Composite Analysis of Polar Mesocyclones over the Western Part of the Sea of Japan

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ABSTRACT

Polar mesocyclones occur frequently over the Sea of Japan during winter in association with cold air outbreaks from the Eurasian continent. In this study, the general characteristics of polar mesocyclones over the western part of the Sea of Japan are examined using composite analysis. The synoptic-scale environment associated with these mesocyclones is characterized by a negative sea level pressure anomaly to the east that causes a cold air outbreak at low levels. There is also a geopotential height trough moving eastward at upper levels. In the cold air outbreak, a convergence zone known as the Japan Sea polar airmass convergence zone (JPCZ), collocated with a thermal ridge, develops on the lee side of the mountains at the root of the Korean Peninsula. These polar mesocyclones are generated when the upper-level trough approaches the JPCZ from the west. However, the behavior of the JPCZ and the movement of the polar mesocyclones differ depending on the location of the upper-level trough. A piecewise potential vorticity inversion analysis revealed that the circulation associated with the upper-level trough modifies the low-level winds, which affects the direction of extension of the JPCZ as well as the genesis location and the movement of the polar mesocyclones.

1. Introduction

Maritime polar mesocyclones are mesoscale cyclonic vortices that develop poleward of the main polar front (Rasmussen and Turner 2003). They form and develop over high-latitude oceans during marine cold air outbreaks over a relatively warm sea (Kolstad 2011). These areas are characterized by large sensible and latent heat fluxes from the sea surface, active cumulus convection, and a shallow baroclinic zone. Note that intense polar mesocyclones are known as polar lows, and produce gale-force winds near the surface (Rasmussen and Turner 2003).

Although the Sea of Japan (Fig. 1) is located at relatively low latitudes (around 40°N), it is one of the regions where polar mesocyclones occur frequently (Asai 1988; Ninomiya 1989). This is because, during winter, cold air outbreaks frequently occur over the Sea of Japan (Iwasaki et al. 2014). In addition, the warm Tsushima Current (a branch of the Kuroshio) causes the sea surface temperature (SST) to be relatively high. When cold air outbreaks occur, the dry-bulb temperature (hereafter “temperature” indicates dry-bulb temperature unless stated otherwise) of the cold polar air from Siberia is typically lower than −10°C. The winter-mean SST of the western Sea of Japan is about 10°C (Fig. 1), although the SST gradually decreases from about 15°C in November to about 9°C in March. The typical temperature difference between the atmosphere and ocean during a cold air outbreak is about 20°C. These factors provide favorable conditions for the genesis and development of polar mesocyclones over the Sea of Japan. Statistical analysis has shown that polar mesocyclones are concentrated in two main regions: the western and northeastern areas of the Sea of Japan (Asai 1988; Yanase et al. 2016; Watanabe et al. 2016).

Some kind of triggering mechanism is required to initiate a polar mesocyclone. An upper-level potential vorticity (PV) anomaly commonly triggers a polar mesocyclone, although not always. Montgomery and Farrell (1992) showed that a mobile upper-level PV anomaly can rapidly spin up an incipient surface cyclonic circulation, and such a process has been observed in several case studies (Claud et al. 2004; Bracegirdle and Gray 2009; Shimada et al. 2014). At low levels, on the other hand, mesoscale structures in a cold air outbreak, such as a shallow front located at the outer edge of a cold air outbreak (Føre et al. 2011), an occluded front (Bond and Shapiro 1991), a topographically
induced cyclonic circulation (Moore and Vachon 2002; Kristjánsson et al. 2011), and a shear line associated with a convergence zone (Bresch et al. 1997; Laird et al. 2001; Sergeev et al. 2017), are also important for initiating polar mesocyclones.

Over the western Sea of Japan, a majority of polar mesocyclones tend to form within a convergence zone referred to as the Japan Sea polar airmass convergence zone (JPCZ; Asai 1988). The JPCZ forms between low-level northerly and westerly winds on its northeastern and southwestern sides, respectively, so that it is accompanied by cyclonic horizontal shear and active cumulus convection in addition to strong convergence. Previous studies (Nagata et al. 1986; Nagata 1991) suggest that the land–sea thermal contrast between the Korean Peninsula and the Sea of Japan, the blocking effect of the Changbai Mountains, which rise to an altitude of more than 2000 m at the northern root of the Korean Peninsula (Fig. 1), and the large SST gradient around 40°N (Fig. 1) contribute to the formation of the JPCZ. The JPCZ usually extends from the northwestern corner of the Sea of Japan in a direction that varies between southerly and easterly. In addition, the JPCZ sometimes shifts to the north or south, and intensifies or weakens in association with the passage of the upper-level geopotential height trough (hereafter “trough” indicates geopotential height trough unless stated otherwise) (Ohigashi and Tsuboki 2007).

The genesis location of polar mesocyclones varies from case to case (Watanabe et al. 2016) between the northwestern Sea of Japan (e.g., Nagata 1993; Watanabe and Niino 2014) and the southwestern Sea of Japan (e.g., Shimada et al. 2014; Choi et al. 2018), which correspond to the start and end points of the JPCZ, respectively. Occasionally, several polar mesocyclones appear simultaneously within the JPCZ (Ninomiya and Hoshino 1990; Nagata 1993; Tsuboki and Asai 2004). In such cases, they often move in different directions: some to the east-northeast and others to the southeast. Using serial satellite images, Ookubo (1995) demonstrated that polar mesocyclones moving east-northeast are located at the point where the JPCZ bends to the north, whereas those moving southeastward are located at the point where the JPCZ bends to the south. The movement of polar mesocyclones and the location of the JPCZ are an important problem in forecasting severe weather because they bring heavy snowfall. However, what causes these variations of the JPCZ and associated polar mesocyclones remains to be clarified. In the present study, we address these variations in terms of their environment using composite analyses.

The horizontal scale of the polar mesocyclones observed within the JPCZ ranges from the meso-β scale (20–200 km) to the meso-α scale (200–2000 km). Meso-β-scale polar mesocyclones within the JPCZ are thought to develop mainly through barotropic instability (Asai and Miura 1981; Nagata 1993; Tsuboki and Asai 2004; Watanabe and Niino 2014), whereas meso-α-scale polar mesocyclones over the Sea of Japan develop mainly through baroclinic instability (Yamagishi et al. 1992; Shimada et al. 2014). Note that condensational heating is important for both meso-α-scale and meso-β-scale polar mesocyclones (Shimada et al. 2014; Watanabe and Niino 2014). Meso-α-scale polar mesocyclones tend to be generated in the southwestern Sea of Japan (Yamagishi et al. 1992; Shimada et al. 2014). On the other hand, a meso-β-scale polar mesocyclone generated in the northwestern Sea of Japan sometimes develops into a meso-α-scale polar mesocyclone as it moves southeastward (Lee et al. 1998). Maejima and Iga (2012) showed that the horizontal scale and development mechanism of the mesocyclones embedded within an idealized frontal structure similar to a JPCZ depend on the static stability. Thus, examining the environment in which polar mesocyclones develop is important to our understanding of the differences in their dynamics and structure.

Most of the previous studies mentioned above are case studies. However, the characteristics of polar mesocyclones (as well as the JPCZs) in these studies vary from case to case, which makes it difficult to obtain a comprehensive understanding of the general characteristics of polar mesocyclones in the Sea of Japan.
A promising approach to establishing these general characteristics is to examine the average characteristics of many polar mesocyclones (Businger 1985; Bracegirdle and Gray 2008; Blechschmidt et al. 2009; Kolstad 2011; Mallet et al. 2013; Rojo et al. 2015; Yanase et al. 2016; Terpstra et al. 2016; Watanabe et al. 2017). Although Yanase et al. (2016) analyzed the environment of meso-α-scale polar lows over the Sea of Japan, they did not consider meso-β-scale polar mesocyclones, which are often observed over the western Sea of Japan (Watanabe et al. 2016). Recently, Watanabe et al. (2017) analyzed polar mesocyclones observed over the northeastern Sea of Japan and described their environmental field using composite analysis. However, the polar mesocyclones over the western Sea of Japan are strongly affected by the JPCZ and differ from those over the northeastern Sea of Japan. Here, we apply a similar approach to polar mesocyclones over the western Sea of Japan with the aim of revealing their general characteristics, including their environment and their relationship with the JPCZ.

The remainder of the paper is organized as follows. Our data and methodology are described in section 2, the results of our composite analyses are described in section 3, and section 4 contains a discussion and our conclusions.

2. Data and methodology

a. Data

We used the Japan Meteorological Agency’s (JMA) mesoscale analysis and global analysis (JMA 2013) to track the polar mesocyclones and for the composite analysis. The mesoscale analysis is based on the JMA Mesoscale Model (JMA 2013) with four-dimensional variational data assimilation. This model has a rectangular domain of 3600 km × 2880 km that covers the Japanese Islands and surrounding areas. The original data have a horizontal grid spacing of 5 km on a Lambert conformal conic map projection with 50 vertical levels, and a temporal resolution of 3 h. For our analysis, we used a dataset prepared by linearly interpolating the original data onto geographic coordinates with a horizontal grid spacing of 0.1° in latitude, 0.125° in longitude (approximately 10 km around the Sea of Japan), and 16 pressure levels. Note that JMA releases mesoscale analysis only in this format. This resolution, however, is sufficient to examine the characteristics of most meso-β-scale polar mesocyclones (Watanabe et al. 2016).

As the mesoscale analysis does not include diabatic heating as a variable, we used the formula proposed by Emanuel et al. (1987) to calculate the diabatic heating:

\[ \frac{d\theta}{dt} = \omega \left( \frac{\partial \theta}{\partial p} - \frac{\Gamma_m}{\Gamma_d} \frac{\theta}{\partial p} \right), \]

where \( \theta \) is potential temperature, \( \theta_e \) is equivalent potential temperature, \( p \) is pressure, \( \omega \) is the vertical \( p \) velocity, \( \Gamma_m \) is the moist adiabatic lapse rate, and \( \Gamma_d \) is the dry adiabatic lapse rate. The diabatic heating is calculated in the region of relative humidity > 90%, and it is zero otherwise because this equation is only valid for saturated air.
The global analysis, which is used as the outer boundary conditions for the mesoscale analysis, uses the JMA Global Spectral Model and four-dimensional variational data assimilation. It has a TL959 spectral horizontal resolution (approximate grid spacing of 20 km), and 60 vertical levels up to 0.1 hPa. The temporal interval is 6 h. Here, we used a dataset prepared by interpolating the original data into geographic coordinates with a horizontal grid spacing of 0.5° in both latitude and longitude (approximately 50 km around the Sea of Japan).

b. Tracking of polar mesocyclones

The polar mesocyclones were objectively detected and tracked from the mesoscale analysis using the same tracking algorithm as Watanabe et al. (2017), which is an improved version of the algorithm proposed by Watanabe et al. (2016) [the difference between the results obtained using the algorithm of Watanabe et al. (2016) and those of the present study are described in appendix A]. The algorithm tracks maxima in the vorticity (hereafter “vorticity” indicates vertical relative vorticity unless stated otherwise) field at 950 hPa and detects polar mesocyclones with a horizontal scale of about 50–300 km. Most of the polar mesocyclones detected by this algorithm are smaller than the polar lows studied by Yanase et al. (2016) and they differ in the location where they attain their maximum intensity and their typical direction of movement (Watanabe et al. 2016). Note that some polar mesocyclones with a horizontal scale smaller than 50 km are not captured by this tracking algorithm and that some polar mesocyclones with a horizontal scale larger than 300 km are excluded from the present analysis. However, using satellite images, we have confirmed that a majority of polar mesocyclones over the Sea of Japan can be captured by this algorithm (Watanabe et al. 2016). In the present study, their intensities are measured by their vorticity at 950 hPa. We analyzed those polar mesocyclones that developed during the cold season (November–March) between November 2009 and March 2015.

c. Composite analysis

In the present study, we focus on polar mesocyclones over the western Sea of Japan. These polar mesocyclones
correspond to those in the WJ region defined by Watanabe et al. (2016, p. 2510) as “the western part of the Sea of Japan from the base of the Korean Peninsula to the center of Honshu Island.” In the present study, to analyze polar mesocyclones associated with the JPCZ, we selected two categories of typical polar mesocyclone in this region based on their genesis location and movement. The first category represents polar mesocyclones that formed over the northwestern Sea of Japan (the area denoted by NW in Fig. 1). Near Honshu Island they move southeastward or eastward (Tsuboki and Asai 2004; Watanabe et al. 2016), reaching regions SE or E, respectively (Fig. 1). We refer to these as NW_SEs and NW_Es, respectively. The second category represents polar mesocyclones that formed over the southwestern Sea of Japan (the area labeled SW in Fig. 1) and attained their maximum intensity during their lifetime to the west of 134°E and south of 37.5°N. These polar mesocyclones are hereafter referred to as SWs. Note that a synoptic-scale low was sometimes erroneously detected as a SW when only a part of the synoptic-scale low was included within the tracking area, but these false detections were manually excluded.

We calculated composite fields in geographic coordinates so that polar mesocyclones are at similar locations in the composite field. We first defined a key time (t = 0 h) for each polar mesocyclone to adjust geographical location and movement. The key time for the NW_SEs and NW_Es was defined as the time when the polar mesocyclones were located closest to the dashed lines in Figs. 2a and 2b, respectively, which were selected to be around the middle of the tracks of the polar mesocyclones. The key time for SWs was taken to be when the polar mesocyclones attained their maximum intensity, because their movement is often complicated. However, it turned out that they attain their maximum intensity at similar locations (Fig. 2c). Two kinds of composite field were calculated using the global analysis and mesoscale analysis. Hereafter, they are referred to as synoptic-scale composite and mesoscale composite, respectively. The synoptic-scale low was included within the tracking area, but these false detections were manually excluded. The synoptic-scale composite and anomaly fields of SLP and total-composite field of the horizontal wind vectors at 850 hPa (green arrows; m s⁻¹) for the NW_SEs. The total-composite fields are shown by contours with an interval of 4 hPa and the anomaly fields by color shading (hPa). (b),(d) As in (a),(c), but for temperature at 850 hPa. The total-composite fields are shown by contours with an interval of 5°C and the anomaly fields by color shading (°C): (a),(b) at t = −24 h and (c),(d) at t = 0 h. Black dots indicate statistical significance at the 0.05 level for the anomaly. The red dots indicate the locations of polar mesocyclones.
composite is used to characterize the synoptic-scale environmental field, whereas the mesoscale composite depicts mesoscale structures including the JPCZ. Although the composite averages do not represent an actual situation but a mean over many cases, in the following we treat the composite average as an actual situation for ease of description. The anomaly fields of these composites were defined by subtracting the 14-day average between $t = -168$ and $t = +168$ h from the total composite at each time step. Hereafter, variables in the total composite are modified by “total composite” in contrast to the anomaly field. The statistical significance of the anomaly is evaluated by Welch’s $t$ test.

3. Overview of the polar mesocyclones

A total of 35 NW_SEs, 29 NW_Es, and 10 SWs were selected from 196 polar mesocyclones in the western Sea of Japan (Fig. 2). The frequency of each type of polar mesocyclone by month is shown in Fig. 3. NW_SEs and NW_Es are most frequent in January and December, respectively, while most SWs occur in February. We evaluated the wind speed around the polar mesocyclones. A total of 18 NW_SEs, 17 NW_Es, and 6 SWs were accompanied by gale-force winds near the surface within...
150 km from the center. Thus, about half of the polar mesocyclones satisfied the criteria of a polar low. In satellite images, polar mesocyclones have either of two typical cloud structures: some polar mesocyclones were embedded in the cloud band of a JPCZ (Fig. 4a), while others had a distinct comma-shaped cloud (Fig. 4b). Most of the NW_SEs and NW_Es were embedded in the JPCZ (34 NW_SEs and 24 NW_Es). On the other hand, more than half of the SWs (7 SWs) were accompanied by a comma-shaped cloud pattern.

4. Results

a. NW_SEs

We will first examine the low-level synoptic-scale environment for the NW_SEs. The total-composite sea level pressure (SLP) field (left column in Fig. 5) shows a distinct zonal pressure gradient between a high pressure system over Siberia (the Siberian high) and a low pressure system over the North Pacific Ocean (the Aleutian low). This is a typical pressure pattern for East Asia during a northwesterly winter monsoon.

Traveling positive and negative SLP anomalies, which are statistically significant at the 0.05 level, are evident in the anomaly field. At $t = 0$ h, the zonal gradient of total-composite SLP over the Sea of Japan is smaller than the mean field, resulting in a relatively weak northwesterly winter monsoon. Then, a synoptic-scale negative SLP anomaly, which corresponds to a composite average of synoptic-scale cyclones, starts to intensify over Japan at $t = -48$ h (Fig. 5a). At this time, the cold air mass remains over the Eurasian continent (Fig. 5b). The negative SLP anomaly moves to the east of Japan by $t = 0$ h (Fig. 5c). To the
west of the negative SLP anomaly, a positive SLP anomaly exists over the Eurasian continent. Both the pressure gradient and the northwesterly winter monsoon over the Sea of Japan intensify between these negative and positive SLP anomalies. The intense northwesterly winter monsoon causes a cold air outbreak over the Sea of Japan (Fig. 5d). The NW_SEs appear and develop between $t = -24$ and $t = 0$ h in a weak SLP trough in the total-composite field that extends westward from the center of the negative SLP anomaly (Figs. 5a and 5c). They are located at the eastern edge of the cold air (Fig. 5d). Finally, they move southeastward, make landfall on Honshu Island, and dissipate, while the negative SLP anomaly continues to develop and move farther east and the cold northwesterly spreads over the Sea of Japan (not shown).

A characteristic feature at 500 hPa is a synoptic-scale trough and ridge pair moving eastward (Fig. 6). At $t = -24$ h (Fig. 6a), a positive geopotential height anomaly is located over the Pacific Ocean, whereas a negative geopotential height anomaly associated with a trough is present over the Eurasian continent and they are accompanied by warm and cold air, respectively (Fig. 6b). These anomalies are statistically significant at the 0.05 level. The negative geopotential height anomaly intensifies and moves eastward (Fig. 6c). The center of
the negative geopotential height anomaly passes over the root of the Korean Peninsula between \( t = -24 \) and \( 0 \) h. The NW_SEs develop directly beneath the upper-level trough and cold air at \( t = 0 \) h (Figs. 6c and 6d). Finally, the trough moves over the Pacific Ocean by \( t = +24 \) (not shown). Note that the negative SLP anomaly intensifies to the east of the upper-level trough, indicating baroclinic intensification of synoptic-scale cyclones (Fig. 5c).

Figure 7 shows the difference in the total-composite SST and total-composite potential temperature at 700 hPa, which is an index of the cold air outbreak (Bracegirdle and Gray 2008), at \( t = 0 \) h. This index is positive over the Sea of Japan, indicating a strong cold air outbreak and large heat flux from the sea surface. The NW_SEs develop where the cold air outbreak is strongest.

Figure 8 shows the total-composite low-level divergence fields and the anomaly fields of thickness between 950 and 850 hPa. A notable characteristic is a convergence zone that develops from the root of the Korean Peninsula. This corresponds to the JPCZ (Asai 1988). The cold air mass, which remains over the Eurasian continent before \( t = -12 \) h, gradually intrudes into the Sea of Japan (not shown). However, it is blocked by the Changbai Mountains (shaded area to the north of the Korean Peninsula) and splits into two streams: a north-northwesterly on the northern side and a west-northwesterly on the southern side. These two streams converge to form the JPCZ, which extends from the northwestern corner of the Sea of Japan (Figs. 8a and 8c), and is collocated with the SLP trough. As the two cold streams pass over the sea, they are warmed up by sea surface fluxes and eventually form a thermal ridge that corresponds to the SLP trough (Figs. 8b and 8d).

The NW_SEs are generated in the JPCZ where there is strong convergence and strong horizontal wind shear. At \( t = -6 \) h, the JPCZ extends almost directly southeast (Fig. 8a); however, it gradually starts to bend in the middle and the NW_SEs are now located at a bending point where the southeastern part of the JPCZ extends in a more zonal direction (Fig. 8c). The thermal ridge and SLP trough also curve in the same manner as the JPCZ (Fig. 8d). The bending point of the JPCZ moves south together with the movement of the NW_SEs (not shown). After the NW_SEs make landfall on Honshu Island, a strong northwesterly prevails over the Sea of Japan, and the JPCZ and thermal ridge gradually become indistinct (not shown).

At upper levels, a cyclonic vorticity maximum associated with the upper-level trough is seen when NW_SEs are observed. Figure 9a shows the total-composite vorticity field at 500 and 850 hPa. When the upper-level cyclonic vorticity maximum approaches the northwestern edge of the Sea of Japan at \( t = -12 \) h, the NW_SEs are generated in the eastern part of this region. Over the following 12 h, the upper-level cyclonic vorticity maximum and the NW_SEs move together. According to quasigeostrophic theory, the traveling upper-level cyclonic vorticity maximum induces dynamic updrafts ahead of it. NW_SEs move under the area of the updraft caused by the upper-level trough, implying that upper-level forcing contributes to the generation and development of the NW_SEs.

The cold air that accompanies the upper-level trough also contributes to reducing the stability of the atmosphere over the Sea of Japan. Figure 9b shows the mesoscale composite of the total-composite lapse rate averaged between 900 and 500 hPa, as well as the total-composite vertical velocity and diabatic heating at 850 hPa. As the warm air remains along the JPCZ because of the blocking of the cold air by the Changbai
Mountains, a local stability minimum forms along the JPCZ at low levels. Moreover, strong dynamical convergence also exists at low levels. These environments are conducive to cumulus convection. Indeed, strong updrafts exist in association with the condensational heating around the NW_SEs (Fig. 9b).

It is noteworthy that some NW_SEs attain their maximum intensity at an early stage of their lifetime, whereas the rest of the NW_SEs intensify during their lifetime and reach peak intensity near Honshu Island (Fig. 2a). Hereafter, NW_SEs that reached their peak intensity at or after \( t = 0 \) h will be referred to as “late-developing NW_SEs” and those that reached their peak intensity before \( t = 0 \) h as “early-developing NW_SEs.” The numbers of late- and early-developing NW_SEs were 15 and 20, respectively. Composite analysis was applied to each group to examine the factors that cause this difference in the timing of their development.

The major difference between the late- and early-developing NW_SEs is the path of the upper-level trough. The paths of the minima of the geopotential height anomaly at 500 hPa for late- and early-developing NW_SEs are shown in Fig. 10. The geopotential height anomaly minimum for the late-developing NW_SEs takes a more southerly path than that of the early-developing NW_SEs. At \( t = 0 \) h, the center of the upper-level cyclonic vorticity maximum of the late-developing NW_SEs is located to the west or northwest of the polar mesocyclones (Fig. 10a), whereas that for the early-developing NW_SEs is located to the north of the polar mesocyclones (Fig. 10b). As the traveling upper-level cyclonic vorticity maximum induces dynamic updrafts ahead of it, the configuration for late-developing NW_SEs is more favorable for the development of the polar mesocyclones.

### b. NW_Es

The synoptic-scale environment for the NW_Es is fairly similar to that of the NW_SEs, including the path of the upper-level trough and low-level negative SLP anomaly. If we examine the environment in more detail, however, we find that the location of the polar mesocyclones relative to the synoptic-scale system, including the negative SLP anomaly and the negative geopotential height anomaly, differs between the NW_SEs and the NW_Es. At \( t = 0 \) h, the center of the negative SLP anomaly for the NW_Es is over the Sea of Japan (Fig. 11a), while that for the NW_SEs is over the Pacific Ocean (Fig. 5c), indicating that NW_Es are closer to the synoptic-scale cyclone. Similarly, at \( t = 0 \) h the minimum of the negative geopotential height anomaly at 500 hPa for the NW_Es is located to the northwest of the polar mesocyclones (Fig. 11b), while that for the NW_SEs is to the north of the polar mesocyclones (Fig. 6c).

The behavior of the polar mesocyclones with respect to the JPCZ also differs between the NW_Es and NW_SEs. Before \( t = -12 \) h, the JPCZ for the NW_Es extends in a slightly more meridional direction than that of the NW_SEs (not shown). At \( t = -6 \) h (Fig. 12a), however, the western part of the JPCZ starts to rotate cyclonically and bends southward in the middle, and the NW_Es are located slightly to the west of the bending point. This direction of bending of the JPCZ is opposite to that associated with the NW_SEs. The eastern part of the JPCZ then starts to move northeastward so that the JPCZ extends directly toward the east-southeast (Fig. 12b). The NW_Es move east along the JPCZ. These relationships between the bending direction of the JPCZ and the direction of
movement of polar mesocyclones are consistent with the findings of Ookubo (1995).

In summary, although the behavior of the synoptic-scale structures of the NW_SEs and NW_Es is almost the same, the synoptic-scale cyclone and the upper-level trough of the NW_SEs are located farther to the east than those of the NW_Es. It is interesting to note that the direction of the JPCZ for the NW_Es at $t = 0$ h is similar to that for the NW_SEs at $t = -6$ h (see Figs. 8a and 12b). These results indicate that given a similar synoptic-scale environment, the NW_Es seem to form earlier than the NW_SEs. Indeed, these two types of polar mesocyclone sometimes appear simultaneously, and the eastern polar mesocyclone moves east or east-northeastward, while the other moves to the southeast (Ookubo 1995; Tsuboki and Asai 2004). Six pairs of NW_SEs and NW_Es that occurred simultaneously within a 24-h period were found during the analysis period.

c. SWs

Similar to the NW_SEs and NW_Es, a traveling negative SLP anomaly and an upper-level trough that are statistically significant at the 0.05 level are evident in the synoptic-scale environment associated with the SWs (Fig. 13). However, their paths differ from those of the NW_SEs and NW_Es. The center of the upper-level negative geopotential height anomaly for the SWs passes over the southern Korean Peninsula (Fig. 13b). In association with the southern path of the upper-level negative geopotential height anomaly, the negative synoptic-scale SLP anomaly appears farther south than that of the NW_SEs and NW_Es and moves east along the southern coast of Honshu Island (Fig. 13a; where it is known as the “southern-coast low”), resulting in northerly winds over the Sea of Japan. The SWs are located in the northwestern quadrant of the negative synoptic-scale SLP anomaly, and the upper-level negative geopotential height anomaly is located directly above the SWs at $t = 0$ h (Fig. 13b). After $t = 0$ h, the SWs make landfall on Honshu Island and dissipate, while the negative synoptic-scale SLP anomaly and the upper-level trough continue to move east (not shown).

A cold air outbreak is also seen in the mesoscale composite for the SWs, and it is strongest over the southwestern Sea of Japan where SWs develop (not shown). A JPCZ is also present in the cold air outbreak (Fig. 14). However, the behavior of the JPCZ associated with the SWs differs substantially from that associated with the NW_SEs and NW_Es. At $t = -12$ h (Fig. 14a), a strong northerly prevails over the East China Sea, while there is a weak northeasterly over the western Sea of Japan. A stationary JPCZ extends in the meridional direction between these northerly and northeasterly winds, corresponding to an SLP trough. This SLP trough is collocated with a thermal ridge between the cold air over the East China Sea and the northwestern Sea of Japan (not shown). The SWs appear and develop in the southern part of the JPCZ (Fig. 14b). After the SWs make landfall on Honshu Island, the JPCZ dissipates and there is a strong northerly over the Sea of Japan (not shown).

As seen in the synoptic-scale composite, the upper-level trough for the SWs passes over the southern part of the Korean Peninsula (Fig. 15a). The SWs begin to develop in association with the approach of the upper-level cyclonic vorticity maximum. The upper-level cyclonic vorticity maximum is located directly above the SWs at $t = 0$ h, which is when the SWs reach their maximum intensity. However, the upper-level vortex continues to move east and leaves the SWs after $t = 0$ h, resulting in a separation of the upper-level trough and low-level polar mesocyclones.
Similar to the situation for the NW_SEs, a less stable atmosphere and strong updrafts associated with condensational heating exist around the SWs (Fig. 15b). Comparing the environment associated with the NW_SEs with that of the SWs, we find that the static stability around the SWs is smaller than that around the NW_SEs (Figs. 9b and 15b). At low levels, the air around the SWs has traveled a longer distance over the Sea of Japan, implying that it has been warmed by the sensible and latent heat fluxes from the sea surface. Furthermore, the SST in the southwestern Sea of Japan is about 15°C and considerably higher than that of the northwestern Sea of Japan (Fig. 1). On the other hand, the cold air at upper levels of the SWs intrudes farther south over the Sea of Japan in association with the southern path of the upper-level trough. Thus, a less stable environment is formed around the SWs.

d. Relationship between large-scale environment and JPCZ

The composite analysis in the previous section demonstrated that the polar mesocyclones that develop over the western part of the Sea of Japan are closely related to the JPCZ. However, the direction of the JPCZ differs among the NW_Es, NW_SEs, and SWs. At t = 0 h, the JPCZ for the NW_Es extends east-southeastward, whereas that for the SWs extends south-southeastward. The JPCZ for the NW_SEs extends in an intermediate direction between the NW_Es and SWs (Figs. 8c, 12b, and 14b). Figure 16 shows the low-level average wind around the Changbai Mountains and the western part of the Sea of Japan at t = 0 h for the NW_SEs, NW_Es, and SWs, where the average is taken within the box enclosed by 35°–45°N, 125°–135°E, and 950–850 hPa. The directions of the low-level winds at t = 0 h are east-southeastward, southeastward, and south-southeastward for the NW_Es, NW_SEs, and SWs, respectively, which is nearly the same as the direction of the JPCZ. This relationship implies that the direction of the large-scale low-level wind affects the direction of the JPCZ, the genesis locations of the polar mesocyclones, and their directions of movement.

Another difference in the composite fields between the NW_SEs, NW_Es, and SWs is the path of the upper-level trough. Figure 17 shows the path of the center of the negative geopotential height anomaly at 500 hPa together with the location of each polar mesocyclone at t = 0 h. Although the paths of the upper-level trough for the NW_SEs and NW_Es are nearly the same before t = 0 h, the location of the upper-level trough for the NW_SEs is slightly to the east of that for the NW_Es. At t = 0 h, the center of the negative geopotential height anomaly for the NW_SEs is located to the north of the polar mesocyclones, whereas that for the NW_Es lies northwest of the polar mesocyclones (see also Figs. 6c and 11b). After t = 0 h, on the other hand, the upper-level trough above the NW_SEs moves much faster than that associated with the NW_Es. The upper-level trough of the SWs takes a more southward path and is located in the southwestern Sea of Japan at t = 0 h (see also Fig. 13b).

Now we consider the relationship between the low-level wind direction and the synoptic-scale atmospheric systems. To evaluate the effects of the synoptic-scale structures on the low-level atmosphere, we conducted a piecewise PV inversion (PPVI) analysis (Davis and Emanuel 1991; Davis 1992; the detailed inversion technique is described in appendix B). Based on the invertibility principle of PV (e.g., Hoskins et al. 1985), the geopotential height and potential temperature anomaly fields associated with specific PV anomalies and the
potential temperature anomaly at the upper and lower boundaries can be recovered by PPVI. Using PPVI, we evaluate the low-level flow fields associated with each large-scale structure, including the upper-level trough and low-level cold air outbreak. In PPVI analysis, the mean and anomaly fields must be defined. In this study, the 14-day running mean and the anomaly field of the synoptic-scale composite were used for the PPVI. Following Davis and Emanuel (1991), we divided the PV anomaly into three parts: the PV anomaly associated with the upper-level trough (hereafter UL), the PV anomaly at the midlevel (hereafter ML), and the potential temperature anomaly at the surface (hereafter SFPT). Since the upper-level PV anomaly is distinct above 500 hPa (not shown), UL and ML are defined as PV anomalies above 500 hPa and below 600 hPa, respectively. SFPT is defined as the potential temperature anomaly at 963 hPa.

Figures 18–20 show the geopotential height anomalies (color shading), the mean fields (green contours), and the sum of the mean field and geopotential height anomaly (black contours) at 850 hPa, respectively, for UL, SFPT, and the sum of UL, ML, and SFPT. Hereafter, the geopotential height anomalies associated with UL, ML, SFPT, and the sum of all anomalies at 850 hPa will be referred to as $Z_{UL}$, $Z_{ML}$, $Z_{SFPT}$, and $Z_{SUM}$, respectively. The distribution of $Z_{SUM}$ agrees well with the geopotential height anomaly of the synoptic-scale composite (not shown), indicating that the PPVI analysis successfully captures the synoptic-scale structures associated with each PV anomaly. As the effect of the ML makes only a minor contribution (not shown), it is not included in the following discussion.

First, we will examine the environment associated with the NW_SEs and NW_Es. As the temporal development of the synoptic-scale environments of the NW_SEs and NW_Es is almost the same before $t = 0$ h, except that the latter precedes the former by about 12 h (Fig. 17), only the results for the NW_SEs are shown. The directions of isobars in the mean field indicate a west-northwesterly over the Sea of Japan. At $t = -12$ h, a negative $Z_{UL}$ exists over the Eurasian continent (Fig. 18a). The wind anomaly corresponding to this negative $Z_{UL}$ has a southwesterly component over the Sea of Japan. A positive $Z_{SFPT}$ associated with low-level cold air over the Eurasian continent develops slightly west of the negative $Z_{UL}$ and has an opposite effect to the negative $Z_{UL}$ (Fig. 18b). As the negative
$Z_{UL}$ is larger than the positive $Z_{SFPT}$, however, a negative $Z_{SUM}$ occurs in the northwestern part of the Sea of Japan (Fig. 18c). The wind anomaly corresponding to this negative $Z_{SUM}$ is southwesterly over the Sea of Japan. The sum of this southwesterly wind anomaly and the west-northwesterly wind in the mean field results in a westerly over the Sea of Japan, and because of this westerly the JPCZ extends initially eastward. As the upper-level trough of the NW_Es moves slowly (Fig. 17) compared with that of NW_SEs, the westerly wind over the Sea of Japan is maintained for a longer period in the environment associated with the NW_Es and thus they move eastward.

At $t = 0$ h, the center of the negative $Z_{UL}$ moves into the Sea of Japan (Fig. 19a). The positive $Z_{SFPT}$ remains over the Eurasian continent and a weak negative $Z_{SFPT}$ is located over the Pacific Ocean at $t = 0$ h (Fig. 19b). The sum of $Z_{UL}$ and $Z_{SFPT}$ results in a negative $Z_{SUM}$ over Honshu Island (Fig. 19c). The wind anomaly corresponding to this negative $Z_{SUM}$ is northerly over the Sea of Japan, so the resultant wind is more meridional than that in the mean field. As a result, the JPCZ also extends in a northwest–southeast direction and the NW_SEs move southeastward.

Next, we consider the environment associated with the SWs. The wind direction in the mean field for the SWs is west-southeasterly over the Sea of Japan, which is similar to that for the NW_SEs (Fig. 20). In contrast to the NW_SEs and NW_Es, however, the negative $Z_{UL}$ passes over the southern Korean Peninsula (Fig. 20a). Although the positive $Z_{SFPT}$ is located over the Eurasian continent (Fig. 20b), taken together, $Z_{UL}$ and $Z_{SFPT}$ form a negative $Z_{SUM}$ to the south of the Korean Peninsula and a positive $Z_{SUM}$ over the Eurasian continent (Fig. 20c). The wind anomaly corresponding to these two anomalies is easterly over the Sea of Japan. If this easterly wind anomaly is superposed on the northwesterly in the mean field, the wind direction around the Changbai Mountains and the western Sea of Japan becomes northerly and the JPCZ extends in a meridional direction before the upper-level trough reaches the Sea of Japan.

5. Discussion and conclusions

In this study, we examined the general characteristics of the polar mesocyclones that develop over the western part of the Sea of Japan. They were categorized into three groups according to their genesis location and direction of movement: NW_SEs and NW_Es are polar mesocyclones that form over the northwestern Sea of Japan and move in a southeast-erly and easterly direction, respectively, whereas SWs are polar mesocyclones that form and develop over the southwestern Sea of Japan. The general characteristics of the synoptic-scale environment and the mesoscale structure of these polar mesocyclones were examined using composite analysis.
FIG. 18. Geopotential height anomalies (color shading: m), geopotential height of the mean field (green contours; interval: 30 m), and the sum of the anomaly and the mean field (black contours; interval: 30 m) for (a) UL, (b) SFPT, and (c) sum of UL, ML, and SFPT at 850 hPa and $t = -12$ h for the NW_SEs.

FIG. 19. As in Fig. 18, but at $t = 0$ h for the NW_SEs.
The common features of the synoptic-scale environment associated with the NW_SEs, NW_Es, and SWs are an upper-level trough accompanied by cold air and a negative SLP anomaly to the east of the polar mesocyclones, associated with a cold air outbreak at low levels. These features are also found during polar mesocyclone genesis in the northeastern Sea of Japan (Watanabe et al. 2017) and polar low genesis in the Sea of Japan (Yanase et al. 2016) and other oceans (e.g., Kolstad 2011). A notable structure that distinguishes the western Sea of Japan from other regions is the JPCZ, which forms to the lee side of the Changbai Mountains in association with a thermal ridge between two cold streams that pass around either side of the mountains. Our composite analysis demonstrates that the JPCZ is closely related to the initiation of all three types of polar mesocyclone, and that they are likely generated when the upper-level trough approaches the JPCZ. However, the behavior of the JPCZ and the polar mesocyclones differs among the NW_Es, NW_SEs, and SWs. We have demonstrated that the movement of the upper-level trough affects the direction of extension of the JPCZ and the genesis location and motion of the polar mesocyclones.

Figure 21 is a schematic illustration of the relationship between the polar mesocyclones, the JPCZ, and the upper-level trough. The NW_Es and NW_SEs appear in association with the upper-level trough passing over the root of the Korean Peninsula. When this upper-level trough is located over the root of the Korean Peninsula, the circulation associated with the upper-level trough generates a westerly wind over the Sea of Japan. The JPCZ then extends zonally and the westerly wind advects the polar mesocyclone eastward (Fig. 21a). As the upper-level trough of the NW_Es moves slowly, this configuration is maintained during the development of the NW_Es. On the other hand, the upper-level trough of the NW_SEs moves faster toward the center of the Sea of Japan and induces a northerly wind to its west, resulting in the northwesterly wind over the western Sea of Japan (Fig. 21b). Thus, the JPCZ extends southeastward and the polar mesocyclone moves southeastward. This relationship between the passage of the upper-level trough and the movement of the JPCZ is consistent with the case study reported by Ohigashi and Tsuboki (2007).

The SWs appear when the upper-level trough passes over the southern Korean Peninsula (Fig. 21c). The southerly path of the upper-level trough affects both the direction of the JPCZ and the genesis location of the polar mesocyclones. The cyclonic circulation associated with this upper-level trough produces an easterly wind component around the Changbai Mountains and the western Sea of Japan, resulting in northerly winds there.
Thus, the JPCZ extends in a meridional direction before the upper-level trough moves into the Sea of Japan. When the upper-level trough approaches the JPCZ, the polar mesocyclones are sometimes generated in the southwestern Sea of Japan.

These results indicate that the potential for the occurrence of polar mesocyclones can be estimated from the large-scale environment. These relationships might be useful when we estimate the long-term tendency or future change in polar mesocyclones from a low-resolution dataset.

Our analysis also demonstrated that the environment of the southwestern Sea of Japan is less stable than that of the northwestern part. According to Maejima and Iga (2012), meso-\(\alpha\)-scale polar mesocyclones tend to develop in a convergence zone in a less stable environment. Thus, this difference in environmental stability seems to be one of the reasons why meso-\(\alpha\)-scale polar mesocyclones with comma-shaped cloud often appear in the southwestern Sea of Japan and also why meso-\(\beta\)-scale polar mesocyclones generated in the northeastern Sea of Japan sometimes develop into meso-\(\alpha\)-scale polar mesocyclones as they move southeastward.

Polar mesocyclones associated with a low-level convergence zone with large horizontal wind shear develop not only over the Sea of Japan, but also over the Norwegian Sea (Sergeev et al. 2017), Bering Sea, (Bresch et al. 1997), and the Great Lakes (Laird et al. 2001).

Although our study showed that the approach of the upper-level trough to the convergence zone triggers the polar mesocyclone, the mechanism that triggers the mesocyclones remains to be clarified. Such a generation process might also affect the horizontal scale of the polar mesocyclone. In addition, although several studies have shown the relationship between the structure and development mechanism of polar mesocyclones and their environment (e.g., Yanase and Niino 2007; Terpstra et al. 2015), the effect of a convergence zone is not included in these studies. Therefore, we also need to clarify the controls on the horizontal scale and development mechanism of polar mesocyclones developing within a convergence zone in a systematic way that will include idealized numerical experiments.

Furthermore, previous case studies have indicated that diabatic processes, including the condensational heating and sensible heat and moisture fluxes from the sea surface, are important for the formation of both the JPCZ and polar mesocyclones (Tsuboki and Asai 2004; Watanabe and Niino 2014). Although our composite analysis suggested that polar mesocyclones develop in association with condensational heating, it is difficult to evaluate the roles of the diabatic processes using composite analysis. In addition, the effects of the topography around the Sea of Japan, which are considered to be important for the formation of the JPCZ (Nagata 1991), on NW_Es, NW_SEs, and SWs might be different.

**FIG. 21.** Schematic illustration of the relationship between the polar mesocyclones, the JPCZ, and the upper-level geopotential height trough for the (a) NW_Es, (b) NW_SEs, and (c) SWs.
because the directions of the JPCZ differ among these polar mesocyclones. To investigate the effects of these factors, numerical simulations similar to those of Watanabe et al. (2017), in which typical polar mesocyclones were successfully reproduced using the composite fields as the initial and boundary conditions, would be useful. The results of such simulations will be reported in a future paper.

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**APPENDIX A**

**Tracking Algorithm for Polar Mesocyclones**

In the present study, we used the tracking algorithm based on Watanabe et al. (2016) with some minor modifications to ensure the isotropy of the algorithm (Watanabe et al. 2017), as follows: 1) the vorticity field is smoothed using a running mean over a radius of 25 km, 2) the steering wind is defined as the average wind over a radius of 200 km between 1000 and 700 hPa, and 3) the estimated area used for the connection of vortices is defined as the area within 120 km of the estimated position based on the steering wind. Because of this modification, the number of polar mesocyclones in the western Sea of Japan is increased from 146 in Watanabe et al. (2016) to 192 in the present study. However, all selected NW_SEs, NW_Es, and SWs in the present study were also detected by Watanabe et al. (2016).

**APPENDIX B**

**Piecewise PV Inversion**

The Ertel PV is defined as

$$q = \frac{1}{\rho} \boldsymbol{\nabla} \cdot \boldsymbol{\nabla} \theta,$$

where $q$ is PV, $\rho$ is the density of air, $\boldsymbol{\nabla}$ is the absolute vorticity vector, and $\theta$ is potential temperature. If an appropriate boundary condition is given, the balanced flow is recovered from the distribution of $q$ (Hoskins et al. 1985). Charney’s balance equation is used in the inversion method by Davis and Emanuel (1991). Assuming hydrostatic balance and that the magnitude of the irrotational component of the wind is much smaller than the magnitude of the nondivergent component, the divergence equation and the Ertel PV can be rewritten as

$$\nabla^2 \Phi = \nabla \cdot (\nabla \Phi) + \frac{2}{a^2 \cos^2 \phi} \left[ \frac{\partial^2 \Psi}{\partial \lambda^2} \frac{\partial^2 \Psi}{\partial \phi^2} - \left( \frac{\partial^2 \Phi}{\partial \lambda \partial \phi} \right)^2 \right],$$

where $\Phi$ is the geopotential, $\Psi$ is the nondivergent streamfunction, $\lambda$ is longitude, $\phi$ is latitude, $a$ is the radius of Earth, $\kappa = (R/c_p)$ is the Poisson constant, $f$ is the Coriolis parameter, $g$ is gravitational acceleration, $p$ is pressure, and $\pi$ is the Exner function. Potential temperatures of the lower and upper boundaries provide Neumann boundary conditions on the top and bottom boundaries. A homogeneous boundary condition is assumed on the lateral boundary. These equations are linearized in the same manner as in Davis and Emanuel (1991). Flow fields associated with each PV anomaly and potential temperature anomaly on the boundary can be recovered by the PPVI. The complete description of the PPVI may be found in Davis and Emanuel (1991).

**REFERENCES**


Bresch, J. F., R. J. Reed, and M. D. Albright, 1997: A polar-low development over the Bering Sea: Analysis, numerical


