A polar mesocyclone (PMC) observed over the Japan Sea on 30 December 2010 was studied using a nonhydrostatic mesoscale numerical model with a horizontal resolution of 2 km. The numerical simulation successfully reproduced the observed life cycle of the PMC. The results of the numerical simulation suggest that the life cycle of the PMC may be divided into three stages: an early development stage, in which a number of small vortices appear in a shear zone; a late development stage, which is characterized by the merger of vortices and the formation of a few larger vortices; and a mature stage, in which only a single PMC is present. During the early development stage, vortices are generated in the shear zones of strong updrafts in discrete cumulus convection cells. In contrast, during the late development stage, the vortices develop as a result of barotropic instability in the shear zone. A cloud-free eye and spiral cloud bands accompany the mature stage of a simulated PMC. A warm core structure also forms at the center of the PMC on account of adiabatic warming associated with downdrafts. The structures in the PMC during the mature stage resemble those of a tropical cyclone. Sensitivity experiments, in which sensible and latent heat fluxes from the sea surface and condensational heating were switched on/off, demonstrate that condensational heating is critical to the development of the PMC at all stages, and that sensible and latent heat fluxes play secondary roles.

1. Introduction

The migration of cold air masses over the Japan Sea in winter is associated with the development of various types of polar mesocyclones (PMCs). These PMCs have broad spatial scales, ranging from meso-γ to meso-α scales (Asai and Miura 1981; Kawashima and Fujiyoshi 2005; Ninomiya et al. 1990; Ninomiya and Hoshino 1990), and exhibit various types of cloud systems, such as comma-shaped clouds or spiral cloud bands. Occasionally, the spatial scales and cloud patterns of PMCs change during their development (Lee et al. 1998). Such diversity may be a consequence of differences in their developmental mechanisms.

The weather events associated with wintertime PMCs in the Japan Sea, such as heavy snowfall (e.g., Miyazawa 1967; Asai 1988) and strong winds (e.g., Kuroda 1992; Yamagishi et al. 1992), often cause serious problems in coastal nearby regions. However, their prediction remains difficult because of their relatively small spatial and temporal scales. Hence, it is important to clarify PMC generation and development mechanisms, not only to advance our understanding of atmospheric phenomena, but also to help mitigate natural disasters associated with these events.

Meso-α-scale PMCs occurring in polar air masses and accompanied by strong surface winds are called polar lows. These lows develop over high-latitude oceans, such as the North Atlantic Ocean, Barents Sea, and Weddell Sea. Satellite images show that polar lows are accompanied by comma clouds or spiral cloud bands (Rasmussen and Turner 2003). Studies on the generation and development mechanisms of polar lows have indicated that baroclinic instability (Reed and Duncan 1987), thermal instabilities such as conditional instability of the second kind (CISK; Rasmussen 1979) or wind-induced surface heat exchange (WISHE; Emanuel and Rotunno 1989), and their combination (Yanase and Niino 2007), play important roles in polar low development.

Polar mesocyclones with small horizontal scales are often observed over the Japan Sea (Asai 1988) and in the vicinity of the Great Lakes, North America (Hjelmfelt 1990; Laird et al. 2001). A number of
observational studies of PMCs over the Japan Sea have been performed using data from meteorological radars and satellite images. Using satellite images, Asai (1988) analyzed PMC genesis locations over the Japan Sea during the winter of 1983–84. He found that PMCs with horizontal scales of less than 300 km were concentrated in two regions: one from the east of the Korean Peninsula to the San-in and Hokuriku districts of Japan, and the other to the west of Hokkaido Island (Fig. 1). These regions are commonly the sites of formation of Japan Sea polar airmass convergence zones (JPCZ; Asai 1988). JPCZs are characterized by northerly and westerly winds on their northeastern and southwestern sides, respectively, so that they are accompanied by significant convergence, cyclonic shear, and active cumulus convection. As for the generation and development mechanisms of PMCs, a number of researchers have suggested the importance of the barotropic instability of horizontal shear in JPCZs (Asai and Miura 1981; Nagata 1993; Kawashima and Fujiyoshi 2005). Nagata (1993) used a hydrostatic numerical model, in which cumulus convection was parameterized, to study meso-β-scale PMCs in the JPCZ to the east of the Korean Peninsula. He succeeded in reproducing realistic PMCs, with spiral bands and a cloud-free eye structure, and showed, using an energy budget analysis, that they developed through barotropic instability. Using dual-Doppler radar data, Kawashima and Fujiyoshi (2005) analyzed meso-γ-scale PMCs along a snowband off the west coast of Hokkaido Island. Their energy budget analysis showed that the PMCs acquired their energy from low-level horizontal shear flow, and that part of the energy was transported upward by the pressure transport term. Cumulus convection is also very active in JPCZs. Because horizontal shear is only significant in JPCZs at low levels, cumulus convection is likely to contribute to the vertical development of PMCs. Cumulus convection typically develops above horizontal shear zones in JPCZs (Kato 2005). It has been suggested that the formation of vertical hot towers by individual cumulus clouds as well as by organized cumulus convection play important roles in tropical cyclone genesis (Hendricks et al. 2004). A similar mechanism might operate during PMC genesis, as both a vorticity-rich environment and strong cumulus convection are present. Therefore, the impact of cumulus convection on the development of PMCs should be examined using a nonhydrostatic model that does not parameterize cumulus convection. This problem was partially examined by Kato (2005), who performed a numerical simulation of a PMC using a nonhydrostatic model with a horizontal resolution of 1 km. His study showed that the horizontal shear of the JPCZ was confined to low levels, and that vertical shear at high levels was large. He emphasized the importance of energy production associated with strong updrafts of cumulus convection within areas of large vertical shear. However, his study focused only on the mature stage of PMCs, and the role of cumulus convection throughout the lifetime of PMCs remains unclear.

One of the significant characteristics of PMCs is their multiscale structure. Ninomiya et al. (1990) and Ninomiya and Hoshino (1990) investigated a meso-β-scale PMC that developed within a meso-α-scale lows (MAL). While the MAL developed in a region of strong baroclinicity, the meso-β-scale PMC developed in a region with strong cyclonic shear and a deep convective mixed layer—conditions that resulted from the MAL. Tsuboki and Asai (2004) studied a similar multiscale structure of a MAL using a numerical simulation based on a hydrostatic model. They showed that the MAL was intensified by upper-level divergence of ageostrophic winds that were formed by updrafts in a baroclinic environment. Meso-β-scale PMCs embedded in the MAL developed from horizontal shear flows intensified by the MAL, while they contributed to the development of the MAL by causing strong updrafts. This multiscale structure suggests that the generation and development of PMCs is affected by larger-scale environmental factors, as well as cumulus convection, which has a horizontal scale smaller than that of the PMCs. Consequently, phenomena over a wide range of horizontal scales should be considered in the study of PMCs. Furthermore, PMCs sometimes change their size and cloud patterns during their lifetime (Lee et al. 1998), suggesting that different mechanisms influence different stages of PMC development. Partly because of their
complex multiscale structure, and partly because they develop over oceans where observations are sparse, the life cycles of PMCs, especially in their early stages, are not well understood. However, high-resolution numerical simulation, which became possible only recently, provides a powerful tool for investigating the life cycles of PMCs.

The purpose of the present paper is to clarify the genesis and development mechanisms of PMCs through their entire life span, and to examine the role of cumulus convection in these mechanisms. On 30 December 2010, a meso-β-scale PMC appeared and developed over the Japan Sea. This case was unusual in that a JPCZ was not present prior to the development of the PMC, but was generated as the PMC developed. To study the life cycle of the PMC from the early stages of its development, prior to the formation of a JPCZ, to its mature stages, we performed a high-resolution numerical simulation using a nonhydrostatic model. We succeeded in reproducing the observed evolution of the PMC and the JPCZ. Here, we report on the results of the detailed analysis of the structure and development mechanisms of the PMC.

The remainder of the paper is organized as follows. Observational analyses of the PMC and synoptic situation are provided in section 2. In section 3, the numerical model used in the analysis is described. The simulation results are presented in section 4, and the sensitivity experiments are described in section 5. A summary and conclusions are given in section 6.

2. Observational analysis

a. Data

We used the following data provided by the Japan Meteorological Agency: hourly infrared (IR) satellite images from the Japanese Multifunctional Transport Satellite 2 (MTSAT-2), surface and 500-hPa weather maps, and meso-objective analysis (MANAL) (e.g., Honda and Sawada 2010) data with a horizontal resolution of 5 km.

b. Evolution of the PMC and JPCZ

Figure 2 shows MTSAT-2 IR images from 2100 UTC 29 December to 0800 UTC 31 December 2010. At
2100 UTC 29 December, a cloud band extended eastward from 41°N, 130°E (Fig. 2a); the cloud band exhibited a southward bend at 41°N, 134°E, and its southern section was masked by upper-level clouds. No obvious JPCZ was observed at this time. As the cold air outbreak from Siberia intensified, the cloud band moved south. At 0700 UTC 30 December, the cloud band extended east-northeastward from 38.5°N, 130°E (Fig. 2b). Stratiform clouds spread over the northern side of the cloud band and a number of vortices formed along its southern edge. Finally, by 1700 UTC 30 December, these vortices had merged into a PMC located at 36.5°N, 135°E, with a cloud-free eye, and spiral-form clouds (Fig. 2c). A JPCZ started to form at 0700 UTC 30 December at 40°N, 135°E (Fig. 2b) extending southeastward to the PMC (Figs. 2c,d). A cloud system with a horizontal scale of approximately 1000 km was also present in the eastern part of the Japan Sea, which corresponded to a rapidly deepening meso-\(\alpha\)-scale low (Fig. 2c).

According to the analysis of Nakai and Yamaguchi (2012), the PMC and the JPCZ were accompanied by extremely deep convective clouds that extended to the tropopause and brought record snowfall to the San-in district (located on the Japan Sea side of the Chugoku district on the western part of Honshu Island; see Fig. 1).

c. Synoptic situation

Figures 3 and 4 show weather maps at 500 hPa and at the surface, respectively, at 1200 UTC 30 December. The synoptic situation in which the PMC developed was characterized by a cold vortex over the root of the Korean Peninsula containing extremely cold air (less than \(\sim\)42°C at 500 hPa) (Fig. 3). Cold air also extended over the entire Japan Sea (less than \(\sim\)36°C at 500 hPa). Thus, the static stability of the atmosphere over the Japan Sea

Fig. 3. Weather map at 500 hPa on 1200 UTC 30 Dec 2010. The solid lines indicate heights at intervals of 60 m, and the dashed lines indicate temperatures at intervals of 6°C.
was low and was favorable for cumulus convection. A meso-\(\alpha\)-scale low associated with the cold vortex over the root of the Korean Peninsula was present over the Japan Sea (Fig. 4). It developed rapidly, and its central surface pressure decreased from 1004 hPa at 0000 UTC 30 December to 990 hPa at 0000 UTC 31 December. The PMC developed to the west of this low. More detailed surface pressure maps prepared from MANAL (Fig. 5) demonstrate that the meso-\(\alpha\)-scale low was accompanied by a surface trough, which extended west from the center of the low and gradually rotated cyclonically. The trough was collocated with the cloud band described in section 2b, on which the PMC formed. Another surface trough corresponding to the JPCZ extended from the root of the Korean Peninsula to the San-in district (Fig. 5f). These features of the PMC, which developed not in a JPCZ but in a surface trough extending westward from a meso-\(\alpha\)-scale low to the east, are similar to those observed on 11 February 1997 (Fu et al. 2004), except that in the present case, the JPCZ formed after the PMC.

3. Model description

To understand the details of the genesis and development mechanisms of this PMC, we performed numerical simulations using the Japan Meteorological Agency nonhydrostatic model (JMA-NHM). The model domain was 1500 km \(\times\) 1500 km in the horizontal dimension and 15.64 km in the vertical dimension (Fig. 1). The grid resolution in the horizontal dimension was 2 km; in the vertical dimension, 40 grid points were distributed at intervals varying from 40 m at the surface to 823 m at the upper boundary.

The JMA-NHM specifications used in this study are summarized in Table 1. Cumulus parameterization was not used. A three-ice single-moment bulk scheme (Lin et al. 1983) was used for microphysics modeling, and the Mellor–Yamada–Nakanishi–Niino Level-3 (MYNN Level-3) scheme (Nakanishi and Niino 2006) was used for boundary layer parameterization. The initial time data for each 3-h period specified by the JMA mesoscale model were used for the initial and boundary conditions. The numerical simulation started at 0900 UTC 29 December, by which time the cloud band had not formed, and continued for 48 h.

4. Results

a. Evolution of the simulated PMC and JPCZ

Figure 6 shows simulated wind vectors and relative vorticity fields at 20 m above ground level (AGL) from 1800 UTC 29 December to 1700 UTC 30 December. The life cycle of the simulated PMC can be characterized by three stages: an early development stage (2000 UTC 29 December–0200 UTC 30 December; Fig. 6b), a late development stage (0200 UTC 30 December–1000 UTC 30 December; Fig. 6c), and a mature stage (1000 UTC 30 December–2100 UTC 30 December; Fig. 6d).

Prior to PMC formation, a high-vorticity shear zone (SZ-A) appeared over the northwestern part of the Japan Sea (Fig. 6a). The SZ-A formed between north-easterly winds that originated from the northern side of the mountains, and westerly winds that originated from the southern side of the mountains in the north of the Korean Peninsula (Mt-A; see Fig. 1 for geographical locations). The SZ-A corresponds to the cloud band observed in satellite images (Fig. 2a) and the trough in the surface pressure maps (Fig. 5). Subsequently, as the SZ-A moved south, a number of small vortices developed on it during the early development stage (Fig. 6b). Successive mergers of the small vortices resulted in the formation of two vortices (vortex A and vortex B) in the late development stage (Fig. 6c). Vortex B showed three or four vorticity peaks, as the mergers had not yet been completed at this stage. Finally, vortex B merged with vortex A to form a single PMC at the mature stage (Fig. 6d).

The PMC was accompanied by a cloud-free eye and spiral-shaped clouds, and showed an axisymmetric pressure depression at its center (Fig. 7). Although the position of the simulated PMC was approximately 200 km west of the observed PMC, its horizontal scale and cloud features are in good agreement with observations. Note that a shear zone with active cumulus convection that later developed into a JPCZ, developed...
Fig. 5. Sea level pressure: (a) 1500 UTC 29, (b) 2100 UTC 29, (c) 0300 UTC 30, (d) 0900 UTC 30, (e) 1500 UTC 30, and (f) 2100 UTC 30 Dec 2010. The thick broken lines in (b)–(f) indicate the surface trough, and the dotted line in (f) indicates the JPCZ.
in the wake of the formation of the PMC and extended northwestern, which is also in agreement with observations. The structure and development mechanisms of the PMC in each of the three stages will be described in more detail in the following sections.

b. Early development stage

Figure 8 shows close-up views of horizontal wind vectors and relative vorticity fields at 20, 1122, 2147, and 3133 m AGL at 0100 UTC 30 December. At 20 m AGL, the SZ-A was approximately 10 km wide and had bow-shaped indentations (Fig. 8a). On the northern edges of the SZ-A, a number of small vortices with horizontal scales of approximately 10 km and intervals of approximately 30 km developed. At 1122 m AGL, the SZ-A consisted of discrete vortices (Fig. 8b), which were collocated with strong updrafts, while pairs of positive and negative vorticity strips were found along the sides of the updrafts at 3133 m AGL (Fig. 8d). The relative vorticity pattern at 2147 m AGL was intermediate.

---

**Table 1. Specifications of the JMA-NHM.**

<table>
<thead>
<tr>
<th>Specification</th>
<th>Details</th>
</tr>
</thead>
<tbody>
<tr>
<td>Model upper surface</td>
<td>15 640 m</td>
</tr>
<tr>
<td>Horizontal domain</td>
<td>750 × 750 grid points with a 2-km horizontal resolution</td>
</tr>
<tr>
<td>Moist physics</td>
<td>Three-ice single-moment bulk scheme (predicting mixing ratio of cloud water, cloud ice, rain, snow, and graupel; Lin et al. 1983)</td>
</tr>
<tr>
<td>Time integration</td>
<td>Horizontally explicit and vertically implicit for sound waves (Δt = 10 s)</td>
</tr>
<tr>
<td>Turbulent closure</td>
<td>MYNN Level 3 scheme (predicting turbulence kinetic energy, potential temperature variance, covariance of potential temperature and mixing ratio of water vapor, and variance of mixing ratio of water vapor; Nakanishi and Niino 2006)</td>
</tr>
<tr>
<td>Cumulus parameterization</td>
<td>Not used</td>
</tr>
<tr>
<td>Surface</td>
<td>Bulk method based on Beljaars and Holtslag (1991)</td>
</tr>
<tr>
<td>Initial condition</td>
<td>Prepared from MSM initial time data at 0900 UTC 29 Dec 2010</td>
</tr>
<tr>
<td>Boundary condition</td>
<td>Radiative boundary condition nested within MSM</td>
</tr>
<tr>
<td>Integration period</td>
<td>48 h (from 0900 UTC 29 Dec 2010 to 0900 UTC 31 Dec 2010)</td>
</tr>
</tbody>
</table>

---

**FIG. 6.** Simulated wind vector and relative vorticity field (10^{-2} s^{-1}) at 20 m AGL: (a) 1800 UTC 29, (b) 0100 UTC 30, (c) 0900 UTC 30, and (d) 1700 UTC 30 Dec 2010. Note that different regions are shown in each panel.
between those at 1122 and 3133 m AGL (Fig. 8c). Note that the wind direction on the northern side of the SZ-A, which was northeasterly at 20 m AGL, changed to southwesterly at 3133 m AGL.

Barotropic instability is one possible development mechanism for the vortices. However, the intervals of the vortices are shorter than the wavelength of the fastest-growing mode, which, at 80 km, is 8 times as long as the width of the shear zone (SZ-A width, ~10 km) (e.g., Gill 1982). Furthermore, the vortices have little phase tilt across the shear zone, except near the surface. For these reasons, barotropic instability is not a likely mechanism for the development of the vortices.

To examine the development mechanisms of the vortices in more detail, we analyzed the eddy kinetic energy budget in the rectangular region indicated in Fig. 8. For simplicity, a system of anelastic equations (Ogura and Phillips 1962), in which the horizontal mean density is assumed to be the density of the basic state, was used to calculate the budget. The x and y coordinates are defined as the directions parallel to and perpendicular to the SZ-A, respectively; and $u$ and $v$ are the wind components in the x and y directions, respectively. Now, any variable $h$ may be written as $h = h_x + h_y$, where $h_x$ is the average value along the x axis, and $h_y$ is the deviation from the average. A linear trend of potential temperature along the x axis is removed from $u_0$ to exclude low-wavenumber variations. By ignoring frictional terms, the tendency of the averaged eddy kinetic energy $\frac{\partial KE}{\partial t}$ can be expressed as

$$\frac{\partial KE}{\partial t} = -\rho_0 \left( \frac{\partial u'v'}{\partial y} + \frac{\partial v'v'}{\partial y} + \frac{\partial w'w'}{\partial y} \right) \frac{\partial h_x}{\partial y}$$

$$-\rho_0 \left( \frac{\partial u'w'}{\partial z} + \frac{\partial v'w'}{\partial z} + \frac{\partial w'w'}{\partial z} \right) \frac{\partial h_y}{\partial z}$$

$$- \rho_0 \left( \frac{\partial u'w'}{\partial y} + \frac{\partial v'w'}{\partial y} + \frac{\partial w'w'}{\partial y} \right) \frac{\partial h_x}{\partial y}$$

$$- C_p \left( \frac{\partial \rho_0 u''}{\partial x} + \frac{\partial \rho_0 v''}{\partial y} + \frac{\partial \rho_0 w''}{\partial z} \right)$$

$$+ \frac{\rho_0}{\Theta} \frac{\partial \Theta}{\partial z}$$

$$= - \rho_0 \left( \frac{\partial u'v'}{\partial y} + \frac{\partial v'v'}{\partial y} + \frac{\partial w'w'}{\partial y} \right) \frac{\partial h_x}{\partial y}$$

$$- \rho_0 \left( \frac{\partial u'w'}{\partial z} + \frac{\partial v'w'}{\partial z} + \frac{\partial w'w'}{\partial z} \right) \frac{\partial h_y}{\partial z}$$

$$- \rho_0 \left( \frac{\partial u'w'}{\partial y} + \frac{\partial v'w'}{\partial y} + \frac{\partial w'w'}{\partial y} \right) \frac{\partial h_x}{\partial y}$$

$$- C_p \left( \frac{\partial \rho_0 u''}{\partial x} + \frac{\partial \rho_0 v''}{\partial y} + \frac{\partial \rho_0 w''}{\partial z} \right)$$

$$+ \frac{\rho_0}{\Theta} \frac{\partial \Theta}{\partial z}$$

(1)
where \( \rho_0 \) is the horizontal mean density over the rectangular region, \( g \) is the gravitational acceleration, \( \Theta \) is the mean potential temperature, \( C_p \) is the specific heat at constant pressure, and \( \sigma \) is the Exner function. Terms on the right-hand side of Eq. (1) may be classified into a horizontal shear production term (HSP), a vertical shear production term (VSP), an advection term (ADV), an eddy transport term (ET), a pressure transport term (PT), and a buoyancy production term (BP). Among these terms, HSP, VSP, and BP represent the production of KE, and the others represent the transport of KE.

Figure 9 shows vertical cross sections of \( \eta \) and \( \bar{v} \). A strong horizontal shear zone, which is tilted northward below 1500 m, exists in the \( \bar{v} \) field (Fig. 9a). The lower part of this strong shear (below 700 m) corresponds to

---

\(^1\) The \( \chi \) divergences in ET and PT do not vanish because of their fluxes through the east and west boundaries.
the southward protruding part of the SZ-A, while the upper part consists of discrete vortices (Fig. 8). There is also strong vertical shear in the $u$ field at approximately $2500 \text{ m}$. In the $v$ field (Fig. 9a), a convergence peak occurs near the surface at $y = 20 \text{ km}$. This peak shifts northward with increasing height, and reaches $z = 800 \text{ m}$ at $y = 30 \text{ km}$. On the other hand, a divergence peak exists at $z = 2500 \text{ m}$ at $y = 30 \text{ km}$.

Figure 10a shows a vertical cross section of the averaged eddy kinetic energy $\overline{KE}$, and Fig. 10b shows the vertical distributions of each term on the right-hand side of Eq. (1) and their sum $\overline{\partial KE/\partial t}$ averaged along the $y$ axis. Two $\overline{KE}$ peaks and several $\partial \overline{KE}/\partial t$ peaks occur below and above 700 m, respectively. A $\overline{KE}$ peak is present below 700 m at $y = 21 \text{ km}$. This peak is intensified by the barotropic instability of SZ-A, because the HSP term is large below 700 m, which is a region of large shear in the $u$ field of the SZ-A (Fig. 9a) and because the long axes of vortices are oriented northeast–southwest, which is opposite to the profile of $u$ (Fig. 8a). The VSP term is negative below 700 m, where the second term of VSP, which is related to the vertical shear of $u$, is negative; the negative VSP term contributes to the suppression of the growth of $\overline{KE}$. Such suppression of the barotropic instability wave by the vertical shear is similar to the results of an idealized experiment on a cold front conducted by Kawashima (2011). The BP term, which has a large value at approximately 400 m, also contributes to the growth of $\overline{KE}$.

Another peak in $\overline{KE}$ occurs at $y = 26 \text{ km}$, between 2000 and 3500 m. This peak is intensified mainly by the BP and VSP terms, which have large positive values at
approximately 1600 and 2500 m, respectively, and are closely related to cumulus convection. The BP term is generated by condensational heat formed in cumulus convection, while the VSP term is related to vertical momentum transport by cumulus convection within areas of strong vertical shear in the $u$ field at approximately 2500 m. The net contribution of the BP and VSP terms is larger than that of the HSP term, indicating that the vortices in the early development stage acquire their energy mainly through cumulus convection. If this process is viewed in the vorticity equation, vorticity increases due to both stretching of vertical vorticity associated with horizontal shear and tilting of horizontal vorticity associated with vertical shear, by strong updrafts within the cumulus convection. This is consistent with the distributions of vorticity and updrafts (Fig. 8). It should be noted that $K_E$ is transported downward by the PT term because $\pi'$ is negative and $w'$ increases with height below cumulus clouds, in which $\theta'$ is positive and strong updrafts exist.

The depth of cold air mass that breaks out from Siberia is less than 1000 m, which is similar to the depth of the SZ-A. As the vertical shear is strong above the cold air, a suppression of barotropic instability waves by the vertical shear, as suggested by Kawashima (2011), is likely to occur. On the other hand, sensible and latent heat fluxes from the sea surface and the cold vortex aloft bring unstable stratification, which is favorable for cumulus convection. Thus, the cumulus convection contributes more to the development of the vortices than does the barotropic instability, resulting in the vortices having a smaller horizontal wavelength than expected from barotropic instability mechanisms.

c. Late development stage

The merger of vortices continues during the late development stage. Figure 11 shows the tracks of the vortices from 2000 UTC 29 December to 1200 UTC 30 December, where the center of each vortex is determined by a local vorticity maximum. At 2000 UTC 29 December, the vortex that later develops into vortex A was located at 40.2°N, 129.2°E; it first moved southwestward as its maximum vorticity gradually increased.

FIG. 11. Tracks of vortices from 2000 UTC 29 Dec to 1200 UTC 30 Dec 2010. Only the tracks of vortices that can be tracked for more than 2 h are shown. The color shade indicates the maximum vorticity of each vortex. The thin black lines extending from the end of each track indicate that the vortex merges into another vortex. The vorticity field at 20 m AGL at 1200 UTC 30 Dec 2010 is also shown by contours.
Then, it changed its direction of movement to south-eastward at 0100 UTC 30 December, while absorbing several vortices that approached it from the northeast. As the mergers continued, the maximum vorticity fluctuated and the horizontal scale of vortex A gradually increased.

In contrast, the vortex that developed into vortex B is difficult to identify because of repeated complex mergers and frequent changes in the core vortices. In each vortex that had some connection to vortex B, the maximum vorticity first increased, then started to decay, and was finally absorbed into another vortex approaching from the northeast; then the latter vortex started to develop. As these processes were repeated, the horizontal scale of vortex B gradually increased. Similar mergers and development patterns of vortices were also observed in the numerical experiment conducted by Kawashima (2005).

Figure 12 shows close-up views of the horizontal wind vector and relative vorticity fields at 20, 1122, 2147, and 3133 m AGL at 0900 UTC 30 December. The maximum vorticities of vortex A and B exceeded $3 \times 10^{-3}$ s$^{-1}$ (Fig. 12a). The horizontal scale of vortex A and B was 50 km, and the interval between them was 100 km; the vortices were connected by a filament of high vorticity exceeding $0.2 \times 10^{-3}$ s$^{-1}$. Similar features were also observed at 1122 m AGL (Fig. 12b). Although the interval of the vortices is a little shorter than the wavelength of the
fastest growing mode (approximately 160 km), as predicted from linear barotropic instability theory, the long axes of these vortices are tilted in the sense that they gain energy from the barotropic instability of the horizontal shear flow associated with the SZ-A (e.g., Gill 1982). Above 2000 m AGL, vortex A and B became obscure, and pairs of positive and negative vorticity, which are likely to be formed by the tilting of horizontal vorticity associated with vertical shear, developed (Figs. 12c, d).

The eddy kinetic energy budget at 0900 UTC 30 December was also analyzed in the rectangular region shown in Fig. 12. Figures 13a and 13b show vertical cross sections of the $x$-averaged horizontal velocity components of $u$ and $v$. The large horizontal shear of $u$ corresponding to the SZ-A is observed below approximately 1500 m (Fig. 13a), which is deeper than that observed at 0100 UTC (Fig. 9a). The increase in the depth of the SZ-A is due to the development of the convective mixed layer through airmass transformation. In the $v$ field (Fig. 13b), both convergence and divergence were intensified. The divergence was located at approximately 4000 m, which suggests that cumulus convection had become more active.

The distribution of $\overline{KE}$ at 0900 UTC 30 December has a peak extending from the surface to 1500 m at $y = 27$ km (Fig. 14a), which is about 6 times as large as that at 0100 UTC (Fig. 10a). Although $\partial \overline{KE}/\partial t$ is nearly uniform due to vertical transport by the PT and ADV terms, the HSP term corresponding to the large horizontal shear (Fig. 13a) and the horizontal convergence (Fig. 13b) is dominant below 1500 m (Fig. 14b). Although the VSP and BP terms also increased from 0100 UTC, their increases were smaller than that of the HSP term (cf. Fig. 10b). These results indicate that the vortices acquire their energy mainly through barotropic instability processes in the late development stage. The increased depth of the SZ-A makes the relative contribution of

![Fig. 13](image1.png)

![Fig. 14](image2.png)
horizontal shear dominant over that of cumulus convection. Note that the peaks of $\overline{KE}$ and $\partial KE/\partial t$ at 3000 m are not physically meaningful because the flows above 1500 m AGL are far from two dimensional (Figs. 12c,d).

d. Mature stage

Vortex B absorbed vortex A to form a single PMC with a cloud-free eye in the mature stage (Figs. 6d and 7). The sea level pressure (SLP) field and cloud pattern demonstrate that the PMC has a nearly axisymmetric shape. Figure 15 shows vertical cross sections along the meridional line AB (Fig. 7). Strong updrafts extending up to 5000 m, with maximum velocities of 5 m s$^{-1}$ occurring both to the south and north of the center of the PMC (near 36.4°N; Fig. 15a), correspond to cumulus clouds in the spiral cloud bands surrounding the eye. Meridional circulation comprises convergence at lower levels and divergence in upper levels. The thickness of the inflow layer near the surface is approximately 2000 m, which nearly matches the depth of the mixed layer. A weak downdraft, with a maximum value of 0.5 m s$^{-1}$, occurs at the center of the PMC.

The potential temperature field (Fig. 15c) showed a significant warm core structure, with the core 3 K warmer than the surroundings at around 2000 m AGL. The distribution of the water vapor mixing ratio $q_v$ demonstrates that dry air occupies the center of the PMC (Fig. 15d), where downdrafts are dominant. These distributions of vertical velocity, potential temperature, and $q_v$ imply that the warm core and cloud-free eye
structure are formed by adiabatic warming associated with the downdraft. The negative pressure anomaly, which is in gradient wind balance with a rotational wind, increases toward the lower levels where the vorticities are higher (Fig. 15b). The resulting downward pressure gradient force is larger than the upward buoyancy, thus causing the downdraft.

To examine the formation processes of the warm core in more detail, we conducted a backward trajectory analysis. A total of 36 parcels were distributed at 2000 m AGL around the center of the warm core (36.28°–36.48°N, 132.5°–132.75°E) at 1700 UTC 30 December, and their backward trajectories were obtained for a period of 8 h (Fig. 16). The model outputs, generated at intervals of 5 min, were used for the backward integration, where the forward Euler method with a time step of 10 s was adopted. The velocity components at each point on the trajectory were calculated by linearly interpolating the model outputs spatially and temporally. The potential temperatures and equivalent potential temperatures of the parcels were also calculated using the same interpolation method.

The trajectory analysis revealed that the parcels that reached the warm core were originally located in four regions relative to the vortex center: near the vortex center, and on the northeastern, western, and southern sides of the vortex (Fig. 16). Hereafter, the parcels in these regions are referred to as Ps-C, Ps-NE, Ps-W, and Ps-S, respectively. The Ps-C were advected together with the vortex, while other parcels were gradually taken into the vortex and then moved with it. Most of Ps-C, Ps-NE, and Ps-W were located at low levels at 0900 UTC. While they traveled at low levels, their potential temperatures and equivalent potential temperature gradually increased due to sensible and latent heat fluxes from the sea surface (Figs. 17 and 18). Then, they rapidly ascended as they approached the vortex. During this rapid ascent, the potential temperature of the parcels also increased rapidly (Figs. 17a–c), while nearly conserving their equivalent potential temperature (Figs. 18a–c). This implies that

![Fig. 16. Trajectories of parcels relative to the PMC. The plot is constructed so that the center of the PMC remains at (0, 0). Note that all trajectories migrate toward the center. Colors on each trajectory indicate parcel height. (inset) The track of the PMC is shown together with the contours of SLP at 1700 UTC 30 Dec 2010. The dots indicate the position of parcels at 1700 UTC 30 Dec 2010.](image-url)
the rapid ascent was caused by updrafts in cumulus convection that are accompanied by diabatic heating. Parcels raised by cumulus convection eventually descended into the warm core, accounting for approximately two-thirds of all parcels. On the other hand, Ps-S originally located at upper levels descended into the warm core while conserving their potential temperatures and equivalent potential temperatures (Figs. 17d and 18d). These results confirm that the warm core is formed by adiabatic warming associated with downdrafts.

The structure of the PMC during the mature stage depicted above resembles a tropical cyclone or a polar low with little baroclinicity (e.g., the M0 case in Yanase and Niino 2007), which indicates that it may have developed through a mechanism similar to that of a tropical cyclone, namely, cooperative interaction between the vortex flow and cumulus convection. It is known that such a vortex can develop only when the initial vortex is sufficiently strong (Emanuel and Rotunno 1989; Yanase and Niino 2007). In the present case, the PMC appears to change its development mechanism from shear instability to a hurricane-like mechanism after it exceeds some threshold strength. Note that Yokota et al. (2012) reported a similar conclusion regarding tropical cyclogenesis associated with an ITCZ breakdown in which initial vortices caused by barotropic instability, after reaching sufficient strength, eventually developed into a tropical cyclone.

5. Sensitivity experiments

As described in the previous section, the PMC in the present study was generated in a shear zone as an array of small vortices associated with individual cumulus clouds in the early development stage, developed through barotropic instability in the late development stage, and finally developed through cooperative interaction between the vortex and cumulus convection during the mature stage. We infer that the contributions of the physical processes, such as condensational heating and heat fluxes from the sea surface, to the development of the PMC may differ at each stage. To examine which physical processes affect the development of the PMC at each stage, we performed sensitivity experiments.

For the sensitivity experiments, the model setting was the same as that described in section 3 except that sensible and latent heat fluxes from the sea surface and condensational heating were switched on/off. A total of six experiments were performed, including experiment CTL described in section 4 (CTL, DRY, NO_SH, NO_LH, NO_SLH, and NO_SLH_DRY) (Table 2). Condensational heating was switched off in DRY. In
No_SH, No_LH, and No_SLH, sensible heat flux, latent heat flux, and both sensible and latent heat fluxes were removed, respectively. In No_SLH_DRY, all three processes were switched off.

Two types of sensitivity experiments were performed. One was a 48-h experiment, in which the physical processes were switched on/off throughout the entire integration time of 48 h. The other was a 9-h experiment, in which the physical processes were switched on–off for only 9 h after 0900 UTC 30 December (24 h after the start of the integration), which covers the mature stage during which the PMC is over the ocean. As mentioned in Yanase et al. (2004), removal of a certain physical process for a prolonged duration, as in the 48-h experiments, changes not only the vortex itself but also the environment in which the vortex develops. In fact, the meso-α-scale low that was located to the east of the PMC was transformed in the 48-h experiments, so that the cold air outbreak in which the PMC developed was significantly changed. Moreover, in the 48-h experiment, the mature stage of the PMC did not exist in some of the sensitivity experiments. Therefore, we performed the 9-h experiments to examine the effects of physical processes mainly on the PMC at the mature stage, while keeping the change of the environment as small as possible.

### a. 48-h experiments

Figure 19 shows the temporal evolution of the maximum relative vorticity at 20 m AGL in the PMC between 1800 UTC 29 December and 2100 UTC 30 December, and Fig. 20 shows the relative vorticity at 20 m and the SLP field at 1700 UTC 30 December for the 48-h experiments. Shear zones corresponding to the SZ-A in CTL were formed in all experiments by 1800 UTC 29 December (not shown). The maximum relative vorticity in CTL gradually increased in the early development stage, rapidly increased in the late development stage, and fluctuated around the value of $5 \times 10^{-3}$ s$^{-1}$ in the mature stage. The maximum vorticity in

### Table 2. The six sensitivity experiments.

<table>
<thead>
<tr>
<th>Expt name</th>
<th>Physical process considered</th>
</tr>
</thead>
<tbody>
<tr>
<td>CTL</td>
<td>Control run with all the physical processes included</td>
</tr>
<tr>
<td>DRY</td>
<td>No condensational heating</td>
</tr>
<tr>
<td>No_SH</td>
<td>No sensible heat flux from the sea surface</td>
</tr>
<tr>
<td>No_LH</td>
<td>No latent heat flux from the sea surface</td>
</tr>
<tr>
<td>No_SLH</td>
<td>No heat fluxes from the sea surface</td>
</tr>
<tr>
<td>No_SLH_DRY</td>
<td>No condensational heating and no heat flux from the sea surface</td>
</tr>
</tbody>
</table>
No_SH displayed a temporal evolution similar to that in CTL, although exhibiting slightly smaller values. In contrast, the maximum vorticities in DRY and No_LH did not increase during the integration. The vortices and shear zones in No_SLH and No_SLH_DRY disappeared during the integration.

The other characteristics of the vortices in No_SH were also similar to those of CTL: generation of small vortices in a shear zone in the early development stage (not shown), merger of vortices and development of larger vortices in the late development stage (not shown), and a PMC with a pressure depression in the mature stage (Fig. 20c). These results indicate that condensational heating associated with cumulus convection is crucial to the development of the PMC. The fact that the difference between CTL and No_SH is small implies that sensible heat flux plays a secondary role in the development of the PMC, contributing toward making the environment favorable for cumulus convection by destabilizing stratification in the lower layer. An eddy kinetic energy budget analysis on the vortices in No_SH at 0900 UTC 30 December showed that the depth of SZ-A in No_SH at 0900 UTC 30 December was approximately 1000 m, which is less than that in CTL (approximately 1500 m), and the contribution of the HSP term was also smaller than that in CTL (not shown). Therefore, by increasing the depth of the shear zone, sensible heat flux also contributes to the development of an environment favorable for horizontal shear instability.

In No_LH, vorticity did not increase, although condensational heating was included. As the cold air mass from Siberia contains little water vapor, little condensation occurs without a supply of water vapor from the sea surface, resulting in a similar temporal evolution to that in DRY. In these two experiments (NO_LH and DRY), wavelike structures with wavelengths longer than the interval of the small vortices in CTL and No_SH were observed during early and late development stages (not shown). An eddy kinetic energy budget analysis on the wavelike structure in DRY showed that the value of $K_E$ was increased mainly by the HSP term at low levels, and no significant peaks in VSP or in BP were observed at upper levels (not shown). Thus, these wavelike structures are likely to be caused by barotropic instability in the shear zone; they eventually rolled up to form a vortex (Figs. 20b,d), although its maximum vorticity did not significantly increase.

Similar wavelike structures were also observed in No_SLH and No_SLH_DRY, but they, as well as the shear zone, quickly decayed (Figs. 20e,f). The SZ-A forms in a surface pressure trough that is associated with a thermal ridge between two cold air masses that originated in Siberia and the Korean Peninsula. Without sea surface fluxes, however, cold air covers the whole of the Japan Sea, and consequently, the thermal ridge does not form. Therefore, both sensible and latent heat fluxes from the sea surface contribute to the maintenance of the SZ-A.

b. 9-h experiments

Temporal evolution of the central pressure of the PMC and the relative vorticity averaged around the center of the PMC from 0900 UTC 30 December to 1800 UTC 30 December for the six sensitivity experiments are shown in Fig. 21, and the relative vorticity at 20 m AGL and the SLP fields at 1700 UTC 30 December are shown in Fig. 22. Although pressure depressions at the center of the PMCs are observed at 1700 UTC 30 December for the six sensitivity experiments are shown in Fig. 21, and the relative vorticity at 20 m AGL and the SLP fields at 1700 UTC 30 December are shown in Fig. 22. Although pressure depressions at the center of the PMCs are observed at 1700 UTC 30 December in all experiments (Fig. 22), the SLPs in DRY, No_SLH, and No_SLH_DRY do not decrease (Fig. 21a). As SLP can depend on the large-scale pressure field, which differs among the six experiments, we use the averaged vorticity to compare the intensities of vortices. Note that the center position of the PMCs at 1700 UTC varies between the six experiments. This is because the wind fields that advect the PMC vary on account of difference in the structures of the mixed layers.

The PMC in CTL had the largest averaged relative vorticity among the six experiments (Fig. 21b). The averaged relative vorticities of the PMC in No_SH and No_LH also increased with time, although they were somewhat smaller than that in CTL. The patterns of the SLP and vorticity fields in these two experiments were similar to those in CTL (Figs. 22c,d). In No_SLH, the averaged relative vorticity increased initially, but started to decrease at 1400 UTC. The averaged vorticities in DRY and No_SLH_DRY did not increase, although the
SLP field displayed a slight deepening of the central pressure (Figs. 22b,f). In these two experiments, vorticity was only simply advected and deformed, and the SLP field adjusted to the vorticity field to achieve a gradient wind balance.

The PMC did not develop without condensation in the 9-h experiment. This result indicates that condensational heating is also crucial to the development of a PMC in the mature stage. However, in contrast to the 48-h experiment, the PMC in No_LH developed in the 9-h experiment. As there was abundant water vapor around the PMC at the beginning of the 9-h experiment, cumulus convection could be maintained by collecting this ambient water vapor during the relatively short period of 9 h. On the other hand, the development of the PMC in No_SLH stopped 5 h after the heat fluxes were removed. Removing the sensible heat flux in addition to the latent heat flux stabilized the stratification of the lower layer, resulting in a suppression of cumulus convection and the development of the PMC.

6. Summary and conclusions

A meso-β-scale PMC observed over the Japan Sea on 30 December 2010 was studied using observational analysis and numerical simulations. Although PMCs
often form in a preexisting JPCZ, the present PMC formed without the presence of a JPCZ; instead, the JPCZ was generated behind the PMC.

Analysis of observational data acquired from satellite images showed that, prior to the development of the PMC, a chain of vortices appeared in an east–west-oriented cloud band. This cloud band corresponded to a trough that extended westward from a meso-a-scale low that moved east over the Japan Sea. The vortices eventually merged into a PMC with a horizontal scale of about 150 km, a cloud-free eye, and spiral cloud bands. A JPCZ formed behind the PMC.

A numerical simulation using the JMA-NHM model with a 2-km horizontal mesh that started prior to the formation of the east–west-oriented cloud band, successfully reproduced the observed PMC. The results of the numerical simulation suggest that the development phase of the PMC can be divided into three stages: an early development stage, in which a number of small vortices appear in a shear zone; a late development stage, in which the merger of vortices proceeds and a few larger vortices form; and a mature stage, in which only a single PMC is present.

We performed an eddy kinetic energy budget analysis to examine the developmental mechanisms of vortices during the early and late development stages. The analysis revealed that in the early development stage, horizontal shear production (HSP) is confined to low levels and is smaller than buoyancy production (BP) and vertical shear production (VSP) at upper levels, where both of BP and VSP are related to cumulus convection. This result indicates that, during the early development stage, vortices acquire their energy through discrete cumulus convection. During the late development stage, on the other hand, HSP becomes relatively dominant because the depth of the shear zone increases as the mixed layer develops, and the vortices develop through barotropic instability.

The structure of the PMC during the mature stage was also examined in detail. The simulated PMC was accompanied by a cloud-free eye and spiral cloud bands. A warm core structure was observed at the center of the PMC, where a dry downdraft developed. Backward trajectory analysis confirmed that the warm core formed due to adiabatic warming in the downdraft. This PMC structure resembles that of a tropical cyclone.

To examine the effects of physical processes on the development of the PMC, we performed sensitivity experiments in which sensible and latent heat fluxes from the sea surface and condensational heating were switched on–off. Two types of sensitivity experiments were performed, a 48-h experiment, in which the physical processes were switched on–off throughout the whole integration time of 48 h, and a 9-h experiment, in which the physical processes were switched on–off for only 9 h after 0900 UTC 30 December. The 48-h experiments demonstrated that condensational heating is essential to the development of the PMC. Sensible heat flux plays a secondary role, although it contributes to the creation of an environment favorable for cumulus convection by destabilizing stratification in the lower layer, and also for horizontal shear instability by increasing the depth of the shear zone. The latent heat flux moistens the environment and makes it favorable for cumulus convection. The 9-h experiments indicated that condensational heating is also essential for the development of the PMC during the mature stage. The effects of horizontal convergence of preexisting water vapor also play an important role in development of the PMC.

Cumulus convection was essential to the genesis and development of the PMC that developed over the Japan Sea on 30 December 2010. In particular, discrete cumulus convection...
convection generated vortices in the early development stage. This suggests that the resolution of cumulus convection is desirable for an accurate prediction of PMCs. As the height of cumulus convection over the Japan Sea in winter is 5 km or less, a horizontal mesh of at least 2 km is necessary to resolve cumulus convection. Indeed, the PMC is not reproduced in the MANAL data, which uses a 5-km mesh and parameterized cumulus convection (Fig. 5e).
The present study examined only a single PMC in detail. However, the size, structure, and development of PMCs depends on various physical parameters, such as surface fluxes, environmental stratification, cumulus convection, the depth of the shear zone, and so on. To deepen our understanding of the diversity of PMCs over the Japan Sea, additional case studies and idealized experiments must be completed in the future.

Acknowledgments. The authors thank Prof. Keita Iga and Dr. Wataru Yanase for their helpful comments and constructive criticism of this paper, Dr. Akiyoshi Wada of the Meteorological Research Institute for providing the SST data, and Tim Dunkerton and three anonymous reviewers for their insightful comments on this paper. Thanks are also extended to the staff and students of the Dynamic Marine Meteorology Group, Atmosphere and Ocean Research Institute, The University of Tokyo for their kind help. This work was supported in part by a Grant-in-Aid for Scientific Research (A) 24244074 from the Japan Society for the Promotion of Science. The first author was also supported by the Program for Leading Graduate Schools, MEXT, Japan. The computation was carried out using the computer facilities at the Atmosphere and Ocean Research Institute and Information Technology Center, The University of Tokyo. Some figures were prepared using the Grid Analysis and Display System (GrADS) by Brian Doty.

REFERENCES


