Structure and Evolution of Misovortices Observed within a Convective Snowband in the Japan Sea Coastal Region during a Cold-Air Outbreak on 31 December 2007

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Abstract

Cyclonic misovortices with a horizontal scale of 0.4–1.9 km embedded within a convective snowband were observed by two X-band Doppler radars in the Japan Sea coastal region on December 31, 2007, during a cold-air outbreak. All vortices initially developed offshore, subsequently making a landfall. The structure and temporal evolution of these vortices during the landfall were investigated using high-resolution data obtained from two X-band Doppler radars.

The studied vortices developed along a low-level convergence line characterized by cyclonic horizontal shear, suggesting that horizontal shearing instability was responsible for the initial development of the vortices. A detailed investigation was performed on a vortex that passed within a close range (< 10 km) of both radars and almost directly over two surface observation stations. As this vortex approached the coast, it extended upward with time and eventually reached a height greater than half of the echo-top height of the parent snowband. During the landfall, the vortex core diameter contracted markedly and its peak tangential velocity and vertical vorticity increased at lower altitudes. Such a temporal change of low-level vortex was associated with an intensification of low-level convergence around the vortex and the convergence line. These facts suggest that the stretching of the low-level vortex was responsible for the low-level vortex contraction and increase in peak tangential velocity and vertical vorticity during the landfall.

Keywords radar observation; winter misovortex; landfall

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1. Introduction

Severe storms frequently occur in the coastal region of the Japan Sea during cold-air outbreaks from the Eurasian continent that mitigate over its warm sea surface. Wind gusts associated with misoscale vortices (40–4000 m; Fujita 1981) to mesoscale vortices, the former including tornadoes, are one example of such convective phenomena that generally develop over the sea and later move inland. A statistical study of tornadoes and waterspouts in Japan from 1961 to 1993 (Niino et al. 1997) showed that winter monsoon tornadoes accounted for 12% of all tornadoes documented in Japan, and that occurrences were concentrated along the coastline of the Japan Sea. Understanding the mechanisms by which such misoscale vortices (i.e., misovortices) evolve—especially during a landfall—is a critical step towards improving disaster prevention and mitigation plans for wind-related hazards along the Japan Sea coastline.

Winter convective clouds generally have low cloud tops and fast moving speeds, which make fine-scale data acquisition on both winter tornadoes and their parent storms problematic. As such, only a small number of studies of misovortices that develop under winter monsoon situation (hereafter, winter misovortices) in the Japan Sea coastal region have been performed. For example, Kobayashi et al. (2007) observed a tornado within a small snowcloud in the Japan Sea coastal region using both Doppler radar and photographs, and documented its characteristics, funnel cloud structure, and misovortex structure. They documented a landfall process of the tornado by mainly using a series of photographs.

In order to study the fine-scale structure of wind gusts and to develop an automatic strong gust detection system for railroads (Kusunoki et al. 2008), field observations were conducted in the Shonai area, which is located in the Japan Sea coastal region. Their study demonstrated that wind gusts recorded in the area were mainly associated with the passage of misovortices including tornadoes (Kusunoki et al. 2009; Inoue et al. 2011; Kusunoki et al. 2011). Inoue et al. (2011) documented multiple characteristics of winter misovortices and tornadoes, including their low-level velocity and pressure structures, temporal variations during a landfall, and their mutual interactions. However, a lack of volumetric radar data prohibited detailed discussion of their evolution processes during a landfall. Kusunoki et al. (2011) used high-resolution volume-scan Doppler radar data to investigate the time-dependent changes in the vertical structures of a tornadic vortex during a landfall and showed that it tilted downstream and reduced in size at low altitudes during a landfall. Their data also indicated that the vertical vorticities at low altitudes rapidly intensified after the landfall, although the detailed temporal evolution of the vortex during this period was not discussed. Recently, Kato et al. (2015) used volume-scan data to investigate temporal changes in the intensity and structure of misovortices during a landfall in the Japan Sea coastal region and showed that vortices weakened after making the landfall, while additionally increasing their forward tilt with increasing height. Although surface roughness was suggested to be a possible cause of this post-landfall vortex decay, other factors that would affect their behavior were not discussed (e.g., parent storm behavior, modification of low-level wind fields, and interaction with land breeze).

On December 31, 2007, several misovortices developed within a convective snowband that was oriented perpendicular to the prevailing wind direction in the region of the Japan Sea. During a landfall, some vortices showed significant variations in both reflectivity and Doppler velocity fields. Strong wind gusts were also observed during the passage of one of the vortices over the observational area, which allowed two sets of X-band Doppler radar data to be obtained, including close range (<10 km) and high-temporal resolution data and volume-scan data. In this study, we use these data to investigate the structure and evolution of these vortices during a landfall. The vortex, which was observed at close range by both Doppler radars and directly over two surface weather stations, was analyzed in detail to investigate its temporal evolution during the landfall and discuss how parent storm behavior and low-level wind fields around the vortex are related to it.

Section 2 describes the observational instruments and the data analysis procedures implemented in our study. Section 3 presents an overview of the synoptic situation in the region, with the observed structure and temporal evolution of the misovortices described in Section 4. In Section 5, we discuss the behavior of a low-level vortex during the landfall, and in Section 6, we summarize our findings and present our conclusions.

2. Observational settings and analysis procedure

2.1 Observational settings

We conducted field observations in the Shonai area, Yamagata Prefecture, Japan (Fig. 1). The observation area faces the Japan Sea and is suitable for studying
winter monsoon tornadoes and other gusty winds. The major observation facilities are the two X-band Doppler radars and a network of automated surface weather transmitters.

To assess the utility of Doppler radar in operational railroad warning systems, the East Japan Railway Company installed an X-band Doppler radar (JR-E radar) at Amarume Station in the Shonai area (Fig. 1) in March 2007 that has operated continuously since then (Kato et al. 2007). Its maximum observation range is 30 km, and its spatial resolution in the radial direction is 75 m. The antenna is 1.2 m in diameter and the beamwidth is 2.0°. To observe wind gusts with high temporal resolution, JR-E radar is operated in plan position indicator (PPI) mode at 2 rpm and at a low elevation angle (3.0°).

The Meteorological Research Institute (MRI) temporarily installed a portable X-band Doppler radar (XPOD: X-band, POrtable Doppler radar) on a rooftop of Shonai airport during the winter of 2007-2008. To observe the three-dimensional structure of wind gusts and their parent storms in high temporal resolution, the XPOD was operated in multiple PPI (6 elevations) and range height indicator (RHI, 2 azimuths) modes with updates every 3 minutes. The XPOD has a maximum observational range of 60 km and a spatial resolution of 150 m in the radial direction. The antenna is 1.2 m in diameter, and the beamwidth is 2.0°. Table 1 summarizes the characteristics and scanning modes of the JR-E radar and XPOD systems. Their azimuthal beamwidth of 2.0° corresponds to 350 m at a 10 km distance from the radar, which is smaller than the observed vortex core diameter in this study.

Surface weather stations (automated weather stations, AWS) were distributed in the observation field to characterize and validate the fine-scale structure of near-surface wind gusts and storms. Table 2

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**Fig. 1.** Map of the studied region (Shonai area, Yamagata Prefecture, Japan). Plus symbols (+) and large circles represent the JR-E radar and XPOD locations and their 30-km-radius observational ranges, respectively. Dotted circles demarcate the dual-Doppler analysis area, with the exception of the region of overlap. Topographic variation in the region is shown by 100-, 200-, and 500-m contours. Black and gray dots represent the locations of automated surface weather stations, and the black diamond symbol indicates the location of the Japan Meteorological Agency’s Sakata Weather Station.
summarizes the AWS characteristics. Twenty-six AWSs were installed approximately 4 km apart to cover the observational area around JR-E radar and XPOD (Fig. 1). The sensors were mounted on the top of steel poles approximately 5 m high. Each instrument measured wind direction and wind speed at 1-second intervals, and measured precipitation, pressure, temperature, and humidity at 10-second intervals. In this study, we mainly used data from stations B1 and B3 (Fig. 1), because the radar observations clearly showed that one of the vortices passed directly over these two stations.

### 2.2 Radar data processing and vortex detection

JR-E radar and XPOD are non-coherent magnetron-type systems that provide radar reflectivity and Doppler velocity data. The JR-E radar observations were performed using dual pulse repetition frequency (PRF) sampling of 900/1200 pps. Doppler velocity and reflectivity data were estimated with FFT using 64 pulses for each block. With an antenna scanning rate of 2 rpm and a pulse repetition frequency of 900 (1200) pps, a series of 64 pulses yields 0.85° (0.64°) in the azimuthal direction. This results in an azimuthal sampling interval of 0.75°, oversampling the beamwidth by approximately a factor of 3. Dual PRF sampling of 900/1200 pps yields a maximum unfolding velocity of ±27.5 m s⁻¹. A Hybrid Multi-PRI method (Yamauchi et al. 2006) was used to dealias Doppler velocities and further corrections were performed manually.

The XPOD observations were performed using a single PRF sampling of 1000 pps. Doppler velocity and reflectivity data were estimated with FFT using 64 pulses for each block. With an antenna scanning rate of 4 rpm and a pulse repetition frequency of 1000 pps.
The maximum unambiguous velocity was ± 7.64 m s⁻¹. The XPOD Doppler velocity was manually dealiased. We also manually eliminated ground clutter contamination around the vortices by referring to a ground-clutter map and removing the data points with Doppler velocity between −0.1 and 0.1 m s⁻¹. Because we did not calibrate the JR-E and XPOD radar reflectivity data, some differences in absolute echo intensity were found between them. Therefore, we do not provide discussions on absolute intensity. Rather, we use reflectivity data primarily to investigate the relative spatiotemporal change of snowband. After these corrections, we individually analyzed the data obtained by JR-E radar and XPOD to detect micro-scale vortices. We manually detected each vortex by identifying a Doppler velocity maximum ($V_{\text{max}}$) and minimum ($V_{\text{min}}$) couplet and then tracked its size and location on each PPI scan. To estimate the core diameter ($D$), peak tangential velocity ($V_t$), and vertical vorticity ($\omega_z$) of a vortex observed by JR-E radar, we used the method described by Inoue et al. (2011). We averaged the observed azimuthal angular distance $\phi$ between $V_{\text{max}}$ and $V_{\text{min}}$ and one-half the difference between $V_{\text{max}}$ and $V_{\text{min}}$ ($\Delta V/2 = (V_{\text{max}} - V_{\text{min}})/2$) of 5 PPI scans and used them to estimate $D$ and $V_t, \omega_z$ was calculated from the estimated $D$ and $V_t$ assuming solid-body rotation. Given that XPOD was operated at a volume scan update rate of 3 minutes, it is not appropriate to average $\phi$ and $\Delta V/2$ in time by using multiple PPI scans at the same elevation angle, because vortex structures could be significantly changed during the averaged time period. Therefore, to estimate $D, V_t$ and $\omega_z$ of each vortex observed by XPOD, $\phi$ and $\Delta V/2$ were calculated for each PPI scan by using the Rankine vortex model which best fits the observed data based on the least squares fitting analysis.

We also performed a dual-Doppler analysis to derive horizontal wind fields around the vortex. The Doppler radar data were interpolated into a Cartesian grid with a 100-m horizontal grid interval and 200-m vertical grid interval using a Cressman-type weighting function. Horizontal wind fields were obtained by a standard dual-Doppler analysis based on the method of Ishihara et al. (1986). Because only the 3°-elevation scan data were available from the JR-E radar, derivation of vertical winds was not possible. Analyses were constrained to an area in which the radar beam intersection angles are between 30° and 150° (Fig. 1). Time differences between the JR-E radar scan and the XPOD scans were less than 1 minute for each wind synthesis volume. The average moving speed of the vortex was used to adjust the time differences between the JR-E radar and XPOD volumetric scans.

3. Synoptic situation

Figure 2 shows the sea-level pressure and wind-field vectors at 0300 LST on December 31, 2007, approximately one hour before passage of the vortices, obtained from the Japan Meteorological Agency (JMA) mesoscale objective analysis (MANAL) data with a horizontal resolution of 10 km (available every three hours). The surface-pressure pattern indicates a northeast-southwest oriented pressure gradient, which is typical during the winter monsoon in this area. A weak pressure trough was located at the southwest of the observational area (dashed line in Fig. 2).

Southwest–northeast oriented snowbands were continuously observed around the observational area (Fig. 3), although neither graupel nor lightning was observed at the JMA’s Sakata Weather Station (see Fig. 1 for location) during their passage. Among these snowbands, the parent snowband of the vortices (indicated by the arrow in Fig. 3) showed the largest radar reflectivity. Wind profiler observations at the Sakata Weather Station for altitudes < 2 km AGL recorded west-southwesterly winds 30 minutes prior to the passage of the parent snowband, which changed into stronger westerly winds after its passage (not shown).

No upper-air sounding associated with the passage of the convective snowband was available for this study because upper-air soundings of thermodynamic quantities were sparse in both time and space. (The nearest sounding was performed at Akita on 2100 LST on the previous day.) As such, we calculated the environmental thermodynamic sounding quantities from the MANAL data. Typical thermodynamic profiles in the study area at 0300 LST on December 31 showed a small convective instability below 2.5 km (not shown). The convective available potential energy (CAPE) calculated for the parcel with the highest equivalent potential temperature below the 850 hPa level was between 100 and 150 J kg⁻¹. The equilibrium level (EL) was between 4.0 and 4.5 km.

4. Detailed analysis of the vortices

4.1 Characteristics of the parent snowband and the vortices

Figure 4 shows reflectivity and Doppler velocity fields for the parent snowband, which formed ~30 km
offshore and intensified as it approached the coastline. The echo top height reached ~ 4–5 km, which is approximately equal to the EL estimated from the MANAL data (Section 3).

The surface wind exhibited a distinct convergence ($1.2 \times 10^{-2}$ s$^{-1}$) and cyclonic shear ($0.3 \times 10^{-2}$ s$^{-1}$) across the snowband. Weak southwesterly to westerly flow prevailed ahead of it, whereas northwesterly flow prevailed behind it (Fig. 4). The difference in the alongfront wind velocity across the low-level convergence line was estimated to be 3.2 m s$^{-1}$ on average, which is smaller than the lowest end of the differential velocity in warm-season nonsupercell tornadoes (5 to 30 m s$^{-1}$; Lee and Wilhelmson 1997a). RHI observations (Fig. 5) also showed low-level convergence and the existence of an overhang structure in the snowband. The depth of the southwesterly flow ahead of the snowband was estimated to be ~ 1.0 km.

At least five cyclonic vortices were observed along the low-level wind shear line in the snowband (labeled 1 to 5 in Fig. 4), which were spaced ~ 4–9 km apart from one another. Horizontal shearing instabilities have frequently been suggested to cause such vortices to form spontaneously along wind-shear lines (e.g., Carbone 1982; Mueller and Carbone 1987; Wakimoto and Wilson 1989; Lee and Wilhelmson 1997a), and linear stability theory suggests that the fastest growing mode for such an instability should occur at a wavelength of ~ 7.5-times the width of the shear zone (Miles and Howard 1964). In the studied example, RHI observations allowed the estimation of the width of the shear zone as 0.5–1.0 km based on the radial gradient of the Doppler velocity. As such, the most unstable wavelength for vortices would be between 3.8 and 7.5 km, which is consistent with the observed spacing of the vortices (4–9 km).

The characteristics of three vortices that made landfall near the JR-E radar and XPOD (vortex numbers 2, 3, and 4 on Fig. 6) were examined in detail. All had similar lifetimes of 15–23 minutes, moving speeds of ~ 12–14 m s$^{-1}$, core diameters of ~ 400–1900 m, peak tangential velocities of 7–11 m s$^{-1}$, and vertical vorti-
cities on the order of $10^{-2}$ to $10^{-1}$ s$^{-1}$ (Table 3).

4.2 Structure of vortex 4 and its temporal evolution

The vortices were initially detected over the Japan Sea and subsequently moved eastward towards the coastline (Fig. 6). Among them, vortex 4 passed directly over surface stations, and was observed at a close range (< 10 km) by both radars. Therefore, it is possible to analyze a temporal evolution of a fine-scale vertical structure of the vortex in detail.

a. Surface in-situ observations of the vortex

The JR-E radar observations showed that the right-hand side of vortex 4 passed over station B1 and that the center passed over station B3. Because the moving speeds of the vortices were slightly larger than their peak tangential velocities (Table 3), the distribution of ground-relative wind speeds became highly asymmetric. The strongest winds that could cause surface damage occurred on the right-hand side of the vortices relative to their forward motions, where the directions of translation and rotation were the same (Fig. 4, left panels: Doppler velocity distribution of vortices 1–5).

At station B1, strong surface gusts of 25.0 m s$^{-1}$ were recorded as the right-hand side of vortex 4 passed over the station (Fig. 7). An increase in wind speed of ~ 20 m s$^{-1}$ coincided with a pressure drop of ~1 hPa, both of which suggest a vortex passage. The temporal variations associated with the passage of the convergence line, such as an increase of wind speed (see Fig. 7a; prefrontal speed of ~ 5 m s$^{-1}$ around 0400 LST to postfrontal speed of ~ 15 m s$^{-1}$ around 0410 LST), a change of wind direction from westerly to northwesterly, and a slight decrease of pressure, were also superimposed in Fig. 7. At station B3, temporal variations suggestive of a vortex passage, a sudden change of wind speed and direction, and a pressure drop of ~ 0.6 hPa were recorded as the center of vortex 4 passed over the station (not shown in the figure). No significant temperature variation was observed at either station during the vortex passage.

b. Vertical structure and temporal evolution of the vortex

To examine the vertical structure and temporal evolution of vortex 4, we investigated variations in
the vertical distribution of core diameters as observed by XPOD (Fig. 8). The vortex was initially confined to low altitudes of \( \sim 500 \) m ASL between 0350:14 LST and 0354:38 LST, after which it began to extend upwards. At a distance of \( \sim 3 \) km offshore, the vortex extended above 2.5 km ASL (timestamps 0401:30 to 0403:05 LST on Fig. 8), which is the maximum elevation that can be observed by XPOD and is also nearly half of the level of the echo top of the snowband. The vortex initiated along a low-level convergence line and developed upward without any preceding mesocyclones as described in Section 4.1, which is a development pattern similar to that reported in previous radar observations of nonsupercell tornadoes in much
deeper convective clouds (e.g., Wakimoto and Wilson 1989; Roberts and Wilson 1995).

Structural changes, such as the variation of core diameter and an axis tilt accompanied vertical extension of the vortex. Immediately following its growth to at least 2.5 km ASL (timestamps 0401:30 to 0403:05 LST on Fig. 8), the core diameter of the low-level vortex was observed to be slightly larger than that present at higher altitudes and the upper part of the vortex (approximately above 1.5 km ASL) slightly tilted downshear. After the vortex crossed the coastline (0407:08 to 0408:43 LST on Fig. 8), its core diameter below ~ 700 m ASL decreased significantly. As the vortex moved further inland (after 0409:57 LST), its low-level vortex structure started to lean downshear. This is consistent with previous observational studies that have reported an increase in forward tilt with height after making landfall (Kato et al. 2015).

Structural changes of the low-level snowband during vortex development are displayed in the reflectivity field observed around vortex 4 (Fig. 9). As

| Table 3. Characteristics of vortices 2, 3, and 4 (shown in Fig. 4) derived from JR-E radar data. The lifetime (T), average eastward (v_e) and northward (v_n) moving speeds, range of estimated core diameter (D), peak tangential velocity (V_t), and vertical vorticity (ω_z) are given for each vortex. |
|---|---|---|---|
|   | 2    | 3    | 4    |
| T (minutes) | 15   | 15   | 23   |
| v_e (m s^{-1}) | 13.7 | 12.1 | 11.5 |
| v_n (m s^{-1}) | -1.2 | -0.5 | -0.5 |
| D (m) | 700-1900 | 600-1600 | 400-1700 |
| V_t (m s^{-1}) | 7-9   | 8-9   | 8-11  |
| ω_z (10^{-2} s^{-1}) | 1-5   | 2-6   | 2-10  |
| Rotation | cyclonic | cyclonic | cyclonic |

Fig. 5. Radar reflectivity (top) and Doppler velocity (bottom) of the parent snowband observed by XPOD. (a) Vertical cross section obtained at 0352:51 LST and (b) PPI scan performed at 0353:22 LST at a scanning elevation of 4.5°. The approximate location of the vertical cross section in (a) is shown by the dotted line A-A’ in (b). Filled arrows in (a) indicate the location of the low-level convergence. The locations of vortices 3 and 4 are shown by the circles (The dashed line means that the vortex was only detected in the lower PPI scan.).
the vortex developed upward, a kink in the adjacent snowband became progressively more pronounced, eventually developing into a hook-shaped echo that encircled the vortex (Figs. 9a–c, timestamps 0356:20 to 0402:09 LST). The end of the hook-shaped echo was subsequently cut off from the snowband (Fig. 9d, 0403:37 LST) and started to shrink (Figs. 9e, f, 0407:29 to 0409:55 LST).

To explore the temporal evolution of the low-level vortex, we examined the variations in the core diameter, peak tangential velocity, and vertical vorticity estimated by JR-E radar (Fig. 10). Although the beam height of the observed vortex decreased with time as shown in Fig. 10, XPOD observation (Fig. 8) suggests that vortex feature below 700 m ASL did not show much variation with height. Therefore, it is most likely that Fig. 10 represents temporal variations of low-level vortex parameters. Following the vortex reaching a distance of ~ 2 km offshore (~ 0403 LST in Fig. 10), its core diameter started to decrease rapidly.

This reduction continued after making landfall until ~0410 LST, when its core diameter reached one-third of its pre-landfall value. The peak tangential velocity and the vertical vorticity of the vortex increased coevally with this low-level contraction. Given that such an increase in peak tangential velocity and vertical vorticity during a landfall as observed during this event is remarkably different from the decrease in peak tangential velocity reported in Kato et al. (2015), we investigated the time-dependent changes of the low-level wind field around the vortex and the parent storm behavior in order to discern the mechanisms responsible.

Figure 11 shows the horizontal wind vectors relative to vortex 4, the divergence, and the vertical vorticity field pattern derived from dual-Doppler analysis. These data show that low-level convergence around the vortex and convergence line began to increase at approximately the same time when the vortex began to contract (0404:35 LST, Fig. 11b),
with the low-level convergence having increased further during continued vortex contraction (0407:29 LST, Fig. 11c). This intensification of low-level convergence is also evident in the JR-E radar data (not shown here). The 30-sec updates of Doppler velocity field show a clear temporal increase in radial convergence calculated across the convergence line near the vortex 4. Because the radar beam direction crossed nearly perpendicular to the convergence line during this time, we note that the observed increase was not due to the geometry change. These observations suggest that stretching of the low-level vortex associated with the updraft due to the low-level convergence was responsible for the low-level vortex contraction and increase in peak tangential velocity and vertical vorticity in our case.

It is notable that such an intensification of the low-level convergence was observed after the snowband reached its mature stage. Figure 12 shows a time-height plot of the maximum radar reflectivity around vortex 4, which reveals that the isocontour of the maximum radar reflectivity (43 and 45 dBZ) descended just before the vortex began to contract (0358 to 0400 LST), suggesting the emergence of a precipitation-induced downdraft around vortex 4.

5. Discussion

As mentioned in Section 4, the vortex showed significant variation during the landfall. The upper part of the vortex slightly tilted greenshear just before the landfall and the associated hook echo gradually became unclear after that (not shown), suggesting a decay of the upper part of the vortex. In contrast, the low-level vortex exhibited a contraction of the vortex core diameter and an increase in both the peak tangential velocity and vertical vorticity during a
landfall, which is a more rapid change than the decay of the upper part. These findings, especially on the temporal evolutions of the low-level vortex, are markedly different from previous studies (e.g., Kusunoki et al. 2011; Inoue et al. 2011; Kato et al. 2015). Inoue et al. (2011) and Kato et al. (2015) demonstrated that the peak tangential velocity clearly decreased (not increased) during the landfall. Kusunoki et al. (2011) reported shrinkage of low-level vortex without a signature of a coeval increase in peak tangential velocity. Because these studies did not discuss the storm characteristics, we cannot provide in-depth discussions on the key physical processes producing different vortex characteristics by means of comparative analysis. However, it is evident that increasing peak tangential velocity and vertical vorticity with decreasing diameter of the low-level vortex during a landfall is a unique feature found in our case study, and is worth paying more attention in relation to the parent storm dynamics.

As shown in Fig. 11, during the time period of the low-level vortex contraction (0403-0408 LST; see Fig. 10), low-level convergence around the vortex and the convergence line located at the front side of snowband significantly increased (Figs. 11b, c). Prior to this time, the time-height plot in Fig. 12 shows a clear descent of precipitation core at 0358 - 0400 LST, suggesting a downdraft formation. To better understand the wind field structure associated with this core downfall, we show RHI and PPI sections obtained at 0401 LST in Fig. 13. Figure 13a shows RHI data obtained along the beam direction that crosses the snowband at approximately 4 km southwest of the vortex 4. The Doppler velocity field implies that a westerly inflow entrained from the rear side of the snowband at ~ 1.4 km ASL (see the open arrow in the middle panel of Fig. 13a ), tilted downward, and came out from the front side at even lower altitudes (below ~ 0.4 km ASL). Judging from the occurrence time and location of the descending precipitation core (Fig. 12),
Fig. 9. Doppler velocity (top row) and radar reflectivity (bottom row) of vortex 4 (red circle) by JR-E radar at (a) 0356:20 LST, (b) 0359:14 LST, (c) 0402:09 LST, (d) 0403:37 LST, (e) 0407:29 LST and (f) 0409:55 LST. The gray line represents the coastline. The locations of the panels are shifted and centered on vortex 4. (The location of panel (c) is shown in Fig. 4a.)

Fig. 10. Variations in core diameter, peak tangential velocity, and vertical vorticity of vortex 4 derived from JR-E radar observations. The beam height at the time of each scan is also plotted. The times at which vortex 4 passed over stations B1 and B3 are also shown.
Fig. 11. Enhanced views of vortex 4 at (a) 0401:40 LST, (b) 0404:35 LST, and (c) 0407:29 LST. The horizontal cross-section of divergence (shade), vertical vorticity (red contour, outermost contour is $1.0 \times 10^{-2}$ s$^{-1}$, incremented by $0.5 \times 10^{-2}$ s$^{-1}$), and vortex-relative horizontal wind vectors at 400 m ASL. Radar reflectivity of the JR-E radar is shown by the black contour (23 dBZ). The gray line shows the coast line and the dot shows the location of AWS B1. The dashed rectangular region shows the region used for deriving maximum radar reflectivity in Fig. 12.

Fig. 12. Time-height plot of the isocontour of the maximum radar reflectivity around vortex 4 (dashed rectangular region in Fig. 11) observed by XPOD. The dots show the time and height of the observed vortices, plus symbols (+) indicate that the vortex was not observed at that time and height, and cross symbols (×) means that no observation was made due to a beam cut. The hatched area represents the area of no data collection.
the results in Fig. 13a suggest a downward transport of the horizontal momentum of rear inflow caused by the precipitation-induced downdraft. As seen in the reflectivity and radial divergence fields (Fig. 13a, the bottom panel), the lower portion of the snowband was advected forward, and the low-level radial convergence was observed at the leading edge of the outflow. The low-level radial convergence zone is also evident in the PPI map shown in Fig. 13b (the bottom panel), as a cold color region extending along the front side of the snowband to the immediate south of the vortex 4. These facts suggest that this low-level outflow was responsible for the intensification of the low-level convergence at the periphery of the vortex 4. Lee and Wilhelmson (1997b) showed that, in the late mature stage of the nonsupercell tornado, precipitation-induced outflow can increase low-level convergence and the associated vortex stretching. Although their case is not the same phenomena as the event analyzed in the present study (snowband during cold-air outbreaks), it is suggested that the low-level outflow observed in our case is also related to the intensifi-
cation of low-level convergence and the subsequent vortex stretching.

As shown in Fig. 6, the vortex 3 also contracted during the landfall. Because the vortex 3 was located adjacent to the vortex 4 and the low-level convergence at the periphery of the vortex 3 was also large during the landfall (Fig. 11), it is suggested that the contraction of the vortex 3 happened through a similar process. Unlike the vortices 3 and 4, the vortex 1 did not show much variation and the vortex 2 even expanded as they landed (Fig. 6). A remarkable difference of the vortices 1 and 2 from the vortices 3 and 4 was that the low-level reflectivity fields associated with the vortices 1 and 2 were much weaker and broader than those associated with vortices 3 and 4 during the landfall. Although differences in low-level convergence might play a crucial role in producing the discrepancies in vortex properties, further in-depth discussions cannot be made because the vortices 1 and 2 were located at more than 15 km from the radars, that is too far for fine scale analyses.

It is notable that radar reflectivity around the vortex 4 began to increase around 0350 LST (~10 km from the coast) and then the precipitation core descended just before the landfall (see Fig. 12). Such a pre-landfall modification of the parent snowband was also observed by the JMA Niigata radar (not shown). Previous studies showed the intensification of snow clouds near the coast and their dissipation after a landfall. Snow clouds are intensified near the coast due to various factors; the variation in vertical wind shear (Takeda et al. 1982), locally produced convergence by orographically deflected winds (Yoshihara et al. 2004; Ohigashi et al. 2014), and interaction with land-breeze (Ishihara et al. 1989; Tsuboki et al. 1989; Ohigashi and Tsuboki 2005). It is also known that, due to the larger surface roughness and resultant larger cross-isobaric wind component over land, mesoscale convergence zone can be generated along the coast downwind of the prevailing wind (Markowski and Richardson 2010). Similarly, it is possible that the observed modification of parent snowband in our case (intensification at 10 km offshore and dissipation during the landfall) was related to the landfall.

Previous studies suggested that surface roughness could also modify vortex characteristics during a landfall. Kato et al. (2015) reported a decreasing peak tangential velocity, and they posited this modification as a result of the effect of surface roughness. Although such an effect might have also worked to some extent, results obtained here suggest that the increase in low-level convergence was probably more influential on the vortex modification than surface roughness difference in our case. Our results suggest that in addition to the effect of surface roughness difference, other controlling factors such as parent storm behavior and the modification of low-level wind fields should be considered in investigations of vortex modification during a landfall. Higher-resolution observation of both vortex and parent storm characteristics using fast-scanning Doppler radars would be one of the most effective techniques for examining these factors. Future investigation of vortex behavior during a landfall should also consider the application of numerical modeling in order to determine the relative importance of each of these factors.

6. Summary and conclusions

Misovortices including tornadoes are frequently observed in the Japan Sea coastal region during cold-air outbreaks. These vortices develop over the sea and move inland, causing gusty winds and damage. On December 31, 2007, several misovortices were observed within a convective snowband. In order to investigate the structure and evolution of these vortices during a landfall, high temporal and spatial resolution data obtained by two X-band Doppler radars were analyzed. The observed vortices developed along the low-level convergence line with distinct convergence ($1.2 \times 10^{-2} \text{ s}^{-1}$) and cyclonic shear ($0.3 \times 10^{-2} \text{ s}^{-1}$) across the snowband, suggesting that horizontal shearing instability was responsible for the initial development of the vortices. They had lifetimes of 15–23 minutes, moving speeds of ~12–14 m s$^{-1}$, core diameters of ~400–1900 m, peak tangential velocities of 7–11 m s$^{-1}$, and vertical vorticities on the order of $10^{-2}$ to $10^{-1} \text{ s}^{-1}$.

One particular vortex, which was observed at close range by both Doppler radars and directly over two surface weather stations, was analyzed in detail to investigate its temporal evolution during the landfall in relation to the parent storm behavior and low-level wind fields. The vortex showed significant temporal variations during the landfall; initiated at lower altitudes and subsequently grew upward as approaching the coastline. During the landfall, the low-level core diameter significantly decreased, while the peak tangential velocity and vertical vorticity increased. Concurrently, a significantly increased low-level convergence was observed around the vortex and at the low-level convergence line. These facts suggest that updrafts associated with the low-level convergence caused the stretching of low-level vortex and resulted in the observed increase in peak tangen-
tial velocity and vertical vorticity during a landfall. Based on the reflectivity and Doppler velocity data of the parent snowband, it is most likely that the downward transfer of the horizontal momentum of rear inflow caused by the descending precipitation core was responsible for the intensification of the low-level convergence. These findings suggest the importance of studying temporal change of the vortex during a landfall in relation to their parent storm system, paying more attention to low-level wind fields.

Previous studies reported the effect of surface roughness in the vortex modification (Inoue et al. 2011; Kato et al. 2015). In contrast, results obtained in our case study suggest that the increase in low-level convergence was probably more influential on the vortex modification than the surface roughness difference. In addition to the previously mentioned surface roughness difference, other factors that are known to affect vortex behavior (e.g., parent storm behavior, low-level wind fields, and interaction with land breeze) should also be considered in investigations of vortex modification during a landfall. Observation by fast-scanning radar would be extremely useful to elucidate critical physical processes associated with the storm landfall.

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References

Carbone, R. E., 1982: A severe frontal rainband. Part I. Stormwide hydrodynamic structure. J. Atmos. Sci., 39, 258–279.

Fujita, T. T., 1981: Tornadoes and downbursts in the context of generalized planetary scales. J. Atmos. Sci., 38, 1511–1534.

Inoue, H. Y., K. Kusunoki, W. Kato, H. Suzuki, T. Imai, T. Takemi, K. Bessho, M. Nakazato, S. Hoshino, W. Mashiko, S. Hayashi, T. Fukuhara, T. Shibata, H. Yamauchi, and O. Suzuki, 2011: Finescale Doppler radar observation of a tornado and low-level mesocyclones within a winter storm in the Japan Sea coastal region. Mon. Wea. Rev., 139, 351–369.

Ishihara, M., Z. Yanagisawa, H. Sakakibara, K. Matsuura, and J. Aoyagi, 1986: Structure of a typhoon rainband observed by two Doppler radars. J. Meteor. Soc. Japan, 64, 923–939.

Ishihara, M., H. Sakakibara, and Z. Yanagisawa, 1989: Doppler radar analysis of the structure of mesoscale snow bands developed between the winter monsoon and the land breeze. J. Meteor. Soc. Japan, 67, 503–520.

Kato, R., K. Kusunoki, H. Y. Inoue, K. Arai, M. Nishihashi, C. Fujitwara, K. Shimose, W. Mashiko, E. Sato, S. Saito, S. Hayashi, S. Yoshida, and H. Suzuki, 2015: Modification of misovortices during landfall in the Japan Sea coastal region. Atmos. Res., 158–159, 13–23.

Kato, W., H. Suzuki, M. Shimamura, K. Kusunoki, and T. Hayashi, 2007: The design and initial testing of an X-band Doppler radar for monitoring hazardous winds for railroad system. 33rd Conf. on Radar Meteorology, Amer. Meteor. Soc., Cairns, Australia, p.13A.15.

Kobayashi, F., Y. Sugimoto, T. Suzuki, T. Maesaka, and Q. Moteki, 2007: Doppler radar observation of a tornado generated over the Japan Sea coast during a cold air outbreak. J. Meteor. Soc. Japan, 85, 321–334.

Kusunoki, K., T. Imai, H. Suzuki, T. Takemi, K. Bessho, M. Nakazato, W. Mashiko, S. Hayashi, H. Inoue, T. Fukuhara, T. Shibata, and W. Kato, 2008: An overview of the Shonai area railroad weather project and early outcomes. 5th European Conference on Radar in Meteorology and Hydrology, Helsinki, Finland, 12.1.

Kusunoki, K., T. Imai, H. Suzuki, T. Takemi, K. Bessho, M. Nakazato, W. Mashiko, S. Hayashi, H. Inoue, T. Fukuhara, T. Shibata, and W. Kato, 2009: Wind gust and storm evolutions observed during the Shonai area railroad weather project: A preliminary survey. 5th European Conference on severe storms, Eur. Meteor. Soc., Landshut, Germany, 9–15.

Kusunoki, K., H. Inoue, M. Nakazato, K. Bessho, S. Hoshino, W. Mashiko, S. Hayashi, H. Morishima, and K. Adachi, 2011: The vertical structures within a winter tornadic storm during landfall over the Japan Sea area. 6th European Conference on severe storms, Eur. Meteor. Soc., Palma de Mallorca, Balearic Islands, Spain, 213–213.

Lee, B. D., and R. B. Wilhelmson, 1997a: The numerical simulation of non-supercell tornadogenesis. Part I: Initiation and evolution of pretrough mesocyclone circulations along a dry outflow boundary. J. Atmos. Sci., 54, 32–60.

Lee, B. D., and R. B. Wilhelmson, 1997b: The numerical simulation of non-supercell tornadogenesis. Part II: Evolution of a family of tornadoes along a weak outflow boundary. J. Atmos. Sci., 54, 2387–2415.

Markowski, P., and Y. Richardson, 2010: Mesoscale Meteorology in Midlatitudes. Wiley-Blackwell, 430 pp.
Miles, J. W., and L. N. Howard, 1964: Note on heterogeneous shear flow. *J. Fluid Mech.*, 20, 331–336.
Mueller, C. K., and R. E. Carbone, 1987: Dynamics of a thunderstorm outflow. *J. Atmos. Sci.*, 44, 1879–1898.
Niino, H., T. Fujitani, and N. Watanabe, 1997: A statistical study of tornadoes and waterspouts in Japan from 1961 to 1993. *J. Climate*, 10, 1730–1752.
Ohigashi, T., and K. Tsuboki, 2005: Structure and maintenance process of stationary double snowbands along the coastal region. *J. Meteor. Soc. Japan*, 83, 331–349.
Ohigashi, T., K. Tsuboki, Y. Shusse, and H. Uyeda, 2014: An intensification process of a winter broad cloud band on a flank of the mountain region along the Japan-Sea coast. *J. Meteor. Soc. Japan*, 92, 71–93.
Roberts, R. D., and J. W. Wilson, 1995: The genesis of three nonsupercell tornadoes observed with dual-Doppler radar. *Mon. Wea. Rev.*, 123, 3408–3436.
Takeda, T., K. Isono, M. Wada, Y. Ishizaka, K. Okada, Y. Fujiyoshi, M. Maruyama, Y. Izawa, and K. Nagaya, 1982: Modification of convective snow-clouds in landing the Japan Sea coastal region. *J. Meteor. Soc. Japan*, 60, 967–977.
Tsuboki, K., Y. Fujiyoshi, and G. Wakahama, 1989: Structure of a land breeze and snowfall enhancement at the leading edge. *J. Meteor. Soc. Japan*, 67, 757–770.
Wakimoto, R. M., and J. W. Wilson, 1989: Non-supercell tornadoes. *Mon. Wea. Rev.*, 117, 1113–1140.
Yamauchi, H., O. Suzuki, and K. Akaeda, 2006: A Hybrid multi-PRI method to dealias Doppler velocities. *SOLA*, 2, 92–95.
Yoshihara, H., M. Kawashima, K. Arai, J. Inoue, and Y. Fujiyoshi, 2004: Doppler radar study on the successive development of snowbands at a convergent line near the coastal region of Hokuriku district. *J. Meteor. Soc. Japan*, 82, 1057–1079.