Institute, The University of Tokyo, Kashiwa, Japan

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ABSTRACT: The environmental characteristics and formation process of a tornado spawned by a quasi-linear convective system (QLCS) over Kanto Plain, Japan, are examined using observations, a reanalysis dataset, and a high-resolution numerical simulation with a horizontal grid spacing of 50 m. The QLCS environment responsible for tornado genesis was characterized by small convective available potential energy and large storm-relative environmental helicity due to strong vertical shear associated with a low-level jet. The strong low-level jet was associated with a large zonal pressure gradient between two meridionally aligned extratropical cyclones and a synoptic-scale high pressure system to the east. The numerical simulation reproduced the tornado in the central part of the QLCS. Before the tornado genesis, three mesovortices developed that were meridionally aligned at 500-m height, and a rear inflow jet (RIJ) associated with relatively cold air originated from aloft and developed on the west side of the QLCS, while descending from rear to front. Tornado genesis occurred in the southernmost mesovortex at the northern tip of the RIJ. This mesovortex induces strong low-level updrafts through vertical pressure gradient force. A circulation analysis and vorticity budget analysis for the mesovortex show that environmental crosswise vorticity in the forward inflow region east of the QLCS played a significant role in the formation of the mesovortex. The circulation analysis for the tornado shows that frictional effects contribute to the increase of circulation associated with the tornado. Moreover, environmental shear associated with horizontal and vertical shear of the horizontal wind also contribute to the circulation of the tornado.

KEYWORDS: Convective storms/systems; Extratropical cyclones; Tornado genesis; Tornadoes; Convective storms

1. Introduction

A quasi-linear convective system (QLCS) is a type of organized convective system that produces torrential rain, straight-line winds, tornadoes, and hail. In the United States, ~18% of tornadoes are spawned by QLCSs (Trapp et al. 2005). It is noted that about half of the tornadoes in the morning between 0600 and 1200 LST are associated with QLCSs (Ashley et al. 2019), although a considerable fraction of tornadoes in QLCSs are EF0 or EF1 scale. The QLCS tornadoes tend to occur in environments with significantly lower convective available potential energy (CAPE) than the supercells tornadoes (Anderson-Frey et al. 2016), and are significant forecast challenge for issuing tornado warnings (Brotzge et al. 2013) since their formation processes are not well understood.

In Japan, a considerable number of tornadoes are also spawned by QLCSs (e.g., Kobayashi et al. 2007). Kobayashi et al. (2007) examined the characteristics of a QLCS tornado on 20 April 2006 and showed that the width and length of damage path for the tornado were 50 m and 2 km,

respectively. On 8 December 1992, two tornadoes occurred in the central part of a QLCS. One tornado in the north hit Chiyoda town in Ibaraki Prefecture, Japan, and destroyed 43 residential and 86 nonresidential buildings. The width and length of the damage path based on tornado database in Japan Meteorological Agency (JMA) were about 500 m and 9 km, respectively, which are the widest and longest among tornadoes cases associated with QLCSs in Japan. Thus, this historic case is of interest to understand the formation processes of these tornadoes spawned by the QLCS.

Niino et al. (1993) examined the observed characteristics of the two tornadoes and their parent storm on 8 December 1992 based on a damage survey, surface weather map, Doppler radar observation, and surface observations. They showed that the tornadoes occurred in association with a QLCS (Fig. 1c), and two mesoscale vortices were observed by a Doppler radar near the damage paths of the corresponding tornadoes (Fig. 1a). Since the radar was not in full operational mode, and had performed only three scans at the lower elevation angles around the time of the tornado genesis, they were unable to clarify the detailed relationship between the mesovortices and tornado genesis. A maximum surface wind of 35 m s^{-1} (Fig. 1b) with rapid change in wind direction from east to west was observed by an anemometer located at the southern end of the runway at Hyakuri Airport, Ibaraki

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Corresponding author: Eigo Tochimoto, tochimoto@aori.u-tokyo.ac.jp

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FIG. 1. (a) Locations of tornado damage (black solid circles) and mesovortices (red solid circles) detected by the MRI Doppler radar from Niino et al. (1993). (b) Time sequence of surface wind speed observed by an anemometer located at the southern end of the runway at Hyakuri Base [the location B in (a)] from Niino et al. (1993). (c) Reflectivity (dBZ) observed by JMA Mt. Fuji Radar located at (35°21'26"N, 138°43'50"E; red circle) at 0910:00 JST 8 Dec 1992. The numerals near the red solid circles in (a) show the observation times in JST and the point A and B show the locations of anemometer at the northern and southern end of the runway at Hyakuri Base, respectively.

Prefecture, where the southern tornado passed nearby. Note that the other anemometer at the northern end of the same runway, which is 1.9 km away, did not record such a gust (not shown). Thus, it is considered that the southern tornado passed nearby the southern anemometer of the runway. Although reported winds were relatively weak (JEF-0) and the damage path was narrow, the length of the damage exceeded 10 km (Fig. 1a). The

radar image shows that the tornadogenesis occurred in the central part of the QLCS, where reflectivity is large (Fig. 1c).

It has been pointed out that mesovortices with horizontal scale of 2–20 km (meso- γ -scale; Orlanski 1975) in QLCSs are often closely related to a tornadogenesis (e.g., Funk et al. 1999; Atkins et al. 2004; 2005; Flournoy and Coniglio 2019). Funk et al. (1999) showed that mesovortices were observed at

FIG. 2. (a) Domains for numerical simulations NHM20km (within the blue lines), NHM2km (within the red box), NHM350m (within the orange box), and NHM50m (within the black box). (b) View of the calculation domains for NHM50m with topographic map (m).

low levels and strengthened just prior to a tornado occurrence. Atkins et al. (2004) demonstrated that tornadic mesovortices were stronger, deeper, and longer-lasting than nontornadic mesovortices. Rear inflow jet (RIJ) is a typical QLCS structure that is closely related to the formation processes of mesovortices and tornadoes (e.g., Smull and Houze 1987; Weisman and Trapp 2003; Atkins et al. 2005; Wakimoto et al. 2006; Schenkman et al. 2012; Parker et al. 2020). Atkins et al. (2005) suggested that mesovortices tend to be tornadic when they were close to the RIJ possibly due to strong stretching of vertical vorticity.

The formation mechanism of mesovortices has been also investigated by previous studies using numerical simulations and observational data. Tilting of horizontal vorticity generated baroclinically by downdrafts in the RIJ region is shown to be a major source of the vertical vorticity of mesovortices (e.g., Atkins et al. 2005; Wakimoto et al. 2006; Weisman and Trapp 2003; Trapp and Weisman 2003; Flounoy and Coniglio 2019; Parker et al. 2020). In contrast, Xu et al. (2015) demonstrated that frictionally generated horizontal vorticity contributed to the intensification of a mesovortex. Atkins and Laurent (2009) showed that stronger vertical shear of environmental zonal wind leads to the stronger mesovortices. However, these previous studies did not focus on the formation process of tornadoes in a QLCS.

In recent years, development in computer technology has enabled high-resolution numerical simulations to investigate the detailed structure and formation process of tornadoes (Mashiko et al. 2009; Mashiko 2016a,b; Schenkman et al. 2012, 2014; Yokota et al. 2018). Schenkman et al. (2012) performed a high-resolution numerical simulation with a horizontal grid spacing of 100 m and reproduced tornado-like vortices formed in a line-end vortex at the northern end of a QLCS on 8–9 May 2007. They demonstrated that the tornadogenesis occurred near a gust front formed with the strong convection associated with the line-end vortex. They further showed that horizontal rotor intensified by frictional effects played a significant role in generating strong updrafts that led to tornadogenesis. Since a tornadogenesis in the present case occurred in the central part of a QLCS, the formation process of the tornado may differ from that associated with a line-end vortex of QLCSs. Although tornadogenesis also often occurs in mesovortices at the central part of QLCSs (e.g., Atkins et al. 2004), the formation processes of such tornadoes are not yet well understood. Furthermore, the formation mechanism of mesovortices has been suggested to rely on environmental conditions such as vertical shear and thermodynamic stability (e.g., Weisman and Trapp 2003; Wheatley and Trapp 2008). Since the formation mechanism of mesovortices and environmental conditions for QLCS tornadoes may

TABLE 1. Settings of each numerical simulation.

Domain	I. NHM20km	II. NHM2km	III. NHM350m	IV. NHM50m
Dimension (zonal × meridional × vertical)	200 × 200 × 40	500 × 500 × 50	1200 × 1200 × 70	1700 × 1700 × 100
Vertical grid spacing (m)	40–904 m	40–904 m	40–618 m	10–443 m
Model top (m)	22 244	22 244	22 244	22 244
Initial time	2100:00 JST 7 Dec	0300:00 JST 8 Dec	0600:00 JST 8 Dec	0750:00 JST 8 Dec

FIG. 3. (a) Surface weather map at 0900 JST 8 Dec 1992. (b) Skew T -log p diagram and hodograph observed at Tateno at 0900 JST 8 Dec 1992. (c) Temperature drop between 0800 and 0900 JST (contour lines); temperature ($^{\circ}\text{C}$) at the surface AMeDAS observation points of JMA at 0900 JST from Niino et al. (1993). Pennant, barbs, and half barbs show 10, 2, and 1 m s^{-1} , respectively. (d) Doppler velocity pattern observed by MRI radar (solid circle) near Tateno at 0912:54 JST, where the contour lines are drawn for each 4 m s^{-1} . The dash-dotted line shows the coast line of the Lake Kasumigaura.

differ between seasons (e.g., Wheatley and Trapp 2008), the formation processes of such tornadoes could be different between springtime and wintertime.

In the present study, we attempt to understand how the environment of the tornadoes that occurred over Kanto Plain, Japan, on 8 December 1992 was established through synoptic-scale to mesoscale processes, and to clarify the formation mechanism of the tornadoes in the central part of a QLCS using a high-resolution numerical simulation. This contrast with Schenkman et al. (2012) who studied a springtime tornado-like vortex at the northern end of a QLCS in the United States.

The remainder of this paper is organized as follows. Our methodology is described in section 2. Results are presented in section 3, and our findings are summarized and discussed in section 4.

2. Methodology

Data and experimental design

The Japanese 55-year Reanalysis (JRA-55; Kobayashi et al. 2015) with $1.25^{\circ} \times 1.25^{\circ}$ horizontal grid spacing, 37 vertical

FIG. 4. Mt. Fuji radar reflectivity map (dBZ) located at ($35^{\circ}21'26''\text{N}$, $138^{\circ}43'50''\text{E}$) at (a) 0810, (b) 0840, (c) 0910, (d) 0940 JST 8 Dec 1992. The brown dashed ellipse encloses the area of the QLCS, and the green circle encloses the QLCS segment that caused the two mesovortices, which spawned the tornadoes.

levels between 1000 and 1 hPa, and 6-hourly temporal resolution is used to perform a synoptic-scale analysis. JRA-55 covers the period between 1958 and the present.

Observational data used in this study are from the Meteorological Research Institute (MRI) C-band Doppler radar with wavelength of 5.703 cm at Tsukuba, Japan Meteorological Agency (JMA) operational radar at Mt. Fuji, upper-air soundings at Tateno, and the surface automated meteorological data acquisition system (AMeDAS) of the JMA. The radiosondes at Tateno are launched twice a day (2100 and 0900 JST). There are approximately 1300 stations of AMeDAS, which corresponds to horizontal density of one station per 400 km^2 . All 1300 stations measure precipitation amount every 10 min, and approximately 840 of them measure wind and temperature.

To examine mesoscale environments and the formation process of the tornadoes, we perform a numerical simulation

using the JMA nonhydrostatic model (JMANHM; Saito et al. 2006). The JMANHM is based on a finite-difference scheme with fully compressible equations including a map factor. Numerical simulations with JMANHM have been used successfully to study several cases of tornadoes and tornado-like vortices in Japan (Mashiko et al. 2009; Mashiko 2016a,b; Yokota et al. 2018; Tochimoto et al. 2019). We perform a quadrupled one-way nested numerical simulation. The outer three calculation domains have horizontal grid spacings of 20 km, 2 km, and 350 m (Fig. 2; Table 1), and the experiments using these domains will be called NHM20km, NHM2km, and NHM350m, respectively. The innermost calculation domain that resolves tornadoes uses a horizontal grid spacing of 50 m with 100 vertical levels (denoted as NHM50m) and is nested in NHM350m (Fig. 2), which provides initial and boundary conditions for NHM50m every 2 min. Initial and boundary conditions for NHM20km are

1200 JST 07 Dec. 1992

2100 JST 07 Dec. 1992

0900 JST 08 Dec. 1992

FIG. 5. (a) Potential vorticity (color shading; PVU) at 900 hPa and (b) specific humidity at 900 hPa (color shading; g kg^{-1}) at 1200 JST 7 Dec 1992. Contour lines in (a) and (b) indicate geopotential height (m) and southerly wind (m s^{-1}) exceeding 15 m s^{-1} , respectively. Arrows in (a) and (b) indicate horizontal wind (m s^{-1}) and water vapor flux ($\text{m g kg}^{-1} \text{ s}^{-1}$), respectively. (c),(d) As in (a) and (b), respectively, but at 2100 JST 7 Dec 1992. (e),(f) As in (a) and (b), respectively, but at 0900 JST 8 Dec 1992.

FIG. 6. (a) Potential vorticity (color shading; PVU) and (b) horizontal wind speed (color shading; m s^{-1}) at 250 hPa. (c),(e) As in (a), but at 2100 JST 7 Dec and 0900 JST 8 Dec 1992. (d),(f) As in (b), but at 2100 JST 7 Dec and 0900 JST 8 Dec 1992. Contour lines indicate geopotential height (m) and arrows horizontal wind vectors (m s^{-1}).

provided by 6-hourly JRA-55 data at a resolution of TL319L60 (~60-km horizontal resolution with 60 vertical levels).

The JMANHM in the present study adopts a bulk-type 6-class cloud microphysics scheme (Ikawa et al. 1991; Murakami 1990), which was developed based on the Lin microphysics scheme (Lin et al. 1983) and predicts mixing ratios

of water vapor, cloud water, rain, cloud ice, snow, and graupel. This scheme is one-moment for cloud water and rain and two-moment for cloud ice, snow, and graupel. The Mellor–Yamada–Nakanishi–Niino (MYNN) level-3 planetary boundary layer scheme (Nakanishi and Niino 2006) is adopted for NHM20km and NHM2km, and the Deardorff subgrid turbulence scheme (Deardorff 1980) is adopted for

0800 JST 08

FIG. 7. (a) Horizontal distribution of SREH ($\text{m}^2 \text{s}^{-2}$) and (b) MLCAPE (J kg^{-1}) simulated in NHM2km at 0800 JST 8 Dec 1992. The red circle in (a) indicates the sounding location of (c). Green and black contour lines in (a) and (b) respectively indicate rainwater mixing ratio (g kg^{-1} ; 1 g kg^{-1} interval) accompanied by the simulated QLCS at 500-m height. (c) Skew T -log p diagram and hodograph at 35.95°N , 140.15°E [red circle in (a)].

NHM350m and NHM50m. The Kain–Fritsch (Kain and Fritsch 1990; Kain 2004) cumulus convection scheme is used only for NHM20km and NHM2km. For the cumulus parameterization, the trigger function of convection, the reduction rate of CAPE, the autoconversion (into precipitation) in the updraft, and the time scale of convection were modified to make it suitable for mesoscale simulations (Ohmori and Yamada 2004, 2006; Saito et al. 2006, 2007). The surface flux of momentum, sensible heat, and moisture are parameterized by the bulk method (Beljaars and Holtslag 1991). A semislip surface condition, which is same as that in Mashiko et al. (2009), Mashiko (2016a,b), and Yokota et al. (2018), is used in the JMANHM.

3. Case overview

At 0900 JST 8 December 1992, two meridionally aligned extratropical cyclones (ECs) moved eastward over the Japanese islands (Fig. 3a): one EC was located over the Sea of Japan, and the other over Kanto region. The upper-air sounding at Tateno at 0900 JST (note that the sonde was launched at 0830 JST right before the passage of the QLCS.), Ibaraki prefecture (Fig. 3b) showed strong vertical shear and veering winds with the southerly wind exceeding 30 m s^{-1} at 1 km AGL. The storm relative environmental helicity between the surface and 1 km (SREH; Davies-Jones 1984; Davies-Jones et al. 1990) and mean-layer convective available potential energy (MLCAPE) were $440 \text{ m}^2 \text{s}^{-2}$ and 45 J kg^{-1} , respectively, where, in the

calculation of MLCAPE, the thermodynamic quantities of an initial parcel to be lifted were specified as the average over the lowest 500-m layer.

Surface observations from AMeDAS (Fig. 3c) indicate that a rapid temperature drop occurred between 0800 and 0900 JST due to the passage of the QLCS, with a maximum temperature change of $\sim -6^{\circ}\text{C}$, which is similar to the observations of QLCS cold pools in the United States (e.g., McDonald and Weiss 2021). In addition, wind directions changed from southerly to northerly across the region of the temperature change. The C-band Doppler radar at MRI in Tsukuba clearly detected two velocity couplets corresponding to mesovortices at 0912:54 JST (Fig. 3d). As was mentioned in the introduction, however, it was not possible to examine the detailed evolution of these two vortices and tornadoes since the Doppler radar data were only acquired for three scans at 0912, 0913, and 0915 JST at the elevation angle of 0.6° , 1.0° , and 0.1° , respectively.

Figure 4 shows the evolution of radar reflectivity by Mt. Fuji operational radar of JMA. At 0810 JST, a linear precipitation system (hereafter referred to as QLCS) extending from southwest to northeast (brown dashed ellipse in Fig. 4a; 35° – 36.5°N , 138° – 140°E) appeared over Shizuoka Prefecture, and propagated eastward (Fig. 4b), reaching Kanto Plain at 0910 JST (Fig. 4c; 35° – 37°N , 139° – 141°E). In the central part of the QLCS, a region of high reflectivity protruded south-eastward (the dashed green circle in Fig. 4c). The tornado formed near the region of high reflectivity at 0913 JST. The QLCS continued to move eastward, and the region of high reflectivity disappeared by 0940 JST (Fig. 4h).

4. Synoptic-scale situation

To understand how the synoptic-scale environment for the tornadic QLCS was established, we examine the evolution of low-level and upper-level fields. At 1200 JST 7 December 1992, a synoptic-scale high pressure system was located over the Northwestern Pacific (30° – 35°N , 150°E ; Fig. 5a). There were low-level southerly winds over 15 m s^{-1} at the western periphery of the high pressure system (Figs. 5a,b). To the south of the Kyushu Islands (25° – 30°N , 130° – 135°E), there was a region with abundant specific humidity exceeding 10 g kg^{-1} (Fig. 5b) together with high positive potential vorticity (PV; 0.8 – 0.9 PVU ; $1\text{ PVU} = 1.0 \times 10^{-6}\text{ m}^2\text{ s}^{-1}\text{ K kg}^{-1}$) at low levels (Fig. 5a). At 2100 JST 7 December, low-level PV in this region with strong low-level southerly winds (exceeding 20 m s^{-1}) exceeded 1 PVU (Figs. 5c,d) and a developing synoptic-scale EC moved over the Sea of Japan. The abundant low-level water vapor (9 – 10 g kg^{-1}) reached the Shikoku Islands (33° – 34°N , 132° – 134°E) and extended farther north. By 0900 JST 8 December, the region with low-level high PV and abundant low-level water vapor moved eastward over the Kanto region (34° – 36°N , 139° – 141°E ; Figs. 5e,h). The low-level PV maximum corresponds to a developing EC over the Kanto region, as seen in Fig. 3a. Another EC in the Sea of Japan also developed and was located to the west of Hokkaido (40° – 45°N , 133° – 140°E). The zonal pressure gradient associated with these two ECs and the high pressure system was enhanced, and thus

FIG. 8. Horizontal distribution of surface temperature (K) and horizontal wind vectors (m s^{-1}) simulated in NHM2km at 0800 JST 8 Dec 1992. Contour lines indicate vertical velocity (m s^{-1}) at 500-m height, where solid and dashed lines show positive and negative values, respectively.

the southerly wind east of the Japan Islands strengthened (exceeding 30 m s^{-1}). It is suggested that these two ECs contributed to the enhancement of the low-level jet, resulting in the large northward transport of low-level moisture and enhancement of the vertical shear of the horizontal wind. Although it is well known that tornadogenesis often occurs in the warm sector of ECs (e.g., Newton 1967; Johns and Doswell 1992; Tochimoto and Niino 2016, 2017, 2018), few studies have examined the environment of tornadogenesis associated with two ECs aligned meridionally (e.g., Niino et al. 1993), which are known as “Futatsudama cyclones” in Japanese.

Figure 6 shows the time series of upper-level PV and horizontal winds at 250 hPa. A region of high PV exceeding 5 PVU was located over east and northwest China (30° – 40°N , 110° – 115°E) at 1200 JST 7 December 1992, and propagated eastward, reaching west of the Kanto Plain at 0900 JST 8 December 1992 (Figs. 6a,c,e). The upper-level jet stream accompanying the trough with high PV extended from China to Kyushu Island at 1200 JST 7 December, and was moving eastward (Figs. 6b,d). At 0900 JST 8 December, upper-level winds exceeding 80 m s^{-1} extended from the East China Sea to the south of the Japan islands (Fig. 6f). The approaching upper-level jet streak enhanced the vertical shear of horizontal winds over the Kanto Plain. Since the Kanto Plain was located at the left exit region of the jet streak and east of the upper-level high-PV region, it is likely that there was dynamically forced large-scale ascent there.

5. Mesoscale situation

Here we examine the mesoscale environments reproduced by NHM2km over the Kanto Plain at 0800 JST 8 December

FIG. 9. Horizontal distribution of reflectivity (dBZ) at 500-m height simulated in NHM350m at (a) 0716, (b) 0736, (c) 0756, and (d) 0816 JST. The thick dashed circle shows the region where a part of the QLCS sticks out southeastward.

(Fig. 7). NHM2km reproduced the QLCS over the Kanto Plain reasonably well (Fig. 7), although the simulated QLCS developed about an hour earlier than observed. Values of SREH calculated between 0 and 1 km exceeding $400 \text{ m}^2 \text{ s}^{-2}$ were found in the forward inflow region east of the QLCS at 0800 JST (Fig. 7a). These values of SREH are consistent with the observed value of $440 \text{ m}^2 \text{ s}^{-2}$ at Tateno (see section 3a). In this region, veering wind and the strong low-level jet exceeding 30 m s^{-1} at 1 km height (Fig. 3c) were also well reproduced, resulting in the strong directional vertical shear and large values of SREH. Although values of MLCAPE exceeding 700 J kg^{-1} were found to the southwest of the QLCS (34° – 35°N , 138° – 139.5°E ; Fig. 7b), the values near the

QLCS are only 50 – 100 J kg^{-1} , which again compares well with the observed value of 44 J kg^{-1} at 0900 JST at Tateno station (see also section 3a). Comparison of skew T - $\log p$ diagram and hodograph near the simulated QLCS (Fig. 7c) shows that thermodynamic and kinematic profiles were also reasonably well reproduced in NHM2km, although near surface temperature and upper-level humidity in NHM2km were warmer and drier than those in the sounding (Fig. 3b).

Figure 8 shows the horizontal distribution of temperature near the surface. Temperature to the east of the QLCS in forward inflow region is 1 – 2 K warmer than that to the west of the QLCS. The simulated difference of the potential temperature across the QLCS was considerably smaller than that

FIG. 10. Horizontal distribution of (a) rainwater mixing ratio (g kg^{-1}) and (b) vertical vorticity (10^{-2} s^{-1}) at 500-m height simulated in NHM50m at 0813:30 JST. Contour lines and vectors in (b) indicate pressure anomaly and horizontal wind, respectively, and M_A , M_B , and M_C in (b) are mesovortices. (c) Vertical vorticity (s^{-1}) at 30-m height and sea level pressure (hPa). (d) Vertical cross section of vertical vorticity and pressure anomaly along A–A'.

observed ($\sim 5 \text{ K}$; Fig. 3c) and previous studies in the United States (e.g., Flournoy and Coniglio 2019; Parker et al. 2020).

NHM350m successfully reproduced the QLCS. At 0716 JST, the simulated QLCS extended from southwest to northeast ($35.5^\circ\text{--}36^\circ\text{N}$, $139.2^\circ\text{--}139.7^\circ\text{E}$). The simulated QLCS propagated rapidly east–northeastward and started to exhibit a line-echo wave pattern (e.g., Nolen 1959; Fujita 1978) at 0736 and 0756 JST (Figs. 9b,c). At 0816 JST, the simulated QLCS was located over Kanto Plain and its central region started to stick out to the southeast (black dashed circle in Fig. 9), showing characteristics similar to the observations (red dashed circle in Fig. 4g).

6. Tornado formation process

a. Characteristics of the simulated tornado and its parent quasi-linear convective system

NHM50m successfully reproduced a tornado associated with the QLCS. Here we first show the characteristics of the tornado and then its relation to the QLCS. The distribution of rainwater mixing ratio at 500-m height displays a hook-like

pattern at 0813:30 JST (Fig. 10a). At this time, there were three vortices along the QLCS (Fig. 10b). Since the horizontal scale of these vortices is about 1–2 km, they are denoted as “mesovortex A (M_A),” “mesovortex B (M_B),” and “mesovortex C (M_C)” from north to south. M_B had a structure much deeper than 1 km and may be described as a mesocyclone, but did not spawn a vortex of tornado strength near the surface. The depth of M_C was about 1 km. Tornado genesis near the surface occurred in M_C near the hook-echo region (Figs. 10a,c), where a tornado is defined by a vortex that has vertical vorticity larger than 0.4 s^{-1} near the surface (15 or 30-m height). The tornado lasted more than 7 min, which satisfies the definition of a tornado (100–1000 s; AMS glossary). The simulated vertical vorticity of the tornado exceeded 0.6 s^{-1} and the associated minimum sea level pressure was $<998 \text{ hPa}$ at 0813:30 JST (Fig. 10c). At this time, the region of strong vertical vorticity exceeding 0.05 s^{-1} extended from the surface to about 1 km AGL. The maxima of vertical vorticity near the surface and of horizontal wind speed over the NHM50m domain exceeded 0.7 s^{-1} and 50 m s^{-1} , respectively, at 0817:30 JST (Fig. 11a). The tornado continued to exist but eventually weakened by $\sim 0820 \text{ JST}$. There are

FIG. 11. (a) Time series of maxima of vertical vorticity and horizontal wind speed at 30-m height. (b) The simulated track of the tornado by NHM50m (blue line) and observed tornado damage paths (red lines). The green square and red circle indicate the location of the simulated tornado at 0812:40 and 0817:40 JST, respectively. Color shading indicates height of topography (m), where green colors denote lake areas. (c) Model-derived Doppler velocity (m s^{-1}) at 1.0° elevation angle. The green circle in (c) indicates the assumed location of Doppler radar.

additional peaks of the maximum vertical vorticity and horizontal velocity at 0832:20 JST. They correspond to the other tornado-genesis over Lake Kasumigaura (not shown). However, we will focus on the tornado that formed and developed over land between 0813 and 0820 JST.

The tracks of the simulated tornado and observed tornado damage paths are shown in Fig. 11b. Note that the location of the tornado is defined as the maximum of vertical vorticity at 30-m height within a $500 \text{ m} \times 500 \text{ m}$ square around the M_C center, which is in turn defined as the maximum of vertical

vorticity averaged over a $5 \text{ km} \times 5 \text{ km}$ square at 500-m height. The direction of movement of the simulated tornado is consistent with that of the observed tornado damage path. The path of the simulated tornado was about 20 km south of the observed damage path, although this discrepancy might be expected unless radar data assimilation is employed (e.g., Schenkman et al. 2014; Yokota et al. 2016, 2018). The simulated Doppler velocity pattern was also reasonably reproduced in NHM50m (Fig. 11c). Although the present simulated tornado is shallower than other simulated supercell tornadoes (e.g.,

FIG. 12. Time–height plot of (a) maximum vertical vorticity (s^{-1}), (b) minimum pressure anomaly (hPa), and (c) maximum vertical velocity (m s^{-1}) within $500 \text{ m} \times 500 \text{ m}$ square around the maximum of vertical vorticity smoothed by taking average over $500 \text{ m} \times 500 \text{ m}$ square at 500-m height.

Schenkman et al. 2014; Mashiko 2016a,b), the depth of the tornado is comparable to the minisupercell tornadoes spawned by tropical cyclones (about 1 km depth; e.g., Mashiko et al. 2009). Moreover, the minimum pressure anomaly of $\sim -12 \text{ hPa}$ (Fig. 12b) for the simulated near-surface vortex is comparable with that of -12.7 hPa for a QLCS tornado reproduced in Schenkman et al. (2012). Thus, we think that the simulated vortex satisfies the definition of a tornado.

b. Temporal evolution

To examine how the tornadogenesis occurs, we show time–height plots of maximum vertical velocity, minimum pressure anomaly, and maximum updrafts within a $500 \text{ m} \times 500 \text{ m}$ square around the M_C center (Fig. 12). The tornadogenesis in the simulation occurred between 0812 and 0813 JST. Between 0810 and 0811 JST, immediately before the tornadogenesis, the updraft intensified between 600- and 1200-m height, then the maximum vertical vorticity (about $0.2\text{--}0.3 \text{ s}^{-1}$) and updrafts between 300 and 500 m, which corresponds to M_C , were enhanced. This enhancement was also accompanied by a pressure drop. Thus, it appears that updrafts between 300 and

500 m were accelerated by the vertical perturbation pressure gradient force (VPPGF) associated with the strong vertical vorticity of M_C . A more detailed analysis will be given in the next subsection. The vertical vorticity associated with the tornado intensified near the surface at 0812:40 JST, and exceeded 0.7 s^{-1} , and maximum vertical vorticity greater than 0.3 s^{-1} extended from near the surface to 600 m height at 0813:30 JST. To examine the formation process of the tornado, we focus particularly on the period between 0809:30 and 0813:30 JST, although a higher peak of vertical vorticity is observed at 0817:30 JST.

Figure 13 shows the evolution of horizontal distribution of rainwater mixing ratio, vertical velocity, and vertical vorticity at 500-m height across three output times. At 0809:30 JST, the hook-shaped precipitation pattern had not yet developed. In region A of Fig. 13b, there were strong updrafts ($>25 \text{ m s}^{-1}$) and large vertical vorticity ($>0.2 \text{ s}^{-1}$; Fig. 13c) associated with a vortex, M_A , which was defined in section 3a (Fig. 10b). At 0811:30 JST, a hook-shaped precipitation pattern started to appear in the southern part of region B (Fig. 13d). Updrafts in region B (Fig. 13e) intensified with two maxima exceeding 20 m s^{-1} (Fig. 13e). Corresponding to these two updraft maxima, two mesovortices, M_B and M_C as defined in section 3a (Fig. 10b), developed (Fig. 13f). Tornadogenesis near the surface occurred below M_C . At 0813:30 JST, a bounded weak echo region appeared in the region where M_C and its associated updrafts developed (Figs. 13g–i).

To examine the relationship between the acceleration of the updrafts and the mesovortices, we diagnose terms in the vertical momentum equation. The Boussinesq version of the vertical momentum equation may be written as follows:

$$\frac{dw}{dt} = -\frac{1}{\bar{\rho}} \frac{\partial p'}{\partial z} - \frac{1}{\bar{\rho}} \rho' g,$$

where w is vertical velocity, g is the acceleration due to gravity, ρ' is the density deviation from $\bar{\rho}$ ($\bar{\rho}$ is horizontal average over a $5 \text{ km} \times 5 \text{ km}$ square), and p' is the deviation from the basic-state pressure calculated from a horizontal average over a $5 \text{ km} \times 5 \text{ km}$ square at each grid point. The first term on the right-hand side represents VPPGF and the second term represents buoyancy. Figures 14a and 14b show the contribution of VPPGF to the acceleration of updrafts at 400-m height. At 0810:30 JST, the VPPGF term was positive near M_B , which contributed to the acceleration of updrafts (Fig. 14a). Near M_C , however, the VPPGF term was only slightly positive since the M_C had not fully developed at this time. The VPPGF term became more clearly positive at 0811:30 JST as M_C intensified (Fig. 14b). In contrast, the buoyancy term was everywhere less than 0.02 m s^{-2} and was notably smaller than the VPPGF term (not shown). Cross section of VPPGF along A–A' in Fig. 14b shows that the positive VPPGF term extended to near the surface (Fig. 14c). Thus, the development of the southern M_C at 500-m height induced strong VPPGF, resulting in strong updrafts below.

We next examine the tornadogenesis process by looking at the evolution of meteorological variables near the surface (Fig. 15). At 0811:30 JST, the tornado had not yet appeared

FIG. 13. Horizontal distributions of (a) rainwater mixing ratio (g kg^{-1}), (b) vertical velocity (m s^{-1}), and (c) vertical vorticity (10^{-2} s^{-1}) at 500-m height at 0809:30 JST. (d)–(f), (g)–(i) As in (a)–(c), but at 0811:30 and 0813:30 JST, respectively. Arrows in (b), (c), (e), (f), (h), and (i) show horizontal wind vectors at 500-m height. Contour lines in (c), (f), and (i) indicate pressure anomaly (hPa).

near the surface, although large vorticity exceeding 0.1 s^{-1} was found in the region of large horizontal wind shear (Fig. 15a). Around this region, an RIJ associated with relatively cold air (1–2 K colder than that in the forward inflow region) started to intrude from the west of the QLCS (Fig. 15b). At 0812:30 JST, vertical vorticity in this region exceeded 0.3 s^{-1} (Fig. 15d), which corresponds to the tornado. Tornadogenesis occurred at the northern tip of the RIJ associated with cold air (Fig. 15e). At this time, the updraft exceeded 4 m s^{-1} near the surface (Fig. 15f) and the maximum updraft above 300-m height was more than 17 m s^{-1} (Fig. 12c), suggesting large stretching of vertical vorticity. Subsequently, the tornado further developed with a notable pressure decrease (Fig. 15g): the minimum perturbation of pressure near the surface exceeded at least 10 hPa at 0813:30 JST (Fig. 12b). At this time, the maximum updraft above 300-m height exceeded 20 m s^{-1} (Fig. 12c). The RIJ further strengthened and strong updrafts in the tornadic region were maintained (Figs. 15h,i). Cross section of vertical velocity (along A–A' in Fig. 15e) in the RIJ region (Fig. 16) indicates that strong

downrafts extend from about 2 km height to near the surface and strong updrafts occurred at the eastern edge of the RIJ.

The relationship between the development of the RIJ and horizontal shear is shown in Figs. 17a–d. Since the strong zonal wind is responsible for the RIJ, the evolution of meridional shear of zonal wind ($-du/dy$) is examined. Between 0807:30 and 0813:30 JST, $-du/dy$ increased as the RIJ indicated by transparent green colors in the westerly develops. Thus, the RIJ enhanced the horizontal wind shear, resulting in stronger cyclonic circulation (Figs. 17a,b). To examine this effect, the circulation along a circle of 1-km radius centered at each grid point is shown in Figs. 17e–h. A region with cyclonic circulation exceeding $20000 \text{ m}^2 \text{ s}^{-1}$ exists in the northern half of the RIJ at 0807:30 JST (black dashed circle in Fig. 17e). Two minutes later (0809:30 JST; Fig. 17f), the cyclonic circulation strengthens and is maintained on the north boundary of the RIJ region. At that time, there were also strong southerly winds in the forward inflow region. At 0811:30 and 0813:30 JST, the cyclonic

FIG. 14. Horizontal distributions of vertical pressure perturbation gradient force (color shading; m s^{-2}) term at 400-m height at (a) 08:10:30 and (b) 08:11:30 JST. Contour lines indicate vertical velocity (m s^{-1}), and arrows horizontal wind vectors (m s^{-1}). (c) Vertical cross section of vertical pressure perturbation gradient force (color shading; m s^{-2}) and vertical velocity (contour lines; m s^{-1}) along A–A' in (b).

circulation further intensified with the enhancement of the RIJ and its associated horizontal shear (Figs. 17c,d), and the tornado was generated (Figs. 17g,h). Thus, strong westerly winds in the RIJ region contributed to the strong cyclonic shear around the region of tornadogenesis. These results are consistent with those of the observational studies in the United States (e.g., Pfost and Gerard 1997; Funk et al. 1999; Atkins et al. 2004, 2005).

c. Source of rotation for the mesovortex and the tornado

To understand the source of rotation in M_C and the tornado, we examine the evolution of circulation (C):

$$C = \oint \mathbf{v} \cdot d\mathbf{l}, \quad (1)$$

along a material circuit, where $\mathbf{v}$ is the wind vector, and $d\mathbf{l}$ is the line element vector. A circuit consisting of 1600 parcels was placed around a region of strong vertical vorticity associated with M_C and the tornado at 08:09:20 and 08:12:40 JST, respectively (Fig. 18a and 20a), and then traced backward in time. The circulation analysis adopted in the present study is similar to that in previous studies (e.g., Mashiko 2016a,b;

Yokota et al. 2018; Tochimoto et al. 2019). If adjacent parcels on the circuit become separated by more than 50 m, a new parcel is added to the circuit at the midpoint between the two adjacent parcels. Backward trajectories were calculated using a fourth-order Runge–Kutta scheme with a time step of 0.4 s. The model output every 0.8 s was used to calculate backward trajectories since the circulation analysis requires a velocity field with fine spatiotemporal resolution to reduce numerical errors (Dahl et al. 2012). A logarithmic wind profile was assumed at heights below $z^* = 5$ m which is the lowest model level. The tendency of the circulation can be written as

$$\frac{d}{dt} C(t) = - \oint \frac{dp}{\rho} + \oint \mathbf{F} \cdot d\mathbf{l}, \quad (2)$$

where ρ is density, p is pressure, $\mathbf{F}$ is the friction vector, and $d\mathbf{l}$ is the line element vector. The first term on the rhs is the baroclinic term and the second term is the friction term. In the present study, the value of $\mathbf{F}$ below the lowest model height was assumed to have the same value as that at the lowest model level. The effects of these two terms were directly calculated using the 0.8-s output. To check the accuracy of the calculation of the circulation, the evolution of the circulation

FIG. 15. Horizontal distributions of (a) vertical vorticity (10^{-2} s^{-1}), (b) potential temperature, and (c) vertical velocity (m s^{-1}) at 30-m height at 0811:30 JST. (d)–(f), (g)–(i) As in (a)–(c), but at 0812:30 and 0813:30 JST, respectively. Arrows show horizontal wind vectors (m s^{-1}) at 30-m height, contour lines indicate pressure anomaly with 2-hPa interval (hPa). The minimum perturbation pressure is approximately 10.7 hPa at the center of the tornado at 0813:30 JST.

calculated along the circuit from Eq. (1) will be compared with that calculated by integrating the sum of the rhs of Eq. (2).

1) ORIGIN OF ROTATION IN THE MESOVORTEX

The initial material circuit used for the backward trajectory analysis and evolution of circulation for M_C is shown in Fig. 18a. The initial circuit surrounding the region with strong vertical vorticity associated with the M_C was set at 500-m height. Figures 18b and 18c show the circuits at 0809:20 and 0806:20 JST, respectively, obtained from the backward trajectory analysis. Most of the parcels constituting the circuit came through low levels (below 100-m height) from the warmer side of the QLCS. The evolution of the circulation (Fig. 18d) shows that the circulation at 0803:20 JST was $45000 \text{ m}^2 \text{ s}^{-1}$ and decreased gradually with time. Since the circulation estimated

by integration of the friction and baroclinic terms reasonably agrees with that directly calculated using Eq. (1) at each time, we believe that the results of the circulation analysis are reliable. The contribution of the baroclinic term was negligibly small, and that of the friction term was responsible for the decrease in circulation. However, during the 8 min for which the circulation analysis was made, the circulation decreased by only 12% by 0811:20 JST.

We have seen that most parcels in the circuit at 0806:20 and 0809:20 JST were located in the forward inflow region east of the QLCS (Figs. 18b,c). While most of the parcels were located below 40-m height, the parcels in the southeastern part of the circuit were at 100–200-m height. In the inflow region, crosswise horizontal vorticity of environments associated with vertical shear of the horizontal winds was notably large. The contribution of horizontal vorticity in a

FIG. 16. Cross section of vertical velocity (color shading; m s^{-1}) along A–A' line in Fig. 15e. Green arrows indicate wind vectors projected onto the figure. Note that vertical component of vectors were multiplied by 20 to emphasize the vertical motions.

forward inflow region to the mesovortices in QLCSs is also suggested in previous studies (e.g., Wheatley and Trapp 2008; Flournoy and Coniglio 2019). Because the contribution of frictional generation to overall circulation is relatively small, we will now proceed to diagnose terms in the time tendency equation of the vertical vorticity to quantify the relative contributions of nonfrictional vorticity sources including crosswise horizontal vorticity.

To obtain additional information on the origin of the rotation in M_C , we perform a vorticity budget analysis in the Lagrangian frame (Lilly 1982; Adlerman et al. 1999; Mashiko et al. 2009; Schenkman et al. 2014; Roberts et al. 2016; Roberts and Xue 2017). Vorticity budget equations for streamwise vorticity, crosswise vorticity, and vertical vorticity may be written as follows:

$$\frac{D\omega_s}{Dt} = -\omega_s \left(V_H \frac{\partial \psi}{\partial n} + \frac{\partial w}{\partial z} \right) + \omega_n \frac{\partial V_H}{\partial n} + \zeta \frac{\partial V_H}{\partial z} + \frac{1}{\rho^2} \left(\frac{\partial p}{\partial n} \frac{\partial p}{\partial z} - \frac{\partial p}{\partial z} \frac{\partial p}{\partial n} \right) + \frac{1}{\rho} \left(\frac{\partial F_z}{\partial n} - \frac{\partial F_n}{\partial z} - F_s \frac{\partial \psi}{\partial z} \right) + \omega_n \frac{D\psi}{Dt}, \quad (3a)$$

$$\frac{D\omega_n}{Dt} = -\omega_n \left(\frac{\partial V_H}{\partial s} + \frac{\partial w}{\partial z} \right) + \omega_s V_H \frac{\partial \psi}{\partial s} + \zeta V_H \frac{\partial \psi}{\partial z} + \frac{1}{\rho^2} \left(\frac{\partial p}{\partial z} \frac{\partial p}{\partial s} - \frac{\partial p}{\partial s} \frac{\partial p}{\partial z} \right) + \frac{1}{\rho} \left(\frac{\partial F_s}{\partial z} - \frac{\partial F_z}{\partial s} - F_n \frac{\partial \psi}{\partial z} \right) - \omega_s \frac{D\psi}{Dt}, \quad (3b)$$

$$\frac{D\zeta}{Dt} = -\zeta \left(\frac{\partial V_H}{\partial s} + V_H \frac{\partial \psi}{\partial n} \right) + \omega_s \frac{\partial w}{\partial s} + \omega_n \frac{\partial w}{\partial n} + \frac{1}{\rho^2} \left(\frac{\partial p}{\partial s} \frac{\partial p}{\partial n} - \frac{\partial p}{\partial n} \frac{\partial p}{\partial s} \right) + \frac{1}{\rho} \left(\frac{\partial F_n}{\partial s} - \frac{\partial F_s}{\partial n} - F_s \frac{\partial \psi}{\partial s} + F_n \frac{\partial \psi}{\partial n} \right), \quad (3c)$$

where the subscripts s , n , and z denote streamwise, crosswise (normal), and vertical directions, respectively; V_H is the magnitude of the ground-relative horizontal wind; $\Psi = \tan^{-1}(v/u)$ is the direction of the ground-relative wind; and $\mathbf{F} = (F_s, F_n, F_z)$ is the frictional force due to the streamwise, crosswise, and vertical components of subgrid turbulence, respectively. The first term on the rhs of each

equation represents convergence. The second and third terms represent tilting. The fourth term represents baroclinic generation. The fifth term is frictional generation. The sixth term in the equations of streamwise and crosswise vorticity represents exchange between streamwise and crosswise vorticity.

We first perform a backward trajectory analysis for 100 parcels (blue and green dots in Fig. 18a), which were distributed over 10×10 grids with a 30-m interval between points around the region with strong vertical vorticity. Most of the parcels originated from the forward inflow region (Fig. 19a). Here we select a representative parcel (green dot in Fig. 18a), in which vorticity components were accurately estimated by integration of budget terms with respect to those directly calculated at parcel locations at each time. The evolutions of vorticity components for the selected parcel are shown in Fig. 19b. The selected parcel was located at the forward inflow region at 0801:20 JST (Fig. 19a) when the QLCS was located around 140.16°E . It moved northwestward, turned northward and then northeastward while ascending. Finally, the parcel reached the region of relatively high vertical vorticity in M_C (Fig. 19a). At 0801:20 JST, the parcel had large positive crosswise ($\sim 0.14 \text{ s}^{-1}$) and streamwise ($\sim 0.04 \text{ s}^{-1}$) vorticities (Fig. 19b). Between 0808:40 and 0811:20 JST, the horizontal vorticities decreased rapidly, while the vertical vorticity increased rapidly. The time series of terms in the vertical vorticity equation shows that the tilting of crosswise vorticity was the main contributor to vertical vorticity between 0808:40 and 0809:40 JST (Fig. 19c). Then, the stretching of vertical vorticity rapidly intensified the vertical vorticity. Since the parcel originated from the forward inflow region and had large crosswise vorticity at 0801:20 JST, this vorticity budget analysis also confirms that vertical shear of horizontal wind in the forward inflow region played a significant role in enhancing vertical vorticity. Other parcels in which vorticities were accurately estimated by integration of budget terms with respect to vorticities directly calculated at parcel locations at each time also showed qualitatively similar results (not shown).

2) ORIGIN OF ROTATION IN THE TORNADO

To examine the origin of the rotation in the tornado, an initial circuit enclosing the region of large vertical vorticity of about 0.1 s^{-1} in the tornado was set at $z = 100 \text{ m}$ (Fig. 20a) at 0812:40 JST, and the backward Lagrangian circulation analysis was performed. Material circuits at 0810:40 and 0807:40 JST are shown in Figs. 20b and 20c, respectively. The location of the circuit shows that parcels surrounding the tornado came from both the rear and forward inflow regions. Figure 20d shows the evolution of the circulation along the circuit. Since the circulation estimated by integrating the friction and baroclinic terms (hereafter referred to as integrated circulation) with time resulted in (quantitatively) reasonable agreement with the circulation directly calculated along the circuit at each time (the difference is less than 10% between 0803:40 and 0812:40 JST), we believe that the results of the circulation analysis are reliable. At 0805:40 JST, the circulation was about $15000 \text{ m}^2 \text{ s}^{-2}$,

FIG. 17. Horizontal distributions of meridional gradient of zonal wind (color shading; 10^{-2} s^{-1}), horizontal wind speed (green contour lines; m s^{-1}), and horizontal wind vectors (arrows; m s^{-1}) at (a) 0807:30, (b) 0809:30, (c) 0811:30, and (d) 0813:30 JST. Transparent green indicates the regions where horizontal wind speed exceed 21 m s^{-1} . (e)–(h) As in (a)–(d), but for circulation ($\text{m}^2 \text{ s}^{-1}$) computed around 1-km-radius rings centered on each grid point (color shading) and sea level pressure (hPa) at 30-m height (contour lines). The center of dashed black circle corresponds to a local maximum of the circulation.

FIG. 18. Backward trajectory analysis for examining the origin of rotation in M_C . (a) Initial circuit (thick black line) enclosing the region of strong vertical vorticity in M_C at 500-m height at 0811:20 JST. (b) Horizontal projection of the circuit at 0809:20 JST. (c) As in (b), but at 0806:20 JST. Color shading, contour lines, and arrows in (a) show the vertical vorticity (s^{-1}), vertical velocity ($m s^{-1}$), and horizontal wind vectors ($m s^{-1}$), respectively. Blue dots and a green dot in (a) indicate initial positions of backward trajectories. The green dot shows a selected representative parcel. Color shading, black vectors, and light green vectors in (b) and (c) indicate height of parcels (m), horizontal wind ($m s^{-1}$), and horizontal vorticity vectors (s^{-1}) at 50-m height, respectively. (d) Time series of circulation (black line; $m^2 s^{-1}$), baroclinicity (blue line; $m^2 s^{-2}$) and frictional (yellow line; $m^2 s^{-2}$) vorticity generation terms, and circulation estimated by temporally integrating these two terms (orange line; $m^2 s^{-2}$).

which is 56% of the circulation enclosing the strong vorticity of the tornado at 0812:40 JST (Fig. 20b). The circulation increased between 0805:40 and 0811:00 JST, where this increase is mainly contributed by the friction term, but little by the baroclinic term.

Wind fields between the RIJ and forward inflow were characterized by both convergence and strong cyclonic horizontal shear, the latter of which is likely to have contributed to the circulation originally associated with the circuit. Horizontal convergence and vertical vorticity at 30-m height at 0810:40 and 0807:40 JST are shown in Figs. 20b and 20c, respectively. Strong horizontal convergence is found in the region across the RIJ and forward inflow region, and strong vertical vorticity associated with horizontal shear is found in the RIJ region. If we go back further before 0802:40 JST, the difference between the integrated circulation and the actual circulation

along the circuit at each time becomes large, making it difficult to reliably evaluate the circulation.

7. Summary and discussion

In the present study, the synoptic and mesoscale environments of a tornado spawned by a quasi-linear convective system (QLCS) over Kanto Plain, Japan on 8 December 1992 and its formation process have been examined by analysis of observations and a reanalysis dataset, and a high-resolution numerical simulation. At 0900 JST 8 December 1992, two meridionally aligned extratropical cyclones, one in the Sea of Japan and the other over Kanto Plain, were moving eastward. In the upper troposphere, a high-PV anomaly was located to the west of Kanto Plain. A strong southerly low-level jet exceeding $30 m s^{-1}$ extended over the southeast part of Kanto

FIG. 19. Backward trajectory analysis of a selected representative parcel for examining origin of rotation in M_C . (a) Trajectories of parcels as shown by blue solid circles in Fig. 18a and of the representative parcel as shown by green solid circles in Fig. 18a together with horizontal wind vectors (green arrows; m s^{-1}) at 500-m height at 0811:20 JST. (b) Time series of vertical vorticity (red line; s^{-1}), streamwise vorticity (orange line; s^{-1}), and crosswise vorticity (blue line; s^{-1}). Dashed lines indicate estimated vorticity (s^{-1}) by integrating the rhs of terms in Eqs. (2)–(4) for every time step. (c) Time series of terms in vertical vorticity equation (s^{-2}): blue, black, orange, yellow, and light blue lines indicate stretching, tilting of streamwise vorticity, tilting of crosswise vorticity, baroclinicity, and friction terms, respectively. The color in the trajectory in (a) shows the height. Red dots in (a) represent the position of representative parcels every 1 min.

Plain, resulting in large vertical shear of horizontal wind. In addition, the low-level jet brought abundant low-level moisture over Kanto Plain. Yamamoto (2012) studied a case of two meridionally aligned ECs in which enhanced low-level moisture transport caused heavy rainfall in Japan. Thus, special attention needs to be paid to meridionally aligned EC pairs, which are observed at a rate of 15 pairs per year (Hirata and Kusaka 2013), since they have a potential to provide favorable environments for extreme events in Japan.

A high-resolution numerical simulation with 50-m horizontal grid spacing using JMANHM has been performed to elucidate the formation process of the tornado. The numerical simulation successfully reproduced the QLCS, as observed by the JMA radar and an associated tornado, although the tornadogenesis occurred 60 min earlier and 20 km farther south than in the observations. Such a difference is also found for previous high-resolution simulations of specific tornadoes

(e.g., Dawson et al. 2015). At the early stage, three mesovortices M_A , M_B , and M_C developed along the QLCS. As the rear inflow jet (RIJ) developed, tornadogenesis occurred at the northern tip of the RIJ beneath the M_C . The circulation of the M_C was strongest at 500-m height and was associated with strong updrafts that contribute to the formation and development of the tornado.

A circulation analysis and vorticity budget analysis indicate that rotation of the M_C mainly originated from the horizontal vorticity of the environmental vertical shear in the forward inflow region. Crosswise horizontal vorticity associated with the environmental vertical shear was tilted into vertical vorticity, which in turn was amplified by stretching. The vertical shear in the forward inflow region appears to be related to the southerly low-level jet associated with the EC located over Kanto Plain at 0900 JST 8 December 1992. The detailed relationship between the EC and the low-level jet that was related

FIG. 20. Backward trajectory analysis for examining the origin of circulation of the tornado. (a) Initial circuit (thick black line) enclosing strong vertical vorticity at 100-m height at 0812:40 JST. The color shading, contour lines, and arrows in (a) show vertical vorticity (s^{-1}), pressure (hPa), and horizontal wind vectors ($m s^{-1}$) at 30-m height. (b) Horizontal projection of the circuit at 0810:40 JST. (c) As in (b), but for 0807:40 JST. The color on the circuit shows height (m), and black contour lines and gray arrows show vertical vorticity (s^{-1}) and horizontal wind vectors ($m s^{-1}$) at 30-m height. Color shading shows horizontal convergence (s^{-1}). (d) Time series of circulation (the orange line; $m^2 s^{-1}$) together with that of baroclinic (yellow) and frictional (purple) terms, where the blue line indicates integrated circulation.

to the tornadogenesis needs to be clarified in a future study. Flournoy and Coniglio (2019) have also shown that environmental parcels with large crosswise vorticity contribute to the formation of a mesovortex. Unlike previous studies (e.g., Weisman and Trapp 2003; Xu et al. 2015), however, the major source of the vorticity in the present mesovortex is not either horizontal vorticity baroclinically or frictionally generated, but is low-level vertical shear of horizontal wind, which is notably stronger than other cases (e.g., Coniglio et al. 2011). Meanwhile, it is noted that our simulation underestimates the strength of horizontal temperature gradient around gust front so that the contribution of baroclinically generated horizontal vorticity in the observed mesovortex may be larger.

The circulation analysis for the tornado shows that frictional effects contribute to the increase in circulation associated with the tornado. Moreover, it is suggested that environmental circulation such as horizontal shear and the vertical shear of horizontal wind also contribute to the

circulation of the tornado. The effects of friction on tornadogenesis spawned by supercells have been examined extensively in previous studies (Schenkman et al. 2014; Roberts et al. 2016; Markowski 2016; Mashiko 2016b), which uses similar type of surface condition. They showed that frictionally generated horizontal vorticity positively contributed to the tornadogenesis in their early development. Meanwhile, the contribution of friction may depend on the particular parcels selected (Markowski 2016; Yokota et al. 2018). We have also examined vorticity budgets along several other backward trajectories in the present case and found that the contribution of friction to the vorticity evolution depends on the choice of parcels (See supplementary information and Figs. S1–S3 in the online supplemental material).

Since the present study is a case study in which a tornado associated with a QLCS is numerically simulated with the highest horizontal resolution (50 m) and the origin of the rotation in the tornado is studied, additional studies for other

cases of tornadogenesis in a QLCS are desired for future work. In addition, an idealized numerical simulation that reproduces a QLCS and associated tornado is useful for obtaining further understanding of the process of tornadogenesis in a QLCS.

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Data availability statement. Since the code of the JMA nonhydrostatic model is owned by the JMA, permission has to be sought from the JMA on the basis of individual requests. JRA-55 is available from JRA project (https://jra.kishou.go.jp/JRA-55/index_en.html).

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