Inner Structures of Snow Clouds over the Sea of Japan Observed by Instrumented Aircraft: A Review

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Abstract

Heavy snowfall, caused by snow clouds over the Sea of Japan, can severely affect social and economic activities in Japan. Therefore, snow clouds, which form and develop mainly over the ocean and bring heavy snowfall to populated coastal plains, have been extensively studied from the perspective of disaster prediction and prevention. Most studies analyzed data acquired by aerological, meteorological satellite, and radar observations, or conducted numerical simulations. Because of the difficulties involved in accessing cloud systems over the ocean, however, few in situ observation data have been available, and up until the middle 1990s, many problems remained unsolved or their analysis and simulation results had not been validated. Here we review knowledge gained from instrumented aircraft observations, made from the middle 1990s through the early 2000s, particularly the development of a convectively mixed boundary layer and the inner structures of longitudinal-mode cloud bands, Japan-Sea polar-air mass convergence zone cloud bands, and a polar low. Unsolved problems relating to the inner structures and precipitation mechanisms of snow clouds and the expected contributions of aircraft observations to further progress in these areas of atmospheric science are also briefly discussed.

Keywords convectively mixed boundary layer; longitudinal mode cloud band; JPCZ cloud band; polar low

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1. Introduction

Snow clouds are often observed over the Sea of Japan in winter, when East Asia is dominated by the winter monsoon. These snow clouds can cover a vast area, on the order of $10^6$ km$^2$, including the entire Sea of Japan, parts of the East China Sea, and sometimes even parts of the Pacific Ocean. Therefore, it is very important to investigate the radiation effects of these clouds from a climate change perspective. In addition, these snow clouds, which form with heat and moisture supplied from the Sea of Japan, bring several hundred millimeters of precipitation to western coastal areas of the Japanese islands during winter. Therefore, studying them is also important for understanding the regional water cycle (water budget).

In winter, especially during strong cold air mass outbreaks, heavy snow often falls on the windward (the Sea of Japan) side of the Japanese islands. While a cold air mass is crossing the Sea of Japan, a convective boundary layer capped by a temperature inversion develops because of the strong heat and moisture supply from the relatively warm sea surface. Stratocumuli start to form and deepen as the distance traveled by the air mass from the eastern coast of the Eurasian continent lengths. Although cloud top heights are usually not very high (typically about 2.0 km near the
western coasts of Hokkaido and northern Tohoku and about 3.0 km near the Hokuriku coast; see the supplementary figure, Fig. S1 for a regional map), the clouds bring snowfall to offshore and onshore areas along the western coasts of Japan. Air masses together with the remaining clouds are forced upward by the mountain ranges that form the backbone of the Japanese islands, producing additional heavy snowfall in the mountains. When coupled with an upper cold trough and/or low-level convergence due to topography, snow clouds sometimes develop with cloud top heights of 5 km or higher. Asai (1988) has proposed the formation mechanisms of snow clouds causing heavy snowfall, combining air mass transformation, upper cold trough, and convergence due to topography and baroclinicity.

Many types of clouds form over the Sea of Japan during cold air mass outbreaks: isolated snow clouds, longitudinal-mode (L-mode; wind-parallel) and transverse-mode (T-mode; wind-normal) snow bands, snow bands that form or intensify under the influence of coastal topography and a land breeze, convergent cloud bands associated with low-level convergence of two differently modified airflows, and spiral- or comma-shaped cloud bands associated with vortical disturbances of various scales.

Snowfall frequently occurs during cold air mass outbreaks and characterizes winter weather on the windward side of the Japanese islands. These cloud systems often bring heavy snowfall to the populated western coastal plains of Japan, where they can severely affect social and economic activities. Therefore, their formation and intensification over the sea have been extensively studied from a disaster prediction and prevention perspective. In contrast, fewer studies have investigated orographic snow clouds in Japan, although snowfall in mountain areas is important from the viewpoint of water resources (Nakai and Endoh 1995; Saito et al. 1996; Kodama et al. 1999; Harimaya and Nakai 1999; Murakami et al. 2001; Kusunoki et al. 2004, 2005). This review focuses mainly on research results for snow clouds over the ocean and coastal areas; orographic snow clouds will be discussed on another occasion.

Many researchers (Matsumoto et al. 1965; Fukuda 1965, 1966; Fujita 1966; Ishihara 1968; Kurashima 1968; Miyazawa 1968; Akiyama 1981a, b) have investigated synoptic-scale conditions leading to heavy snowfall from the viewpoint of weather forecasting. In addition, air mass transformation caused by the supply of great amounts of heat and moisture from the warm surface of the Sea of Japan, a fundamental process leading to snowfall during cold air mass outbreaks, has been studied since the 1950s, mainly by analyzing heat and moisture budgets and water substance budgets based on aerological data (Manabe 1957, 1958; Ninomiya 1964, 1968) and numerical model simulations (Asai 1965; Asai and Nakamura 1978; Nakamura and Asai 1985; Yamagishi 1980; Murakami et al. 1994a; Yoshizaki et al. 2004).

During the Heavy Snow Storm Project, conducted during 1963–1967 in the Hokuriku area (Japan Meteorological Agency 1968), intensive maritime cloud observations using aerial photographs, along with radar, aerological (rawinsonde and dropsonde), and surface observations, were carried out to examine air mass modification and cumulus convection over the Sea of Japan. The results showed that the causal mechanisms of heavy snowfall in the region were mesoscale disturbances such as low-level convergence and middle-level cold trough or cold vortex formation over the sea (Matsumoto et al. 1965, 1967; Matsumoto and Ninomiya 1969).

In the 1960s, overviews of cloud systems in satellite images revealed that snow clouds organized at mesoscale over the Sea of Japan during cold air mass outbreaks were typically characterized by L-mode and T-mode snow bands and convergent cloud bands. Tsuchiya (1969) showed that cumulus activity during cold air mass outbreaks depended on the length of the fetch over the warm sea (i.e., the distance from the Asian continent) as well as atmospheric conditions. Okabayashi (1969, 1972) and Okabayashi and Satomi (1971) reported that two convergent cloud bands are associated with periods of heavy snowfall, one extending from the base of the Korean Peninsula to Hokuriku, and the other extending from the Mamiya Strait to the western coast of Hokkaido.

Okabayashi (1969) showed that these convergent cloud bands form in a large-scale convergence zone between two polar air masses which undergo different transformations as they travel over the Sea of Japan. These cloud bands cause heavy snowfall when they reach the western coast of the Japanese islands. Later, Asai (1988) named this convergence zone the Japan-Sea polar-air mass convergence zone (JPCZ), by analogy with the intertropical convergence zone. In this review, such convergent cloud bands are referred to as JPCZ cloud bands.

Since the 1970s, various types of clouds over the Sea of Japan have been extensively studied by many researchers. L-mode and T-mode cloud bands, which are most frequently seen during cold air mass outbreaks, are self-organized into bands in the absence of significant mesoscale convergence. However,
other cloud bands often form under the influence of mesoscale convergence on the order of $10^{-4}$ s$^{-1}$ before or after cold air mass outbreaks (Yamada et al. 1996; Murakami 2005b). Typically, L-mode and T-mode cloud bands form in a fairly shallow, convectively mixed layer during cold air mass outbreaks. Owing to their prevalence and persistence, these types of cloud bands are important to both the energy and water budgets, and their occurrence frequencies and contributions to total wintertime precipitation have been investigated (Sakakibara et al. 1988a; Mizuno 2005). Mizuno (2005) reported that over the sea off western Tohoku, which is not significantly influenced by JPCZ cloud bands, convective clouds (i.e., isolated snow clouds and L-mode and T-mode snow clouds) account for 60% of total cloud occurrence and 40% of total precipitation.

Asai (1970a, b, 1972), Yagi (1985), and Yagi et al. (1986) investigated the formation mechanisms of L-mode and T-mode cloud bands. From a linear theory for dry convection, Asai (1972) pointed out that convective rolls tend to develop preferentially with their axes oriented parallel to the vertical shear vector of the horizontal wind within a cloud layer. Yagi et al. (1986) analyzed satellite images, along with radar and aerological data, and reported that the relationship pointed out by Asai (1972) held for most L-mode and T-mode snow bands over the Sea of Japan. Miura (1986) investigated aspect ratios of L-mode snow bands in satellite images and reported that the roll convection was rather flat; that is, the aspect ratios of these snow bands were much larger than those predicted by the dry convection theory.

Cloud bands form and intensify offshore and/or onshore along the western coasts of the Japanese islands under the influence of coastal topography and land breezes. These cloud bands, which bring heavy snowfall to the coastal and plain areas, have been investigated primarily by using Doppler radar (Sakakibara et al. 1988b; Ishihara et al. 1989; Fujiyoshi et al. 1992; Yoshimoto et al. 2000; Ohigashi and Tsuboki 2005; Ohigashi et al. 2014).

The structures and formation mechanisms of JPCZ cloud bands have also been studied by analyzing stereoscopic aerial photographs (Hozumi and Magono 1984) and aerological soundings made by an observation vessel traversing the cloud bands (Arakawa et al. 1988), and by conducting numerical simulations (Nagata et al. 1986; Nagata 1987) in addition to satellite image analyses (e.g., Okabayashi 1969; Uchida 1979, 1982; Yagi et al. 1986). The JPCZ cloud bands consist of a line of convective clouds developing along their southwestern edge (referred to as “Cu-Cb line” in Yagi et al. 1986) and cloud bands normal to the wind direction of the winter monsoon (referred to as “transversal cloud bands” in Yagi 1985). Numerical simulations with hydrostatic models have shown that these snow bands form in the JPCZ between two air masses having different characteristics. The formation mechanism of the horizontal convergence of the two air masses has been attributed to a combination of three low-level boundary forcings: the land–sea thermal contrast between the Korean Peninsula and the Sea of Japan; the blocking effect of the mountains north of the Korean Peninsula; and the characteristic distribution of sea surface temperature in the Sea of Japan (Nagata et al. 1986; Nagata 1987, 1991). Endoh et al. (1984) showed by a statistical analysis that JPCZ cloud bands can appear over a wide area that covers almost all of the Sea of Japan south of 40°N latitude. JPCZ cloud bands usually extend southeastward from the base of the Korean Peninsula; however, its downstream part sometimes shifts from the western San-in District to the southern Tohoku District and changes its orientation according to change in the large-scale field (Endoh et al. 1984; Uemura 1980). Nagata (1992) and Ohigashi and Tsuboki (2007) conducted numerical simulations with a hydrostatic model and a cloud-resolving non-hydrostatic model (NHM), respectively, to investigate the movement and development of JPCZ cloud bands during the passage of a short-wave trough accompanied by a mid-level cold core.

Vortical disturbances appear at various scales during cold air mass outbreaks over the Sea of Japan and the northwestern Pacific. Asai and Miura (1981) studied the distributions of cloud vortices with diameters of about 200, 100, and 50 km over the Sea of Japan during winter 1983/1984 in geostationary meteorological satellite (GMS) images, and showed that vortices frequently occurred east of the Korean Peninsula and west of Hokkaido (northern Japan), coinciding with the JPCZ. Meso-β- and meso-γ-scale disturbances that form in the JPCZ have been studied mainly in satellite imagery and numerical simulations (Okabayashi and Satomi 1971; Okabayashi 1972; Nagata 1993; Lee and Park 1998; Tsuboki and Asai 2004). The lifetime of such vortical disturbances is typically several hours, but heavy, concentrated snowfall sometimes occurs along their tracks. The JMA operational meteorological radar observation network has observed remarkable meso-β-scale spiral- or ring-shaped radar echoes corresponding to vortical disturbances (e.g., Miyazawa 1967; Magono 1971; Shimizu and Uchida 1974; Muramatsu et al. 1975; Nyuda et al. 1976;
Asai and Miura 1981; Ninomiya and Hoshino 1990; Ninomiya 1994). These disturbances usually develop offshore near the Sea of Japan coast and then move across the coastal plains before disappearing in the mountains. Meso-\(\alpha\)-scale vortical disturbances have also been studied by Takeuchi and Uchiyama (1985), Ninomiya (1989, 1991), Ninomiya et al. (1990, 1993), Yanase et al. (2002), Fu et al. (2004), and Yanase and Niino (2004). Ninomiya (1989) examined the distribution of cloud vortices in GMS cloud images and reported that, during winter 1986/1987, many meso-\(\alpha\)-scale cloud vortices were concentrated between 50° and 40°N over the northwestern Pacific and between 45° and 35°N over the Sea of Japan; a few were also observed between 40° and 30°N over the northwestern Pacific. Meso-\(\alpha\)-scale cloud vortices over the Sea of Japan were situated in the NW quadrant of a large-scale parent low over the northwestern Pacific.

Recently, 3-dimensional NHMs have been used to investigate the structures and formation mechanisms of cloud bands under the influence of coastal topography, JPCZ cloud bands, and polar lows formed over the Sea of Japan (Murakami 2005a; Ohigashi and Tsuboki 2007; Ohtake et al. 2009; Eito et al. 2010; Yanase et al. 2002; Yanase and Niino 2004). Although these studies revealed the main features of those cloud systems over the Sea of Japan (by analyzing meteorological satellite data and radar observations in addition to performing numerical simulations), few in situ observation data are available to validate the results. In general, few in situ measurements of the inner structures of snow clouds and air mass transformation processes that produce the atmospheric environment for snow cloud formation over the Sea of Japan, which are not strongly affected by coastal topography, are available, although some observations of the inner structures of snow clouds were obtained by special sondes launched from the coast or a small offshore island (Magono et al. 1983; Murakami et al. 1994b; Takahashi et al. 1999; Watanabe et al. 2014).

In this paper, the inner structures of several types of snow clouds, which were observed mainly by instrumented aircraft in the 1990s and early 2000s, are reviewed and examined (Murakami 2005a; Murakami et al. 2005b, c). The two aircraft used and their instrumentation are described in Section 2; observational results for air mass transformation are presented in Section 3; the inner structures of L-mode cloud bands, one of the typical cloud types associated with cold air mass outbreaks, are described in Section 4; and the inner structures of JPCZ cloud bands and polar lows are described in Sections 5 and 6, respectively. In Section 7, remaining issues regarding the inner structures and precipitation mechanisms of snow clouds are discussed.

### 2. Aircraft observations

The aircraft data presented in Section 4 were collected by an instrumented aircraft (King Air B200T, University of Wyoming; Fig. S2) during the 1993 field campaign of the “Study of Precipitation Formation in Snow Clouds and Feasibility of Snow Cloud Modification by Seeding” (Murakami et al. 1994b, 2003). The instrumented aircraft was deployed in lieu of obtaining observations by hydrometeor videonsondes (HYVIS; Murakami and Matsuo 1990) launched from Tobishima Island and hydrometeor video dropsondes (HYDROS; Murakami et al. 1994b) released from an aircraft in the vicinity of Tobishima Island, as was done during the first four seasons of the five-year project. The instrumented aircraft provided data on microphysical, thermodynamic, and kinematic structures in and around snow clouds. The cloud microphysics instruments, which were installed on the wing-tip pylons of the aircraft, included a Particle Measuring Systems (PMS) forward scattering spectrometer probe (FSSP), PMS 1-dimensional and 2-dimensional cloud particle optical array probes (1D-C, size range 12.5–185.5 μm; 2D-C, size range 25–800 μm), a PMS 2-dimensional precipitation particle optical array probe (2D-P, size range 200–6400 μm), a Johnson–Williams liquid water content (LWC) probe, and a CSIRO King LWC probe. A dew point hygrometer (Cambridge 137C3) was used to measure the water vapor content.

The three components of wind vectors relative to the ground \((u, v, w)\) were measured at high temporal resolution by using a Rosemount gust probe (858A3J) with pressure transducers and an inertial navigation system (Honeywell Laserf SM). In addition to the aircraft observations, dual Doppler radar observations were obtained by X-band Doppler radars operated by the Meteorological Research Institute (MRI) and the National Research Institute for Earth Science and Disaster Prevention (NIED). Both radar systems had a wavelength of 3 cm and a detection range of 64 km and were deployed in the coastal area with a base line length of \(\sim 30 \text{ km}\). The aircraft data were collected within the range of the dual Doppler radar observations.

The aircraft data presented in Sections 3, 5, and 6 were collected by an instrumented Gulfstream II (G-II) aircraft (Diamond Air Service Inc., Fig. 1) during field campaigns of the “Winter MCSs Observations over the Japan Sea” project (hereafter, WMO; Yoshizaki...
et al. 2001, 2003, 2004, Kobayashi et al. 2003). This aircraft was equipped with various cloud microphysics and ordinary meteorological instruments, which were installed on the wing-tip pylons. The cloud microphysics instruments included a FSSP, a 2D-C, a 2D-P, a Droplet Measurement Technologies (DMT) cloud and aerosol particle spectrometer, two King LWC probes (KLWC-5), a Nevzorov total water content/LWC probe, and a Gerber particle volume monitor (PVM-100) probe. For measuring the water vapor content, two dew point hygrometers (EG&G 137) and an AIR Lyman-Alpha hygrometer were used. The three components of wind relative to the ground \((u, v, w)\) were measured, with high temporal resolution, through a five-hole radome with pressure transducers used in combination with an inertial navigation system. The aircraft was also equipped with a Communications Research Laboratory W-band cloud radar (SPIDER; Horie et al. 2000), a Radiometrics dual wavelength MWR-1100 microwave radiometer, and a Vaisala GPS dropsonde system. The cloud radar and microwave radiometer were installed in pods on the left- and right-hand sides of the fuselage, respectively. The W-band cloud radar provided reflectivity, Doppler velocity, and polarization parameters, and the microwave radiometer provided the vertically integrated liquid water amount (i.e., the liquid water path: LWP).

A Citation V aircraft provided GPS dropsonde data in the vicinity of the snow cloud systems observed by the G-II. In addition, ground-based observations were obtained during these campaigns by using four portable X-band Doppler radars, two wind profilers, and three additional aerological stations. Three JMA observation ships also carried out upper air soundings every three or six hours. In this review, however, mainly data collected by the B200T and G-II instrumented aircraft are used to document the microscale and mesoscale structures of snow clouds.

3. Boundary layer development

Air mass transformation by the large amounts of heat and moisture supplied from the warm surface of
the Sea of Japan is a fundamental process leading to snowfall during cold air mass outbreaks. However, most studies of this phenomenon have analyzed heat and moisture budgets and water substance budgets by using aerological data (Manabe 1957, 1958; Ninomiya 1964, 1968) and regional model simulations (Yamagishi 1980; Murakami et al. 1994a; Yoshizaki et al. 2004) (see Section 1). There were no studies using in situ observations of heat and moisture fluxes [except Nakamura and Asai (1995), who measured heat and moisture fluxes over the East China Sea], convectively mixed boundary layer development, and stratocumulus (shallow convective cloud) formation before the 2002 WMO campaign, during which the observations described in this section were carried out.

In this section, the development of a convectively mixed boundary layer, as a result of heat and moisture fluxes from the warm sea surface and cloud formation over the Sea of Japan on 12 February 2002, are documented, mostly based on in situ measurements, cloud radar observations, and GPS dropsonde soundings made by the G-II instrumented aircraft.

3.1 Synoptic features and observation flight

At the time of the aircraft observations, from 12 to 16 JST on 12 February 2002, the winter monsoon pressure pattern (i.e., a high pressure system to the west and a low pressure system to the east) had already begun to weaken (Fig. 2, upper panel), and shallow convective clouds, which are typical during late stages of a cold air outbreak, prevailed.

The flight path of the aircraft was west–northwest at 6.5 km altitude and covered a distance of more than 300 km. During this level flight, W-band cloud radar measurements were made in nadir-looking mode, and GPS dropsondes were released at four points along the flight path (Fig. 2, lower panel). On the return (east–southeast) flight, in situ measurements were made in three vertical cross sections (stacks), which were oriented nearly cross-wind; in each stack, measurements were made while flying at three or four levels. Between each adjacent pair of vertical stacks, the aircraft flew at 300 m above sea level, and cloud radar measurements were made in horizontal-looking mode. In general, the cloud tops along the flight path were bumpy, but sometimes very thin cloud layers were observed just above them (Murakami et al. 2005a, 2006).

3.2 Evolution of a convectively mixed boundary layer

The echo top height of the stratocumuli increased from 1.5 to 1.9 km, and the cloud top temperature decreased from −12.5 to −16°C, over a travel distance of 230 km from west–northwest to east–southeast (from X = −80 km to X = 120 km) (Fig. 3). The dropsonde sounding at X = 230 km indicated a rapid increase in the convectively mixed boundary layer height due to the upwind influence of coastal topography (Fig. 4).
Fig. 3. Vertical cross section of cloud radar reflectivity and horizontal distributions of the LWP, measured with a microwave radiometer, and air and dew point temperatures at altitudes of 6500 m and 300 m along the flight track. The X-distance indicates positions along the flight track projected onto a latitude line; X = 0 corresponds to 136.55°E (from Murakami et al. 2005a).

Fig. 4. Vertical profiles of potential temperature (Θ) and the water vapor mixing ratio measured with GPS dropsondes along the flight track (see Fig. 2, lower panel). The most upwind sounding is indicated by the blue line, and the most downwind sounding is indicated by the red line (from Murakami et al. 2005a).
Radar echoes showed not only an increase in reflectivity but also many more convective features with travel distance. The LWP gradually increased from 0.2 to 0.4 mm as the travel distance increased, although it was much less than the adiabatic condensation amount, except in a few places where values of more than 1 mm (equivalent to 1000 g m\(^{-2}\)) were observed. The air temperature at 6500 m was about \(-40^\circ\text{C}\), and the air was very dry. At 300 m, however, the air temperature increased from \(-3^\circ\text{C}\) to \(-0.5^\circ\text{C}\) over a travel distance of 230 km. Air and dew point temperatures at 300 m were much more turbulent and fluctuated more with time than those at 6500 m (Fig. 3).

Vertical profiles of potential temperature and the water vapor mixing ratio indicate that the thickness of the convectively mixed boundary layer increased from 1.5 km (847 hPa) to 1.9 km (803 hPa) as the air mass traveled over the distance of 230 km. The whole convective boundary layer became warmer by 3 K and more moist by 0.3 g kg\(^{-1}\) from the westernmost (X = \(-80 \text{ km}\)) to the second easternmost (X = 120 km) dropsonde release points (Fig. 4).

### 3.3 Heat and moisture fluxes

The sensible heat flux was positive (upward) below the cloud base and was almost zero within the cloud layer. It is well known that the sensible heat flux increases linearly with decreasing height below the cloud base (Stull 1988). On the basis of this relationship, the surface sensible heat flux was estimated by extrapolating the observed vertical change of the sensible heat flux below the cloud base. The estimated sensible heat fluxes ranged from 100 to 130 W m\(^{-2}\) (Fig. 5a).

Sensible heat fluxes near the cloud tops were negative in STACK 1 and positive in STACKs 2 and 3. In general, in the case of a convectively mixed, cloud-free boundary layer capped with a temperature inversion layer, the sensible heat flux just above the boundary layer top is negative, a reflection of the vertical temperature profile. However, in the case of a cloud-capped boundary layer, shallow penetration by an aircraft is expected to result in the upward transport of colder air caused by overshooting of the cloud top, together with the downward transport of warmer air due to the compensating downdraft; thus, the sensible heat flux is negative. The expected positive sensible heat flux near or just below the boundary layer top reflects a combination of the upward transport of warmer air in the cloud parcel and the downward transport of colder air chilled by evaporative cooling as a result of mixing with entrained, dryer air from the cloud top. Therefore, a change in the polarity of the sensible heat flux might be observed as a result of just a slight change in the flight level near the cloud top. Moreover, because the aircraft made a limited number of cloud top penetrations, the statistical sample was small, and the flux calculation may include large uncertainty.

The largest contributor to the net moisture flux was the water vapor flux. Vapor fluxes below the cloud base were almost constant and corresponded to latent heat fluxes of 150–250 W m\(^{-2}\). In the cloud layer, a part (20–30 %) of the upward vapor flux was converted to an upward flux of liquid water. In the upper and middle parts of the cloud layer, about one-half of the upward vapor flux was converted to a downward flux of ice water. However, most of the ice water evaporated below the cloud base; only 10–20 % of the upward vapor flux returned to the sea surface in the form of precipitation. Most water vapor and latent heat fluxes were consumed by moistening and heating of the convectively mixed boundary layer and increasing the ice water content in the cloud layer (Figs. 5b–f).

Water vapor fluxes near the cloud tops were positive in STACK 1 and negative in STACKs 2 and 3. The upward water vapor flux in STACK 1 is easily explained as a physical phenomenon. However, the downward fluxes in STACKs 2 and 3 have yet to be explained, although several studies have reported downward water vapor fluxes near cloud tops (Nicholls 1984; Sykes et al. 1990; Kristovich and Braham 1998). Temporal changes in the boundary layer structure during the observation flight, the influence of meso-scale convection, and small sample sizes because of the short flight legs mean that the flux measurements are susceptible to both random and systematic errors.

Momentum fluxes of the \(u\) wind component were negative at all levels except near the cloud top, where they were close to zero. Momentum fluxes of the \(v\) component were close to zero at the cloud top and in the lowest level, and they were negative in middle levels (within the cloud layer) (Fig. S3). Because both the \(u\) and \(v\) components of the horizontal wind have a tendency to increase with altitude, these observational results indicate that the kinetic energy of the mean wind was converted to one of the disturbances (convection). This conversion is consistent with the linear theory of roll convection with horizontal axes parallel to the vertical shear of horizontal wind (Asai 1970a; Yamada et al. 2010; Eito et al. 2010).

There have been no aircraft observations providing data on boundary layer development and cloud formation across the entire Sea of Japan. However,
composites of aircraft observations made upwind and downwind of the Sea of Japan under similar synoptic weather conditions can provide valuable insights. Conditions over the Sea of Japan on 2 February 2001 (Inoue et al. 2005) were very similar to those on 12 February 2002; in both cases, the winter monsoon pressure pattern was weak and satellite images indicated similar cloud features, although the 2 February 2001 case provides us with kinematic and thermodynamic structures, but not cloud microphysical structures. The results reported by Inoue et al. (2005), together with the instrumented aircraft observations reviewed here, allow air mass transformation processes in the boundary layer to be examined over the

Fig. 5. Vertical profiles of temperature flux (a), cloud water (solid lines) and snow water (dashed lines) contents (b), water vapor flux (c), liquid water flux (d), ice water flux (e) and net flux (f) averaged over each level flight leg in STACK 1, STACK 2, and STACK 3 as a function of height normalized by the height of the boundary layer top. Temperature and water fluxes were calculated by eddy correlation methods using 1 Hz data. Sensible and latent heat fluxes are defined by $\rho_a C_p T'w'$ and $\rho_a L v Q'w'$, where $\rho_a$, $C_p$ and $L_v$ are air density, specific heat capacity of air at constant pressure and latent heat of evaporation of water, respectively (from Murakami et al. 2005a).
entire distance traveled by air masses from the eastern coast of the Asian continent to the western coasts of the Japanese islands. The sensible heat flux, which was 200–300 W m\(^{-2}\) upwind near the coast of the continent, decreased to 100–130 W m\(^{-2}\) downwind near the coast of Japan, whereas the latent heat flux increased from 100–200 W m\(^{-2}\) (upwind) to 150–250 W m\(^{-2}\) (downwind). Consequently, Bowen’s ratio (i.e., the ratio of the sensible to the latent heat flux) showed a decreasing trend from 1.7 (upwind coast) to 0.7 (downwind coast). This result is qualitatively consistent with the analytical results of Ninomiya (1968) and the 2-D numerical simulation results of Murakami et al. (1994a).

3.4 Micro- and macrostructures of shallow convective clouds

Along the aircraft flight path, the maximum number concentration of cloud droplets decreased from 1000 to 900 cm\(^{-3}\) over the distance of 230 km from windward to leeward. The maximum cloud water content did not change very much with travel distance. Consistent with the cloud radar observations, the 2D-C cloud particle concentration increased from 10 to 40 particles L\(^{-1}\), and the 2D-P precipitation particle concentration increased from 0.6 to 3 particles L\(^{-1}\). A plausible explanation for the increase in ice and snow particle concentrations is that the decrease of cloud top temperatures promoted ice nucleation. The typical precipitation particle shape changed from aggregates (dendritic) to rimed crystals and thence to heavily rimed crystals over the travel distance.

Solid cloud bases were located at a normalized height of around 0.5 (i.e., the height normalized by the height of the boundary layer top) in each case, although small amounts of cloud water were observed at lower levels in a few places during the level flight. The snow water content tended to increase monotonically with the travel distance. However, the cloud water content did not indicate a corresponding monotonic increase; instead, it decreased from STACK 2 to STACK 3 owing to intensive depletion of cloud water due to the increasing snow particle concentration. The snow water content decreased rapidly below the cloud base in all three vertical stacks. The reason for the rapid decrease in snow water content was intense evaporation (sublimation) in the dry air below the cloud base (Fig. 5b). The relative humidity at 300 m above sea level was around 60 % in all vertical stacks (Fig. 54).

Horizontal cross sections of cloud radar reflectivity at an altitude of 300 m showed a primary band-like structure parallel to the mean wind direction in the cloud layer, but the cloud bands were not well organized. The orientation of a secondary band-like structure deviated from the mean wind direction by 50° to the left (Murakami et al. 2005a).

3.5 Conceptual model

As the air mass traveled over a distance of 230 km during the late stage of the cold air outbreak, the convectively mixed boundary layer deepened from 1.5 to 1.9 km, the entire mixed layer was heated by 3 K and moistened by 0.3 g kg\(^{-1}\) and stratocumulus clouds developed within it.

The sensible heat flux from the sea surface was 100–130 W m\(^{-2}\), and the latent heat flux was 150–250 W m\(^{-2}\). The ice water flux (precipitation flux) near the sea surface was 10–20 % of the water vapor flux from the sea surface, although the maximum ice water flux in the cloud layer was about one-half the surface water vapor flux. This finding means that most of the heat and water vapor fluxes were used to develop the convectively mixed boundary layer and increase in the cloud particle concentration of the stratocumulus clouds.

The stratocumulus clouds were loosely organized into cloud bands (also called cloud streets). The orientation of the primary band-like structure was almost parallel to the mean wind in the cloud layer, whereas the orientation of the secondary band-like structure deviated from the mean wind direction by 50° to the left. The LWP gradually increased downwind, although most LWP values were much lower than the adiabatic condensation amount. Ice and snow particle concentrations increased downwind as a result of a decrease in cloud top temperature by 3.5 K. These ice and snow particles depleted the cloud water and mostly canceled the LWP increase by water condensation in the convectively mixed boundary layer. The precipitation particle shape changed downwind from aggregate (dendritic) to rimed crystals and thence to heavily rimed crystals.

4. L-mode cloud bands (shallow convective snow bands)

Asai (1972) pointed out that, theoretically, roll convection develops preferentially with orientations parallel to the vertical shear vector of horizontal winds within the cloud layer. This explains the mechanisms of formation and maintenance of band-shaped snow clouds (cloud streets). This theoretical prediction was supported by Yagi et al.’s (1986) analyses of satellite images, radar data, and aerological data of L-mode
and T-mode snow bands over the Sea of Japan. However, Yamada (2005) reported that the orientation of snow bands does not necessarily coincide with the vertical shear vector of horizontal wind, and Miura (1986) showed that the aspect ratios of roll convection are several times larger than those expected from dry convection theory.

Linear theories of dry convection, which have been proposed to explain the formation of horizontal roll convection (Asai 1970a; Brown 1970; Kuettner 1971), are supported by observations of roll-like convection without cloud formation (Kropfli and Kohn 1978; Reinking et al. 1981; Rabin et al. 1982) and cloud streets without precipitation (LeMone 1973; Christian and Wakimoto 1989) in the boundary layer. However, in the case of clouds accompanied by precipitation, the results of numerical simulations suggest that the airflow structures of horizontal roll-like convection predicted by linear theories may be altered by the latent heat of cloud condensation, evaporation of precipitation particles in the sub-cloud layer, and hydrometeor loading (Chlond 1992; Rao and Agee 1996). These studies suggest that the airflow structures of precipitating cloud bands differ from those of cloud-free horizontal roll vortices and non-precipitating shallow cloud streets.

To understand the formation mechanism of such L-mode and T-mode snow bands and subsequent precipitation processes, comprehensive studies of microphysical, thermodynamic, and kinematic structures of snow bands (isolated, L-mode, and T-mode snow clouds) are needed; previous studies that addressed these issues were not comprehensive (Murakami et al. 1994a; Yamada et al. 1994; Murakami et al. 1996; Yoshimoto et al. 2000; Fujiyoshi et al. 1998).

In this section, microphysical, thermodynamic, and kinematic structures of shallow convective snow bands are described on the basis of instrumented aircraft and dual Doppler radar observations made during the 1993 field campaign of the “Study of Precipitation Formation in Snow Clouds and Feasibility of Snow Cloud Modification by Seeding” (see Section 2).

4.1 Synoptic and mesoscale features

At 0900 JST on 29 January 1993, a well-developed synoptic low with a minimum pressure of 970 hPa was located at about 44°N, 145°E, and pressure contours over the observation area were close together and aligned NW–SE (Fig. 6, upper panel). In the upper levels (850, 700, and 500 hPa; not shown), the temperature gradient in the wind direction was very small, indicating that cold air advection had weakened by this time. Snow clouds had started to form over the Sea of Japan about 300 km offshore of the Asian continent, but they lacked distinct band-like structures even near the Japanese islands (Fig. 6, lower panel).

Thermodynamic structures of atmosphere were characterized by a strong temperature inversion at ~770 hPa, which confined convection below that level. Above that level, the air was very dry because of anticyclonic subsidence. From vertical profiles of temperature and dew point temperature, the cloud top height (temperature) and cloud base height (tem-
temperature) were estimated to be ~ 2.5 km (−15°C) and ~ 1.0 km (−7°C), respectively. The vertical profile of equivalent potential temperature showed a convectively unstable layer at lower levels. The prevailing wind was west–northwest within the entire cloud layer, and wind speed decreased slightly with increasing height. The wind shear vector within the cloud layer was directed NW with small magnitude (Murakami et al. 1996; Murakami 2005c).

4.2 Observation results
During 1100–1300 JST 29 January 1993, three snow bands were examined consecutively by the University of Wyoming’s King Air B200T instrumented aircraft, and the ground-based dual Doppler radars (MRI and NIED X-band Doppler radars). Radar reflectivity plan position indicators (PPIs) for the three cases showed individual convective cells loosely organized in bands (Fig. 7). In cases 1 and 2, the snow band orientations deviated from the mean wind direction by 30–40° to the right, whereas in case 3, the snow bands were oriented parallel to the mean wind direction. Thus, during the aircraft observations, the snow bands in cases 1 and 2 were primary mode snow bands and those in case 3 were secondary mode snow bands.

Because the microphysical and dynamical structures of the snow bands were similar in cases 1 and 2, the following sections document the snow bands in cases 2 and 3.

a. Case 2 (snow band with primary mode of orientation)
The radar echo with the maximum reflectivity of 20 dBZ stood almost erect and its top reached an altitude of 2.7 km. An updraft region located in the center of the precipitation echo had a maximum velocity of 3 m s$^{-1}$. In vertical cross section, the airflow was symmetric, except at lower levels there was inflow from the SW. Horizontal convergence was observed below an altitude of 1.8 km, whereas horizontal divergence was dominant above that level (Fig. 8).

Constant altitude PPIs (CAPPIs) of radar reflectivities and system-relative horizontal winds showed that airflow converged toward the center of the echo cell at lower levels, whereas the airflow radiated outward.

Fig. 7. PPIs of radar reflectivity for snow bands: (a) case 1 at 1148; (b) case 2 at 1217; (c) case 3 at 1246 (from Murakami 2005c).
from the center in all directions at an altitude of 1.8 km and above. Thus, convective cells embedded in the snow bands had axisymmetric airflow structures, except at the lowest level (Fig. S5).

Aircraft observations also showed a symmetrical airflow structure in a vertical cross section perpendicular to the snow band orientation along BB’ in Fig. S5, except in the lower level, where the inflow came from the SW. The maximum updraft velocity was 3–4 m s\(^{-1}\) at an altitude of about 2 km, and it transformed to divergent outflows above this level. In the cross section, weak downdrafts were observed on both sides of the main updraft region, with weak updrafts at the right-most and left-most edges.

The distribution of regions with high cloud water content corresponded to updraft regions (Fig. 9b). In the central cloud water region, the maximum cloud water content was 0.7–0.8 g m\(^{-3}\). High concentrations of ice crystals were observed above and on both sides of the main updraft region, but in the updraft core, ice crystal concentrations were generally very low (Fig. 9d). Note, however, that relatively high concentrations of ice crystals were sometimes found in updraft cores below the cloud base. The spatial distribution of snow particles mostly corresponded to that of the ice crystals (Fig. 9e). In contrast, in the weak updraft regions at both edges of the cross section, concentrations of cloud water, ice crystals, and snow particles were all high (Fig. 9).

Within the updraft core, low number concentrations of large (up to 5–6 mm) snow particles (graupel) were observed (Fig. 10); this result is consistent with the dual Doppler radar observations, which showed that the precipitation core and updraft core were almost collocated (Fig. 8).

b. Case 3 (snow band with secondary mode of orientation)

Case 3 was a snow band oriented parallel to the mean wind direction observed during 1230–1249 JST on 29 January (Fig. 7c). Dual Doppler radar observations and in situ aircraft observations showed a dynamical and microphysical structure similar to the case 2 snow band (not shown). The major difference was that the asymmetric inflow stream at lower levels observed in case 2 was not observed in case 3 (see details in Murakami 2005d).

4.3 Discussion

a. Snow band orientation

The snow bands’ orientations in cases 1 and 2 deviated from the mean wind direction within the cloud layer by 30–40° to the right (primary mode), whereas the snow bands of case 3 were aligned parallel to the mean wind direction (secondary mode). During the period of aircraft observation, the horizontal distribution of temperature below 700 hPa around the study area, which was under the influence of the circulation of a well-developed low pressure system situated over eastern Hokkaido (Fig. 6, upper panel), showed an unusual thermal structure gradually relaxed and completely disappeared as the low moved slowly eastward.

In cases 1 and 2, the vertical wind shear vectors were almost parallel to the band orientation (direction 310°, magnitude ~ 3 m s\(^{-1}\)), consistent with the thermal wind relationship. Asai (1972) examined the relationship between the orientations of convective roll and vertical wind shear vectors in terms of the linear theory for dry convection, and Yagi (1985) reported a good agreement between theory and observation in analyses of L- and T-mode snow bands based on radar echoes, satellite images, and aerological data. However, to the author’s knowledge, a snow band forming in the boundary layer with wind speeds decreasing with height and with an orientation parallel to the wind shear vectors has not been reported previously. The unusual snow band orientation with respect to wind shear vectors of cases 1 and 2 can be attributed to the unusual horizontal distribution of temperature described above.
Fig. 9. Vertical cross sections along BB’ in Fig. S5: (a) $vw$ components of 6-s-averaged system-relative wind; (b) cloud water content measured with King’s hot wire probe; (c) cloud droplet number concentration measured with the FSSP; (d) ice crystal concentration measured with the 2D-C probe; and (e) precipitation particle concentration measured with the 2D-P probe. In panel (a), barbs along the flight track indicate wind speed and direction, where wind direction is from the flight track to the tip of each barb. The scale in the upper right corner indicates a wind speed of 1 m s$^{-1}$ (from Murakami 2005c).
Fig. 10. 2D-P images in the center and on both sides of the updraft core at three different levels. See Fig. 9e for the image locations (from Murakami 2005c).
In contrast, in case 3, the wind shear vectors were weaker than in cases 1 and 2, and they were directed toward ~ 280°. The case 3 snow band was thus a typical L-mode snow band, with an orientation parallel to the mean wind within the cloud layer.

Snow bands with orientations parallel to the vertical wind shear vector but deviating considerably from the mean wind direction, such as those of cases 1 and 2, have also been reported in lake effect snow clouds (Kelly 1982). However, snow band orientation sometimes does not coincide even with the vertical wind shear vector (e.g., Yamada 2005; Kelly 1984), and a formation mechanism for such snow bands has not been clarified yet.

b. Snow band aspect ratios

Based on radar echoes, the aspect ratios of the snow bands in the three cases ranged from 4:1 to 8:1, the spacing between the snow bands was 10–20 km, and echo top heights were 2.5–3.0 km. These results are consistent with satellite image analyses (Miura 1986). Similar aspect ratios have been reported for snow bands over the Bering Sea (Walter 1980) and Lake Michigan (Kelly 1982).

The aircraft observations showed that updraft cores were embedded in the snow bands every 4–6 km perpendicular to the band orientation. Thus, in terms of airflow structures, the aspect ratios of the snow bands were between 1:1.5 and 1:2.

None of the snow bands documented here showed either 2-dimensional roll structures or a simple structure of convective cells aligned at regular intervals. Instead, convective cells were distributed irregularly within each snow band.

c. Recirculation and large graupel formation

Aircraft observations showed that large graupel particles of up to 5 mm were produced in snow clouds with thicknesses of 1–1.5 km and updraft velocities of 3–4 m s\(^{-1}\) (Figs. 9, 10). In updraft cores near cloud base levels, many small graupel particles with fall velocities smaller than the updraft velocities were observed. This finding suggests that recirculation of precipitation particles played an important role in producing large graupel particles (diameter, 5–6 mm) in shallow convective snow clouds. According to this scenario, graupel embryos (heavily rimed snow crystals) that initiate and grow in the main updraft core are transported to surrounding regions by the divergent outflow in the upper parts of the updraft cores; as a result, concentrations of ice and snow particles are high around the updraft cores. Then, some of these graupel embryos re-enter the updraft core near the cloud base, allowing them to grow even larger.

Matsuo et al. (1994) used a one-dimensional cloud model with detailed microphysical parameterization to demonstrate that graupel particles can grow to, at most, 1–2 mm in size during a single ascent and descent in the updraft core. Therefore, the recirculation of graupel particles must play an important role in producing large (5–6 mm) graupel particles in shallow snow clouds.

Harimaya (1976, 1988) investigated graupel particles in thin section and showed that heavily rimed snow crystals and frozen drizzle drops or raindrops can serve as graupel embryos in the snow clouds that form during cold air mass outbreaks over the Sea of Japan. However, aircraft observations did not indicate the presence of drizzle drops (diameter 100 μm or larger) in the snow clouds, although the 2D-C probe, which has a resolution of 25 μm, may have missed small drizzle drops present at low concentrations.

4.4 Conceptual model

Two types of snow bands were observed in shallow convective snow clouds (stratocumulus clouds) formed during a cold air mass outbreak. One type had orientations deviating from the mean wind direction by 30–40° to the right (primary mode), and the other type was parallel to the mean wind direction (secondary mode), although both types of snow bands were almost parallel to the vertical wind shear in the cloud layer. Both types of snow bands were loosely organized: they had neither 2-dimensional roll structures nor a simple convective cell structure arranged at regular intervals along a line; instead, convective cells were distributed irregularly in the snow bands. Radar observations showed that the snow bands had aspect ratios ranging from 4:1 to 8:1, and that they were spaced 10–20 km apart with echo top heights of 2.5–3.0 km. These observations are consistent with satellite image analysis by Miura (1986). However, aircraft observations indicated that updraft cores were embedded in the bands every 4–6 km in a direction perpendicular to band orientation and that the aspect ratios of airflow structures in the snow bands were between 1:1.5 and 1:2.

In both types of snow bands, the structure of the system-relative wind in each convective cloud was nearly axis symmetrical, except for airflow from the southwest at lower levels in the primary mode snow bands. In updraft cores (6-s-averaged updraft velocity of 3–4 m s\(^{-1}\)), the total number concentrations of ice crystals and precipitation particles were often seen
to be low, but many large graupel particles were observed. This result is consistent with dual Doppler radar observations showing that updraft cores mostly corresponded to high-reflectivity regions. The many small graupel particles near the cloud base strongly suggest that recirculation of precipitation particles played an important role in producing large graupel particles (5–6 mm) in these shallow convective snow clouds.

5. JPCZ cloud bands

In 1969, satellite images first showed that broad (JPCZ) cloud bands extending southeastward from the base of the Korean Peninsula sometimes bring heavy snowfall to the San-in and Hokuriku districts (Oka-bayashi 1969), and many studies of JPCZ cloud bands were carried out in the 1980s and 1990s.

Some conceptual models of mesoscale structures and the mechanisms of formation and development of JPCZ cloud bands have been proposed on the basis of observation and simulation results (Hozumi and Magono 1984; Arakawa et al. 1988; Nagata 1987), but a major point of controversy has been the low-level thermal structure. Hozumi and Magono (1984) suggested that WNW airflow on the southwestern side of the JPCZ is colder than the NNW airflow on the northeastern side, but Arakawa et al. (1988) suggested an opposite temperature contrast between the two air masses. Nagata (1987) inferred a warm-core structure with cold air on both sides based on an analysis of simulated fields and observation data. Nagata (1992) showed that colder air is present on the left-hand side of the JPCZ cloud bands, looking downstream, when the orientation of the JPCZ cloud bands is mainly W–E, whereas it is on the right-hand side when their orientation is more N–S.

Another major controversial point has been the formation mechanism of transverse mode clouds, which extend northeastward from the well-developed convective clouds on the southwestern edge of the JPCZ cloud bands. Hozumi and Magono (1984) suggested that the transverse-mode clouds originate from parts of the cloud mass that were detached from the well-developed convective clouds by winds. In contrast, Arakawa et al. (1988) suggested that airflow with warmer temperatures from the southwestern side of JPCZ glides up over the stable layer, suppressing shallow convective clouds on the northeastern side and forming the transverse mode clouds. Nagata (1987) showed by a numerical simulation with a hydrostatic numerical model that a wet region is located in the middle level on the northeastern side of the line of active convection and suggested that the transverse-mode cloud bands are stratiform clouds generated in the wet region and maintained by a gentle upgliding motion. However, the numerical model used by Nagata (1987) did not explicitly simulate hydrometeors.

On the basis of numerical simulation results, Nagata (1987) suggested that JPCZ cloud bands have some additional characteristic structures: suppressed cumulus convection adjacent to the Cu-Cb line on the northeast is a result of vertical circulation, enhanced by a concentrated release of latent heat along the Cu-Cb line, and a zone of weak wind along the Cu-Cb line and in a middle-level zone downstream of the Cu-Cb line.

However, because these snow cloud systems form over the ocean and are difficult to measure directly, it has been difficult to carry out observational studies of mesoscale and microscale structures of JPCZ cloud bands. For this reason, there have been some major controversial points in the conceptual models proposed on the basis of the few available observations, and the status of the additional characteristic structures inferred from the results of numerical simulations has remained unsettled.

To investigate the inner structures of mesoscale cloud systems and to clarify their formation and development mechanisms, WMO field campaigns were carried out during three winters, 2000/2001, 2001/2002, and 2002/2003 (Yoshizaki et al. 2001; Kobayashi et al. 2003; Yoshizaki et al. 2003; Yoshizaki et al. 2004; see Section 2). During intensive observation periods of the three seasons, seven JPCZ cloud bands were observed with instrumented aircraft over the Sea of Japan. A typical, but relatively less developed case of a JPCZ cloud band with a NW–SE orientation was observed on 14 January 2001 and simulated with a cloud-resolving NHM (Eito et al. 2010; Murakami et al. 2005b).

In this section, inner structures of well-developed JPCZ cloud bands with a W–E orientation, which were investigated off the Hokuriku coast on 29 January 2003, are documented mainly on the basis of in situ measurements, cloud radar observations, and dropsonde soundings from instrumented aircraft. The characteristic features of JPCZ cloud bands are also examined in the seven JPCZ cloud bands observed by instrumented aircraft during the three winters.

5.1 Synoptic and mesoscale features

At 0900 JST on 29 January 2003, a well-developed low with a minimum pressure of 976 hPa was centered
in the vicinity of Sakhalin Island, and NW–SE oriented isobars were packed over the Sea of Japan (Fig. 11, upper panel). Cold air was advecting strongly from the northwest at 850 and 700 hPa over the observation area (not shown).

At 500 hPa, a trough (cut-off low) was passing over the Sea of Japan, and cold air of −42°C had moved to the south over the observation area by this time (Fig. 11, middle panel). Wide, arch-shaped snow cloud bands extending E–W, which were observed by instrumented aircraft, had higher cloud tops than the surrounding clouds (the cloud top temperature was therefore lower). Anvil clouds had spread northward from the tops of the developed cloud bands, and T-mode cloud bands were observed beneath them. To the south of these developed cloud bands, L-mode cloud bands were observed, and other L-mode cloud bands were observed far to the north in visible GMS images (Fig. 11, lower panel).

The wide, arch-shaped cloud bands moved slowly southward from the morning of 29 January; this movement corresponded to the movement of the leading edge of cold air (−42°C near 500 hPa). During the aircraft observation period (1200–1500 JST), the movement stalled near the Noto Peninsula, and the cloud band structure gradually became obscure.

5.2 Observational results

The G-II instrumented aircraft observed JPCZ cloud bands in a vertical cross section along longitude 135.5°E between 36.5°N and 38.5°N latitude. Cloud radar observations were made in downward-looking mode, and three GPS dropsondes were released during the northbound level flight at 7.5 km altitude (~ −40°C). Subsequently, level flights were made at five different altitudes, 4.9 km (~ −40°C), 3.0 km (~ −27°C), 2.1 km (~ −18°C), 1.2 km (~ −13°C), and 0.5 km (~ −6°C), and in situ measurements of kinematic, thermodynamic, and microphysical structures in and

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Fig. 11. Surface weather map at 0900 JST on 29 January 2003, 3 hours before the aircraft observations (upper panel), 500 hPa weather map at 0900 JST on 29 January 2003 (middle panel) and GMS-5 visible image at 1300 JST on 29 January 2003 (lower panel). Solid and dashed contours in middle panel indicate geopotential height and temperature, respectively. The vertical green line in the lower panel indicates the aircraft flight track.
around the cloud system were made.

Tall convective clouds about 5 km high stood vertically erect around the midway point of the flight (37.5°N), and some L-mode convective clouds about 4 km high were observed in places south of 37.5°N, where a distinct contrast between cloud and cloud-free regions could be seen.

North of 37.5°N, stratiform clouds (shown by light blue radar reflectivity in the upper panel of Fig. 12) blown out from the tall convective clouds near the midway point were observed above T-mode convective clouds with a top height of about 3.5 km. Doppler velocity (sum of the vertical air velocity and the falling velocity of hydrometeors; Fig. 12, middle panel) indicated updrafts of up to 10 m s⁻¹ and downdrafts of several meters per second in the central and southern parts of the snow clouds.

In the vicinity of 37.5°N and to the south of that latitude, strong updrafts penetrated close to the cloud tops, whereas they reached a height of only 3.5 km north of 37.5°N. The LWP peaked at about 1.3 mm (equivalent to 1300 g m⁻²) around 37.5°N; south of that latitude, peak LWP values were from 0.5 to 1.0 mm. From 37.5°N to 38.2°N, LWP values were near zero, and further north, peaks of about 0.5 mm were observed (Fig. 12, lower panel).

GPS dropsonde soundings indicated that, in the vicinity of 38.5°N, the relative humidity became slightly lower at an altitude of 3.5 km or higher, and the wind also became strong and southwesterly. The convective available potential energy ranged from 150 to 215 J kg⁻¹, which is high for the atmosphere over the Sea of Japan in winter. The convective inhibition was zero; thus, atmospheric stratification was unstable enough for convection to initiate anywhere and anytime. The cloud base, as estimated from the lifting condensation level, was about 500 m, and it increased to up to 1 km northward. Wind shear was weak in the south, and a weak eastward wind shear was seen at upper levels. Wind shear was also weak in the central area, and a weak wind region was observed at upper levels. In the north, a strong wind shear was directed toward the northeast (Fig. S6).

Convective clouds developed as a result of strong convergence below 1.2 km height between NW winds in the north and WNW winds in the south. The momentum at the lower levels was transported to the upper levels by strong updrafts. Although the WNW winds in the south gradually became westerly with height, convergence with the NW winds at altitudes

![Fig. 12. Vertical cross sections of cloud radar reflectivity (upper panel) and Doppler velocity (middle panel), measured by the W-band Doppler radar, and the horizontal distribution of the LWP (lower panel) measured by microwave radiometer from an altitude of 7.5 km along longitude 135.5°E (from Murakami et al. 2005b).](image-url)
up to 3 km was observed in the north. A remarkable divergence was observed between westerly winds in the south and the strong SW winds in the north at a height of 5 km (Fig. 13).

The thermal contrast in the north–south direction was not very strong; the horizontal gradient of the equivalent potential temperature was 2 to 3 K per 200 km (Fig. 14, upper left panel). At altitudes of 3 km or lower, the equivalent potential temperature in the vicinity of 37.5°N, where well-developed convective clouds were observed, was slightly higher than in the surrounding area. The dew point temperature showed that the air was almost uniformly humid, although it tended to be drier north of 37.5°N at heights of 4.9 and 0.5 km; in the south, a distinct contrast was observed between humid and dry areas.

At a height of about 7.5 km, there was a strong southwesterly jet in the vicinity of 36.5°N. At a height of 4.9 km, the westerly component was weaker in the vicinity of 37.5°N, where well-developed convective clouds were observed, and the southerly component rapidly increased northward from there. Although the data are not shown because severe icing caused a large error in the differential pressure measurement of the 5-hole radome, updrafts of 7–8 m s⁻¹ and downdrafts of 4–5 m s⁻¹ were observed in some places in the upper and middle levels of the clouds before the onset of severe icing; this finding is consistent with the Doppler velocity measurements by cloud radar. The difference between the pressure altitude and the GPS altitude (corresponding to the relative change of geopotential height on the isobaric surface, although with opposite sign) indicated the relative geopotential height decreased northward at rates of 50 m/200 km in the lower layer and 150 m/200 km in the upper layer (Fig. 14, upper right panel).

In terms of the cloud microphysical structure (Fig. 14, lower right panel), at an altitude of 4.9 km, small ice crystals of 200 μm or less were distributed from the midpoint northward, and in the well-developed convective cloud region, there were high concentrations of ice crystals reaching 1000 L⁻¹ (0.3 g m⁻³). Because the air temperature was as low as ~40°C, there was no supercooled cloud water. Below an altitude of 3 km, snow aggregates and lightly rimed snow particles predominated in the north, whereas graupel and heavily rimed snow particles were observed from the center southward, with concentrations of about 100 L⁻¹ and a maximum size of 5 mm.

5.3 Formation and maintenance processes of JPCZ snow clouds

The systematic air mass (airflow) convergence and systematic water vapor convergence at an altitude of 2.1 km or lower observed north of 37.3°N is inferred to contribute to the formation and maintenance of the well-developed convective clouds in the central area, and to the convective clouds with a top height of 3.5 km or less to their north (Fig. S7).

Significant air mass divergence was observed near the cloud top. Considering the presence of ice crystals with a maximum ice water content of 0.3 g m⁻³ (the actual ice water content was estimated to be 0.6–0.9 g m⁻³ after correction for the ice particle collection efficiency of the Nevzorov TWL/LWC probe with a shallow cone; Korolev et al. 2013) in the cloud top vicinity, a considerable amount of water vapor converged in the middle and lower levels, where it contributed to the formation of well-developed convective clouds and anvil-shaped ice clouds in the north, enhancing the precipitation efficiency of the convective clouds with top heights of 3.5 km or less through a seeder–feeder mechanism. At lower levels, the increase in water vapor density as the air mass...
traveled southward is attributed to water vapor supplied from the sea surface and evaporation of snow particles.

To the south of 37.3°N, however, no systematic air mass or water vapor convergence was observed, except for a small-scale, self-induced air mass and water vapor convergence produced by the lower-level downdraft as a result of mass loading and evaporation and sublimation of snow particles, which are typically seen below L- and T-mode snow cloud bands (or streets) (Kelly 1982; Yamada 2005).

5.4 Conceptual model of the well-developed arch-shaped cloud band

A conceptual model of the well-developed, arch-shaped cloud bands associated with the JPCZ is shown in Fig. 15. The cloud band was formed and maintained by the convergence of the northerly component of NW
winds in the north, destabilization of the cloud layer by the upper cold air mass, and the supply of heat and moisture from the warm sea surface and the moisture supply from evaporation (sublimation) of precipitation particles. Air temperature and equivalent potential temperature increased southward by 2–3 K/200 km, although the maximum temperatures were seen in the central area because of the heat flux from the warm sea surface and diabatic heating.

The JPCZ cloud bands consisted of well-developed convective clouds along their southern edge, strati -form ice clouds (anvil-like clouds) spreading north -ward from their upper parts, and shallow T-mode convective clouds beneath the anvil-like clouds. L-mode convective cloud bands occurred to the south of the well-developed convective clouds, and shallow L-mode convective cloud bands occurred to the far north.

LWP had a maximum value of 1.3 mm in the central area, peak values of 0.5–1.0 mm in the south, values of almost zero beneath the anvil-like clouds, and values of 0.5 mm in the far north. The maximum ice water path ranged from 2 to 3 mm. Under the anvil-like clouds, a seeder–feeder mechanism operated, and ice crystals from the anvil-like clouds grew mainly through vapor deposition. From the center southward and in the far north, the main growth mechanism was accretion of supercooled cloud droplets, and the predominant precipitation particle types were graupel and heavily rimed snow particles.

5.5 General characteristics of JPCZ cloud bands

During the three WMO seasons, a total of seven JPCZ cloud bands were observed. The features of each JPCZ cloud band are summarized in Table 1. The existence of T-mode cloud bands (streets) northeast (or north or east) of the well-developed convective clouds on the southwestern (or southern or western) edge of the JPCZ cloud bands was confirmed by infrared imaging (T (IR)) in five of the seven JPCZ cloud bands, in visible images (T (VIS)) in six, and in radar echoes (T (radar)) in six.

In six of the cloud bands, a strengthening of the southwesterly wind (SWinc), just above the top of shallow convective clouds on the northeast (or north or east) side of the well-developed convective clouds, was observed. A middle and upper level weak wind zone (WWZ) in the well-developed convective clouds was conspicuous in two cases, and a tendency toward weak winds in the middle and upper levels was observed in three cases. The orientation of the JPCZ cloud bands (the well-developed convective cloud line) was northwest–southeast in three cases and west–east in two cases, and it changed with time in two cases.

The moving speed (Moving Vel.) of the JPCZ cloud bands was, at most, about 70 km h⁻¹, and it was

| Table 1. Characteristics of seven JPCZ cloud bands (from Murakami et al. 2005b). |
|-----------------|-----------------|-----------------|-----------------|-----------------|-----------------|-----------------|-----------------|-----------------|
| T (IR)          | T (VIS)         | T (radar)       | dT/dL           | SWinc           | WWZ             | Orientation    | Moving Vel.     |
| 14 January 2001 | ○               | ○               | ○               | −2.5°C/170 km   | (u)             | △              | NW             | 60 km h⁻¹       |
| 15 January 2001 | ○               | ○               | ○               | −5.0°C/100 km   | (u, v)          | ×              | E              | 70 km h⁻¹       |
| 16 January 2001 | ○               | ○               | ○               | −2.6°C/170 km   | (v)             | ○              | NW             | 30 km h⁻¹       |
| 11 February 2002| ×               | ×               | ×               | +1.4°C/300 km   | ×               | ×              | N > NW         |                 |
| 28 January 2003 | ○               | ○               | ○               | −0.8°C/300 km   | (u, v)          | ○              | NW > E         |                 |
| 29 January 2003 | ○               | ○               | ○               | −1.2°C/220 km   | (v)             | △              | N              |                 |
| 30 January 2003 | ○               | ○               | ○               | −2.3°C/330 km   | (u)             | △              | NW             |                 |

“T (IR),” “T (VIS),” and “T (radar)” are abbreviations of T-mode cloud bands seen in satellite infrared images, satellite visible images and radar images, respectively. “dT/dL” is abbreviation of the maximum horizontal temperature gradient, “SWinc” a strengthening of the southwesterly wind, “WWZ” a middle and upper level weak wind zone, “Orientation” an orientation of a well-developed convective cloud line, and “Moving Vel.” a moving speed (refer to the details in the text).
negligibly small in four cases. The maximum horizontal temperature gradient (dT/dL) in the lower level between the two converging air masses was 5°C/100 km, and the minimum was 0.3°C/100 km. In six of the cloud bands, the air temperature was warmer on the southwestern (or south or west) side and colder on the northeastern (or north or east) side. The only exception was on 11 February 2002, when cold air protruded from the Korean Peninsula; on that date, the air temperature was colder on the west side than on the east side.

The strengthening of the southwesterly wind in the upper cloud layer appeared to be closely related to the presence of T-mode stratiform clouds in the upper layer, which were clearly seen in infrared satellite images, as well as the formation of the T-mode convection clouds in the middle and lower levels. They were also seen in visible satellite images. In most JPCZ cloud bands, strengthening of southwesterly wind was seen in the upper part of the clouds on the northeast or north side of the developed convective clouds; as a result, the vertical shear vector of horizontal风 was seen in the upper part of the clouds on the JPCZ cloud bands, strengthening of southwesterly wind was also seen in visible satellite images. In most cases, but in the region of the L-mode cloud bands (streets), the orientation of the cloud bands tended to coincide with the vertical shear vector in most cases, but in the region of the L-mode cloud bands (streets), the cloud band orientation and the vertical shear vectors often did not coincide.

The second hypothesis is that the band orientation is determined by interaction with snow particles falling from above the cloud band, as suggested by Fujiyoshi et al. (1998). Interestingly, an abrupt change, rather than a continuous change, is commonly observed in band orientation near the boundary between the shallow T-mode convective cloud band region on the northeast side of the well-developed convective cloud line of JPCZ cloud bands and the shallow L-mode convective cloud band region further to the northeast. Another interesting observation is that the T-mode convective cloud bands in the middle and lower levels emanate from the well-developed convective cloud line, and the width of the T-mode cloud bands is often large on the southwest side, tapering toward the northeast side. To elucidate the formation mechanisms of the T-mode convective cloud bands, more intensive research with a NHM that explicitly handles cloud microphysical processes is needed.

6. Polar low

Mesoscale vortical disturbances, commonly seen over the Sea of Japan, can be classified into three types according to their generation mechanism. Type (1) meso-α- or meso-β-scale vortical disturbances are triggered by a cold vortex aloft and develop in the central part of the Sea of Japan by low-level frontogenesis (cyclogenesis). The frontogenesis results from the confluence of, and differential thermal advection between, the polar air stream from the continent to the W–SW of a synoptic low and a warm air stream of Pacific origin originated near the center of the synoptic low (Ninomiya 1994). Type (2) meso-β-scale vortical disturbances form in the JPCZ by barotropic (Nagata 1993) or baroclinic (Tsuboki and Asai 2004) shear instability. Type (3) meso-β-scale vortical disturbances form off the west coast of Japan from interaction between the flow field of a synoptic low east of the Japanese islands and the topography of the Japanese islands (Hayashi et al. 2001; Hayashi 2005).

Although polar lows have been defined in several ways, all three types of vortical disturbances described above are considered to be polar lows in this review. This is in accordance with Fu et al. (2004), who defined polar lows as intense mesoscale cyclones with a horizontal scale of less than 1000 km that form poleward of the main polar front and develop over a high-latitude ocean in winter. They are usually characterized by comma (or spiral) cloud patterns on satellite images.

In this section, a vortical disturbance of the third type is described. Hayashi (2005) used an NHM to simulate this vortical disturbance and suggested, based on sensitivity tests for the Japanese island topography, that it formed as a result of interaction between the flow
field of a synoptic low and the topography of Japanese islands. He also suggested that the superposition of a weak low at 700 hPa, which traveled from northwest to southeast, triggering a vortical disturbance.

Inner structures of this polar low, which appeared over the Sea of Japan near Toyama Bay on 27 January 2001, are documented here, mainly on the basis of in situ measurements, cloud radar observations, and dropsonde soundings from the G-II instrumented aircraft, along with additional dropsonde soundings from the Citation V aircraft and radiosonde soundings from JMA observation vessels, along with special observation sites in coastal areas.

6.1 Synoptic and mesoscale features

A polar low formed between the Noto Peninsula and Sado Island over the Sea of Japan when a synoptic-scale low passed along the Pacific coast of the Japanese islands during the daytime on 27 January 2001. After forming, the polar low moved slowly northeastward and reached a point north of Sado Island by 2100 JST on 27 January (Fig. 16, upper panel).

The timing of the polar low formation coincided with the superposition of a low-pressure region at the 700 hPa level in the formation area. It was moving from the western Sea of Japan southeastward toward the Pacific Ocean. Echo areas associated with the polar low were circular or spiral-shaped, and they were weak in intensity and quasi-stationary during the aircraft observation period (1230 to 1600 JST).

6.2 Observational results

The instrumented aircraft flight path had four legs, each at a different height (10.3, 3.5, 1.5, and 0.3 km), along the 137.5°E meridian, and two legs at heights of 10.2 and 1.5 km along the 37.7°N parallel.

Upper-level clouds associated with the synoptic-scale low extended widely over the observation area, and the polar low was not identified in the satellite images during the aircraft observation period (Fig. 16, lower panel).

At 10.3 km, the horizontal winds were uniformly southwesterly with a speed of ~30 m s$^{-1}$. Cloud radar indicated upper level stratiform clouds associated with the synoptic-scale low between 4 and 9 km; beneath the upper-level clouds, lower-level clouds associated with the polar low were observed, and their tops reached a height of 3–4 km (Fig. 17). Horizontal winds showed a remarkable spatial change below 1.5 km. Especially at 0.3 km, easterly winds with a speed greater than 10 m s$^{-1}$ were observed to the north, and westerly winds were observed to the south. A similar cyclonic wind pattern was not observed at a height of 3.5 km (Fig. S9).

A warm core was observed in the central part (distance = −30 to −100 km) of the polar low. The temperature contrast was most prominent at 1.5 km, and equivalent potential temperature in the warm core was higher by 2–3 K than in the surrounding area at this level (Fig. 18, upper panel).

On average, a weak updraft (0.1–0.3 m s$^{-1}$) was observed in the central part of the polar low below 1.5
km, although the actual vertical wind field was a mingling of updrafts and downdrafts at horizontal scales of 10–20 km (Fig. 18, lower panel).

At altitudes of 0.3 and 1.5 km, in situ measurements of thermodynamic and kinematic structures suggested the air temperature in the warm core increased because of the descending air motion, as inferred from the fact that the difference between the air temperature and the dew point temperature was large in the high-temperature region. In addition, there was good correspondence between the region of mixed moderate updrafts and downdrafts (within ±3 m s⁻¹) and the region of high equivalent potential temperature. (Note that the absolute values of vertical velocity in Fig. 18 are underestimates, because the vertical cross section was drawn by interpolation using data averaged over horizontal distances of 5 km.) A pressure dip of 1–2 hPa in the central part of the polar low at 0.3 km was inferred from the difference between pressure altitudes and GPS altitudes (Fig. S10).

Horizontal distributions of equivalent potential temperature and horizontal winds at 0.3 and 1.5 km show that an air mass with high equivalent potential temperature associated with the circulation of the synoptic-scale low entered the domain across its southeast boundary and exited the domain across its southwest boundary. This result suggests that the part of the circulation associated with the synoptic-scale low formed a secondary closed cyclonic circulation, and the warm core was located near the center of that secondary circulation. At an altitude of 3.5 km, no noticeable cyclonic circulation was observed, but the equivalent potential temperature was high, similar to that in the lower layers, because of the vertical transport of air with high equivalent potential temperature from the lower levels and diabatic heating, which
resulted from water vapor condensation caused by the convective activity (Fig. 19).

A thick stratiform cloud layer, associated with the synoptic low that was traveling along the southeastern coast of the Japanese islands, was above the precipitating clouds that accompanied the polar low (Fig. 17). Snow particles falling from the stratiform clouds grew through the consumption of excess water vapor and supercooled cloud droplets by a seeder–feeder mechanism.

Fairly high concentrations of ice crystals and snow particles, measured with the 2D-C probe, were observed north and south of the warm core, where the horizontal gradient of equivalent potential temperature was large (Fig. S11). Cloud droplet regions with low water content (at most 0.1–0.2 g m\(^{-3}\)) were approximately collocated with regions with high concentrations of snow particles (distance = −10 to −20 km and distance = −90 to −110 km), except in the north (distance ≈ 80 km).

As shown in the 2D-C image (Fig. S12), the snow particles observed in the polar low consisted mainly of dendritic crystals, grown mostly by vapor deposition, lightly rimed dendritic crystals, and aggregates of those crystals.

Excess water vapor, produced by the circulation of the polar low, was consumed by the depositional growth of ice crystals seeded from the upper cloud deck associated with the synoptic-scale low. Therefore, cloud droplet regions were confined to well below 3.5 km and above the cloud base (~0.5 km), except in the north (distance = 60–80 km), where concentrations of both ice and snow particles were very low.

6.3 Conceptual model

The mesoscale structures of the polar low, which formed between the Noto Peninsula and Sado Island

Fig. 18. Vertical cross section of equivalent potential temperature (THETA\(E\)) (upper panel) and vertical wind velocity (lower panel) along longitude 137.5°E. The distance increases northward (from Murakami et al. 2005c).
over the Sea of Japan on 27 January 2001 by the interaction between the flow field of a synoptic low and the topography of the Japanese islands, are schematically shown in Fig. 20. The cloud system had a horizontal scale of 100–200 km and a vertical scale of 3–4 km. Characteristic wind fields were detected below a height of 1.5 km. Remarkable cyclonic wind patterns and a pressure dip of 1–2 hPa were observed in the lowest level (0.3 km). The warm core was most remarkable at the 850 hPa level, and equivalent potential temperature was 2–3 K higher in the warm core than in the surrounding area. Neither cyclonic circulation nor a warm core was observed at the 700 hPa level.

The observation results suggest that the causes of the warm core were 1) advection of an air mass with high equivalent potential temperature accompanying the circulation of the synoptic-scale low passing along the southeastern coast of the Japanese islands, 2) diabatic heating accompanied by condensation, and 3) an adiabatic temperature increase caused by the downdraft.

Most of the excess water vapor produced by the polar low circulation was consumed through depositional growth of ice crystals seeded from the upper-level clouds associated with the synoptic-scale low passing along the Pacific coast of the Japanese islands.
(Seeder–feeder mechanism). As a result, cloud droplet regions were spatially and temporally limited, and their water contents were, at most, 0.1–0.2 g m\(^{-3}\). The primary growth mechanism of precipitation particles was depositional growth of ice and snow particles; riming growth of snow particles was a secondary growth mechanism in this cloud system.

7. Concluding remarks

Observations made from instrumented aircraft have considerably advanced our understanding of the microphysical characteristics and cloud- and meso-scale structures of snow clouds over the Sea of Japan.

However, observation data of heat and water vapor fluxes from the sea surface, the development of a convectively mixed boundary layer, and the formation of stratocumuli with precipitation, which are fundamental processes of snow cloud formation over the Sea of Japan during the winter monsoon, are available only for the late stages of a cold air mass outbreak. Thus, additional observation data need to be collected, especially data relating to the water substance budget. Uncertainty also remains regarding the absolute value of the ice water content, owing to limitations of instrument accuracy in past observations; therefore, more accurate observations of ice water content should be obtained by the latest measuring methods. Because the water contents of ice crystals and snow particles are also important parameters for validating numerical models and remote sensing algorithms, it is highly desirable to acquire data using highly accurate measuring instruments from various types of snow clouds.

To date, quantitative discussion of conditions leading to the transition from isolated convective clouds to L-mode or T-mode cloud bands (streets) has been insufficient. The transition from isolated convective clouds in an atmospheric environment characterized by weak horizontal winds or weak vertical shear to L-mode cloud bands in an atmospheric environment characterized by strong horizontal winds or their strong vertical speed shear, or to T-mode cloud bands in an atmospheric environment characterized by strong vertical directional shear of the horizontal wind, should be investigated in greater detail.

With regard to the organization into the L-mode and T-mode cloud bands, the band orientation sometimes deviates from the direction of vertical shear of the horizontal wind, which is inconsistent with the theory of dry convection (Asai 1970a, b, 1972). Moreover, primary and secondary band orientations sometimes occur simultaneously. Also, convective cells are frequently observed to be randomly distributed in cloud bands, instead of arranged in a row. Below the cloud base, downdrafts of colder air caused by sublimation of snow particles and consequent intensification of the horizontal wind at lower levels are frequently observed, which suggests that horizontal convergence influences the organization of the cloud bands (Inoue et al. 2005; Murakami 2005a). Thus, wet convection influence on the orientation of cloud bands can be inferred, but the mechanism of this influence has not been sufficiently investigated.

In snow cloud bands, both dry and moist convection are observed. In moist convection, the latent heat released by the phase change of water substances and the water substance loading may have a large impact on the convection structure. These effects must be considered to elucidate the mechanism that organizes convective snow clouds. Further, it is necessary to examine whether roll convection (dry convection) parallel to the vertical shear of horizontal wind adequately explains the formation of shallow T-mode cloud bands to the northeast of JPCZ snow clouds, or whether the effect of mass loading and latent heat release caused by sublimation of water substances needs to be considered as well.

The small-scale polar low documented here formed under the strong influence of topography (the Japanese islands). In situ observations of the larger scale polar lows that typically occur in the central part of the Sea of Japan have not yet been carried out. Therefore, observation data on those polar lows should be acquired to clarify their structures, formation, and maintenance mechanisms.

To acquire high-precision observation data required to tackle these remaining problems, strategic observations by aircraft equipped with high-accuracy, in situ measuring instruments and multi-parameter radar systems capable of simultaneous multidirectional measurements are needed. To observe meteorological disturbances such as JPCZ snow clouds and polar lows, which have a spatial scale of several hundred kilometers, it may be necessary to develop an observation strategy using multiple aircraft.

To carry out innovative research that addresses the unsolved problems in this field, it is necessary (1) to understand the kinds of data that can be obtained by aircraft observations, (2) to review strategic observation flight plans, (3) to develop and deploy new equipment if necessary, and (4) to deploy the aircraft to carry out the observations.

The results of aircraft observations carried out from the early 1990s through the early 2000s have been reviewed in this paper. Needless to say, the further
accumulation of aircraft observation data is essential, because of the limited aircraft observation data currently available in Japan. However, it may be difficult to achieve breakthrough results simply by conducting aircraft observations.

The 1990s and the early 2000s were the heyday of hydrostatic models; NHMs able to represent clouds and precipitation were only beginning to be developed, and they could not yet adequately reproduce natural phenomena. Therefore, at that time, aircraft observation data had a large role to play in the elucidation of atmospheric phenomena and model validation. Since then, however, NHMs have advanced remarkably, although further improvement is still needed. Recently, NHMs have shown quite reasonable reproducibility of natural phenomena, providing detailed four-dimensional data. With regard to satellite observations, multi-frequency data with high temporal and spatial resolution and data acquired by both active and passive sensors have become available.

These developments must be considered when planning and carrying out aircraft observations to contribute to atmospheric science progress.

Supplements

The supplement includes a map around Japan with the area division in Japan and geography (Fig. S1), a photograph of the University of Wyoming’s King Air B200T instrumented aircraft (Fig. S2), vertical profiles of the u and v components of the momentum fluxes in convectively mixed boundary layer capped with stratocumuli (Fig. S3), vertical profiles of relative humidity (R.H.) in convectively mixed boundary layer capped with stratocumuli (Fig. S4), CAPPIs of radar reflectivity and system-relative winds at 0.8, 1.3, 1.8, and 2.3 km altitude in L-mode cloud bands (Fig. S5), photos of JPCZ cloud band tops taken from the G-II and 2.1, 3.0, 4.9, and 7.5 km across JPCZ cloud bands (Fig. S7), shear vectors of horizontal wind between the cloud base and cloud top, obtained from dropsonde observations from the G-II and Citation-V aircraft and rawinsonde observations from observation vessels and ground-based observation sites, superimposed on a visible GSM image (Fig. S8), wind barbs at 0.3 km, 1.5 km, and 3.5 km measured by the G-II, obtained from ground-based and ship-borne radiosonde measurements overlaid on a CAPPI of the JMA C-band radar composite at 2-km height (Fig. S9), horizontal distributions of the relative change of geopotential height on the isobaric surfaces in polar low (Fig. S10), vertical cross section of snow particle concentrations in JPCZ cloud bands (Fig. S11), cloud and precipitation particle images acquired at altitudes of 0.3, 1.5, and 3.5 km in polar low (Fig. S12).

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