Mesoscale and Microscale Structures of Snow Clouds over the Sea of Japan

Part I: Evolution of Microphysical Structures in Short-lived Convective Snow Clouds

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(Manuscript received 14 December 1993, in revised form 27 July 1994)

Abstract

Convective snow clouds over the Sea of Japan were observed with hydrometeor vide sondes (HYVIS), hydrometeor video dropsondes (HYDROS), Doppler radars and a microwave radiometer in four winter seasons from 1989 through 1992. This paper describes the evolution of microphysical structures in short-lived convective snow clouds with a cloud top temperature of $-20^{\circ}\text{C} \pm 3^{\circ}\text{C}$, based on a composite of separate snow clouds at various stages.

At developing stages, moderate updraft ($w_{\text{max}} \sim 4 \text{ m s}^{-1}$) produced a high concentration of supercooled cloud water (close to the adiabatic liquid water content) throughout a cloud. Number concentrations of ice crystals ($d < 200 \mu\text{m}$) were about $10 \text{ particles L}^{-1}$ and few precipitation particles (snow and graupel; $d \geq 200 \mu\text{m}$) were present. At mature stages, the maximum ice crystal concentrations sometimes exceeded $100 \text{ particles L}^{-1}$ and concentrations of precipitation particles consisting of graupel and heavily rimed snow crystals were $\sim 10 \text{ particles L}^{-1}$. These solid particles depleted a material amount of supercooled cloud water through depositional and accretional growth. By decaying stages, almost all supercooled cloud water had been depleted by snow and graupel particles. Only unrimed and lightly rimed snow crystals were still suspended in upper and middle layers.

Ice crystals, especially isometric ones such as thick plates and columns, initiated in coexistence with supercooled cloud water (a likely nucleation mechanism is the freezing of cloud droplets), continue depositional and accretional growth and finally produce graupel particles. On the other hand, supercooled and frozen drops were also found in snow clouds at developing and mature stages, which suggests that the warm rain process (precipitation formation through collision and coalescence among cloud droplets) also operates and frozen drops serve as graupel embryos, although these contributions are considered to be small because of low concentration of such drops and their spatially and temporally limited distributions in snow clouds.

The cloud water budgets are computed and indicate that, at the developing stage, excess vapor production due to adiabatic ascent of air parcels is predominant and cloud water contents in snow clouds become close to adiabatic values. However, once precipitation is well developed (at the mature stage), clouds (even with an updraft velocity of 1 m s^{-1}) can not maintain a steady (equilibrium) state and precipitation particles quickly deplete the cloud water through their depositional and accretional growth.

1. Introduction

Heavy snowfall often occurs over the coastal regions facing the Sea of Japan as well as the mountain regions during winter monsoon seasons in Japan. The cold air mass from the Eurasian Continent receives a great amount of heat and moisture from the relatively warm sea surface as it crosses the Sea of Japan. Consequently, active cumulus convections develop in the moist and unstable stratification and cause the intense snowfall in downstream coastal regions of the Sea of Japan. A large amount of snow-

fall thus produced sometimes does damage to traffic and housing.

Understanding of precipitation mechanisms in the snow clouds is essential for improvement of short-range forecasting and the establishment of snow cloud modification that would lead to a lessening of this damage. As for short range forecasting of mesoscale precipitation phenomena, the type of precipitation particles or, more strictly speaking, the fall velocity of precipitation particles is as important a factor as the kinematic and thermodynamic structures in clouds, because it controls positions of surface precipitations. Assuming snow clouds with

Fig. 1. Observation area (the square in the upper right panel) and schematic of the observational facilities used in the Cooperative Japan Sea Snow Cloud Project. MD is the meteorological dropsonde, HD the hydrometeor video dropsonde, RS the rawinsonde, HV hydrometeor video sonde, DR the Döppler radar, PR the dual polarization radar, and SO the surface snowfall observation.

a typical cloud top height (~ 3 km) and mean wind velocity ($\sim 15 \text{ m s}^{-1}$) in a mixed-layer, snow crystals and aggregates with fall velocities of 1 m s^{-1} or less are drifted several tens kilometers leeward, while graupel particles with fall velocities of $\sim 4 \text{ m s}^{-1}$ are drifted only ten kilometers or so. The separation of surface precipitation areas for graupel and snow cases would be a few tens kilometers or more. Types of precipitation particles produced in snow clouds are determined by the amount of supercooled cloud water and by concentrations and spatial and temporal distributions of ice particles. Also, for snow cloud modification, where one attempts to alter types of hydrometeors by cloud seeding and consequently the position and amount of surface precipitations, it goes without saying that a deep understanding of these parameters is essential.

Many studies (Ninomiya, 1968; Matsumoto *et al.*, 1965; *etc.*) have been made on synoptic and mesoscale features of snow clouds over the Sea of Japan. However, from the viewpoints of microphysical structures of snow clouds and precipitation mechanisms, limited studies (Isono *et al.*, 1966; Magono and Lee, 1973) have been made to date. To identify meso- and microscale structures of the snow clouds and their precipitation mechanisms, and to

evaluate the feasibility of snow cloud modification by seeding, the Cooperative Japan Sea Snow Cloud Project was carried out over the coastal regions facing the Sea of Japan in Tohoku District from 1989 through 1993 (see the upper right panel of Fig. 1). Three national institutes and three universities participated in this project and a variety of instruments such as 3 Döppler radars, a dual polarization radar, a microwave radiometer, hydrometeor videosondes, hydrometeor video dropsondes were deployed.

In this paper, we present microphysical and thermodynamic structures of convective snow clouds investigated through about forty soundings of hydrometeor videosondes and hydrometeor video dropsondes, which were made in four winter seasons from 1989 to 1992. Especially, we focus on the behavior of supercooled cloud water and ice particles in short-lived convective snow clouds. Airflow structures in isolated snow clouds are presented by Yamada *et al.* (1994), although the clouds described in this paper are not identical with those in their paper.

Table 1. Characteristics of the snow clouds observed in the 1989, 1990, 1991 and 1992 field projects.

TIME	DATE	MONTH	YEAR	SONDE	CTYPE	STAGE	Ttop	Htop	LCL	Ncmax	LWCmax	Nrmax	RWCmax	Nimmax	IWCmax	Nsmax	SWCmax	REMARKS
10:14:00	2	FEB	1989	HYVIS	Isolated	Mature	-21	3	0.63	8.1	0.13	0	0	228	0.031	5.3	0.12	
16:24:00	2	FEB	1989	HYVIS	Isolated(T)	Mature	-25.1	3.2	0.3	4.2	0.03	0.25	0.016	24.2	0.013	3.3	0.033	
17:43:00	2	FEB	1989	HYVIS	Isolated(T)	Mature	-25.2	3.5	0.3	1.3	0.02			93.5	0.03			*1
10:55:00	3	FEB	1989	HYVIS	Isolated	unknown	-20.5	2.25	0.6	9.9	0.16	0	0	51.6	0.015	3.9	0.095	*2
15:26:00	3	FEB	1989	HYVIS	Isolated	Decaying	-19.1	2.5	0.75	0	0	0	0	17.1	0.0075	15	0.2	
13:43:00	4	FEB	1989	HYVIS	Isolated	Mature	-21.9	3.25	0.84	9.4	0.14	0.07	0.0059	37.8	0.014	33.6	0.315	
15:01:00	4	FEB	1989	HYVIS	Isolated	Mature	-20.9	3	0.8	14.7	0.12	0.036	0.011	31.3	0.059	10	0.21	
14:43:00	9	FEB	1989	HYDROS	Isolated	Unknown	-5.5	1.34	0.4	175	0.29	0	0	0	0	0	0	
15:58:00	3	FEB	1990	HYVIS	C-F	Maintained	-20.2	3.59	0.38	118	0.4	17.8	0.83	44	0.0048	29.4	0.14	*2
10:57:00	7	FEB	1990	HYVIS	Isolated(T)	Mature	-14.5	3	0.67	9.7	0.15	2.8	0.067	11	0.0044	4.6	0.011	*2
11:58:00	7	FEB	1990	HYVIS	C-F	Maintained	-20.1	4	0.85	28.3	0.24	0.34	0.012	11.7	0.0061	2.1	0.067	
14:45:00	7	FEB	1990	HYVIS	post C-F	Maintained	-18	3.5	0.6	31.2	0.21	0.18	0.0054	9.1	0.0028	2.4	0.031	
8:14:00	12	FEB	1990	HYVIS	Intermediate	Maintained	-23	4.5	0.6	10	0.2	5.9	0.12	43.2	0.008	7.9	0.16	
14:51:00	12	FEB	1990	HYDROS	Band	Maintained	-23.1	3.76	0.7	10.2	0.18			622	0.2			*3
6:49:00	13	FEB	1990	HYVIS	Isolated(T)	Mature	-23.7	3.75	0.8	13.3	0.2	0	0	67.5	0.0075	4.7	0.066	
8:03:00	13	FEB	1990	HYVIS	Isolated	Decaying	-20.4	3.5	1.5	0	0	0	0	78	0.0087	4.3	0.097	
9:09:00	13	FEB	1990	HYVIS	Isolated	Mature	-20.3	3	1.7	2	0.03	0	0	11.7	0.00047	1.7	0.03	
10:36:00	13	FEB	1990	HYVIS	Isolated	Decaying	-19.2	3.25	1.8	0	0	0	0	13.4	0.011	2.5	0.11	
15:41:00	2	FEB	1991	HYVIS	Isolated	Mature	-26.5	3.5	0.48	0	0	0	0	306	0.0045	3.2	0.024	
14:36:00	3	FEB	1991	HYVIS	Isolated	Mature	-18.3	2.5	1.2	15.6	0.06	0	0	12.9	0.0024	0.78	0.0086	*2
5:24:00	5	FEB	1991	HYVIS	Band(T)	Maintained	-25.3	3.25	0.45			0.038	0.0022			15.1	0.34	*4
10:58:00	5	FEB	1991	HYDROS	Isolated	Mature	-23.5	2.65	0.45	97	0.13			28.6	0.013			
11:43:00	5	FEB	1991	HYDROS	Isolated(T)	Mature	-16.1	2.04	0.45	56.3	0.06			26.3	0.014			
11:53:00	5	FEB	1991	HYVIS	Isolated(T)	Decaying	-28.2	3.75	0.7	0	0	0	0	55.4	0.0011	11.2	0.089	
15:41:00	5	FEB	1991	HYDROS	Isolated(L)	Developing	-17	2.25	0.45	95.5	0.67			14.3	0.013			
14:56:00	8	FEB	1991	HYVIS	Isolated	Mature	-24.2	3.25	0.35	29.9	0.23	0	0	25.7	0.0019	15.8	0.077	
19:04:00	8	FEB	1991	HYVIS	Isolated(T)	Decaying	-32.4	4.25	0.35	0	0	0.31	5.00e-05	7.7	0.011	23.2	0.24	
15:12:00	10	FEB	1991	HYVIS	W-F	Maintained	-36.8	7.25	0.65	89.7	0.08	4.8	0.0017	66.9	0.0087	4.3	0.067	*5
17:04:00	1	FEB	1992	HYVIS	Isolated(L)	Mature	-16.4	2	0.4	0	0	0	0.0017	0.00073	11.5	0.0036	2.7	0.02
15:09:00	3	FEB	1992	HYVIS	W-F	Maintained	-20.1	2.55	0.45	6.5	0.08	5.9	0.23	8.4	0.00314	4.2	0.27	
16:42:00	8	FEB	1992	HYVIS	Isolated(T)	Mature	-22.8	2.75	1.08	55	0.15	0.18	0.00028	61.1	0.011	4.1	0.04	
21:46:00	8	FEB	1992	HYVIS	Isolated(T)	Mature	-23.6	3	0.8	48.6	0.15	0	0	114	0.0098	13.4	0.17	
23:06:00	8	FEB	1992	HYVIS	Isolated(T)	Mature	-24.8	3.25	0.9	29.7	0.16	0	0	182	0.023	8.4	0.2	
9:36:00	9	FEB	1992	HYDROS	Isolated	Developing			0.5									*6
10:16:00	9	FEB	1992	HYDROS	Isolated	Developing	-17.2	2.04	0.5	257	1.15			13.2	0.0034			
11:59:00	9	FEB	1992	HYVIS	Band(T)	Maintained	-17.4	2.25	0.4	29.4	0.08	0.09	0.03	22.7	0.002	2.5	0.037	*2
13:56:00	9	FEB	1992	HYVIS	Band(T)	Maintained	-17.8	2.75	0.2	0	0	0.02	0.00095	16.2	0.0039	17.4	0.1	
14:41:00	9	FEB	1992	HYVIS	Band(T)	Maintained	-17.6	2.5	0.12	38.5	0.13	0	0	63.6	0.0074	6	0.085	

*1 No data on precipitaton particles

Ttop: cloud top temperature (C)

*2 Natural seeding from upper clouds

Htop: cloud top height (km)

*3 Double layer structures

LCL: lifting condensation level (km)

*4 No data on cloud particles

Ncmax: the maximum number concentration of cloud droplets (cm^{-3})

*5 HYVIS grazed the cloud

LWCmax: the maximum cloud water content (gm^{-3})

*6 Quality of HYVIS images is poor

Nrmax: the maximum number concentration of drizzle drops (L^{-1})RWCmax: the maximum water content of drizzle drops (gm^{-3})Nimmax: the maximum number concentration of ice crystals (L^{-1})IWCmax: the maximum ice water content (gm^{-3})Nsmax: the maximum number concentration of snow crystals (L^{-1})SWCmax: the maximum snow water content (gm^{-3})

Fig. 2. Schematic of the microphysical structures in snow clouds with cloud top temperatures of $-20^{\circ}\text{C} \pm 3^{\circ}\text{C}$ at their developing, mature and decaying stages. Wave lines and shaded areas indicate cloud top and supercooled cloud droplet regions, respectively. Solid circles and *f.d.* indicate supercooled drizzle drops and frozen drizzle drops. Classification of observed snow crystals is based on Magono and Lee (1966), which is, however, simplified to some extent. In addition, levels of 0°C , -10°C and -20°C are shown in the figures.

2. Facilities in the Cooperative Japan Sea Snow Cloud Project

2-1. Disposition of observation facilities

In this project, we used various instruments to investigate microscale as well as mesoscale features of snow clouds. The facilities used in this project and their dispositions are schematically shown in Fig. 1.

To investigate microphysical and thermodynamic structures of snow clouds, we have developed a hydrometeor videoprobe (HYVIS; Murakami and Matsuo, 1990) and its dropsonde version, hydrometeor video dropsonde (HYDROS). We had not made any direct measurements by aircraft for the first four winter seasons, although we finally flew an instrumented aircraft into snow clouds for the fifth winter season of the project. This was because we did not have sufficient information about the spatial extent of supercooled cloud water and its maximum water content in snow clouds forming over the Sea of Japan, which is needed to evaluate the safety of aircraft penetration.

We chose Tobishima Island for a HYVIS observation station because its smallness (less than 2 km in size) would not affect snow clouds passing over the island and because it is located 30 km offshore and within a dual Doppler radar observation area. At this site, vertically integrated cloud water contents

were measured by using the microwave radiometer (Radiometric Corporation, WVR-1000). For observations of "developing stage" snow clouds without a significant radar reflectivity, the microwave radiometer provided a guidance for a timing of an HYVIS launch.

The airplane (Fairchild Marine IV) was used to release HYDROS and meteorological dropsondes and to receive the signal from them. An on-board GPS (satellite-based global positioning system) provided us with accurate positions of dropsonde release.

Four Doppler radars were deployed in this project. Three of them operated at 3 cm wavelength: Doppler radars operated by the Meteorological Research Institute (MRI), the National Research Institute for Earth Science and Disaster Prevention (NIED), and the Institute of Low Temperature Science (ILTS), Hokkaido University. The fourth radar, operated by the Public Works Research Institute (PWRI), is a Doppler and dual polarization radar operating at a wavelength of 5 cm. Dual Doppler radar observations were made using MRI and ILTS radars (MRI and NIED radars in 1991 and 1992). Both radars (MRI radar at Kisakata and ILTS radar at Sakata) were located on the coastal area with a base line length of ~ 30 km. At the both sites, the radars were able to obtain an unobstructed view of snow clouds over the Sea of Japan. ILTS and NIED

Fig. 3. PPI of radar reflectivity at 1545 5 Feb. 1991. The cross mark indicates the position of HYDROS.

radars, which were located on the windward and leeward sides of Dewa Hill were used to investigate snow cloud modification over the hill.

Rawinsonde measurements were made at Tobishima Island, Sakata and Shinjo (leeward side of Dewa Hill). Surface measurements were made of snow crystal habits, size distribution, and fall velocity at Tobishima Island, Sakata and Karikawa (windward slope of Dewa Hill).

To investigate meso- and synoptic scale features, surface temperature, wind and precipitation measurements from the Automated Meteorological Data Acquisition System (AMeDAS) operated by Japan Meteorological Agency (JMA), routine rawinsonde measurements and satellite data were available.

2-2. Hydrometeor videopsonde and hydrometeor video dropsonde

The HYVIS has two small video cameras with different magnifications to take pictures of hydrometeors from 7 μm to 2 cm in size. The HYVIS and a conventional rawinsonde are attached to the same balloon and launched into clouds. They transmit hydrometeor images and meteorological data over 1,687 and 1,680 MHz microwave link to a ground station (see details in Murakami and Matsuo, 1990).

To increase mobility in microphysical measurements, we improved on the HYVIS and have made a hydrometeor video dropsonde that is released from aircraft from above cloud tops. One of main differences is that the meteorological signal is superposed on the image signal (*i.e.*, the HYDROS uses only one microwave frequency to transmit hydrometeor images and meteorological data). This allows us

Fig. 4. Vertical profiles of temperature, relative humidity (solid and dashed lines in the left panel), snow crystals observed aloft (the center panel) and potential temperature, equivalent potential temperature and saturated equivalent potential temperature (solid, dashed and dash-dotted lines in the right panel) at 1541 5 Feb. 1991.

to make multiple HYDROS soundings at the same time. The other is the way to collect hydrometeor particles. The HYDROS collects cloud particles on a film facing downward by inertial impaction and precipitation particles in a rear wake onto a film facing upward by sedimentation, while the HYVIS collects all hydrometeors on a film facing upward. The hydrometeor images and meteorological data can be received at the aircraft which was equipped with an auto-tracking antenna, or at the ground station (the details will be presented in a future paper).

Snow and graupel particles sometimes bounce off or are blown off the collecting film surface. However, no corrections for these effects were applied to HYVIS data because we did not have any quantitative information about the effects. Therefore, the number concentration of precipitation particles (snow and graupel) measured by the HYVIS might be underestimated to some extent. On the contrary, precipitation particles, especially rimed parti-

cles sometimes produce a significant number of fragments of particles on their collection onto a film surface. In this study, therefore, only particles smaller than 200 μm whose crystal habits were discernible were counted as ice crystals, and heavily rimed particles and fragments of snow crystals were rejected.

3. Results

3-1. General features of snow clouds

We made about forty HYVIS and HYDROS soundings in four winter seasons from 1989 through 1992. The observed snow clouds were in forms of L-mode, T-mode or isolated convective snow clouds, or snow bands formed by mesoscale convergence, and were at various stages of their life cycles. The whole life cycle of individual snow clouds has not been successfully observed yet. General and microphysical features of the snow clouds are summarized in Table 1. All soundings except for the cases on 12 Feb. 1990, 10 Feb. 1991 and 3 Feb. 1992, which are

Fig. 5. Vertical changes in hydrometeor concentrations. (A) water content of cloud droplets (Q_c , rectangles) and ice crystals (Q_i , dotted line); (B) number concentration of cloud droplets (N_c) and ice crystals (N_i).

stratiform snow clouds associated with warm fronts and extra-tropical cyclones, were from snow clouds associated with cold air outbreaks that occurred under pressure patterns of the winter monsoon (west-high and east-low). Their cloud-top temperatures (heights) ranged from -6°C to -32°C (from 1.3 km to 4.25 km). The cloud-base heights (temperatures) that were estimated from lifting condensation levels were between 0.1 km and 3.9 km (between 1°C and -14°C), and cloud thickness was between 0.9 km and 3.9 km. Cloud water contents and number concentrations of cloud droplets in snow clouds ranged from 0 to 1.2 g m^{-3} and 0 to 260 cm^{-3} , depending on the stages of cloud life cycles. Number concentrations of ice crystals ($d < 200 \mu\text{m}$) and snow particles ($d \geq 200 \mu\text{m}$) ranged from 0 to 620 L^{-1} and from 0 to 30 L^{-1} , respectively. Supercooled cloud droplets usually existed in upper and middle parts of clouds although they did not necessarily show single layer structures but multiple layer structures. A schematic of all the snow clouds listed in Table 1 are shown in the Appendix. Note that all the observed snow clouds were associated with normal or mild cold air outbreaks, not with extremely cold air

outbreaks.

The observed convective snow clouds consisted of two types of clouds. One is the short-lived snow cloud whose radar echoes are easily recognized as a convective cell and shows distinct evolution (~ 1 hr of lifetime) corresponding to the cloud's life cycle. The other is the well-organized snow band that is self-maintained and long-lasting ('maintained' in the seventh column of Table 1). In this chapter, evolution of microphysical structures in the former convective snow clouds is investigated. Observed short-lived convective snow clouds were classified into 3 stages of life cycles based on radar echo evolution (PPIs at elevation angles between 1° and 4°) around HYVIS and HYDROS launch time: At the developing stage, radar reflectivities of snow clouds are increasing and maximum reflectivities do not yet exceed 15 dBZ. This means that precipitation formation has just begun and the concentration of precipitation particles in the snow clouds is still low. Precipitation particles may not yet reach the ground. At the mature stage, their radar reflectivities are still increasing but maximum reflectivities already exceed 15 dBZ. Alternatively, their radar reflectivi-

Fig. 6. Examples of hydrometeor images taken at -14 , -13 and -10°C levels. The width of photos is approximately 1.5 mm.

ties do not show any significant increase or decrease. This means that active precipitation formation continues to occur in the snow clouds. Precipitation particles may already reach the ground. At the decaying stage, their radar reflectivities are decreasing. In this stage, precipitation formation weakens or ceases and the main portion of precipitation particles may have already settled out of clouds. For convenience, we use the definition mentioned above for stages of life cycle, although this definition may not be universally accepted.

Microphysical structures of snow clouds are expected to depend upon cloud top temperatures as well as on the cloud's dynamical structures. Therefore, choosing convective snow clouds with their cloud top temperatures of $-20^{\circ}\text{C} \pm 3^{\circ}\text{C}$ and compositing observational results of separate snow clouds at various stages, we present the evolution of microphysical structures of snow clouds below (see

the chosen snow clouds at each stage in Fig. 2).

3-2. Snow cloud at the developing stage

The microphysical structure of the snow cloud observed with the HYDROS at 1541 5 Feb. 1991, as an example of snow cloud in its developing stage, is described. At that time, snow cloud streets of longitudinal mode prevailed over the observation area, as seen in PPI echoes of Fig. 3. Individual convective snow clouds composing L-mode cloud streets formed around 50 km offshore and developed as they approached the shore. The airplane was directed to one of the cloud streets by a radar operator and launched the HYDROS into one of the newly developing convective snow clouds with a bulging cloud top (shown by the cross mark in Fig. 3).

Vertical structures of thermodynamic and microphysical features are shown in Figs. 4 and 5. The cloud top height (temperature) and cloud base

Fig. 7. Vertical changes in the size distributions of cloud droplets.

height (temperature) were 2.3 km (-17°C) and 1.3 km (-10°C), respectively. Rapid change in temperature and humidity profiles (solid and dashed line in the left-most panel) showed a good correspondence with the top of supercooled cloud droplet layer. After subtracting the artificial temperature increase due to latent heat release by the freezing of cloud droplets on a temperature sensor between 1.8 and 2.3 km, the sounding shows that a whole layer in the snow cloud except for upper parts of the cloud was convectively unstable. Across the cloud top,

there was also a rapid decrease in equivalent potential temperature, even after subtracting the artificial temperature increase. This rapid decrease in θ_e across the cloud top could be related to cloud top entrainments and downdraft initiation in snow clouds. These issues will be discussed elsewhere. Comparison among successive dropsonde soundings inside and outside the snow cloud showed that temperatures in the convective cloud were $1\text{--}2^{\circ}\text{C}$ higher than those outside convective clouds. The maximum updraft velocity was estimated to be $4\text{--}5\text{ m s}^{-1}$ from

Fig. 8. PPI of radar reflectivity at 1503 4 Feb. 1989. The cross mark indicates the position of the HYVIS.

the descent velocity of the HYDROS, although as pointed out by Asai (1968), errors of $\sim \pm 1 \text{ m s}^{-1}$ were inherent in the estimates because of changes in descent (ascent) velocity due to deformations of the parachutes (balloons) and the loading of hydrometeors on them.

Figure 6 shows an example of hydrometeor images taken at -14 , -13 and -10°C . The width of each panel is approximately 1.5 mm. Numerous black dots and circles with a white opening in the center are supercooled cloud droplets. In addition, a stellar crystal (in upper left corner of Panel a), a columnar crystal (in middle left of Panel c) and an hexagonal plate (in lower left corner of Panel d) are seen.

The maximum cloud water content was 0.7 g m^{-3} and existed near the cloud top, while ice water contents were around 0.01 g m^{-3} and confined to supercooled cloud droplet regions. Here, ice water contents were calculated using the mass ver-

sus maximum dimension relations derived by Davis (1974). The observed cloud water content showed good agreement with the adiabatic liquid water content (0.7 g m^{-3}). Also, the observed cloud base height was coincident with the calculated lifting condensation level. Number concentrations of cloud droplets were almost uniform and approximately $100 \text{ particles cm}^{-3}$. Their size distribution had a peak around $15\text{--}20 \mu\text{m}$ in diameter and a broad tail extending up to $50 \mu\text{m}$ (Fig. 7). The average number concentration of ice crystals was several particles L^{-1} and was one order of magnitude greater than that expected from Fletcher's (1962) equation. However, large precipitation particles ($d \geq 0.2 \text{ mm}$) were not observed in this cloud (at least along the HYDROS descent path).

Two case studies (including the case described above) showed that microphysical and thermodynamic structures of snow clouds at developing stages

Fig. 9. The same as Fig. 4, except for at 1501 4 Feb. 1989.

were characterized by (1) the highest supercooled cloud water content ($\sim 1 \text{ g m}^{-3}$, which is close to adiabatic values) through the lifetime of individual clouds, (2) a significant number (several particles per liter) of ice crystals which have not yet grown to precipitation size, (3) a large extent of convectively unstable layer in the clouds, (4) a moderate updraft velocity of $\sim 4 \text{ m s}^{-1}$.

3-3. Snow cloud in the mature stage

Snow cloud observed with the HYVIS at 1501 4 Feb. 1989 showed a typical microphysical structure of convective snow clouds at the mature stage. Figure 8 shows the PPI of radar reflectivities at around the HYVIS launch. The cross mark in the figure indicates the position of the HYVIS. When the HYVIS was launched, radar reflectivities of the snow cloud were rapidly increasing and exceeded 25 dBZ immediately after the launch. At the same time, graupel started to reach the ground surface. From these radar and surface observations, the snow cloud observed with the HYVIS is considered to be at early mature stage.

As seen from Fig. 9, the cloud top height and temperature were $\sim 2.9 \text{ km}$ and -21°C . By this stage, convectively unstable layer had disappeared and the stratification in the whole cloud became stable. The temperature dip just above the cloud top was due to sublimation of rime on a temperature sensor. Even after subtracting this effect, however, there still seems to be a slight decrease in equivalent potential temperature across the cloud top. Sounding data outside the snow cloud were not available for this case, but comparisons between inside and outside snow clouds for other cases showed that temperatures inside snow clouds were still warmer than outside by $1\text{--}2^\circ\text{C}$ above cloud base and colder than outside below cloud bases. The cold air beneath cloud bases may have been caused by the evaporative cooling of precipitation particles and/or by the downdraft originating from upper-level ambient air with lower equivalent potential temperatures near the cloud tops. The maximum updraft velocity estimated from the ascent velocity of the HYVIS was $1\text{--}2 \text{ m s}^{-1}$, although this value was marginal because of rather large errors ($\sim \pm 1 \text{ m s}^{-1}$) in vertical

Fig. 10. Images of various types of hydrometeors taken in the snow cloud at the mature stage. The width of Photo (a) is approximately 2 cm while that of Photos (b), (c) and (d) is 1.5 mm.

velocity estimates.

Figure 10 shows hydrometeor images taken by the HYVIS. Panel (a) shows graupel particle observed in the lower parts of the snow cloud, Panel (b) a frozen drop, Panel (c) supercooled cloud droplets and pristine crystals. Panel (c) provides evidence that columns and thick plates have already started to rime before their size reaches $100\ \mu\text{m}$.

Cloud water contents (histograms in the left panel of Fig. 11) were roughly $0.1\ \text{g m}^{-3}$ in the upper and middle parts of the snow cloud, which is much less than the adiabatic cloud water content ($\sim 1.2\ \text{g m}^{-3}$). The precipitation processes started to erode away the lower portion of cloud water and the observed cloud base height of 1.1 km was significantly higher than the calculated lifting condensation level (800 m). Ice water contents (ice crystals $d < 200\ \mu\text{m}$; dotted line in the left panel) were roughly $0.05\ \text{g m}^{-3}$. High ice water contents existed in regions with supercooled cloud water. Snow water contents (dashed line in the left panel) in-

creased with decreasing height and their maximum value reached $0.2\ \text{g m}^{-3}$. Supercooled and frozen drops of $100\text{--}200\ \mu\text{m}$ in diameter were observed at heights of 200–400 m. These drops were considered to form through a collision-coalescence process among cloud droplets. Such supercooled drizzle drops were found also in upslope clouds (Rasmussen *et al.*, 1992), maritime stratocumulus (Rangno and Hobbs, 1991), orographic storms (Politovich, 1989), and warm-frontal clouds (Murakami *et al.*, 1992).

The number concentrations of cloud droplets (histograms in the right panel of Fig. 11) were roughly $10\ \text{particles cm}^{-3}$, while number concentrations of ice crystals were high in the supercooled cloud droplet layer and their maximum value was $\sim 300\ \text{particles L}^{-1}$. Number concentrations of snow particles ($d \geq 200\ \mu\text{m}$) were $\sim 10\ \text{particles L}^{-1}$, and those of drizzle drops were less than $100\ \text{particles m}^{-3}$. Figure 12 shows size distributions of cloud droplets and snow particles at upper, middle and lower levels of the snow cloud. Cloud droplets had a broad

Fig. 11. Vertical changes in hydrometeor concentrations. (A) water content of cloud droplets (Q_c , rectangles), ice crystals (Q_i , dotted line), snow particles (Q_s , dashed line) and drizzle drops (Q_r , solid line); (B) number concentration of cloud droplets (N_c), ice crystals (N_i), snow particles (N_s) and drizzle drops (N_r) for the snow cloud at mature stage.

size distribution with a peak near $20 \mu\text{m}$ in diameter. No marked difference among size distributions at the three levels was found. On the other hand, precipitation particles continued accretional and depositional growth at the expense of cloud droplets and their size distributions broadened toward larger size with decreasing height. The larger particles consisted of graupel and heavily rimed snow particles.

General microphysical and thermodynamic structures of convective snow clouds at mature stages derived from this case and others were characterized by (1) a rapid decrease in cloud water contents (down to $0.1\text{--}0.3 \text{ g m}^{-3}$) and partial eroding of lower portions of cloud water regions due to depositional and accretional growth of ice crystals and snow particles, (2) an active ice nucleation and consequent high concentrations of ice crystals in upper parts of snow clouds, (3) snow particles with large fall velocities such as graupel particles (up to $\sim 5 \text{ mm}$) and heavily rimed snow particles, (4) no convectively unstable layer in snow clouds, (5) a weak updraft velocity of $1\text{--}2 \text{ m s}^{-1}$.

3-4. Snow clouds at the decaying stage

As an example of snow clouds at the decaying stage, we describe the microphysical and thermodynamic structures of the snow cloud observed at 1526 3 Feb. 1989.

Cloud top height and temperature were $\sim 2 \text{ km}$ and -20°C . The stratification was stable in the whole cloud except for the lowest layer. A convectively unstable layer was seen in the lowest layer, 500 m deep, where sensible and latent heat fluxes from sea surface restarted to destabilize the layer. The temperature excess in the cloud was thus eroded, making the temperature inversion above the cloud top more definite (see Fig. 13).

Water contents (left panel) and number concentrations (right panel) of ice crystals and snow particles are shown in Fig. 14. All supercooled cloud droplets were depleted through depositional and accretional growth of ice crystals and snow particles. Snow water contents were rather high in the upper and middle parts of snow clouds, and their maximum value was $\sim 0.2 \text{ g m}^{-3}$. Here, the number concentrations were $\sim 10 \text{ particles L}^{-1}$, with only

Fig. 12. Vertical changes in size distributions of cloud droplets (left) and snow particles (right) for the snow cloud at its mature stage.

a few particles L^{-1} in the middle and lower parts. These snow particles consisted of unrimed dendrites and hexagonal plates with small fall velocities. Ice water contents were less than 0.01 g m^{-3} and negligibly small. This means that active ice nucleation had ceased.

General features of snow clouds in their decaying stages, derived from three cases (including the case described above), are characterized by (1) absence or slight presence of supercooled cloud water, (2) unrimed or slightly rimed snow crystals with small fall velocities, (3) re-commencement of destabilization in the lowest layer through heat and moisture supply from sea surface (in the late decaying stage).

4. Discussion

4-1. Ice nucleation

Ice nucleation is a very important process in understanding snow cloud evolution because the position, timing and rate of ice initiation in clouds determine the dominant shapes of snow particles

and the timing and position of snowfall reaching the ground. However, mechanisms of ice nucleation are not well understood qualitatively nor quantitatively, even now. Although we did not measure ice nuclei in our field project, we approach this problem by analyzing the spatial distributions of pristine ice crystals ($d < 200 \mu\text{m}$).

Figure 15 shows the relationship between the number concentrations of pristine crystals ($d < 200 \mu\text{m}$) in clouds and the air temperature. Solid and open circles indicate the concentration of ice crystals in and out of supercooled cloud water regions, respectively. High concentrations of ice crystals were observed in supercooled cloud water regions, although they do not show any significant temperature dependency. However, Figure 15 does not deny the well-known relationship between the maximum ice crystal concentration in clouds and cloud top temperature. Looking at this relationship (not shown here), our data showed a significant dependency of ice crystal concentrations on cloud top temperatures, that is, an increase of ice crystal concentration with decreasing cloud top temperature. Thus, ice nucleation was a function of both cloud water content and air temperature but appeared to be more strongly dependent upon the former.

These observational results suggest that dominant mechanism in the snow clouds was freezing of cloud droplets (see Murakami *et al.*, 1994). In the clouds studied here, the Hallett and Mossop's (1974) rim-splintering mechanism did not seem to apply because most parts of the clouds were in temperature regions colder than -8°C . However, there still remains the possibility of some secondary nucleation associated with the riming process, or fragmentation resulting from collisions between solid precipitation particles.

To investigate ice nucleation mechanisms in more detail, we need to accumulate the data on spatial and temporal distributions of pristine ice crystals in snow clouds and to make *in situ* measurements of ice nuclei.

4-2. Precipitation mechanisms

Graupel formation is generally the most efficient way to remove water substance from snow clouds. Also, in the snow clouds studied, accretional (riming) growth of snow particles leading to graupel particles was a dominant precipitation mechanism at mature stages. On the other hand, there was also some evidence that the warm rain process (drizzle formation due to coalescence of cloud droplets) worked in some snow clouds. However, its contribution to the total precipitation seemed to be small (less than 10 %) according to the HYVIS observations.

Drizzle drops also play an important role in graupel formation. It is well known that both rimed crys-

Fig. 13. The same as Fig. 4, except for the snow cloud at its decaying stage.

tals and frozen drops can become graupel embryos, depending on the atmospheric conditions. Harimaya (1977, 1988) examined thin sections of graupel particles that reached the ground in the Hokkaido and Hokuriku districts and confirmed that graupel particles include both types of embryos. However, no *in situ* measurements have been made to identify graupel embryos in snow clouds over the Sea of Japan, to our knowledge. Hydrometeor ($d < 200 \mu\text{m}$) images taken in graupel-forming snow clouds showed that columns and thick plates (so-called isometric crystals) were dominant (see Fig. 3 of Matsuo *et al.*, 1994) and they already started to rime before their size reached $100 \mu\text{m}$. The number of supercooled and frozen drops observed was much less than that of these isometric crystals. These observational results suggested that rimed columnar and thick plate crystals were the most promising candidates for graupel embryos. However, drizzle drops are supposed to form in supercooled cloud water regions and fall out immediately after their formation before depositional and accretional growth of ice crystals becomes effective. Thus, the existence of supercooled and frozen drops is limited in space and time, and we may have missed high concentrations of these drops in HYVIS and HYDROS observations. Therefore we reserve the possibility for frozen drops to act as the primary graupel embryos.

In snow clouds in their decaying stages, unrimed and lightly rimed dendritic and plate-like crystals were predominant and aggregates of these crystals were also commonly found. However, these aggre-

gates were not large ones but small ones, consisting of several crystals. Since the fall velocities for such small aggregates are, in general, not appreciably different from those for the individual crystals that comprise them (see Kajikawa, 1974 and 1989), it is considered that the precipitation rate was not significantly enhanced by the aggregation process.

4-3. Production and depletion of supercooled cloud water

As mentioned previously, we have not successfully observed the whole life cycle of individual snow clouds. Therefore, from only the composite of separate snow clouds at different stages of life cycles described in Section 3, it is hard to tell when the significant depletion of supercooled cloud water starts and how fast it proceeds. Such detailed features of the depletion are analytically investigated in this sub-section.

Considering an air parcel ascending vertically and neglecting the effects of turbulent diffusion and entrainment, the rate of change of the cloud water mixing ratio in snow clouds is expressed by the following equation:

$$\frac{\partial q_c}{\partial t} = -\frac{dq_{vs}}{dt} - w \frac{\partial q_c}{\partial z} - \frac{1}{\rho_a} VD - q_c CVOL. \quad (1)$$

Here, q_{vs} is the saturated water vapor mixing ratio, w the vertical velocity, ρ_a the density of air, VD the depositional growth rate of solid particles, and $CVOL$ the sweep rate by settling solid particles. The first term of the right hand side is the

Fig. 14. Vertical changes in hydrometeor concentrations. (left panel) water content of ice crystals (Q_i , dotted line) and snow (Q_s , dashed line); (right panel) number concentration of ice crystals (N_i) and snow (N_s).

production of excess vapor due to ascending motion (adiabatic cooling), the second the vertical advection of cloud water, the third the depletion of excess vapor through depositional growth of solid particles (ice crystal, snow and graupel), and the fourth the depletion of cloud water due to accretional growth of solid particles. The first and third terms should be considered together and the sum of these two terms is the net production rate of excess vapor. If it is positive, new cloud water formation (condensation) occurs in the parcel. On the contrary, if it is negative, cloud water (droplet) evaporation occurs.

Figures 16 and 18 show the vertical distributions of excess vapor depletion rates due to depositional growth VD of solid particles for snow clouds at developing stages (2 cases) and mature stages (6 cases), respectively. Figures 17 and 19 show the sweep rates $CVOL$ by settling solid particles (fraction of air volume to be swept by solid particles in unit time) at the developing and the mature stages, respectively. These rates are calculated based on the observed size distribution of solid particles averaged over 250 m in the vertical. In these figures, the vertical axis is a non-dimensional height normalized by cloud top height. Thick and thin dotted lines in Figs. 16 and 18 indicate excess vapor production rates due to ascending motions of air parcel with a maximum updraft velocity $w_{\max}$ of 1 m s^{-1}

and 4 m s^{-1} , respectively. As seen from Figs. 18 and 19, at the mature stage, the excess vapor depletion rates and sweep rates ranged from 1×10^{-3} to $4 \times 10^{-3} \text{ g m}^{-3} \text{ s}^{-1}$ and from 3×10^{-4} to 3×10^{-3} , respectively. On calculating the excess vapor production rate, an imaginary snow cloud with a cloud top temperature (height) of -20°C (2.5 km), a cloud base height of 0.8 km, and a surface temperature of 0.3°C is considered. The vertical profile of updraft velocity in the cloud is assumed to be a parabolic shape with a maximum updraft velocity at a height of 1.25 km and $w = 0$ at both the cloud top and the ground surface. Assuming that q_c increases linearly with height except near cloud top, the rate of change of the supercooled cloud water content in middle levels, which are below the height of maximum q_c , would be approximated by

$$\frac{\partial q_c}{\partial t} = -\frac{dq_{vs}}{dt} - w \frac{q_c}{\Delta h} - \frac{1}{\rho_a} VD - q_c CVOL, \quad (2)$$

where the density of air ρ_a and Δh are specified to be 1 kg m^{-3} and 1000 m, respectively. The calculations are performed for four sub-stages; developing, transition, early mature and late mature stages. Values of $w_{\max}$, dq_{vs}/dt , VD and $CVOL$ at each sub-stage are specified as shown in the third to fifth columns of Table 2 on the basis of observational results. Here a developing sub-stage is defined as the period from

Fig. 15. Relationship between air temperatures and concentrations of ice crystals ($d < 200 \mu\text{m}$) in (solid circles) and out (open circles) of supercooled cloud droplet regions. Dashed and dotted lines indicate linear regressions in and out of the supercooled cloud droplet region.

the time when adiabatic cloud water contents are established in a whole cloud, with a thermal reaching the cloud top, to the time when a significant depletion of q_c due to depositional and accretional growth of solid particles starts. Equation (2) is solved using values in the second to fifth columns of Table 2, where these values are assumed to be constant during each sub-stage. Results of calculations are also shown in the sixth and seventh columns. Here, t_c is the characteristic time (e-folding time) with which q_c approaches an asymptotical value $q_{c \rightarrow \infty}$ shown in the right-most column. Δt is the time needed for q_c to be completely depleted. For reference, time changes in q_c through four sub-stages are calculated, starting with 1 g kg^{-1} at the developing sub-stage. No observational data regarding time periods of each sub-stage are available, because the whole life cycle of individual snow clouds has not been successfully observed yet. In time integration, therefore, the characteristic time t_c is tentatively used as the time duration of each sub-stage. This choice is not far away from the consideration of the growth time of solid particles or the observed radar echo's growth time. However, the results of calculations must be interpreted within the framework of sampling limitation and assumptions mentioned above.

In the developing sub-stage, the first term on the right hand side of Eq. (2) is much greater than the third and fourth terms so that the depletion of q_c due to depositional and accretional growth of solid particles is negligibly small. Therefore, vertical distributions of cloud water contents are predominantly determined by the adiabatic condensation process

due to ascending motion of an air parcel (except near cloud tops, where q_c decreases with height due to entrainment of dry environmental air). In other words, q_c distributes vertically so as to satisfy the following equation:

$$-\frac{dq_{vs}}{dt} = w \frac{\partial q_c}{\partial z} \left(\text{or } -\frac{\partial q_{vs}}{\partial z} = \frac{\partial q_c}{\partial z} \right). \quad (3)$$

In the transition sub-stage, the first term is still the dominant factor and the net excess vapor production rate (the first term plus the third term) is positive. Therefore, cloud water is newly produced in the ascending air parcel. On the other hand, unlike the developing sub-stage, the depletion of q_c through depositional and accretional growth of solid particles can not be ignored. Here, q_c approaches a new asymptotical value of 0.7 g kg^{-1} with a time constant of $\sim 230 \text{ sec}$ and would reach 0.8 g kg^{-1} in 230 sec .

In the early mature sub-stage, the contributions of the first and third terms become equal and the net production rate of excess vapor in the ascending air parcel is zero, which means that q_c production through condensation no longer occurs. Thus q_c decreases and approaches 0.0 g kg^{-1} with a time constant of $\sim 330 \text{ sec}$. In 330 sec , q_c would decrease from 0.8 to 0.2 g kg^{-1} .

In the late mature sub-stage, the third and fourth terms are dominant factors and the net production rate of excess vapor becomes negative. That is, the ascending air parcel can not maintain the water-saturation condition because of rapid depositional

Fig. 16. Vertical distributions of excess vapor depletion rates due to depositional growth of ice crystals and precipitation particles at the developing stages.

Fig. 17. Vertical distributions of sweep rates by ice crystals and precipitation particles at the developing stage.

growth of solid particles. Evaporation of cloud water and accretional growth of solid particles cause a rapid decrease (with time constant of 250 sec) in q_c and cloud water would be completely depleted in ~ 60 sec.

As seen from the rough estimates mentioned above, in snow clouds with a cloud top temperature of -20°C , it takes $\sim 1 \times 10^3$ sec for q_c that has attained its maximum value to be completely depleted through depositional and accretional growth of solid particles. It is noteworthy that even in the existence of a weak updraft ($w_{\max} = 1 \text{ m s}^{-1}$), cloud water is completely depleted, that is, eroding of cloud water commences.

5. Conclusions

We had studied convective snow clouds over the Sea of Japan for four winter seasons from 1989 through 1992. Snow clouds of various forms (L-mode, T-mode, isolated, *etc.*) at various stages of their life cycles were observed by using the HYVIS, the HYDROS, multiple Döppler radars, and a microwave radiometer.

Based on the HYVIS and HYDROS observations, microphysical and thermodynamic structures of short-lived convective snow clouds with cloud top temperatures of $-20^\circ\text{C} \pm 3^\circ\text{C}$ at their developing, mature and decaying stages are summarized as follows. (see Fig. 20)

At their developing stages, a high concentration

Fig. 18. The same as Fig. 16, except for at mature stages.

Fig. 19. The same as Fig. 17, except for at mature stages.

of supercooled cloud droplets occupied the whole cloud. Their maximum water contents of $\sim 1 \text{ g m}^{-3}$ were close to the adiabatic liquid water content and their number concentrations were $\sim 100 \text{ particles cm}^{-3}$. Number concentrations of ice crystals ($d < 200 \mu\text{m}$) were several particles L^{-1} but few precipitation particles (snow and graupel; $d \geq 200 \mu\text{m}$) had yet appeared.

At their mature stages, ice nucleation processes (probably freezing of cloud droplets) were still active and ice crystal concentrations sometimes exceeded $100 \text{ particles L}^{-1}$. Subsequent depositional and accretional growth had produced heavily rimed snow crystals and graupel particles of $\sim 10 \text{ particles L}^{-1}$. These precipitation particles and numerous ice crystals had rapidly depleted a large amount of su-

percooled cloud droplets and their water contents and number concentrations were $\sim 0.2 \text{ g m}^{-3}$ and $\sim 10 \text{ particles cm}^{-3}$. Supercooled and frozen drops ($100 \mu\text{m} < d < 200 \mu\text{m}$) sometimes coexisted with heavily rimed snow crystals and graupel particles.

At their decaying stages, almost all supercooled cloud droplets were depleted, whereas a considerable amount of unrimed and lightly rimed snow crystals were still suspended in the upper and middle parts of clouds.

Kinematic and thermodynamic structures in snow clouds also evolved with life cycles. Moderate updrafts of $\sim 4 \text{ m s}^{-1}$ at the developing stage weakened to $1\text{--}2 \text{ m s}^{-1}$ by the mature stage. The excess temperature of $1\text{--}2^\circ\text{C}$ in the convection cores at developing and mature stages became eroded by

Table 2. Parameters for cloud water budgets and the depletion rates of cloud water contents.

Sub-stage	w (m s^{-1})	$\frac{dq_{vs}}{dt}$ ($\text{g kg}^{-1} \text{s}^{-1}$)	VD ($\text{g m}^3 \text{s}^{-1}$)	$CVOL$	$t_c(\text{sec})$ ($\Delta t(\text{sec})$)	$q_{ct \rightarrow \infty}$ (g kg^{-1}) $q_{ct} - q_{ct+t_c}$ (g kg^{-1})
Developing	4	4×10^{-3}	1×10^{-4}	2×10^{-5}	250	0.98 ($1 - 0.97$)
Transition	4	4×10^{-3}	1×10^{-3}	3×10^{-4}	230	0.7 ($0.97 - 0.78$)
Early mature	2	2×10^{-3}	2×10^{-3}	1×10^{-3}	330	0.0 ($0.78 - 0.2$)
Late mature	1	1×10^{-3}	4×10^{-3}	3×10^{-3}	250 (60)	-0.75 ($0.2 - 0.0$)

$q_{ct \rightarrow \infty}$: an asymptotical value of q_c

$q_{ct} - q_{ct+t_c}$: time change in q_c during characteristic time t_c or Δt

Fig. 20. Schematic of the evolution of microphysical structures in snow clouds.

the decaying stage. Convectively unstable stratification through a whole cloud at its developing stage changed to stable stratification at the mature stage, followed by destabilization of the lower layer through heat and moisture supply from the sea surface at the late decaying stage.

High concentrations of ice crystals were observed in supercooled cloud water regions, which suggests that ice nucleation in the snow clouds was strongly related to the existence of supercooled cloud droplets and the most likely mechanism was the freezing of cloud droplets. These ice crystals continued depositional and accretional growth at the

expense of supercooled cloud droplets and produced heavily rimed snow crystals and graupel particles. This is the primary precipitation mechanism in the snow clouds. However, the existence of supercooled and frozen drizzle drops in snow clouds at their developing and mature stages indicates the possibility of a warm rain process (drizzle formation through collision-coalescence between cloud droplets) and of frozen drizzle drops serving as graupel embryos, although their contribution seemed to be small because of their low concentrations and limited spatial and temporal distributions in snow clouds.

Computations of cloud water budgets indicate

(a)

(b)

Fig. A1. The same as Fig. 2, except for snow clouds observed in 1989(a), 1990(b), 1991(c) and 1992(d) field projects. (a) indicates aggregates of snow crystals.

(c)

(d)

Fig. A1. (Continued)

that, at the developing stage, the excess vapor production term due to adiabatic ascent of air parcel is much greater than any other depletion terms and cloud water contents become close to the adiabatic liquid water content. As the updraft decreases and the size and concentration of ice crystals and precipitation particles increases with time, snow clouds approach a new steady (equilibrium) state with lower water contents. Once precipitation is well developed, clouds, even with an updraft velocity of 1 m s^{-1} , can not maintain a steady state and precipitation particles deplete the supercooled cloud water away within a thousand seconds.

Acknowledgments

This study is a part of "Study of Precipitation Formation in Snow Clouds and Feasibility of Snow Cloud Modification by Seeding" that is financially supported by the Japan Science and Technology Agency. They are also indebted to Prof. T. Harimaya, Hokkaido University, and three anonymous reviewers for their constructive comments.

Appendix

A schematic of all snow clouds listed in Table 1 is shown in Fig. A1. Since the erosion of supercooled cloud water starts from lower parts of clouds (cloud base), supercooled cloud droplets usually existed in upper and middle parts of clouds. However, they did not necessarily show single-layer structures but multiple layer structures. This may reflect the multicell structures of snow clouds or complicated inner structures of snow clouds due to entrainment of dry ambient air. Also note that supercooled cloud water in well-organized snow bands (*e.g.*, 1158 and 1445 7 Feb. 1990, 1451 12 Feb. 1990 and 1441 9 Feb. 1992) distributed in a deep layer although their water contents were not so high ($< 0.5 \text{ g m}^{-3}$). Detailed microphysical structures and precipitation mechanisms in such well-organized snow bands will be presented in future papers.

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日本海降雪雲のメソスケール及びマイクロスケールの構造 第一部：寿命の短い対流性降雪雲の微物理構造の変化

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1989年から1992年の4冬期間に、雲粒子ゾンデ・雲粒子ドロップゾンデ・ドップラーレーダ・マイクロ波放射計を用いて、日本海上の対流性降雪雲の観測を行った。この論文では、雲の一生の中でステージの異なる雪雲の観測例を多数コンポジットし、雲頂温度 $-20 \pm 3^\circ\text{C}$ の比較的寿命の短い対流性降雪雲内の微物理構造の変化を調べた。

発達期には中程度($\sim 4\text{m/s}$)の上昇流によって雲全体に断熱凝結量に近い高濃度の過冷却雲水が生成される。このとき氷晶数濃度($200\text{ }\mu\text{m}$ 以下)は10個/リットル程度で降雪粒子($200\text{ }\mu\text{m}$ 以上)はまだ形成されていない。最盛期には、氷晶は時々100個/リットルを越える高濃度となり、アラレや濃密雲粒付雪結晶からなる降雪粒子ができ、その濃度は10個/リットル程度となる。これらの降雪粒子の昇華及びライミング成長により、過冷却雲水のかなりの部分が消費される。衰退期までには、ほとんどの過冷却雲水が消費され、雲内には全くライミングしていないか、又は軽くライミングした雪結晶が残る。

降雪機構としては、過冷却雲粒の共存下で発生した氷晶(特に厚角板や角柱などの軸比が1に近い結晶)が、昇華・ライミング成長を続け、最終的にアラレになる機構が主なものである。一方、暖かい雨の形成機構が働いていることや、凍結水滴がアラレの芽となっている可能性を示唆する結果も得られているが、過冷却及び凍結水滴の数濃度が低いこと、また分布が時空間的に限定されていることから、その寄与は小さいものとみられる。

雲水量の収支計算から、発達期には、気塊の断熱上昇による過剰水蒸気生成項が卓越しており、雲水量は断熱凝結量に近い値となるが、一端降雪が強くなると、 1m/s 程度の上昇流を含む雲でさえ定常状態を維持できなくなり、降水粒子の昇華・ライミング生成に消費され、過冷却雲水は急速に減少・消失することが示された。