Structural and Environmental Characteristics of Extratropical Cyclones Associated with Tornado Outbreaks in the Warm Sector: An Idealized Numerical Study

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ABSTRACT

To clarify the effects of the horizontal shear of the jet stream on the structure and environment of extratropical cyclones that are accompanied by tornado outbreaks (OCs) and those that are not (NOCs), two idealized numerical experiments are performed. The experiments (OC-CTL and NOC-CTL) adopt the basic states taken from the corresponding composites of reanalysis data (JRA-55), except that the humidity field in both cases is taken from the OC composite.

The simulated cyclone in OC-CTL exhibits a more meridionally elongated structure and stronger low-level wind in the southeast quadrant of the cyclone center, resulting in larger values of storm relative environmental helicity (SREH) than those in NOC-CTL. These results are consistent with the characteristics of the cyclones found for OCs and NOCs in the authors’ composite study. The distributions of surface-based convective available potential energy (SBCAPE) show no notable differences between OC-CTL and NOC-CTL, while those of CAPE based on the most unstable air parcel (MUCAPE) show some differences.

A sensitivity experiment without moist processes such as condensation heating and evaporative cooling shows that the differences in the cyclone structure and environmental parameters between OCs and NOCs can be qualitatively explained by the dry dynamics. However, inclusion of moist processes results in notably larger differences.

1. Introduction

In the United States, tornado outbreaks often occur in the warm sector of extratropical cyclones, where the atmosphere is thermodynamically unstable and the horizontal winds have strong vertical shear (e.g., Carr 1952; Fujita et al. 1970; Galway 1975, 1977; Grazulis 1993). In the warm sector, strong low-level southerly winds advect warm and moist air, and upper-level strong westerly winds bring cold and dry air above (e.g., Newton 1967; Johns and Doswell 1992; Hamill et al. 2005), resulting in an environment favorable for severe convective storms. Thus, synoptic conditions are closely related to the occurrence of tornado outbreaks.

Many studies have clarified the relationship between tornado outbreaks and synoptic features (e.g., Miller 1972; Johns and Doswell 1992; Stensrud et al. 1997; Roebber et al. 2002; Gaffin and Parker 2006; Mercer et al. 2009, 2012; Shafer et al. 2009, 2010; Corfidi et al. 2010). Gaffin and Parker (2006) studied synoptic conditions and terrain effects on significant (F2 or greater) tornadoes across the southern Appalachian region. They revealed that wind direction and dewpoint were closely related to the surface boundaries rather than to large-scale terrain features. Corfidi et al. (2010) examined synoptic features for the Super Outbreak over the central and eastern United States on 3–4 April 1974. They indicated that the Super Outbreak was accompanied by a surface cyclone associated with a strong jet streak in the upper troposphere. The strong upper-level jet created the conditions favorable for cyclogenesis east of the Rockies, and provided the vertical shear necessary for intense supercells. Mercer et al. (2009, 2012) examined synoptic patterns that are accompanied by tornadic outbreaks (TOs) and by nontornadic outbreaks (NTOs). They showed that stronger upper-level troughs and low-level thermal advection favor TOs.

Tochimoto and Niino (2016, hereafter TN16) examined differences in the structural and environmental characteristics between extratropical cyclones (ECs)
that were accompanied by tornado outbreaks [hereafter referred to as outbreak cyclones (OCs)] and those that were not [referred to as nonoutbreak cyclones (NOCs)] in April and May between 1995 and 2012 in the United States. They defined OCs as ECs accompanying 15 or more tornadoes and NOCs as those accompanying 5 or fewer tornadoes. Based on a composite analysis using an objective analysis dataset [the Japanese 55-year Reanalysis (JRA-55); Kobayashi et al. (2015)], they found that the 0–1-km storm relative environmental helicity (SREH; Davies-Jones 1984) and convective available potential energy (CAPE) for OCs are larger than those for NOCs, and the region in which their values are significant is wider for OCs. The larger SREH values for OCs are shown to be due to the stronger low-level southerly winds, which are associated with the structure of OCs in the lower to middle troposphere. 

TN16 also suggested that the stronger southerly winds in OCs are due to larger horizontal anticyclonic shear of the jet stream in which the ECs are embedded. As shown by Wernli et al. (1998), background barotropic shear affects the structure of ECs and their associated frontal systems. The stronger anticyclonic shear results in ECs and frontal structures that are elongated in the northeast–southwest direction. This is due to the deformation fields associated with the jet structures: anticyclonic shear has a deformation axis aligned from southwest to northeast (Wernli et al. 1998). The results of TN16 indicated that the anticyclonic shear of the environmental jet stream for OCs is stronger than that for NOCs, and OCs have a more meridionally elongated structure than NOCs. Thus, the zonal pressure gradient (PG) southeast of the cyclone center for OCs is stronger than that for NOCs, resulting in the stronger southerly winds and larger SREH in the warm sector for OCs (Fig. 1).

In the present study, we attempt to clarify the relationship between the background horizontal shear of the jet stream and the EC structure, including associated environmental convective parameters. We are particularly interested in dynamical processes that contribute to environmental parameters favorable for
tornadogenesis. We perform idealized numerical experiments to identify the role of the jet stream in EC structure. Since the jet structure for OCs and NOCs obtained from the composite analyses in TN16 will be used as the basic states for the numerical experiments, it is expected that average features of ECs that influence a tornado outbreak may be reproduced in the numerical experiments.

The remainder of this paper is organized as follows. The model used in this study and the experimental design are described in section 2, and the results are outlined in section 3. A discussion is presented in section 4, followed by a summary in section 5.

2. Numerical model and experimental design

a. Numerical model

The Weather Research and Forecasting (WRF) Model, version 3.4, is used in the present study. The model domain is 12000 km in the zonal (x) direction, 6000 km in the meridional (y) direction, and 25 km in the vertical (z) direction. The horizontal grid spacing is 20 km. The domain has 50 vertical levels, with the vertical spacing varying from about 20 m near the surface to about 1000 m at the model top at 25 km. The WRF single-moment 6-class (WSM6) explicit moisture scheme (Hong and Lim 2006), the Kain–Fritch cumulus scheme (Kain 2004), the Noah land surface model (Chen and Dudhia 2001), and the Mellor–Yamada–Nakanishi–Niino (MYNN) level-2.5 planetary boundary scheme (Nakanishi and Niino 2006) are used in the experiments. Periodic and symmetric conditions are adopted on the zonal and meridional boundaries, respectively. The lower boundary is flat and is set as grassland with a fixed surface temperature that is equal to the initial surface air temperature at each grid. There are no sensible or latent heat fluxes from the surface. A Rayleigh damping layer of 4-km depth is placed near the upper boundary. The experiments including all these settings are denoted as CTLs.

b. Basic fields and initial disturbance

In TN16, OCs and NOCs with similar intensity of vorticity and sea level pressure (SLP) were selected (see appendix A in TN16). The numbers of OCs and NOCs are 32 and 21, respectively. The basic fields for the numerical experiments are obtained from the three-dimensional JRA-55 (Ebita et al. 2011; Kobayashi et al. 2015) through the following procedure. First, we take 5-day mean fields centered on the time (t) 18 h before key-time (KT) for OCs and NOCs, where KT is defined as the time when an EC is accompanied by the largest number of tornadoes within 24 h (see TN16). Next, for OCs and NOCs the meridional–height profiles of the 5-day mean temperature and pressure fields along the longitude of each cyclone center are produced, and then these profiles are superposed with respect to each cyclone center to obtain composite fields at t = KT − 18 h. Finally, the composite fields are specified to be zonally uniform for both OCs and NOCs, with zonal winds set to be in geostrophic balance with the meridionally varying pressure fields. Thus, the basic fields vary in the y and z directions, but are constant in the x direction. The numerical experiments performed for the OC and NOC environments thus obtained are denoted as OC-CTL and NOC-CTL, respectively. Note that the water vapor field for the OCs is used for both OC-CTL and NOC-CTL, to isolate the effects of the jet structure on the structures of OCs and NOCs.

Figure 2a shows vertical cross sections of basic-state zonal winds for both OC-CTL and NOC-CTL, and Fig. 2c shows the meridional distributions of these winds averaged between 1000 and 250 hPa. In the southern part of the zonal jet, westerly winds for OC-CTL decrease more sharply southward than for NOC-CTL (Figs. 2a,c). In other words, the westerly winds for OC-CTL have stronger anticyclonic shear south of the jet axis (Figs. 2c,d). This feature is consistent with the analyses from JRA-55 in TN16. On the other hand, in the northern part of the zonal jet, the westerly winds for NOC-CTL have stronger cyclonic horizontal shear. However, the vertical shear of the westerly winds between OC-CTL and NOC-CTL differs little below 10 km (Fig. 2a). The vertical cross section of the water vapor field is shown in Fig. 2b. The specific humidity has its maximum value of about 0.018 kg kg⁻¹ near the surface and decreases with height. We will perform idealized numerical experiments for these basic states. To start the experiments, a temperature perturbation that has zonal and meridional horizontal width of 3000 km and maximum amplitude of 2 K is placed at a height of about 7 km, at the center of the calculation domain. Similar initial disturbances were chosen in previous studies (e.g., Schultz and Zhang 2007).

3. Results

a. Temporal evolution of extratropical cyclones

As a starting point, we show the evolution of the minimum SLP for OC-CTL and NOC-CTL (Fig. 3). In OC-CTL, the cyclone starts deepening from about t = 60 h, and attains the minimum pressure of about 985 hPa at t = 300 h.
In NOC-CTL, the cyclone deepens more slowly than in OC-CTL from \( t = 120 \) to \( 240 \) h. After \( t = 240 \) h, however, the deepening rate of SLP is similar to that for OC-CTL. The cyclone attains its minimum pressure of \( \sim 989 \) hPa at \( t = 320 \) h, which is slightly higher than that for OC-CTL.

b. Structural characteristics of simulated extratropical cyclones

Here, we examine the horizontal structures of ECs for both OC-CTL and NOC-CTL. To this end, it is desirable to compare the ECs of similar intensity between OC-CTL and NOC-CTL. Figure 3 shows that the EC at \( t = 264 \) h for OC-CTL and that at \( 306 \) h for NOC-CTL have the same central pressure of 990 hPa. Thus, the horizontal maps for NOC-CTL presented below will be 42 h later than those for OC-CTL. Figure 4 shows the evolution of SLP and potential vorticity (PV) at 900 hPa for OC-CTL and NOC-CTL. The strongest cyclones in both OC-CTL and NOC-CTL are located between \( x = -1500 \) and \( x = 1500 \) km. Hereafter, we focus on these cyclones.

The EC at \( t = 264 \) h for OC-CTL (Fig. 4a) and that at \( t = 306 \) h for NOC-CTL (Fig. 4b) exhibit remarkable characteristics of the ECs in both experiments. The central SLP of the ECs is about 990 hPa. For OC-CTL, a low pressure trough extends southwestward from the minimum SLP center and the low-level PV exceeding 1.5 PVU extends southward. In NOC-CTL, on the other hand, a low pressure trough and low-level strong PV field extend southeastward from the minimum SLP center. After these times, the differences in the structures between OC-CTL and NOC-CTL become more evident. The cyclones for both OC-CTL and NOC-CTL deepen with time, and the cyclone for OC-CTL becomes elongated southwestward and that for NOC-CTL southeastward, respectively. These structures for OC-CTL and NOC-CTL are maintained in the developing period (Figs. 4c,d).
Low-level wind fields show notable differences between OC-CTL and NOC-CTL (Fig. 5). At $t = 240\text{ h}$, the region in which low-level winds exceed $15\text{ m s}^{-1}$ in the warm sector for OC-CTL is wider than that for NOC-CTL at $t = 282\text{ h}$, even though the EC for OC-CTL is notably weaker than for NOC-CTL. As time passes, low-level winds in OC-CTL further intensify with the cyclone’s development (Figs. 5a,c,e,g). For OC-CTL, the low-level winds in the southeast quadrant of the cyclone center exceed $21\text{ m s}^{-1}$ at $t = 264\text{ h}$, while those for NOC-CTL are less than $21\text{ m s}^{-1}$ at $t = 306\text{ h}$; by $t = 288\text{ h}$, the maximum low-level wind for OC-CTL reaches $27\text{ m s}^{-1}$ and the region in which the wind speed exceeds $21\text{ m s}^{-1}$ extends $\sim 700\text{ km}$ southward. On the other hand, there is no notable intensification of the low-level winds in the southeast quadrant of the cyclone center in NOC-CTL. Although the low-level winds exceed $21\text{ m s}^{-1}$ in a region located southeast of the cyclone center at $t = 330\text{ h}$, the winds are weaker and the region is narrower than in OC-CTL.

These differences between the characteristics of OC-CTL and NOC-CTL are explained by the structure of the jet stream. The jet stream in OC-CTL in the basic state has stronger anticyclonic shear on the southern side of the jet axis than in the NOC-CTL. Stronger anticyclonic horizontal shear of a jet stream...
The detailed development of the fronts is now examined using \( \mathbf{Q} \) vectors, whose along- and across-isentropic components are given by \( Q_n \) and \( Q_s \), respectively (e.g., Keyser et al. 1992; Martin 1999). The \( Q_n \) effects changes in the magnitude of \( \nabla \theta \), while \( Q_s \) affects its direction. Figures 6a and 6b show distributions of \( Q_n \) and \( Q_s \) and their magnitudes, together with distributions of surface pressure and temperature around the ECs for OC-CTL, and Figs. 7a and 7b show those for NOC-CTL. In OC-CTL, the contribution of \( Q_n \) dominates near the cyclone center, while \( Q_s \) contributes to the cold frontogenesis extending southwestward (Figs. 6a,b). In NOC-CTL, on the other hand, the contribution of \( Q_n \) to the cold frontogenesis is smaller than in OC-CTL and is located southeast of the cyclone center. Again \( Q_s \) is concentrated around the cyclone center in NOC-CTL (Figs. 7a,b). Thus, the structures of OCs become meridionally more elongated with time, resulting in a stronger PG in the south-southeast region of the cyclone center.

The stronger PG in the south-southeast of the cyclone center is likely to result in stronger geostrophic winds there in OC-CTL. To confirm this, wind fields are decomposed into geostrophic and ageostrophic components as shown in Figs. 6c,d and 7c,d. It is clear that the geostrophic component is dominant for ECs in OC-CTL and NOC-CTL, while the ageostrophic contributions to the total winds are small. However, the ageostrophic winds play an important role in the acceleration of geostrophic winds. The tendency of meridional geostrophic winds on an \( f \) plane is given by advection due to geostrophic winds (Figs. 6e and 7e) and the Coriolis force acting on the ageostrophic wind (Figs. 6f and 7f) (Holton 2004). The positive contributions due to the ageostrophic term dominate the advection term, and their magnitude is comparable to the tendency of southerly geostrophic winds. The ageostrophic winds that contribute to the acceleration of geostrophic winds have an easterly component (Figs. 6d and 7d), which is associated with zonal PG. Thus, the stronger zonal PG in the south-southeast of the cyclone results in stronger southerly winds in OC-CTL than in NOC-CTL.

c. Upper-level features

In this subsection, we describe the characteristics of the evolution of upper-level jets and troughs at 300 hPa (Figs. 8a,b). Upper-level troughs are located to the west of the surface cyclone centers in both OC-CTL and NOC-CTL. During the simulations, the jet stream has a stronger anticyclonic horizontal shear in
FIG. 6. (a) Horizontal distributions of $\mathbf{Q}_n$ vectors (arrows) and their magnitudes (color shading; $10^{-13}$ m$^2$ kg$^{-1}$ s$^{-1}$) at 950 hPa for OC-CTL at $t = 258$ h. (b) As in (a), but for $\mathbf{Q}_s$.

Black contour lines in (a),(b) indicate potential temperature (K), and blue contour lines indicate SLP (hPa). (c) Geostrophic wind vectors (arrows) and their magnitude (color shading; m s$^{-1}$) at 900 hPa for OC-CTL at $t = 264$ h. (d) As in (c), but for ageostrophic wind. Yellow contour lines in (c),(d) indicate SLP (hPa). (e),(f) Horizontal distributions in the tendency equation of southerly geostrophic winds [color shading; m s$^{-1}$ (12 h)$^{-1}$] at 900 hPa for OC-CTL at $t = 264$ h: (e) ageostrophic term and (f) advection term. Contour lines in (e),(f) indicate the tendency of geostrophic winds [m (12 h)$^{-1}$].
FIG. 7. As in Fig. 6, but for NOC-CTL and for different times: (a),(b) $t = 270$ and (c)–(f) $t = 306$ h.
OC-CTL. The horizontal axes of the troughs in both OC-CTL and NOC-CTL are tilted northwest–southeast; the tilt is greater in NOC-CTL than in OC-CTL. In OC-CTL, the jet core region with wind speeds exceeding 33 m s$^{-1}$ is located to the east of the upper-level trough. In contrast, the region of the jet core in NOC-CTL is found west of the trough and the core wind is weaker than in OC-CTL. These results suggest that the upper-level forcing for upward motion in OC-CTL is likely to be stronger than in NOC-CTL and contributes to stronger cyclone development.

The stronger jet in the upper layer in OC-CTL may also be due to the deformation fields, which are associated with the structure of the jet in the basic fields. Figures 8c and 8d show the distributions of $Q_n$ and geopotential height anomaly from the basic fields for OC-CTL and NOC-CTL. Both positive and negative
geopotential height anomalies for OC-CTL are elongated southwestward between \( y = -1000 \) and \( y = 500 \) km, and the region of large \( \nabla^2 \phi \) is located in the southern part of the region of negative geopotential height anomalies (Fig. 8e). On the other hand, those for NOC-CTL are also elongated (though southeastward), but the values of \( \nabla^2 \phi \) are smaller than for OC-CTL (Fig. 8d). The southwestward-elongated structure of geopotential height anomalies likely contributes to the stronger jet in OC-CTL through the enhancement of southwesterlies.

d. Mesoscale environmental parameters

Environmental parameters such as the SREH defined between 0 and 1 km, surface-based CAPE (SBCAPE), and the most unstable CAPE (MUCAPE) in the storm environment are examined to identify differences between OC-CTL and NOC-CTL. The storm motion vectors used to calculate SREH are estimated following Bunkers et al. (2000). Figure 9a shows the time sequence of the maximum SREH in both OC-CTL and NOC-CTL. In both experiments SREH is smaller than \( 100 \) m\(^2\) s\(^{-2}\) up to \( t = 120 \) h. After this time, however, SREH in OC-CTL gradually increases with deepening of the cyclone and exceeds \( 100 \) m\(^2\) s\(^{-2}\) by \( t = 200 \) h, while that in NOC-CTL remains less than \( 100 \) m\(^2\) s\(^{-2}\). From \( t = 200 \) to 280 h, SREH in OC-CTL rapidly increases to \( 490 \) m\(^2\) s\(^{-2}\) during the rapid cyclone deepening. Although SREH for NOC-CTL rapidly increases from \( t = 240 \) to 300 h, its maximum remains about \( 390 \) m\(^2\) s\(^{-2}\), which is smaller than in OC-CTL. It might be argued that the difference of the maxima between OC-CTL and NOC-CTL could be partly due to the difference in cyclone strength. However, SREH for OC-CTL is larger than that for NOC-CTL even if they are compared at times when they have similar SLPs (Fig. 9a). Additionally, the area of SREH exceeding \( 150 \) m\(^2\) s\(^{-2}\) in OC-CTL is much wider than that in NOC-CTL throughout the developing period of the cyclone (solid lines in Fig. 9b): the area exceeds \( 35 \times 10^4 \) km\(^2\) at 300 h in OC-CTL, while it is less than \( 15 \times 10^4 \) km\(^2\) at 330 h in NOC-CTL. The area of SREH exceeding \( 200 \) m\(^2\) s\(^{-2}\) shows more notable difference between OC-CTL and NOC-CTL (dashed lines in Fig. 9b). The area of SREH > \( 200 \) m\(^2\) s\(^{-2}\) for NOC-CTL (\( 5 \times 10^4 \) km\(^2\) at 330 h) is much narrower than that for OC-CTL (\( 15 \times 10^4 \) km\(^2\) at 300 h), which indicates greater extent of larger values of SREH for OC-CTL.

The spatial distribution of SREH is shown in Fig. 10. At 264 h in OC-CTL, large SREH exceeding \( 300 \) m\(^2\) s\(^{-2}\) is found in the region east-southeast of the cyclone center (Fig. 10a). This region extends southward as the cyclone develops. At 288 h, the region of SREH exceeding \( 300 \) m\(^2\) s\(^{-2}\) extends more than 500 km to the south of the cyclone center and the maximum value increases to more than \( 400 \) m\(^2\) s\(^{-2}\) (Fig. 10c).

The value of SREH in NOC-CTL is notably smaller than in OC-CTL and the region of large SREH is notably narrower. At \( t = 306 \) h, a region with SREH larger than \( 150 \) m\(^2\) s\(^{-2}\) extends southeastward from the north of the cyclone center (Fig. 10b). This region expands southeastward as the cyclone develops. Although the region of SREH exceeding \( 200 \) m\(^2\) s\(^{-2}\) is found from the north to the east of the cyclone center at \( t = 330 \) h, it is narrower than in OC-CTL and values of SREH are notably smaller (Fig. 10d).

These differences in SREH between OC-CTL and NOC-CTL are strongly associated with the strength of low-level winds. The region with large SREH corresponds to that with strong low-level winds (Fig. 5). The low-level winds in OC-CTL are stronger and the region of strong winds is wider than in NOC-CTL. Figure 11a shows a vertical cross section of horizontal wind speeds along A–A in Fig. 10c, and Fig. 11b that along B–B’ line in Fig. 10d for NOC-CTL. The vertical shear of winds below 1 km for OC-CTL is stronger than that for NOC-CTL, resulting in larger SREH for OC-CTL. Note that the wind direction below 1 km is predominantly southerly (Figs. 11a,b). Since the storm motion vectors estimated with the method of Bunkers et al. (2000) reflect both the vertical wind distributions between the surface and 6-km height and assumed internal supercell storm dynamics, they deviate to the right of the mean winds between 0 and 1 km (not shown) and SREH can have large values even for unidirectional 0–1-km shear. These results are consistent with TN16, who pointed out the importance of the low-level shear of southerly winds for large SREH. Stronger midlevel winds for OC-CTL result in faster storm motion, which also contributes to larger SREH.

Unlike the dynamic parameter SREH, there is no notable difference in a thermodynamic parameter, SBCAPE, between OC-CTL and NOC-CTL (Figs. 12a and 12b). In both OC-CTL and NOC-CTL, SBCAPE of \( 100–300 \) m\(^2\) s\(^{-2}\) is found at more than 500 km to the south of the primary cyclone center. Although SBCAPE in the composite analysis for OCs exceeds at least \( 1000 \) m\(^2\) s\(^{-2}\) (e.g., Fig. 5 in TN16), it is notably smaller in the present idealized experiments. This is likely due to the absence of shortwave radiative heating in the experiments, as will be discussed in section 4. Thus, it is not easy to examine thermodynamic instability for the numerical experiments. However, the MUCAPE calculated by lifting the most unstable air parcel give some information about the instability.
above the boundary layer. In fact, the distributions of 
MUCAPE show some differences between OC-CTL and 
NOC-CTL (Figs. 12c and 12d). In the southeast of 
the cyclone center for OC-CTL, the value of MUCAPE 
exceeds 1100 m$^2$s$^{-2}$, which is larger than that for NOC-CTL, and the region in which MUCAPE exceeds 900 m$^2$s$^{-2}$ is wider.

These differences in MUCAPE contribute to differences between OC-CTL and NOC-CTL in an energy helicity index (EHI), which is defined by a combination of MUCAPE and SREH. Although the levels of 0–1-km SREH and the most unstable parcel (900–850 hPa) do not overlap vertically, it is likely that the level of the most unstable parcel may be lowered when the convective mixed layer develops under the daytime shortwave radiation heating. Thus, it may be meaningful to examine the distribution of the EHI. The EHI for OC-CTL exceeds 1 in the southeast of the cyclone center, while it is less than 0.8 for NOC-CTL (Figs. 13e,f). The larger EHI for OC-CTL is reflected by the horizontal overlap of the regions of larger SREH and MUCAPE. Note that Figs. 12 and 13 compares SBCAPE, MUCAPE, and EHI between OC-CTL and NOC-CTL with a 36-h offset. This is because MUCAPE and EHI are largest at $t = 270$ h for OC-CTL and at $t = 306$ h for NOC-CTL. The 36-h offset could result in a high bias of the EHI values for OC-CTL since the cyclone for OC-CTL rapidly develops between $t = 264$ and 270 h.

The differences in MUCAPE between OC-CTL and NOC-CTL are strongly associated with the low-level water vapor and temperature fields. The distributions of specific humidity and potential temperature at 850 hPa are shown in Fig. 13. A region of specific humidity exceeding 0.011 kg kg$^{-1}$ extends over the southeast region of the cyclone center in OC-CTL (Fig. 13a), while such a region is not found in NOC-CTL (Fig. 13b). The warm sector of the EC center for OC-CTL is also warmer than that in NOC-CTL. Warm air with potential temperature above 302 K intrudes to around $y = 500$ km in OC-CTL (Fig. 13c), while the potential temperature around the same region is below 300 K in NOC-CTL (Fig. 13d). Thus, the higher low-level water vapor and warmer temperature advected from the south-southwest contribute to the larger MUCAPE in OC-CTL.

The vertical profiles of temperature, moisture, and winds at the positions of the maxima of EHI are examined using skew $T$–log$P$ diagrams and hodographs (Fig. 14). In both OC-CTL and NOC-CTL, a nearly saturated layer with the maximum temperature of about 287–300 K, which results from meridional advection of temperature and specific humidity, exists around 900–850 hPa. In contrast, there is a shallow stable layer between 900 and 1000 hPa due to the absence of radiative heating and surface fluxes. Thus, the most unstable air with MUCAPE of 880 m$^2$s$^{-2}$ is located at 850 hPa in OC-CTL, while that with MUCAPE of 802 m$^2$s$^{-2}$ is located at 875 hPa in NOC-CTL. The temperature in the 900–850 hPa layer in OC-CTL is higher than that in NOC-CTL. The hodographs show that wind speeds near the surface and at 1-km height for OC-CTL (NOC-CTL) are about 6 (5) m s$^{-1}$ and 18 (15) m s$^{-1}$, respectively. Additionally, the storm motion speeds for OC-CTL and NOC-CTL are 18 and 12 m s$^{-1}$, respectively. Thus, the larger southwesterly shear between 0 and 1 km, and faster storm motion in OC-CTL give SREH of 194 m$^2$s$^{-2}$, which is larger than 154 m$^2$s$^{-2}$ in NOC-CTL. The hodographs change their shape while going northward toward the cyclone center within the warm sector: the low-level southwesterly (south-southeasterly) wind vectors between 0 and 1 km in the south for OC-CTL (NOC-CTL) gradually turn to southerly (southeasterly), and then to southeasterly...
(east-southeasterly) in the north. Thus, the hodographs near the cyclone centers are characterized by stronger backing of the near-surface winds, which are dominated by southeasterly vertical shear (not shown). On the other hand, hodographs above the lower layer continue to have veering shear due to the upper-level westerly.

4. Discussion

Wernli et al. (1998) have shown that the horizontal structure of the jet stream affects the structure and dynamics of an extratropical cyclone in a general context. We have examined the structures of ECs that develop in the composited jet streams for OCs and NOCs in idealized numerical experiments, and have shown that the structures differ notably between OC-CTL and NOC-CTL. Furthermore, we have shown that the different structure of the EC results in different mesoscale environment characterized by parameters such as SREH and MUCAPE.

In OC-CTL, the EC and its associated front have a meridionally elongated structure and stronger low-level winds in the east-southeast region of the cyclone.
center, resulting in larger SREH. In NOC-CTL, on the other hand, the EC and its associated front have a zonally elongated structure and weaker low-level winds in the region east-southeast of the cyclone center, resulting in smaller SREH. These differences are due to the structure of the jet stream. Since the environmental anticyclonic horizontal shear in OC-CTL is stronger than in NOC-CTL, the horizontal structure of the cyclone in OC-CTL is stretched meridionally, resulting in the stronger low-level winds in the region east-southeast of the cyclone center. These results are consistent with the features found in our composite analysis (TN16). In the composite analysis, OCs have a meridionally more elongated structure and stronger low-level winds in the warm sector than those in NOCs. The SREH is also larger in OCs in the east-southeast quadrant of the cyclone center. Although previous studies also showed the importance of the 0–1-km SREH (e.g., Stensrud et al. 1997; Mercer et al. 2009, 2012; Shafer et al. 2009), the present study has revealed the relationship between the cyclone structure and the 0–1-km SREH. The results of the present study also suggest that the stronger upper-level jet in OC-CTL results in faster storm motion, thus contributing to larger SREH.

In the idealized experiments, the values of SBCAPE are notably smaller than those in composite analyses. It is likely that the difference is caused by the absence of shortwave radiation in the idealized experiments. As shown in TN16, SBCAPE in both OCs and NOCs is largest at 1800 or 0000 UTC. This means that daytime shortwave radiation is likely to contribute to the larger SBCAPE (e.g., Thompson and Edwards 2000). It seems necessary to perform experiments to include the effects of the radiation in future work.

Since the basic field of specific humidity in OC-CTL is the same as in NOC-CTL in the present experiments, the differences in EC structure between OC-CTL and NOC-CTL contribute to their differences in thermodynamic instability. The result that MUCAPE in OC-CTL is somewhat larger than in NOC-CTL suggests that OCs have a greater potential for creating more unstable stratification in the troposphere as a consequence of advection of warmer and more humid air from the south or southwest into the warm sector by south-southwesterly winds.

The results of our experiments show that upper-level troughs in OC-CTL and NOC-CTL have a northwest–southeast tilt, with a larger tilt in NOC-CTL. Johns and Doswell (1992) and Boustead et al. (2013) showed that upper-level troughs accompanied by warm sector tornadoes have negatively tilted structures, as found in both OC-CTL and NOC-CTL. The present simulations also show that the warm sector is located near the exit region of a jet streak in OC-CTL and NOC-CTL, similar to those in Boustead et al. (2013).

In TN16, it was argued that the distance between a low-level cyclone and a preexisting upper-level high PV associated with an upper-level trough is related to the strength and distribution of upward motion in the warm sector. The results of the composite analysis suggest that since the upper-level high PV for OCs is closer to the low-level cyclone than that for NOCs, the vertical velocity from lower to middle levels is stronger and the region of upward motion is wider for OCs. However, the present idealized experiments do not simulate the effects of preexisting upper-level PV. Thus, additional experiments with initial upper-level disturbances would be useful in future studies to compare the simulation results with the observations in detail.
We also performed dry experiments (OC-DRY and NOC-DRY) and found differences in the structure of simulated ECs between these experiments (see appendix A). The EC in OC-DRY has a meridionally elongated structure similar to that in OC-CTL, while the EC in NOC-DRY has a zonally elongated structure similar to that in NOC-CTL. The low-level winds and SREH in the east-southeast region of the cyclone center in OC-DRY are larger than those in NOC-DRY, consistent with those in OC-CTL and NOC-CTL. However, the magnitudes of low-level winds and SREH in OC-CTL and NOC-CTL are larger than those in their dry
counterparts, partly because the ECs in OC-CTL and NOC-CTL are more intense.

5. Summary

We have performed idealized numerical experiments on extratropical cyclones that develop in zonally uniform jet streams, which are typical for environments of outbreak cyclones (OCs) and nonoutbreak cyclones (NOCs). The meridional–height structures of the jet streams are taken from the composites for OCs and NOCs in TN16. The experiments for the OC environment and NOC environment are denoted OC-CTL and NOC-CTL, respectively.
The results of the experiments show that the horizontal shear of the environmental jet stream has large impacts on the structure of extratropical cyclones (ECs). The EC for OC-CTL, in which the anticyclonic horizontal shear of the environmental jet is stronger, has a more meridionally elongated structure and stronger low-level winds in the southeast region of the cyclone center. On the other hand, the EC for NOC-CTL, in which the anticyclonic horizontal shear is weaker, has a more zonally elongated structure and weaker low-level winds in the southeast region of the cyclone center than in OC-CTL. These differences are consistent with the results of the composite analysis in TN16.

The strength of the horizontal anticyclonic shear of the jet stream also affects the convective environmental parameters in the ECs. A dynamic parameter, SREH, attains greater values in the east-southeast quadrant of the cyclone center for OC-CTL than for NOC-CTL. This difference is mainly due to the differing strength of the low-level southerly wind, which reflects the structure of the ECs. These differences of the EC structure and SREH between OCs and NOCs can be found even in dry experiments, although low-level winds are weaker and SREH smaller than those in the corresponding CTLs. In the idealized experiments a thermodynamic parameter, MUCAPE, and low-level specific humidity also show some differences between OC-CTL and NOC-CTL, which both use the same water vapor field for the basic state. Larger MUCAPE for OC-CTL is due to greater northward advection of low-level moisture.

The present study has revealed the impact of the horizontal structure of the jet stream on environmental parameters, such as SREH, in ECs. However, several important factors, such as shortwave radiation, upper-level disturbances, and zonal inhomogeneity of basic fields, are not considered in the experiments. Since these factors may play important roles in the occurrence and development of severe convective storms, additional experiments that take account of these factors are desired in the future. Furthermore, the present experiments do not have enough resolution to reproduce supercells or tornadoes. Thus, the
downscaling of experiments to resolve supercells may also contribute to our understanding of the relationship between the dynamics of extratropical cyclones and supercells.

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FIG. B1. Horizontal distribution of low-level wind speed at 900 hPa (m s$^{-1}$) for (a) OC-WN5 and (b) NOC-WN5, SREH (m$^2$ s$^{-2}$) for (c) OC-WN5 and (d) NOC-WN5, and CAPE for (e) OC-WN5 and (f) NOC-WN5. Contour lines indicate SLP (hPa). Note that the time for OC-WN5 is 216 h and that for NOC-WN5 is 240 h.
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APPENDIX A

Dry Simulations

To examine the effects of the moist processes on the structure of ECs and environmental parameters, we have carried out dry simulations corresponding to OC-CTL and NOC-CTL (hereafter these experiments are called OC-DRY and NOC-DRY, respectively) and have compared their results with those of the CTLs. In the dry experiments, moist processes such as cumulus heating and latent heating associated with microphysics are turned off but specific humidity is retained. It turned out that the cyclone in NOC-CTL develops much slower than that in OC-DRY. Figure A1 shows the horizontal structure of ECs in OC-DRY and NOC-DRY at the times when the ECs have similar intensity, where the time offset is 90 h. The cyclones in the dry simulations have larger meridional scales than those in the moist simulations (e.g., Tokioka 1973; Yanase and Niino 2004, 2007). In OC-DRY (NOC-DRY), the maximum low-level wind and SREH are about 20 (17) m s\(^{-1}\) and 250 (200) m\(^2\) s\(^{-2}\) at \(t=288\) (\(t=378\)) h, respectively. The cyclones for both OC-DRY and NOC-DRY have weaker low-level wind and smaller SREH in the south-southeast quadrant of the cyclone center than those for OC-CTL and NOC-CTL, respectively (Figs. 5 and 10). However, differences in the cyclone structure and horizontal distribution of SREH between OC-DRY and NOC-DRY are evident. The cyclones for OC-DRY and NOC-DRY have meridionally and zonally elongated structures, respectively. This difference between OC-DRY and NOC-DRY results in the differences in low-level winds and SREH, which are similar to those between OC-CTL and NOC-CTL. Therefore, the dry dynamics seem to explain the qualitative differences in the cyclone structure and associated SREH.

APPENDIX B

Sensitivity to Initial Perturbation

The results in section 3 are obtained for an initial temperature perturbation as described at the end of section 2. To examine the robustness of the results, additional experiments with zonally sinusoidal initial temperature perturbations of wavenumber 5 and maximum amplitude of 2 K at the meridional center are performed for the same basic states as OC-CTL and NOC-CTL. These experiments, corresponding to OC-CTL and NOC-CTL, are denoted as OC-WN5 and NOC-WN5, respectively. Figure B1 shows the horizontal structure of low-level winds, SREH, and CAPE for OC-WN5 at 216 h and NOC-WN5 at 240 h. The 24-h offset is chosen to make the cyclone intensities in both experiments similar. The simulated cyclones for OC-WN5 have a meridionally more elongated structure, while those for NOC-WN5 have a zonally more elongated structure. Low-level winds in the east-southeast quadrant of the cyclone center for OC-WN5 are stronger than those for NOC-WN5, resulting in larger SREH for OC-WN5. On the other hand, there is little difference in CAPE between OC-WN5 and NOC-WN5. These features are quite similar to those for OC-CTL and NOC-CTL.

REFERENCES


