The Response of 36- and 89-GHz Microwave Channels to Convective Snow Clouds over Ocean: Observation and Modeling

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(Manuscript received 29 May 1999, in final form 21 March 2000)

ABSTRACT

The first study in both observing and modeling radiative properties of snow clouds in the microwave frequencies is described in this paper. Snow clouds over ocean were observed simultaneously using an airborne microwave radiometer and an X-band Doppler radar. Results show that brightness temperatures at 36- and 89-GHz microwave channels responded well to the horizontal variations of precipitation particles and to the cloud dynamic structures determined by the Doppler radar, which reflect the development stages of convective cells. For the quantitative validation, physical retrievals of liquid water and snow water amounts were performed using a radiative transfer model. The retrieved snow water amount agrees well with the observed snow water amount that was converted from observed radar reflectivity. In the retrieval method, the model-simulated brightness temperatures were able to match the observed values within 3 K per channel for the most part. The ambiguities of the retrieved parameters that depend on some assumptions are also examined.

1. Introduction

Physical parameters measured from satellites have been increasingly utilized in many areas of meteorology. Among various satellite-borne sensors, microwave radiometers are particularly notable because of the potential in measuring such physical parameters as air temperature, precipitation, and cloud water amount. The recent installation of the Special Sensor Microwave Imager (SSM/I) aboard Defense Meteorological Satellite Program (DMSP) satellites has achieved many successes in quantitatively estimating the amounts and distributions of rain, cloud water, and surface wind speed (e.g., Liu and Curry 1992; Kummerow and Giglio 1994a,b; Prigent et al. 1994).

The major technical improvements of SSM/I compared to its predecessors include more accurate temperature calibration and the addition of two 85-GHz channels. While accurate calibration is essential in satellite measurements, the additional 85-GHz channels are particularly important to cloud and precipitation retrievals because of their high sensibly to scattering by precipitation-size particles and their high spatial resolution [the effective field of view (EFOV) is 12.5 km]. The scattering signatures are useful in investigating the structure of mesoscale cloud systems, especially the upper part of the clouds, which consists of ice-phase particles that can hardly be detected by microwave radiometers at lower frequencies (≤36 GHz).

Ice-phase processes play important roles in cloud physics by influencing the vertical redistribution of energy and water, especially in the middle and upper troposphere in the tropical and extratropical regions. For example, the “seeder–feeder mechanism” (Hobbs et al. 1980; etc.) is maintained by the seeding of ice-phase particles and the feeding through riming of supercooled cloud droplets: both mechanisms involve ice-phase pro-
cesses. Ice-phase processes are also important in the stratiform part of a mesoscale convective system (MCS), in which precipitation particles are generated above the melting layer. Takahashi and Uyeda (1995) studied convective systems observed during Tropical Ocean and Global Atmosphere Coupled Ocean–Atmosphere Response Experiment (TOGA COARE) Intensive Observation Period (IOP), and showed that the reinforcement and maintenance processes above freezing level also take place in the convective portion of a tropical MCS, not only in the stratiform portion. These studies suggest that understanding of the overall structure of the MCSs requires detailed knowledge of the processes occurred in the frozen precipitation layer (the layer above freezing level).

Microwave radiometry measurements may contribute to these studies by determining the spatial and temporal distributions of hydrometeors (snow, graupel, liquid droplets, etc.). However, many uncertainties remain in the radiative transfer models used in physical retrievals of these hydrometeors: the model contains various assumptions about the effect of a frozen precipitation layer on microwave radiation. The Cooperative Huntsville Meteorological Experiment (COHMEX) was designed to resolve such problems (Vivekanandan et al., 1990; Fulton and Heymsfield 1991; Adler and Hakkarinen 1991; etc.). Results of the experiment yielded good agreements between the upward microwave radiation measured by airborne microwave radiometers and measurements obtained by a dual-polarized, dual-wavelength radar system.

In this study, we examine microwave radiometer and radar data observed for snow clouds over ocean. The clouds in COHMEX are different from those in the current study in two ways: COHMEX clouds were over land while our clouds are over ocean; COHMEX clouds included layers below freezing level while our clouds are entirely above freezing level.

On the former difference, the upward radiation from ocean at an oblique viewing angle is strongly polarized while from dry land it is weakly polarized. Depolarization then occurs due to the overlying atmosphere on the oceanic background. Observations showed that the depolarization reflects the existence of aqueous substances (Spencer 1986; Petty and Katsaros 1990; etc.). Additionally, convective clouds observed in COHMEX are not simple enough to allow us to study the frozen precipitation layer alone. They include a melting layer and a rain layer below it. Consequently, the upward radiation was also contributed by the raindrops and cloud water in the warm (>0°C) layer.

Snow clouds such as those observed in this study are ideal clouds for studying the effect of the frozen precipitation layer because they consist of only a frozen precipitation layer; the upward radiation transferred through them has not been affected by melting or raindrops layers. Katsumata et al. (1998) showed that brightness temperatures at 85-GHz channels have clear skill in reflecting the horizontal distributions of cloud water and precipitation in snow clouds over oceanic areas. In this study, we examine qualitative and quantitative characteristics of snow clouds for the microwave radiation by observation and modeling; the observation was carried out simultaneously using an airborne microwave radiometer and an X-band Doppler radar, and the physical retrievals using radiative transfer simulation are performed to assist in interpreting the observational results. Instrumentation and data for the observation are described in section 2, and the results of the observation are discussed in section 3. Section 4 describes the result of the physical retrieval.

2. Instruments and data

a. Airborne microwave radiometer

The Airborne Microwave Radiometer (AMR) is a prototype simulator of the Advanced Microwave Scanning Radiometer (AMSR) aboard the Advanced Earth Observation Satellite-II (ADEOS-II), which is scheduled to be launched in 2001. AMR was built for algorithm developing and testing in prelaunch time frame. Brightness temperatures \( T_B \) are measured at 12 channels: horizontally and vertically polarized channels at six frequencies (6.925, 10.65, 18.7, 23.8, 36.5, and 89.0 GHz). These channels are to be used in AMSR. The four higher frequencies are also similar to those in the SSM/I.

The antenna horns of all the channels are fixed at a nadir angle of 54° to the back of the aircraft. This angle is analogous to the zenith angles in the SSM/I and the AMSR/ADEOS-II. The data collected are upward radiances along the line trailing the aircraft’s flight track. The size of the EFOV varies with frequency, flight altitude, and flight speed. In the case of the present study, for example, the aircraft flew at an altitude of 3.8 km and a speed of about 100 m s\(^{-1}\), resulting in an EFOV of about 1.5 km at sea level for both 36.5- and 89.0-GHz channels. Detailed specifications of AMR are listed in Table 1.

The first product of the observed parameter is in the form of \( T_B \). In addition to the \( T_B \), two simple parameters are introduced in this study for the recognition of the cloud property: polarization ratio \( P \) and scattering index \( S \), proposed by Petty (1994). The former \( P \) is determined as

\[
P_a = \frac{T_{BAV} - T_{RAH}}{T_{BAV,O} - T_{RAH,O}}.
\]

The subscript \( \lambda \) means frequency. The subscripts \( V \) and \( H \) means the vertically and horizontally polarization. The \( T_{BAV} \) with subscript \( O \) is the hypothetical brightness temperature for the same scene without cloud, that is, liquid and solid (ice) hydrometeor. In this study, these background \( T_{BAV} \) are adopted from the observed data on the cloud-free area. This parameter \( P_a \) is the indicator...
TABLE 1. Specifications of AMR.

<table>
<thead>
<tr>
<th></th>
<th>12 channels</th>
</tr>
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<tbody>
<tr>
<td>Number of channels</td>
<td>12</td>
</tr>
<tr>
<td>Frequency (GHz)</td>
<td>6.925</td>
</tr>
<tr>
<td></td>
<td>10.65</td>
</tr>
<tr>
<td>Passband (MHz)</td>
<td>18.7</td>
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<tr>
<td></td>
<td>23.8</td>
</tr>
<tr>
<td>Integration period (s)</td>
<td>36.5</td>
</tr>
<tr>
<td></td>
<td>39.0</td>
</tr>
<tr>
<td>Temperature resolution (K)</td>
<td>1.33</td>
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<tr>
<td>Polarization</td>
<td>Horizontal/vertical</td>
</tr>
<tr>
<td>Beamwidth (V, 3 dB) (°)</td>
<td>6.6</td>
</tr>
<tr>
<td></td>
<td>5.5</td>
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<td></td>
<td>6.3</td>
</tr>
<tr>
<td></td>
<td>6.0</td>
</tr>
<tr>
<td>Beamwidth (H, 3 dB) (°)</td>
<td>7.9</td>
</tr>
<tr>
<td></td>
<td>6.3</td>
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<tr>
<td></td>
<td>7.2</td>
</tr>
<tr>
<td></td>
<td>6.9</td>
</tr>
<tr>
<td>Sampling interval (s)</td>
<td>3.2</td>
</tr>
<tr>
<td>Dynamic range</td>
<td>30–340 K</td>
</tr>
<tr>
<td>Absolute accuracy of $T_B$</td>
<td>&lt;1 K</td>
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of the visibility of the sea surface for the frequency $\lambda$. The $P$ is equal to 1 in the cloud-free area and 0 where the cloud is thick enough to obstruct the emission from and the reflection by the sea surface. On the other hand, the latter parameter $S$ is determined as

$$S_\lambda = P_\lambda T_{BAV} + (1 - P_\lambda) T_C - T_{BAV},$$

where $T_C$ is the limiting brightness temperature as a hypothetical non-scattering liquid water layer becomes optically thick. In this study we adopted 247.5 K as the $T_C$, which is the weighted mean temperature by the ideal maximum liquid water content for each height (see section 4a for details to obtain the profile). This parameter $S_\lambda$ is used as the indicator of the degree of the scattering, which is caused by the ice-phase particles such as aggregates, graupel, and so on.

For the snow clouds on the Sea of Japan, Katsumata et al. (1998) used only 85-GHz channels of SSM/I because of the variation of the spatial resolution. In this study, we use both 89- and 36-GHz channels. The EFOV at 36-GHz channels is almost the same as that at 89-GHz channels for AMR, while the EFOV and sampling interval at similar frequencies in SSM/I are largely different. This allows us to assume that $T_{BA}$s at these two AMR frequencies come from the same volume of atmosphere and clouds. Using the two frequencies, the effect of the scattering and the emission could separate more clearly.

b. X-band Doppler radar

The X-band Doppler radar system of the National Institute of Earth Science and Disaster Prevention of Japan (NIED) was used for the observation. The exact frequency of the radar is 9.445 GHz. It measures reflectivity and Doppler velocity over a range of 60 km with a range gate of 250 m and a beamwidth of 1.25°. The antenna rotation speed and the azimuthal sampling interval were set to 1 rpm and 0.70°, respectively, in plan position indicator (PPI) scan mode. In range height indicator (RHI) scan mode, these two parameters were set to be one-half of those in PPI scan mode.

The radar was installed at the tip of a peninsula that projects into the Bay of Ishikari (see Fig. 1). The coverage of the NIED radar was within the range of the Kenashi-yama Radar (250-km range) of the Japan Meteorological Agency (JMA). To the north of the NIED radar site it was an oceanic area, being suitable for the purpose of this study. Moreover, the area around the Bay of Ishikari is a place where snow clouds occur very frequently during winter (Kikuchi et al. 1989; Yamada et al. 1996; etc.). The radar covered most of the bay area.

3. Observational results

a. Outline of the observation

The aircraft observation was carried out around noon LST 16 February 1996, which is 9 h ahead of UTC. On that day, the northern part of the Sea of Japan was covered by streak clouds (Fig. 2) that are a common cloud type during the period of cold air outbreak from the Eurasian continent during the winter monsoon.

The radar and aircraft observations were carried out extensively on one of the streaks, which extended across

![Fig. 1. Map of the area where observation was conducted. Land area is shaded. The coverage of the NIED radar (cross) over the sea is filled by lines. The Kenashi-yama radar (square) covers all of the area in this figure. Star indicates the location of Sapporo District Meteorological Observatory, which is the nearby aerological observation site.](image-url)
the targeted area from west to east. Figure 3, the close-up image of the observational area in Fig. 2, shows that the width of the targeted cloud was about 20 km. The aircraft flew six paths over the cloud both parallel and transverse to the streak as shown in Fig. 3. The aircraft flew over the cloud system at an altitude of 4.3 km, resulting in an EFOV of about 1.5 km at sea level.

The radar took a vertical cross section along the line by RHI scan, when the flight path crossing the streak was a straight line to or from the radar site. For the other paths, the radar was operated in PPI scan mode at multielevations (volume scan) to detect the three-dimensional distribution of the reflectivity and Doppler velocity.

b. Direct comparison between AMR data and RHI images

The aircraft reported that the maximum cloud-top height was about 3.5 km. On the temperature profile (Fig. 4), which was observed at 0000 UTC at the Sapporo District Meteorological Observatory (see Fig. 1 for the location), the cloud-top temperature was higher than 240 K, while the surface air temperature was about 266 K. In this air temperature, the precipitation particles were frozen, while the cloud water might still exist in the cloud.

These snowing streak clouds generally do not have
Fig. 5. Time series of JMA Kenashi-yama radar reflectivity fields from 0247 to 0404 UTC 16 Feb 1996. Crosses and circles are the location and observation range of NIED radar, respectively. Arrows in the middle panels are the direction of vertical cross section, shown in Fig. 7 (path 3) and Fig. 6 (path 4), respectively. Here, C1 and C2 mark the individual convective cells.

The uniform structure in detail along its direction of extension (see, e.g., Fujiyoshi et al. 1992; Fujiyoshi et al. 1998; etc.). The streaks in this case also did not have a two-dimensional structure. Figure 5 reveals that the convective cells aligned along the streak clouds at a horizontal scale of about 15 km, and moved from west to east at a speed of about 15 m s\(^{-1}\). In addition, the figure shows that the radar echo masses had an isolated and circular form in the western portion of the NIED radar observation area, while the form was unclear in the east. This suggests that the convective cells were developing in the western portion of the observation range of the NIED radar, while they were dissipating in the eastern portion.

These two characteristic states of convective cells are illustrated in the two cross sections (paths 3 and 4) of simultaneous observation by AMR and NIED radar. The positions of the two paths are shown in Fig. 5: path 3 is for the cell in the eastern portion, while path 4 is for the cell in the middle of the western and eastern portion. The AMR observed parameters and the corresponding vertical cross sections of radar data are shown in Figs. 6 and 7 for paths 4 and 3, respectively.

In the cross section for path 4 (Fig. 6), the radar reflectivity core (the area with the highest radar reflectivity) is located in the upper part of the echo area (at an altitude of 1.8 km) at a distance of 11 km from the radar. Around the echo core, the Doppler velocity field shows a horizontal convergence pattern below 1.0 km and a horizontal divergence pattern above, as shown by arrows in Fig. 6d. The pattern suggests that a strong updraft existed in the cell. These features, an updraft signature and a reflectivity core in the upper part of a cell, are characteristic of a convective cell in its developing stage.

The corresponding AMR data shows the symmetric pattern, with the peak around 15 km from the radar site. Because the high \(T_B\) and \(P\) do not correspond to the position of the convective cell, this pattern is likely to be caused by the background radiation, that is, the emission from and the reflection by the sea surface. However, a local but maximum drop of \(T_B\) (about 8 K) was observed in the upward radiation through the echo core (the blue broken line A indicates the point of maximum drop in Figs. 6a,b. In Fig. 6c it indicates the radiometer’s beam path). This drop well reflects the extinction of upwelling microwave energy caused by scattering. The scattering index \(S_{89}\) also shows a maximum value (about 10 K) at this point. The maximum radar reflectivity is about 25 dBZ at the core.

On \(P_{89}\), its local peak is at a distance of 11 km from the radar site (indicated by orange broken line B in Figs. 6a,b,c), right next to the line A. The local peak is also shown in \(P_{36}\) in the same place. This depolarization without the scattering signal on \(S\) or \(T_B\) indicates the existence of a large quantity of liquid water and possibly graupel, which is made among the cloud-water-rich part. The signal corresponds well to the existence of updraft in the cell, which is estimated using the Doppler velocity field. The asymmetric distribution of the parameters within the cell also suggests that the cloud water, and the updraft that makes the cloud water, existed only in the north (farther from the radar site) of the convective cell.

In comparison, the vertical cross section of the radar data on path 3 (Figs. 7c,d) shows that the reflectivity core, which is located 8–13 km from the radar site, was in the lower part of the convective cell. Around the core, the Doppler velocity field does not show a coupled pattern of low-level convergence and upper-level divergence. This indicates that the updraft in the cell had diminished, and most of the precipitation particles were falling to the sea surface. These characteristics correspond to that of a convective cell in the dissipating stage.

On the AMR data, it should be mentioned that the southern part of the radar echo core does not accompany the significant values in \(T_B\) and \(P\). This should be caused by the moving of the system (about 15 m s\(^{-1}\)) perpendicular to the aircraft track. The \(P_{89}\) shows a symmetric pattern like in path 4, with the peak at 12 km from the radar site. The trends of \(T_B\) and \(S_{89}\) are also the same.
around the convective cell; significant drop of $T_{89}$ were observed at 8–18 km from the radar site, with the peak at 12 km. The $S_{89}$ shows scattering signal in the same area. These suggest that the depolarization in the convective cell is caused principally by the scattering by precipitation particles, not by the absorption and the emission by liquid water. These characteristics correspond well to that observed by the Doppler radar in which no updraft was present to produce liquid water. The $S_{89}$ and the depression of $T_{89}$ from the value over the cloud-free area are 16 and 12 K, respectively. These values are larger than those at line A in path 4. The
maximum radar reflectivity (27 dBZ) is also larger. On 89-GHz data, the AMR observed parameter corresponds well with the radar reflectivity.

On the other hand, the 36-GHz signals show no significant peak at line C, where the parameters of 89-GHz channels show peak values. The peaks of $P_{36}$ and $S_{36}$ are not on the line C, though $P_{36}$ and $S_{36}$ around the echo core differs from that of cloud-free area. This frequency dependence is shown more clearly in Fig. 8, which shows the scatter plots of $T_B$, $S$, and $P$ for data collected during all 6 paths. In the figure, the parameters on line A and C are almost the same in 36-GHz channels, while they are different in 89-GHz channels. Because $S_{89}$ is similar to the depression of $T_{89HV}$, it is likely that the frequency dependence is largely caused by the scattering by ice-phase particles: much smaller ice-phase particles that are only effective to scatter 89-GHz radiation, not 36-GHz, existed in line C than in line A. The solid ice sphere with the diameter of about 1 mm have such characteristics from...
Katsumata et al. 2329

FIG. 8. Scatter diagram of observed parameters for the all paths on 16 Feb 1996, for (a) $T_B$, (b) $S$, and (c) $P$. Each dot in the diagram donates an average of 1.33 s of data. The data corresponding to the locations of lines A, B, C in Figs. 6 and 7 and cloud-free part described in the previous subsection are indicated by the shaded circle. Data collected over land or with large deviation of the antenna horn nadir angle (>1.0°) are excluded.

Grody (1993). It is reasonable that the convective cell in the dissipating stage, which mostly consists of aggregates and cloud ice, includes much more small ice particles than in the cell in the developing stage, which mostly consists of graupel and supercooled cloud waters.

On the other hand, the data at B show zero $S$ and low $P$ in both 89 and 36 GHz. This is the ideal response only to the cloud water. However, Fig. 6c shows that the radar echo existed along line B. This implies some difficulties on the direct evaluation of snow water amount from these parameters, though the stage of the development of the convection could be estimated from these parameters.

4. Validation using the physical retrieval

To apply the observed $T_b$ to the quantitative estimation of the snow water and the cloud water in snow clouds, a physical retrieval method is performed. The key factors of the method that include radiative transfer simulation are the initial condition of and the forcing to the vertical profile of hydrometeors. In this study, we assume that the rough estimation of the vertical profiles could be done already by using the analyses in the previous section. We use $T_b$ in 36- and 89-GHz channels for retrieval because of their high sensitivity to snow clouds and their smaller EFOV size as mentioned previously.

a. Description of the simulation and retrieval

Retrieval was performed by minimizing the sum of square errors between observed and calculated $T_b$ at the four channels in 36- and 89-GHz channels, using a quasi-Newtonian method of iteration. The calculated $T_b$ are the output from the radiative transfer simulation using a given pattern of the vertical profile of the hydrometeors. The amounts of hydrometers are adjusted in each iteration step to minimize error.

The radiative transfer calculation scheme proposed by Liu (1998) is adopted. The scheme uses a plane-parallel atmosphere and includes the effect of O$_2$ and H$_2$O for each atmospheric layer. The H$_2$O is categorized in the five forms: vapor, cloud water, cloud ice, rain, and snow. The category “snow” includes the ice-phase precipitation particles, that is, snow, graupel, and so on.

In this study for the snow clouds, the amounts of snow and liquid water are considered as the variables to be obtained by the retrieval. For the simplification, cloud ice is excluded from the calculation. Rain is also excluded because of the air temperature as shown in Fig. 4. The water vapor amount is fixed as the observed value by nearby radiosonde at the Sapporo District Meteorological Observatory (see Fig. 1 for the location) on 0000 UTC.

The radiative transfer model adopted in this study uses a four-stream discrete ordinate solution as the source term.
Fig. 9. Result of radiative transfer simulation used in the retrieval, using aerological profile above Sapporo at 0000 UTC 16 Feb 1996. The variation of (a) $T_{B_{89}}$, (b) $P_{89}$, (c) $T_{B_{36}}$, and (d) $P_{36}$ with the variation of liquid water path and snow water path are shown.

for scattering to simplify the calculation. In simulating snow clouds, however, scattering should be treated as one of the most important factors. On the accuracy of the model, Liu (1998) concluded that the difference between this model and the polarized 32-stream discrete ordinate model is less than 3 K. We believe that this accuracy is adequate for the application of this study. All hydrometeor particles are assumed to be sphericals to further simplify the calculation of scattering. The density of the snow particles is set at 0.1 g cm$^{-3}$.

The atmosphere is divided into some layers in the calculations. The geometrical thickness of each layer is set to 250 m below and 1000 m above the cloud top. The cloud top height is set at 3000 m. Temperature and pressure of each layer are determined by vertical interpolation of data collected at the nearest aerological observation, the same as for the amount of water vapor. The sea surface temperature is fixed to 278 K, which is adopted from the half-month average of the observed value, reported by the Japanese Maritime Safety Agency. The wind speed that affects the surface radiative characteristics is set to 10 m s$^{-1}$. The value is adopted from the average of the observed values at surface observation points nearby the coast.

Under the conditions described above, values of the vertically integrated amount of two hydrometeors, the
liquid water path (LWP) and the snow water path (SWP), are permuted to find the solution having the least error between observed and calculated $T_b$. In the calculations, LWP and SWP is distributed to each layer while keeping the vertical distribution pattern unchanged for an individual convective cell. For liquid water, the vertical distribution pattern is set to the adiabatic liquid water content (LWC) profile with the temperature forcing. The adiabatic LWC is determined by assuming that an air parcel with a relative humidity of 70% is lifted from the surface. The value of relative humidity is adopted from the average value observed around the snow clouds in the Sea of Japan by M. Murakami (personal communication, 1998). The value determines the cloud base at the fourth layer from the surface, which corresponds to the height of 750 m. The air temperature $T$ also forces the LWC as

$$LWC_{\text{maximum}}(Z, T) = \begin{cases} 
LWC_{\text{adiabatic}}(Z) & T > 253 \text{ K} \\
\frac{T - 233}{253 - 233}LWC_{\text{adiabatic}}(Z) & 253 \text{ K} > T > 233 \text{ K} \\
0 & T < 233 \text{ K}. 
\end{cases}$$

Figure 9 shows the simulated $T_b$ and $P$ at both 89- and 36-GHz for the certain range of LWP and SWP using the above specification. The results show that the observed $T_b$ and $P$ do not saturate for LWP and SWP. On the other hand, the 36-GHz parameters, especially $P_{\text{36}}$, have little dependence on snow water, and are not saturated for simulated LWP. These support that the combination of 89- and 36-GHz data could be used enough to retrieve LWP and SWP, especially LWP.

### b. Results of retrieval

The calculated results are shown in Fig. 10. By comparing them to the results of paths 4 and 3 in Figs. 6 and 7, it appears that the characteristics described in section 3 are well simulated by the model. LWP shows peak value in the cell at developing stage at line B of Path 4, while little LWP at line A of path 4 and throughout path 3. The SWP is the largest in the cell at dissipating stage (line C).

As an indicator of the accuracy of the retrievals, the residuals, which are defined as the root-mean-squares of the differences between observed and calculated $T_b$ are also shown in Fig. 10. Except the southern edge of the large convective cell in path 3 where the characteristics of radar and radiometer data did not match, the residuals are less than 3 K for the most part. It means that the residuals are less than the ambiguous value of the radiative transfer scheme itself for the most part. The maximum error value is almost the same as that of Kummerow et al. (1989) or Prigent et al. (1994), though the snowing clouds in this study were geometrically not thick.

For the quantitative validation of the retrieved SWPs, observed radar reflectivity is converted to the snow water content by using the relationship of Sekhon and Srivastava (1970) and is compared with the retrieved
SWPs. The scatter diagram of the computed and observed SWPs (Fig. 11) shows a good correspondence of them. When the retrieved data with the residual lower than 3.0 K are adopted, the correlation coefficient is 0.73, and the root-mean-square difference (rmsd) is 0.047 kg m\(^{-2}\).

On the retrieved LWP, it is not clear about the accuracy of computed LWP because of the lack of LWP measurement in the other method. On the maximum value (at line B), the retrieved LWP is about 10% of the adiabatic LWP. The ratio is smaller than the computed results on the one-dimensional models, summarized by Cotton (1975). On the absolute value (0.12 kg m\(^{-2}\) in LWP and 0.07 g m\(^{-3}\) in the maximum LWC), the amounts are much smaller than that reported by Murakami et al. (1994) (1.15 g m\(^{-3}\) in the case of maximum and 0.2 g m\(^{-3}\) in the cell on mature stage) for the isolated winter snow clouds in the Sea of Japan. However, the computed LWP might be overestimated because the background upward radiation possibly depolarized more than in the modeled upward radiation, as mentioned in the previous section. For the quantitative validation, the in situ data such as aircraft observation data are required.

c. Sensitivity on the fixed parameters

Though the calculated SWPs correspond well to the observed value, the retrieval algorithm includes many assumptions. The present retrieval estimates only two variables, with many fixed (assumed) conditions. To examine the sensitivity of the retrieved SWP and LWP to these assumptions, the other calculations with the different conditions and same AMR observed data are performed. The results of these tests are compared with the radar-observed SWP and the “reference” result, which means the result of the retrieval already shown. The statistics are calculated for the data within 15 km from the radar, which include the cloud-free part and characteristic convective cell for each path.

The tested assumptions are the following parameters: density of precipitation particles, vertical profile pattern of snow water, vertical profile pattern of liquid water, vertically integrated water vapor amount, and surface emissivity, which is the function of surface wind speed. The tested results are shown in Table 2, with the label of “snow density,” “vertical snow water profile,” “vertical liquid water profile,” “water vapor amount,” and “sea surface wind speed,” respectively.

The density of precipitation particles are examined with six different particle densities: 0.01, 0.02, 0.05, 0.1, 0.2, and 0.5 g cm\(^{-3}\) (0.1 g cm\(^{-3}\) is adopted in the reference). Among them, the results are almost the same for the density larger than 0.05 g cm\(^{-3}\). The result suggests that the value in reference, 0.1 g cm\(^{-3}\), is not too far from the average value of the event. Fujiyoshi et al. (1990) reported that the density of the precipitating snow for almost the same area to the present study varies 0.02-0.08 g cm\(^{-3}\). The results of the sensitivity test imply that the precipitation particles in the present case were relatively dense for this area.

The tests for the vertical profile pattern of SWCs are examined by alternating the patterns used in the reference. The result shows the significant underestimation of SWP with large rmsd. The LWP values are also larger than the reference when the precipitation particles are assumed to be more in the lower part. The result shows that the alternative vertical profile of SWC contributes to the better estimation of SWP, even by such a simple system.

The larger effects are shown in the tests with altered vertical profile of the liquid water contents. The tests are performed by multiplying another coefficient to the liquid water content, which is used in the reference. The additional restrictions are determined by simplifying the results of the summarized figure in Cotton (1975) for the ratio of liquid water content to the adiabatic value, as

\[
LWC(Z) = LWC_{\text{reference}}(Z)0.5^{(Z-Z_{\text{CB}})},
\]

for the “bottom-heavy” test, and

\[
LWC(Z) = LWC_{\text{reference}}(Z)0.5^{(Z-Z_{\text{CT}})},
\]

for the “top-heavy” test. The parameters \(Z_{\text{CB}}\) and \(Z_{\text{CT}}\) are the cloud-base height and cloud-top height (km), respectively.

On both tests, the results differ largely from the reference LWP and observed SWP, with large residuals. From these results, it seems that the LWC profile in the reference seems to be a suitable one for the cases in the
present study. However, the variations of LWP and SWP, especially LWP, are so large.

The effect of the water vapor, which is another emitting atmospheric medium in the frequencies, is also not small enough to ignore. The tests are performed by adding $-20\%$, $-10\%$, $+10\%$, $+20\%$, and $+30\%$ to relative humidity in the layer below the cloud top. The results show that the SWP gets slightly better in the high SWP area, while the LWP decreases with the increase of water vapor. It implies the difficulties in separating the effect of the water vapor and liquid water by the retrieval using only the vapor window channels. In contrast, it could be interpreted that the retrieved SWP is affected little by the water vapor amount in the snow clouds.

The largest variations are shown by the tests with the different sea surface state. The results with the sea surface wind speed of 0, 5, 10, 15, and 20 m s$^{-1}$ are shown in the table. The test with the surface wind of 0 and 20 m s$^{-1}$ are obviously erroneous as shown by large residuals. It seems that the wind speed of 10 m s$^{-1}$, which is adopted in the reference, is the best choice. However, the residuals are better in the precipitating part than for the whole area in the test with 15 m s$^{-1}$ wind. This suggests that the inhomogeneous surface emission affected the observed $T_B$ through the cloud. In fact, the different winds were observed at the shore in the south and the north of the cloud streak: the wind speed in the north was strong northwesterly, while weak westerly in the south (not shown). Irregular wind under the convective cell, such as downdrafts in the dissipating stage (Shirooka et al. 1989), might also change the sea surface state for the limited area. In addition, the downward radiation, which is reflected by the surface, is also inhomogeneous as implied by the symmetric pattern of observed $P$. The suitable expression of the inhomogeneous surface condition remains for future work.

5. Concluding remarks

Snow clouds over the ocean were observed simultaneously using an airborne microwave radiometer and a Doppler radar. The observed $T_B$ and calculated parameters, $P$ and $S$ in 36- and 89-GHz channels, responded well to the distribution of precipitation particles and to the dynamic structure of convective cells determined from the cross section of radar reflectivity and Doppler velocity. Though this correspondence is still qualitative at this point, it shows that the horizontal distribution pattern of $T_B$ could be applicable to determining vertical structure of convection, as suggested by Kummerow and Giglio (1994a,b). For the further application of microwave radiometer data, a radiative transfer model was used in the retrieval. Based on estimated patterns of vertical profiles of hydrometeors, we performed physical retrievals of liquid water and snow water amounts. Brightness temperatures at 36- and 89-GHz channels were simulated by a radiative transfer model in the retrieval. Values of the

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observed and retrieved SWP are in good agreement with the correlation coefficient of 0.73. The retrieved liquid water and snow water amounts, as well as their distribution patterns, are reasonable when compared with the observed radar reflectivity and Doppler velocity, though we do not have in situ data for direct validation. These results suggest that the data from high-frequency microwave channels could be used to estimate the LWP and SWP above the freezing level, especially for snow clouds. The result could be contributed to the retrieval algorithm for the data of AMSR/ADEOS-II.

To check the dependencies of the retrieval on the assumptions in the radiative transfer model, the other retrievals are performed with a different set of the fixed parameters to check the sensitivity of the retrievals. The results showed that the vertical profiles of liquid water and snow water are the key factors on the retrieval, even within the frozen precipitation layer. On the other hand, the expression of the background emission also could be a problem.

For the improvement of the system, the horizontal/vertical distribution of the hydrometers are desirable. Especially in situ measurement of the liquid water is also needed for the validation of the retrieved value. Observations using airborne probes, dual-polarization radars and/or videosondes are useful. The sea surface emissivity data are also important factors.

This study is the first in both observing and modeling snow clouds over ocean. It shows the possibility for using high-frequency microwaves to sense snow water amount and snowfall remotely. Further investigation of the physical retrieval method including more adequate validations by in situ data remains as future work.

Acknowledgments. The authors would like to express their thanks to Dr. Keiji Imaoka, Earth Observation Research Center of National Space Development Agency of Japan (NASDA/EORC), for the provision of AMR data and his support on handling the data. The authors are also grateful to the staff of Remote Sensing Technology Center (RESTEC) and the crew of Nakanithon Air Lines for their appreciated effort on the aircraft observation.

Special thanks are due to the staff of Sapporo District Meteorological Observatory of Japan Meteorological Agency for the provision of radar data and aerological data. The data of AVHRR/NOAA are from Japan Image Database (JAIDAS) of Tohoku University.

Last, the authors thank to Professor Katsumiro Kikuchi and Dr. Yoshio Asuma, Graduate School of Science, Hokkaido University, and three anonymous reviewers for their valuable comments.

This study was performed as the cooperative research with NASA for the development of retrieval algorithm for AMSR/ADEOS-II. This study was also supported by Grant-in-Aid for Scientific Research of the Ministry of Education, Science, Culture and Sports of Japan (Nos. 08241101 and 09227202) and Research Fellowships of the Japan Society for the Promotion of Science for Young Scientists.

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