A Dual-Doppler Radar Study of Longitudinal-Mode Snowbands.  
Part I: A Three-Dimensional Kinematic Structure of Meso-γ-Scale Convective Cloud Systems within a Longitudinal-Mode Snowband

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ABSTRACT
Observations of snowstorms were carried out around Ishikari Bay, Hokkaido, Japan, from December 1991 to February 1992. On 15 January 1992, a longitudinal-mode snowband was observed by a dual-Doppler radar system. The snowband was composed of meso-γ-scale (~20 km) convective cloud systems. The three-dimensional kinematic structures and organization processes of the meso-γ-scale systems were studied in detail.

Strong band-parallel winds appeared in the upper-north rear of the meso-γ-scale systems. Since their speed was faster than that of the cellular radar echoes, the winds formed rear-to-front currents. Increasing in volume and speed, the rear-to-front currents developed into meso-γ scale and penetrated toward the lower front and lower south of the systems. Consequently, the rear-to-front currents caused strong meso-γ-scale convergence and the enhancement of updrafts at their leading edges. The transport of the band-parallel momentum greatly contributed to the successive development of new convective cells within the systems and the organization of the meso-γ-scale systems in the snowband. Ice/snow particles were transported by the updrafts toward the rear and north of the systems in the overlying stable layer. They then evaporated outside the clouds. The downdraft caused by evaporative cooling played an important role in the transport of band-parallel momentum from the upper to the lower levels and from the outside to the inside of the snowband.

1. Introduction
In winter, many cloud bands frequently form and develop over the Sea of Japan during cold-air outbreaks from the Eurasian landmass. They bring a large volume of snowfall to the coastal regions of Japan. The cloud bands show two patterns: longitudinal (wind-parallel) mode and transverse (wind-normal) mode (e.g., Tsuchiya and Fujita 1967). The longitudinal-mode cloud bands have also been known as “cloud streets.” Since organized flow plays an important role in the vertical transport of heat, moisture, momentum, and chemical substances within the planetary boundary layer (PBL), many observational, theoretical, and numerical studies of cloud streets have been done in association with horizontal roll vortices (e.g., Brown 1980; Etling and Brown 1993; Atkinson and Zhang 1996). Linear theories have proposed dynamic instabilities (inflection point instability and parallel instability) and thermal instability as the formation mechanisms of horizontal roll vortices (e.g., Asai 1970; Brown 1970; Kuettner 1971). These mechanisms have been confirmed by observations (e.g., LeMone 1973). In the cloud free PBL, thermal streets and roll-like circulations were detected using a high power radar (Konrad 1968) and dual-Doppler radars (e.g., Kropfi and Kohn 1978; Rabin et al. 1982). The linear theories also can explain the structure of these roll-like circulations but cannot explain the wide range of the observed aspect ratios (cloud-row spacing to cloud-top height) of cloud streets (Walter 1980; Miura 1986). In particular, the cloud streets developing during cold-air outbreaks over the ocean show larger aspect ratios. Several theories to explain the observed large aspect ratios have been presented (e.g., Walter and Overland 1984; Clark et al. 1986; Sykes et al. 1988). Since most of these studies, however, assumed cloud streets to be two-dimensional, they cannot discuss the role of band-parallel wind in the organized two-dimensional circulation. Cloud streets are composed of cumuli lined up like pearls on a string [as stated by Kuettner (1959)]. Such a cloud pattern has also been reported by many other investigators (e.g., Christian and Wakimoto 1989; Wakimoto and Atkins 1994; Weckwerth et al. 1996). Brown

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streets. This result qualitatively confirmed the observation that the characteristic of horizontal roll vortices change considerably after the condensation of water vapor occurs. Sykes et al. (1990) reported that after the development of convective clouds, three-dimensional circulations associated with individual clouds destroy the dominant roll structure and dominate the airflow structures of the snowbands. Kristovich (1993) investigated longitudinal-mode snowbands using dual-Doppler radar. He reported that roll-like circulations dominated the mean airflow structures of the snowbands. However, these investigators also reported that the reflectivity and wind fields fluctuated considerably along the snowbands. Kristovich emphasized the need for research that studies the interactions between rolls and kilometer-scale convective systems along the snowbands.

Three-dimensional numerical simulations of cloud streets showed that the characteristics of horizontal roll vortices change considerably after the condensation of water vapor occurs. Sykes et al. (1990) reported that after the development of convective clouds, three-dimensional circulations associated with individual clouds destroy the dominant roll structure and dominate the airflow structure of the PBL. LeMone and Pennell (1976) observed that the formation of deep clouds affected the subcloud-layer circulations. Chlond (1992) indicated that the dynamic energy due to strong vertical wind shear is essential in the formation of rolls during the first phase of a cold-air outbreak, but as clouds evolve, thermal energy due to latent heat release becomes more important to the maintenance of cloud streets. This result qualitatively confirmed the observational results reported by Brümmer et al. (1992). Rao and Agee (1996) emphasized that the evaporation of particles in the subcloud layer and particle loading may greatly affect the airflow structures of the rolls. These studies indicate that precipitating cloud bands have different airflow structures from cloud-free horizontal roll vortices and nonprecipitating shallow cloud streets.

In recent years, the kinematic structures of snowstorms have been actively investigated over the Sea of Japan. Sakakibara et al. (1988) observed squall line-like snowbands using a single-Doppler radar. They proposed that the evaporative cooling of ice/snow particles played an important role in the enhancement of mesoscale downdrafts within the snowbands. Shirooka and Uyeda (1990) found the strong downdraft below snow clouds known as “snowburst.” From dual-Doppler radar observations of isolated convective snow clouds, Yamada et al. (1994) reported that downdrafts appeared in the mature and decaying stages of clouds situated in the high-reflectivity regions. They attributed the formation of the downdrafts to particle loading. Maki et al. (1992) observed a longitudinal-mode snowband. They found that a strong downdraft existed in the center of a snow cloud and induced a low-level cold outflow in front of the snow cloud. However, they did not examine how the cold downdraft is formed in the snowband. In addition, Fujiyoshi et al. (1992) reported that a snowband was composed of meso-$\gamma$-scale radar echoes.

These numerical and observational studies suggest that three-dimensional aspects may be important for the formation and maintenance of a snowband. To clarify the maintenance mechanisms of “three-dimensional” longitudinal-mode snowbands, which cannot be explained by horizontal roll vortices only, it is necessary to investigate three-dimensionally the structures and maintenance processes of longitudinal-mode snowbands. However, the three-dimensional dynamic structure of a longitudinal-mode snowband and its temporal change are not yet clear.

To reveal the heavy snowfall mechanism, a special observation was carried out around Ishikari Bay, Hokkaido, Japan, from 25 December 1991 to 10 February 1992. During this period, longitudinal-mode snowbands formed frequently. A typical longitudinal-mode snow-
band was observed by a dual-Doppler radar system on 15 January 1992. The purpose of this paper is to describe the three-dimensional kinematic structures of meso-γ-scale systems composing the snowband and how they change with time. Special attention will be given to the role of a strong band-parallel wind in the maintenance and formation mechanism of the meso-γ-scale systems.

2. Observations and data

The data analyzed in this paper were mainly obtained by the X-band dual-Doppler radar system of the Institute for Hydrospheric–Atmospheric Sciences (IHAS), Nagoya University, Japan. Figure 1 shows the radar sites, the area of quantitative radar observation, and a topographic map of the observation area. The Doppler radars were installed at Otaru (43°14′N, 141°01′E) and Atsuta (43°22′N, 141°26′E). The distance between the two radars was 38 km. Upper-air sounding data observed at the Sapporo Meteorological Observatory were also used.

Table 1 summarizes the characteristics of the dual-Doppler radar system. Both radars have the same characteristics. On 15 January 1992, longitudinal radar-echo bands appeared over Ishikari Bay between approximately 1100 LST (LST = UTC + 9 h) and 1600 LST. During this period, snowfalls (up to 10 cm deep) were observed on Ishikari Plain. Each radar was operated in a coordinated operation mode at an interval of exactly 6 min. This mode consisted of nine velocity–azimuth display (VAD) scans at elevation angles of 0.3° to about 9°, one VAD scan at an elevation angle of 20°, two range–height velocity (RHV) scans at two different azimuth angles, and vertical pointing measurements.

Reflectivity and Doppler velocity data were interpolated on a Cartesian coordinate system with horizontal and vertical grids of 0.5 and 0.25 km, respectively, using a Cressman weighting function (Cressman 1959). The horizontal radius of influence of the weighting function was set at 1.5 km to dampen the convective-scale fluctuations. A correction of radar-echo advection was applied using the mean motion of cellular radar echoes. The horizontal wind field was then synthesized, assuming that the average fall speed of snow particles was 1.0 m s⁻¹. Vertical velocities were derived by integrating the horizontal divergence field from the sea surface with the anelastic mass continuity equation. Where the radar-echo top was higher than the top of the mixed layer, vertical velocities were adjusted to be 0 m s⁻¹ at the sea surface and the echo-top height using a variational function (O’Brien 1970). Strictly speaking, the radar-derived quantity in the snowfall event is the equivalent...
FIG. 4. Vertical profiles of wind direction and wind speed observed at Sapporo at 0900 (dashed line) and 1500 LST (dot-dashed line) on 15 January 1992. The average of these winds is also shown (solid line).

reflectivity factor $Z_e$. However, $Z_e$ will be written as $Z$ in the following sections for simplicity.

3. Atmospheric conditions

Figure 2 shows weather maps at 0900 LST 15 January 1992. A moderate cold-air outbreak from the Eurasian landmass had started the previous day. At the surface (Fig. 2a), the isobars were densely aligned around Hokkaido (shown by the hatched region) and over the Sea of Okhotsk (located to the north of Hokkaido). A cold air mass, the minimum temperature of which was $-42^\circ$C, existed over the continent. These features indicate that the cold-air outbreak continued in this region. At a level of 850 hPa (Fig. 2b), the isometric lines (solid lines) off the west coast of Hokkaido oriented from the west-northwest toward the east-southeast and were almost parallel to the isotherms (broken lines), which suggests that little or no cold-air advection was occurring at this level. At low levels around Hokkaido, these pressure and temperature patterns were also observed at 2100 LST (not shown), indicating that the large-scale atmospheric conditions were maintained during the occurrence of the snowband.

Figure 3 shows the vertical air structure above Sapporo. Since no upper-air soundings were made during the occurrence of the snowband, the profiles at 0900 LST are shown. A typical mixed layer existed from the surface to 2.0 km above mean sea level (MSL). A weak inversion layer existed from 2.0 to 2.2 km. The top and bottom of the inversion layer were characterized by a local minimum and maximum relative humidity. The surface temperature and dewpoint temperatures were $-3.7^\circ$ and $-1.2^\circ$C, respectively, and the lifting condensation level (LCL) was 0.95 km. The vertical profile of temperature at 2100 LST (not shown) resembled the profile at 0900 LST; that is, a capped inversion layer existed between 2.1 and 2.2 km at 2100 LST. In addition, there were no significant changes in the cloud-top temperature over Ishikari Bay, as observed by the Geostationary Meteorological Satellite (GMS) between 0900 and 2100 LST (not shown). Therefore, the inversion

Fig. 5. Radar reflectivity pattern at an altitude of 1.0 km at 1231 LST. The hatched regions indicate ground echoes. Thick solid lines show the seashore. The region (labeled shadow) between two broken lines indicates the region in which the collection of radar data was prevented by the terrain. The boxes outlined by the dot-dashed line and the dotted line correspond with the radar display areas in Fig. 11 and 14, respectively.
Table 2. Summary of the characteristics of two snowbands.

<table>
<thead>
<tr>
<th>Snowband</th>
<th>Average width (km)</th>
<th>Maximum echo-top height (km)</th>
<th>Orientation (deg)</th>
<th>Mean motion of cellular radar echoes (m s(^{-1}), deg)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Band I</td>
<td>10.0</td>
<td>2.5</td>
<td>280</td>
<td>12.5, from 280</td>
</tr>
<tr>
<td>Band II</td>
<td>9.0</td>
<td>3.0</td>
<td>280</td>
<td>12.5, from 280</td>
</tr>
</tbody>
</table>

Figure 4 shows vertical profiles of wind direction and wind speed observed at Sapporo at 0900 and 1500 LST. The average of these wind parameters is also shown. In the mixed layer, the averaged wind shows the following features. Except in the lowest layer, the wind direction was backing with height from the northwest to the west-northwest. The wind speed increased with height. The mean vertical wind shear was \(7 \times 10^{-3}\) s\(^{-1}\) from the surface to 2.0 km. This value meets the range of values reported by Miura (1986). Following the definition of Tsuchiya and Fujita (1967), in which the wind speed at the surface was assumed to be 0 m s\(^{-1}\), the mean vertical wind shear from the surface to 2.0 km was calculated to be \(9 \times 10^{-3}\) s\(^{-1}\). This value satisfies Tsuchiya and Fujita’s criterion for the occurrence of cloud streets. Therefore, the snowband occurred in atmospheric conditions under which cloud streets can develop.

4. General features of snowbands

Figure 5 illustrates a constant-altitude plan position indicator (CAPPI) of radar echoes at an altitude of 1.0 km at 1231 LST. Two snowbands, designated as band I and band II, are clearly seen over the middle and southern parts of Ishikari Bay. Northern Ishikari Bay was studded with cellular radar echoes 5–10 km in size. These cellular radar echoes were also weakly aligned. The characteristics of bands I and II, based on a radar-echo boundary of 10 dBZ, are summarized in Table 2. The two snowbands had similar characteristics. The echo-top levels were apparently higher than the inversion level (2.0 km), suggesting penetration of the updrafts into the overlying stable layer.

Figure 6 shows the NOAA-11 imagery of channel 2 (near-infrared) at 1323 LST. Several cloud streets were identified off the west coast of Hokkaido. Individual cloud streets were difficult to distinguish toward the downstream area (commonly observed during cold-air outbreaks over the Sea of Japan) due to the increase in width of each cloud street (Miura 1986) and the development of anvil-like cloud sheets.

As illustrated in Fig. 5, band I was composed of several high-reflectivity (\(\geq 22\) dBZ) regions approximately 20 km in size with a maximum reflectivity of about 30 dBZ. Assuming that a cellular radar echo less than 10 km in size corresponds with an individual convective cloud (Christian and Wakimoto 1989), these high-reflectivity regions can be regarded as meso-\(\gamma\)-scale convective cloud systems. Figure 7 shows a reflectivity pattern at an altitude of 1.0 km averaged between 1201 and 1337 LST. As denoted in the figure, the axes parallel and perpendicular to the alignment of the snowbands will be referred to as the \(x\) axis and the \(y\) axis, respectively. Velocity components along the \(x\) and \(y\) axes will be referred to as \(u\) and \(v\), respectively. The reflectivity along the \(x\) axis changed little in band I but changed to a large extent in band II. Orographic factors would have affected the development of band II [as reported by
Fujiyoshi et al. (1992)]. This paper will focus on the structure of band I. The structure of band II and the interaction between bands I and II will be reported in a forthcoming paper.

Figure 8 shows a hodograph of the winds observed over Ishikari Bay. Here, the ordinate and the abscissa do not show $u$ and $v$; instead they show the northward and eastward components of the horizontal winds. These were derived by averaging horizontally and temporally the dual-Doppler radar data in the area outlined by a dotted line in Fig. 7. The upper-air sounding data ob-

![Figure 7](image1.png)

**Fig. 7.** Radar reflectivity pattern averaged from 1201 to 1337 LST at an altitude of 1.0 km. The hatched regions represent the ground echoes. Thick solid lines show the seashore. The area (labeled shadow) between two broken lines indicates the region in which the collection of radar data was prevented by the terrain. The $x$–$y$ coordinate system used in this paper is denoted by arrows. The area outlined by the dotted line indicates the region in which the wind data were averaged to derive the hodograph presented in Fig. 8.

![Figure 8](image2.png)

**Fig. 8.** Hodograph of the winds observed over Ishikari Bay. The wind speeds were calculated by horizontally and temporally averaging wind data obtained by the dual-Doppler radar. The averaged area is the region where the angles between the beams of the two Doppler radars are from $70^\circ$ to $110^\circ$ over Ishikari Bay (outlined by the dotted line in Fig. 7). The averaged period is between 1201 and 1337 LST. The solid line with circles shows the wind inside the strong echo regions of the snowband ($\geq 10$ dBZ). The thick broken line with triangles shows the wind outside the strong echo regions of snowband ($<10$ dBZ). The dot–dashed line shows the upper-air sounding data observed at Sapporo. The average of winds observed at 0900 and 1500 LST is shown. The alignment of the snowband ($280^\circ$) is shown by the thin broken line.

![Figure 9](image3.png)

**Fig. 9.** Vertical profiles of horizontal divergence (solid line), $\partial u/\partial x$ (broken line), and $\partial v/\partial y$ (dot-dashed line) averaged horizontally and temporally in the regions of high reflectivity ($\geq 25$ dBZ) in band I. The averaged region is $-10$ km $\leq x < 10$ km. The averaged period is between 1201 and 1255 LST.
served at Sapporo is also shown in this figure. The axis of the snowbands (280°) was parallel to the wind direction at an altitude of 1.5 km over Ishikari Bay. Although the orientation of the snowbands was nearly parallel to the wind shear vector between the altitudes of 0.5 and 2.0 km above Sapporo, it was nearly perpendicular to the wind shear observed over Ishikari Bay. This is mainly attributed to the difference in wind speed between the two regions. Below 1.0 km, the wind speed at Sapporo was less than that observed over Ishikari Bay, but above this altitude it was greater than over Ishikari Bay. In addition, there was a marked difference in wind speed between the inside and outside of the strong echo regions of snowbands. Inside the regions (≥10 dBZ), the wind speed was nearly constant (~11 m s⁻¹) between the altitudes of 0.5 and 1.5 km. In contrast, outside the regions (<10 dBZ), the wind speed increased with height between the altitudes of 0.5 and 1.5 km. These differences in wind profile reflect the degree of vertical momentum transport due to convection. It has been observed that at the ground surface strong winds blow only in snowbands. Considering that no precipitating clouds passed over Sapporo when the upper-air soundings were made, the environmental winds outside the detectable area of the Doppler radars over Ishikari Bay had strong vertical wind shears, as were observed at Sapporo. This point will be discussed in section 7.

5. Radar-echo and airflow structures in vertical cross sections perpendicular to the elongated axis of band I

a. Mean structure

In this subsection, the time-averaged vertical structures of band I will be studied for comparison with the other snowbands reported in previous literature (Kelly 1982; Kristovich 1993). It should be noted that the averaged structure of a snowband does not necessarily show a typical internal structure when the system is time dependent.

Figure 9 shows vertical profiles of mean horizontal divergence and horizontal shear ($\partial \omega / \partial x$ and $\partial \omega / \partial y$) in the

Fig. 10. Mean structure of band I in a vertical cross section ($x = 0$ km) perpendicular to the band axis. The averaged period is between 1201 and 1255 LST. (a) Radar reflectivity and airflow. Reflectivities are contoured every 3 dBZ above 10 dBZ. (b) Vertical wind velocity $w$. The contours are drawn every 0.05 m s⁻¹. The thick lines are drawn every 0.1 m s⁻¹. The broken lines indicate negative values. The heavily stippled and hatched regions indicate $w > 0.05$ m s⁻¹ and $w < -0.05$ m s⁻¹, respectively. (c) Band-normal component (ν) of the horizontal winds (m s⁻¹). The broken lines indicate negative values. (d) Band-parallel component (u) of the horizontal winds (m s⁻¹). The heavily stippled region shows $u > 11.4$ m s⁻¹. In all panels, the lightly stippled region indicates reflectivity greater than or equal to 10 dBZ.
regions of high reflectivity (≥25 dBZ) in band I. Clearly, convergence existed below 1.25 km and divergence existed above 1.5 km. In the convergence layer, the magnitude of $\partial u / \partial x$ was comparable with that of $\partial v / \partial y$, which indicates that the band-parallel winds as well as the band-normal winds contributed to the formation of updrafts in the snowband.

Figure 10 shows the mean structure of band I in a vertical cross section on the $y$ axis (i.e., $x = 0$ km, see Fig. 7), where $v$ winds could be calculated with a high level of accuracy because the band-normal component ($v$) of the horizontal winds was almost directly measured by the Doppler radar at Otaru. Radar reflectivities and horizontal winds in band I were averaged between 1201 and 1255 LST, since band I had begun dissipating after 1301 LST. Figure 10a shows the vertical cross section of mean reflectivity and airflow patterns. The snowband showed an asymmetric radar-echo pattern; that is, the southern (left) side of the snowband had a sharper edge, stronger intensity, and a higher echo top than the northern (right) side. The echo-top level of 1.75 km was slightly lower than the altitude of the mixed-layer top (2.0 km). The width and echo intensities of the snowband changed little below a height of 1.0 km, which agrees well with the LCL at 0900 LST (0.95 km) in Fig. 3. This result indicates that snow particles did not, on average, grow to any great extent or evaporate below the cloud base, as indicated by Fujiyoshi et al. (1992).

Figure 10b shows the distribution of vertical velocities. An updraft core (shown by the darkly stippled region) was situated exactly in the high-reflectivity region. In contrast, downdrafts (shown by the hatched regions) were situated on both sides of the updraft (mainly in the low-reflectivity regions). The maximum velocities of the updraft and downdraft were 0.2 and 0.1 m s$^{-1}$, respectively. The updraft and the northern downdraft formed a clockwise (CW) rotation in the vertical plane. However, different from Kristovich’s (1993) results, no counterclockwise (CCW) rotations were found in this figure.

The vertical distribution of the band-normal wind component ($v$) in Fig. 10c agrees well with that of the upper-air sounding (see Figs. 4 and 8). The isotachs of the $v$ wind are almost horizontal above 0.75 km. In contrast, those of the band-parallel wind component ($u$) are almost vertical (Fig. 10d). The magnitude of the $u$ wind was large inside the main echo areas but particularly small to the far north, which indicates a remarkable contrast in $u$ wind between the inside and outside of the snowband. The maximum difference in $u$ velocities between the upper north and the northern outside ($x \approx 35$ km) of the snowband reached approximately 3 m s$^{-1}$. The axis of the strong $u$ wind slanted with height toward the northern edge of the snowband. At the middle to upper altitudes, the strong wind was situated in the low-reflectivity downdraft regions. In contrast, at low altitudes it was situated in the high-reflectivity updraft regions. As seen in Fig. 9, the convergence of the band-parallel wind along the band axis (i.e., $\partial u / \partial x$)
contributed significantly to the formation of the updrafts in the high-reflectivity regions of the snowband. Therefore, the strong band-parallel wind collocating with the downdraft is closely related to the formation of the updraft and precipitation particles in the snowband. In the following section, special emphasis is given to the formation and evolution of the strong band-parallel wind.
b. Time change of the airflow structure

Figure 11 shows the time series horizontal reflectivity pattern of band I at an altitude of 1.25 km. The area shown in Fig. 11 corresponds with that outlined by the dot–dashed line in Fig. 5. At 1201 LST, the snowband was composed of cellular radar echoes (defined by the 22-dBZ contour) aligned with the band axis at a nearly regular interval of about 7 km. Between 1201 and 1219 LST, several cellular radar echoes made a meso-γ-scale (~20 km) system of high reflectivity (~22 dBZ). At the same time, low reflectivity regions between the meso-γ-scale systems decreased in intensity. An isolated meso-γ-scale system (meso 2) formed at 1231 LST. Meso 1, meso 2, and meso 3 moved parallel to the band axis (toward the right in Fig. 11) at speeds of 13.1, 14.2, and 14.2 m s⁻¹, respectively. Interestingly, all meso-γ-scale systems moved faster than the mean speed of cellular radar echoes (12.5 m s⁻¹) (see discussion in section 8). The organization process of meso 2 will be studied here, because it developed within the observation area of the dual-Doppler radar. The features of the other meso-γ-scale systems were essentially the same as for meso 2, as will be described later.

Figure 12 shows the time series band-normal vertical structure of meso 2 along the vertical broken lines in Fig. 11. The place of the vertical cross section was moved in phase with meso 2. The $u$ component of the winds was superimposed in these figures (dashed lines). Figures 12a and 12b show that the general structures are the same as the mean structure shown in Fig. 10. The southern part (left-hand side in each panel) of the system showed a higher echo top, higher reflectivity, and an updraft. In contrast, the northern part of the system showed lower echo top, lower reflectivity, and downdraft. In the upper levels, the updraft transported ice/snow particles toward the north rather than the south and thus caused the asymmetric reflectivity pattern of the snowband (Fig. 10a). However, it should be noted that the airflows changed significantly with time, as described below.

The meso-γ-scale system showed three important features of airflows, that is, 1) appearance of a strong band-parallel wind, 2) increment of the strong band-parallel wind with time, and 3) subsidence of the strong band-parallel wind with time. At 1219 LST, no remarkable differences were found in the $u$ field of the system. At the next time step, however, a strong band-parallel wind began to appear near the upper northern edge of the system. The wind that has a $u$ wind larger than the mean speed of cellular radar echoes (12.5 m s⁻¹) will be called a “rear-to-front current.” The strong band-parallel wind formed a rear-to-front current (stippled region in Fig. 12), which rapidly increased in width, depth, and speed. Its maximum speeds were 15.5 m s⁻¹ in Figs. 12a and 14.8 m s⁻¹ in Fig. 12b. The region of rear-to-front current corresponded well with that of a downdraft. The rear-to-front current descended like a wedge from the upper northern edge to the lower southern area of the system. The maximum downdraft velocity exceeded 1 m s⁻¹. At low levels, the rear-to-front current reached the highest-reflectivity region by 1237 LST. It then spread toward the southern edge of the system. The
motion of the strong band-parallel wind region explains how the slanted pattern was formed in the mean-\textit{u} field of the snowband (Fig. 10d).

6. Radar-echo and airflow structures in vertical cross sections parallel to the elongated axis of band I

Figure 13a shows time series radar reflectivities and storm-relative airflows in a vertical cross section parallel to the band axis indicated by the horizontal broken line, labeled \textit{GH} in Fig. 11. This cross section was chosen to cut the northern edge of the snowband. The term “storm-relative airflow” means airflow relative to the mean motion of cellular radar echoes. In Fig. 13, large arrows toward the left (i.e., the airflow velocities relative to the surface were small) at the lowest level (\(x = 10-25\) km) were erroneous due to the contamination of scattering from the sea surface (sea clutters).

In meso 2, the rear-to-front current (stippled region) collocating with the downdraft began to appear at 1225 LST in the upper rear of the system (\(x = -13\) km). The current increased in width and depth with time. At 1231 LST, another rear-to-front current appeared in the middle of the system (\(x = -2-4\) km). The rear-to-front currents then aggregated with each other and finally...
evolved up to the same horizontal scale as the meso-γ-scale system (i.e., ~20 km). After 1237 LST, the rear-to-front currents rapidly descended from the upper rear to the lower front of the system. The boundary between meso 2 and meso 3 was not clear at 1225 LST (as shown in Fig. 11). However, the echo intensity between them became weak with the development of the rear-to-front currents, and finally meso 2 was apparently separated from meso 3 at 1255 LST. These aspects are also seen in Fig. 12c. There was also a good correspondence between the descent of the radar-echo top and the development of the rear-to-front current \( y = 23–26 \text{ km at 1243 to 1255 LST in Fig. 12c} \).

In the southern part of the snowband (Fig. 13b), the rear-to-front current associated with meso 2 first appeared at low levels \( x = 8–9 \text{ km at 1237 LST} \) and \( x = 0–6 \text{ km at 1243 LST} \). The current corresponded to the leading edge of the wedge-shaped descending current (as shown in Figs. 12a and 12b). When the rear-to-front current arrived at the low levels, gusts of wind approximately 2 km in width and 0.5–0.75 km in depth formed at the low levels (see arrows toward the right in Fig. 13b, \( x = 0–16 \text{ km and 12–15 km at 1243 LST} \)). Strong low-level convergence occurred at the leading edge of the rear-to-front current, for example, \( x \approx 10 \text{ km at 1237 LST} \) and \( x \approx 6 \text{ km at 1243 LST} \). The maximum magnitude of the low-level convergence attained about \( 2 \times 10^{-3} \text{ s}^{-1} \). Updrafts were enhanced above the convergence regions and their maximum velocity exceeded 1 m s\(^{-1}\).
Fig. 14. Time series radar reflectivities and storm-relative airflows in horizontal cross sections: (a) 0.5 km, (b) 1.25 km, and (c) 2.0 km. The region displayed in this figure is the box outlined by the dotted line in Fig. 5. The contour interval is the same as that of Fig. 12. The areas of reflectivity greater than or equal to 28 dBZ are hatched. The regions of the rear-to-front current are stippled.
at 1249 LST). Strong radar echoes (≥28 dBZ) appeared at the middle levels (x = 20 km at 1249 LST and x = 17–19 km at 1255 LST). The maximum echo-top level reached 2.5 km at x = 14 km at 1255 LST, which was the highest in the period of analysis. These changes in radar-echo and airflow structures clearly show the development of new convective cells in the meso-γ-scale system.

The evolution of the rear-to-front current can be summarized as follows. The rear-to-front current began to
appear in the upper-north rear of the system. Increasing in volume and speed, it penetrated toward the lower front and lower south of the system and developed in meso-γ scale. The appearance and development of the rear-to-front currents were also seen in meso 1 (Fig. 13) and meso 3 after 1301 LST (not shown). Therefore, it is concluded that the penetration of the rear-to-front currents characterized the airflow structures of the meso-γ-scale systems and was associated with the development of new convective cells in the meso-γ-scale systems.
7. Radar-echo and airflow structures in horizontal cross sections at different altitudes

Figure 14 shows horizontal cross sections of radar reflectivities and storm-relative airflows at altitudes of 0.5, 1.25, and 2.0 km. The area shown in Fig. 14 corresponds with that outlined by the dotted line in Fig. 5. At 2.0 km (Fig. 14c), the rear-to-front current associated meso 2 appeared in the northern rear area of the system at 1231 LST ($x = -7$ km, $y = 24 - 25$ km). The current was also seen at an altitude of 1.25 km ($x = -13$ to $-3$ km, $y = 22$ to 28 km at 1231 LST in Fig. 14b). At this altitude, another rear-to-front current existed just behind a strong echo core in the middle of the system ($x = 1$ to 3 km, $y = 22$ to 27 km at 1231 LST). Figure 14b clearly shows the aggregation of rear-to-front currents and their development. Similar features of rear-to-front currents were also seen in meso 1.

The rear-to-front current reached an altitude of 0.5 km at 1231 LST ($x = 3$ km, $y = 22$ to 24 km in Fig. 14a). At this altitude, the contrast between the winds inside and outside the main echo areas of the snowband was evident. Outside the snowband, the relative wind was in good agreement with the upper-air sounding data observed at Sapporo (shown in the upper left of the first panel), which indicates the existence of undisturbed environmental winds (as suggested in section 4). The rear-to-front currents encountered the low-level environmental winds coming from the northern outside of the system and formed the strong convergence at their leading edges, for example, $x = 10$ km, $y = 21$ to 25 km, and $x = -1$ to 0 km, $y = 21$ to 23 km at 1237 LST. The encounter between the two airflows caused the enhancement of the updrafts and the development of new convective cells in the system. The relationship between the evolution of meso-$\gamma$-scale systems and the rear-to-front current is discussed in the next section.

8. Discussion

a. Mean structure and roll convection

It is interesting that a roll-like circulation is seen if the mean $v$ components of the horizontal wind speeds at each altitude (shown in Fig. 10a) are subtracted from the flow (Fig. 15). However, an apparent roll-like circulation only exists above an altitude of 1.0 km. Below 1.0 km, downdrafts exist at both sides of an updraft core. This result suggests that the snowband contained a different airflow to that of the ordinary roll circulation.

The horizontal distribution of vertical air velocities at an altitude of 1.25 km (Fig. 16) shows that the updrafts (dotted region) and downdrafts (hatched region) were jumbled along the band axis. In addition, their distribution considerably changed with time. The perturbed $v$ components of horizontal wind speeds—that is, $v_{\text{measured}} - v_{\text{mean}}$—also showed the same features with those of vertical wind speeds (not shown). These facts indicate that there were no rolls in the snowband. Therefore, it is concluded that the meso-$\gamma$-scale system was maintained by a three-dimensional process.

b. Role of the rear-to-front current

In meso 2, there was an alternating pattern of an updraft tilted toward the upstream side and a downdraft situated in its upstream side before the penetration of the rear-to-front current (see 1225 LST in Fig. 13a and 1231 LST in Fig. 13b). A similar pattern was also seen in meso 1 (see from 1201 LST in Figs. 13a and 13b). This configuration of airflows is preferable for making a long-lasting cloud because the downdraft does not prevent the flow of low-level warm and moist air into the updraft. This configuration was also found in snowbands and snow clouds (e.g., Sakakibara et al. 1988; Maki et al. 1992; Yamada et al. 1994). After the penetration of the rear-to-front current, however, the positions of updrafts and downdrafts changed to be opposite in the northern edge of the system (see 1249 and 1255 LST in Fig. 13a); that is, the updraft was tilted toward the downstream side and the downdraft was situated in its downstream side. If the air flowed only in this two-dimensional plane, the clouds in the snowband should have been short-lived after the penetration. However, meso 2 was maintained for at least 48 min until 1313 LST, when it landed over Ishikari Plain. This suggests that the meso-$\gamma$-scale system was maintained by a different process from the essentially two-dimensional process presented by Sakakibara et al. (1988), Maki et al. (1992), and Yamada et al. (1994).

Figure 13a indicates that the rear-to-front current began to appear in the upper-north rear of the system at 1225 LST when meso 2 formed in the snowband. Several rear-to-front currents formed behind the strong radar echoes that composed the system, indicating that the clouds were in mature stages (see $x = -9$ to $-4$ km and $x = -2$ to 4 km at 1231 LST in Fig. 13a and $x = -13$ to $-3$ km and $x = 2$ to 4 km at 1231 LST in Fig. 14b). They aggregated with each other and finally de-
Fig. 16. Time series vertical velocities of winds in a horizontal cross section at an altitude of 1.25 km. The region displayed in this figure is the same as that in Fig. 14b. The areas of updrafts of at least 0.2 m s\(^{-1}\) are stippled. The areas of downdrafts of at least 0.2 m s\(^{-1}\) are hatched. The 10-dBZ contours of reflectivity are also shown.

dveloped in meso-γ scale (∼20 km) at 1243 LST. At the same time, the currents penetrated to the lower south of the system. They transported the large longitudinal momentum from the upper to lower levels and from the outside to the inside of the system. When the rear-to-front currents arrived near the ground surface, they encountered the low-level environmental winds coming from the northern side of the system (see Figs. 13b and
14a) and caused the strong low-level convergence and enhancement of updrafts at the leading edges of the rear-to-front currents (see Fig. 13b). No remarkable changes in reflectivity were found at the middle to upper levels before the arrival of the currents (see Fig. 13b). Therefore, it is concluded that the strong convergence and the enhancement of updrafts were not attributed to buoyancy generated by latent heat release but to penetration by the rear-to-front currents.

After penetration, the echo-top levels exceeded the inversion level and high-reflectivity ($\geq 28$ dBZ) regions appeared at the middle levels (see Fig. 13b), suggesting the support of large snow particles by strong updrafts. As exemplified in Fig. 13b, these convective cells with high reflectivities and strong updrafts formed and developed ($x = 12-15$ km at 1249 LST) in front of strong radar echoes ($x = 8-12$ km at 1249 LST) where the rear-to-front currents arrived. The rear-to-front currents, which began to appear behind strong radar echoes at the upper levels, arrived in front of strong radar echoes at the low levels because their speeds were faster than the mean motion of the cellular radar echoes. The successive formation of new convective clouds in front of the older clouds could explain why the meso-$\gamma$-scale systems moved faster than the individual cellular radar echoes. In the newly developed clouds, again the updrafts tilted toward the upstream side in the south of the system and downdrafts tilted toward the downstream side in the north of the system (cf. 1249 and 1255 LST in Figs. 13a and 13b). It is to be noted that both downdraft air and updraft air came from the northern outside of the system, as seen in Figs. 12 and 14. This three-dimensional airflow structure is completely different from the two-dimensional ones presented by former researchers.

c. Formation process of the rear-to-front current

Strong band-parallel winds, that is, longitudinal winds, have been commonly observed in association with cloud streets and longitudinal-mode snowbands. Brümmer (1985) reported that the downdraft was mainly connected with strong longitudinal wind and the updraft with weak longitudinal wind. Martin and Bakan (1991) also reported that there was a phase difference of $\pi$ between longitudinal wind and vertical wind. Conversely, Kristovich (1993) reported that the longitudinal wind in the updraft region was larger than that in the downdraft region.

The horizontal wind speed of updraft would differ from that of downdraft under conditions of strong vertical wind shear. In fact, there was a strong vertical wind shear in the mixed layer (as shown in Figs. 4 and 8). As mentioned in section 6, the rear-to-front currents that began to appear at the upper levels collocated with the downdraft. In addition, there was a good correspondence between the enhancement of the rear-to-front currents and the downdrafts (see Figs. 12 and 13). These facts suggest that the downward transport of longitudinal momentum would have caused the strong band-parallel winds in the snowband.

Next, the cause of the downdraft should be discussed. In this case, the strong band-parallel winds suddenly appeared to be associated with downdrafts. Considering the stabilization of the surrounding air caused by the entrainment of the overlying dry and warm air, Sykes et al. (1988) succeeded in simulating cloud streets with aspect ratios exceeding 10. In contrast, Chlond (1992) and Zhang and Atkinson (1995) showed the destabilization of cloud-top air due to the evaporation of cloud droplets. Chou and Ferguson (1991) reported that the downward motions transferred not only dry and warm air in the dry regions but also the wettest and coldest air in the regions between the roll-scale wet updrafts and the dry downdrafts. They attributed the formation of the wet and cold downdrafts to the evaporation of cloud droplets diffused from the updrafts. These results apparently indicate that the destabilization of cloud-top air would occur if a large amount of cloud droplets and/or ice/snow particles evaporated in the entrained dry air.

Figures 12a and 12b show that the band-normal upper-level outflow increased with time. Although radar echoes larger than 10 dBZ did not spread toward the north at the upper levels, weak radar signals were able to be detected there, indicating the existence of small particles. Figure 12c also clearly shows that the dissipation of radar echoes occurred in association with the penetration of the rear-to-front current. These facts strongly suggest that the evaporation of ice/snow particles occurred at the upper levels. Since the updrafts transported ice/snow particles toward the rear and north at the upper levels (Fig. 14c), ice/snow particles would have evaporated outside but near the clouds. The downdraft caused by evaporative cooling would have formed in the rear north of the updraft. This process could explain how the rear-to-front current formed in the rear north and descended to the front south of the meso-$\gamma$-scale systems.

d. Conceptual model

Figure 17 illustrates a conceptual model of the evolution and maintenance of the meso-$\gamma$-scale systems composing the snowband. For simplicity, a meso-$\gamma$-scale system is assumed to be composed of two convective clouds. In the figure, clouds move from right to left, in the opposite direction to the previous figures, in order to show clearly the phenomenon on the northern side of the system.

Updrafts (white arrows) form convective clouds (indicated by solid lines) in a meso-$\gamma$-scale system (stage 1). The updrafts penetrate into the overlying stable layer and transport ice/snow particles toward the rear and north of the system. Ice/snow particles evaporate outside, but near the clouds (stage 2). The evaporation of ice/snow particles cools the air and causes downdrafts at the rear north side of individual clouds. The down-
9. Summary and conclusions

A longitudinal-mode snowband (band 1) was observed over Ishikari Bay, Hokkaido, Japan, by using the dual-Doppler radar system of the IHAS, Nagoya University, Japan. The snowband was composed of meso-γ-scale (~20 km) convective cloud systems. This paper has studied the three-dimensional kinematic structures and organization processes of the meso-γ-scale systems that composed the snowband.

In the early stage, the snowband was composed of cellular radar echoes with horizontal scales of 5–10 km. These cellular radar echoes were aligned with the band axis at an almost regular interval (~7 km). Several cellular radar echoes formed a meso-γ-scale system. Strong band-parallel winds appeared in the upper-north rear of the meso-γ-scale systems. The winds caused rear-to-front currents because their speed was faster than the mean motion of the cellular radar echoes. Each rear-to-front current formed behind each strong radar-echo core that composed the systems.

The currents penetrated toward the lower front and lower south of the systems while they increased in volume and speed and finally developed into meso-γ scale. The penetration of the rear-to-front currents characterized the airflow structures of the meso-γ-scale systems. When the currents arrived near the ground surface, they encountered warm and moist environmental air coming from the northern outside of the systems and caused strong convergence and the enhancement of updrafts at their leading edges. Consequently, new convective cells formed and developed in the systems. Therefore, the downward transport process of the large longitudinal momentum played an important role in the organization and maintenance of the meso-γ-scale systems.

The updrafts in the meso-γ-scale systems penetrated into the overlying stable layer. Ice/snow particles were transported by the updrafts toward the rear and north of the systems and evaporated outside the clouds. The evaporation of ice/snow particles cools the air and would have caused a downdraft in the upper-north rear of the systems. The downdraft had large longitudinal momentum, since the wind speed in the overlying air was much faster than that inside clouds. Therefore, downdraft caused by evaporative cooling would have played an important role in the transport of longitudinal momentum from the upper to the lower levels and from the outside to the inside of the snowband.

It is suggested that the differences in vertical wind shear and humidity between the inside and outside of the snowband are important in determining the strength and size of the rear-to-front current. The dynamic struc-
tures of the rear-to-front current, the snowband, and the environmental air cannot be discussed here due to the lack of temperature and humidity data. Some other dynamic mechanisms, including the areas outside the detectable area of the Doppler radars, might explain the formation of the rear-to-front current. In addition, it is uncertain whether or not the airflow structures of the meso-γ-scale systems analyzed here were affected by airflows around Ishikari Bay. In addition to further multi-Doppler radar observations at various places and under various atmospheric conditions, cooperative studies with research aircraft, the application of methods for retrieval of thermodynamic and microphysical variables, and model calculation are needed to discuss the generality of the process presented in this paper.

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