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The DOI for this manuscript is
DOI:10.2151/jmsj.2020-028

J-STAGE Advance published date: February 24th 2020

The final manuscript after publication will replace the preliminary version at the above DOI once it is available.
Vertical Structure and Dynamical Properties during Snow Events in Middle Latitudes of China from Observations by the C-band Vertically Pointing Radar

Ye Cui

Collaborative Innovation Center on Forecast and Evaluation of Meteorological Disasters, NUIST, Nanjing, China and State Key Laboratory of Severe Weather, Chinese Academy of Meteorological Sciences, Beijing, China

Zheng Ruan

State Key Laboratory of Severe Weather, Chinese Academy of Meteorological Sciences, Beijing, China

Ming Wei

Collaborative Innovation Center on Forecast and Evaluation of Meteorological Disasters, NUIST, Nanjing, China

Feng Li, Runsheng Ge

State Key Laboratory of Severe Weather, Chinese Academy of Meteorological Sciences, Beijing, China

Corresponding author address: Zheng Ruan, State Key Lab of Severe Weather, Chinese Academy of Meteorological Sciences, 46 Zhongguancun South Avenue, Beijing, China.

E-mail: ruanz@cma.gov.cn
ABSTRACT

Data from the continuous observations of 4 shallow snow events (echo top <8 km) and 2 deep events (>10 km) were obtained by the C-band vertically pointing radar with frequency modulation continuous wave technology (VPR-CFMCW) with extremely high resolution during the winter of 2015–2016 in middle latitudes of China. Generating cells (GCs) were found near the cloud top in each event. Reflectivity ($Z$), radial velocity ($V_r$), the vertical gradient of $Z$ ($dZ/dh$, $h$ is the vertical distance) and $V_r$ ($dV_r/dh$) showed different vertical distribution characteristics between the upper GC and lower stratiform (St) regions. The fall streaks (FSs) associated with GCs were embedded in the St regions. In the deep events, the proportions of GC regions were slightly larger, but the average contributions to the growth of $Z$ (33%) were lower than those in the shallow events (42%). The average $dZ/dh$ were usually 2–3 times larger inside GCs and FSs compared to outside. Bimodal Doppler spectra were used to establish the relationships between the $Z$ and the reflectivity-weighted particle fall speed ($V_z$) for the 2 regions. The vertical air velocity ($W_a$) and $V_z$ were then retrieved. The results show that both updraft and downdraft were alternately observed in GC regions. GC locations usually accompanied strong upward air motions, with average speeds mostly distributed around $1.2$ m s$^{-1}$, while downward air motions often appeared between GCs. In the St regions, the speeds of $W_a$ were mainly within $0.5$ m s$^{-1}$. The upper areas of the St regions consisted primarily of weak upward motions, while weak downward motions dominated the lower areas. There was no apparent difference in $W_a$ inside and outside
the FSs. The average $V_z$ was slightly larger inside GCs and FSs compared to outside, with the differences of 0.1–0.3 m s$^{-1}$ and 0.2–0.4 m s$^{-1}$ respectively.

1. Introduction

Since the 1950s, weather radar has been used to study snow. Langille and Thain (1951) found a good correlation between $Z$ and snowfall ($R$) by radar observation data during snow. Marshall and Gunn (1952) carried out further analysis on the data of Langille and Thain (1951), and used $Z$ and $R$ values to estimate the parameters of particle size distribution for snow.

Later, vertically pointing radars were introduced to snow research. Marshall (1953) first found snow-generating cells (GCs) near the cloud top with temperature ($T$) around $-15^\circ$C, and the altitude at which they are located was called the “generating level”. The term GC describes a small region of locally high radar reflectivity at the cloud top from which an enhanced reflectivity trail characteristic of falling snow particles originates (American Meteorological Society 2016). Patterns of the trails often show as virga-like slanted streaks, termed “fall streaks” (FSs). In addition, many studies have observed GCs and FSs in frontal rainbands and non-precipitating regions, such as cirrus uncinus clouds (e.g., Heymsfield 1975; Houze et al. 1981). The horizontal extent of GCs is about 1.6 km, while the vertical extent is slightly smaller (e.g., Gunn et al. 1954; Wexler 1955; Langleben 1956). Rosenow et al. (2014) also found that the GCs had typical depth of 1–2 km and horizontal scales of about 0.5–2 km.
The dynamical process in GCs is also a key research focus. Douglas et al. (1957) studied the dynamics of GC formation, estimating updraft speeds in the cells of 1.0–3.0 m s\(^{-1}\) through theoretical analysis. Several other studies applied radar data alongside theoretical deductions to obtain the vertical air velocity in the GCs. Heymsfield (1975) derived updraft velocities of 1.2 to 1.8 m s\(^{-1}\) by vertically pointing radar. Cronce et al. (2007) calculated that the strongest updrafts were typically between 2.0 and 4.0 m s\(^{-1}\) (based on 915-MHz profiler data), with an uncertainty of about ±0.6 m s\(^{-1}\). Rosenow et al. (2014) reported statistical analyses of GC vertical motions using cloud radar measurements, showing that the GCs had vertical velocities of ±1–2 m s\(^{-1}\) or more. Rauber et al. (2015) reported that the updrafts at the GC level near cloud top ranged from 1 to 3 m s\(^{-1}\). Downdrafts with magnitudes ranging from 0 to 1 m s\(^{-1}\) were present between the GCs. Rauber et al. (2017) even found that the vertical air velocities in the GCs were on the order of ±3–5 m s\(^{-1}\) in a case. Keeler et al. (2017) simulated the GCs by using the idealized WRF model at high resolution. The results of simulations indicated that updrafts within GCs under conditions with radiative forcing were typically ~1–2 m s\(^{-1}\) with maximum values < 4 m s\(^{-1}\). Many studies have shown that the convective air motion can promote ice nucleation and growth (e.g., Houze et al. 1981; Ikeda et al. 2007; Crosier et al. 2014).

During the 2009–2010 Profiling of Winter Storms (PLOWS) project, 14 winter cyclones detected by the airborne Wyoming Cloud Radar (WCR) indicated that the GCs were ubiquitous in the warm-frontal and comma-head regions of midlatitude winter cyclones (Rosenow et al. 2014; Rauber et al. 2014a, b). Keeler et al. (2016a,b,
used models to simulate GCs and examine their origin, their forcing, and their relationship to vertical wind shear, ambient thermal instability, and cloud-top radiative forcing. Plummer et al. (2014) conducted a statistical analysis of the microphysical properties of GCs and analyzed their structure. The measured GC data indicated that there was an enhanced nucleation phenomenon and initial particle growth within GCs, which verified the hypothesis of Houze et al. (1981). The PLOWS observations showed that considerable growth sometimes occurred within GCs. Particles may grow up to 5–6 mm in maximum dimension in the GCs, particularly with the $T$ of $-11 \pm 5^\circ$C. The GCs have also been studied by polarimetric radar during the past few years. Kumjian et al. (2014) found the peak of $Z_{dr}$ was positioned near $-15^\circ$C, corresponding to platelike or dendritic growth. The observations reported by Kumjian and Lombardo (2017) implied that aggregation was efficient at $-15^\circ$C.

GCs usually present atop stratiform clouds, where the FSs embedded, indicating the GC’s impact on precipitation in the stratiform regions (St regions). Weak upward motion associated with frontal-scale forcing and ice supersaturation exist in the stratiform regions, providing conditions for the continued particle growth below GCs (Rauber et al. 2014b; Rosenow et al. 2014; Rauber et al. 2015). Many studies (e.g., Matejka et al. 1980, Syrett et al. 1995, Schultz et al. 2004, Cunningham and Yuter 2014) have used the seeder–feeder concept (Bergeron 1950) to explain the processes that occur in the GCs and the underlying stratiform clouds. Plummer et al. (2015) statistically analyzed the microphysical characteristics in the FSs and found that the majority of ice growth typically occurred below the GC level. There were no obvious
vertical velocity characteristics within FSs. Pfitzenmaier et al. (2017) presented a new algorithm to retrieve FSs within a radar time–height plot based on genuine high-resolution wind information obtained with the Transportable Atmospheric Radar (TARA). Keppas et al. (2018) used the data obtained from dual-polarization radars to provide analyses of the structure, the origin, and the effects of the FSs associated with warm fronts. However, the research on the vertical structure and dynamical characteristics of the entire stratiform region is less than that on GC regions.

The C-band vertically pointing radar with frequency modulation continuous wave technology (VPR-CFMCW) was developed by the Chinese Academy of Meteorological Sciences in 2013 to study the vertical structure, and dynamical processes in precipitating cloud and the microphysical parameters for the retrieval of precipitation data. Compared with rainfall, there are few studies on snowfall based on vertically pointing radar in China, leading to deficiencies in previous understanding of the detailed structure and the dynamical processes in snow clouds over this region.

The VPR-CFMCW with high-resolution is capable of obtaining Z, radial velocity ($V_r$), spectral width (SW) and full Doppler spectra, and has great potential to enhance our understanding of the evolution of vertical structure and the dynamical process in snow clouds.

In this research, 6 snow events in Shou County (a typical region in middle latitudes of China) were analyzed during the winter of 2015–2016 using VPR-CFMCW. The Huai River is the climate boundary between north and south China, with subtropical monsoon climate in the south and temperate monsoon climate in the north. In terms of
temperature, the north of Huai River belongs to warm temperate zone, while the south of Huai River belongs to subtropical zone. In terms of humidity, the north of Huai River belongs to semi-humid region, while the south of Huai River belongs to humid region. Shou County is adjacent to the Huai River and located on the south bank, featured the topography of plain. Although the times of snowfall in the south of Huai River are not frequent, the influence scope is large and the duration is long. The water vapor content in this region is relatively abundant in winter, and under certain weather background, severe snowstorms and freezing disasters would also occur (e.g., Zhou et al. 2009, Wen et al. 2009).

The novel aspects of this study are as follows: (1) the 6 snow events were classified into deep and shallow categories and the snow clouds were divided into GC regions and stratiform regions (St regions) for statistical comparison; (2) the characteristics of GC regions and the increase of Z within GC regions and St regions were quantified. Moreover, the average reflectivity gradients \( \frac{dZ}{dh} \), \( h \) is the vertical distance) inside and outside GCs and FSs were analyzed; and (3) the vertical air velocity \( W_a \) in the snow clouds and reflectivity-weighted fall speed \( V_z \) of snow particles were retrieved more precisely from Doppler spectra data combined with \( Z \) and \( V_r \). The continuous evolution and statistical characteristics of \( W_a \) and \( V_z \) in the entire snow clouds were also obtained, providing a more comprehensive understanding of the 2 regions.

The next section introduces the instrumentation and selected data, and the observed 6 snow events are classified into deep and shallow categories for discussion. The synoptic background and atmospheric stratification characteristics of the typical
events are described in Section 3. In Section 4, the snow clouds are divided into GC
and St regions and the statistics for the characteristics of GC regions along with
average dZ/dh inside and outside GCs and FSs is also presented. Section 5 describes
the retrieval algorithm for $W_a$ and $V_z$ and the retrieved values of the 2 regions in 2
types of snow events are statistically studied. This is followed in Section 6 by a
summary of our research.

2. Instrumentation and Data

2.1. VPR-CFMCW system and other Equipments

Since the pioneering work in the 1960s (Atlas et al. 1973, and references therein),
vertically pointing Doppler radars have been recognized as irreplaceable tools in
cloud physics. Compared with evolution in the horizontal direction of a precipitation
system, that in the vertical direction is more rapid within the scope less than 20 km
between ground and echo top. The vertically pointing radars need to completely
describe the vertical structure of precipitation clouds in the detection range, which
requires the radars with detection performance of high spatial and temporal resolution,
large dynamic range and high sensitivity. Different from pulsed radars, the vertically
pointing radars with frequency modulation continuous wave (FM-CW) technology
have ideal ranging accuracy from the window function with high isolation degree.
The power density spectra of return signal is obtained by coherent spectrum analysis
method, and high accumulative gain improves the system sensitivity, making the
detection information comprehensive and real. Therefore, the meteorological radars
using FM-CW technology have the advantages of precise distance and velocity measurement, small ranging blind area, low peak power, large system dynamic range and high sensitivity.

The VPR-CFMCW, located at the Shou County meteorological station (32°26′N, 116°47′E), utilizes a simple and reliable solid-state transmitter and continuous wave fully-coherent Doppler system. Transmit-receive isolation was realized using a bistatic antenna with double paraboloid, which transmits a linear frequency modulation signal produced by a direct digital synthesizer. The fast Fourier transform algorithm with 512 points is used twice after demodulation to obtain range information and spectral distribution on the range gate and to allow precise detection of the cloud vertical structure. The radar outputs density spectra of power with 512 channels, which can be converted into density spectra of $Z$ via the radar equation, allowing the calculation of three spectral parameters ($Z$, $V_r$ and SW). The photo of VPR-CFMCW in Shou County is shown in Fig. 1.

![Photo of VPR-CFMCW](image)

Fig. 1. Photo of VPR-CFMCW located in Shou County.

The radar was strictly calibrated and compared with S-band weather radars (SWRs), in the nearby Hefei (82 km southeast of Shou County) and Bengbu (70 km to the
northeast of Shou County) cities, Anhui Province before it was put into use in 2013. The comparison results showed that for the uniform stratiform precipitating cloud, the detection data of VPR-CFMCW were close to those of 2 SWRs, with a difference in reflectivity of less than 1 dB. The VPR-CFMCW could give rapid evolution and fine structure details for convective precipitating cloud, and performed better than SWRs. The main performance indexes of the radar system are summarized in Table 1 (Ruan et al. 2015). The maximum usable range of the VPR-CFMCW is 15–24 km, and the minimum measurable signal power is close to −170 dBm. The minimum detectable reflectivity is −20 dBZ at 15 km, −22 dBZ at 12 km and −28 dBZ at 6 km. The radar data was quality-controlled following the method described in Li et al. (2018).

In addition to the VPR-CFMCW, the Shou County meteorological station is also equipped with ground observation equipment, allowing measurements of the snow depth and snowfall. The Fuyang meteorological station, equipped with radiosonde, is located about 95 km northwest of Shou County. During the winter, the mid- and upper-level winds are generally from the northwest. Since Fuyang lies to the northwest of Shou County, it is feasible to use sounding data from Fuyang to analyze the weather conditions of Shou County. The $T$ and humidity of layers at different heights can be acquired using the Fuyang sounding data on corresponding snow days.
Table 1. Main performance indexes of the VPR-CFMCW system.

<table>
<thead>
<tr>
<th>No.</th>
<th>Parameters</th>
<th>Indexes</th>
</tr>
</thead>
<tbody>
<tr>
<td>1</td>
<td>Radar system</td>
<td>Frequency</td>
</tr>
<tr>
<td>2</td>
<td>Center frequency</td>
<td>5530 MHz</td>
</tr>
<tr>
<td>3</td>
<td>Beam width</td>
<td>$\leq 2.6^\circ$</td>
</tr>
<tr>
<td>4</td>
<td>Band width</td>
<td>6 MHz$\pm 3$ MHz$\pm 3.5$ MHz</td>
</tr>
<tr>
<td>5</td>
<td>isolation between antennas</td>
<td>90 dB</td>
</tr>
<tr>
<td>6</td>
<td>Transmit power</td>
<td>$\geq 150$ W</td>
</tr>
<tr>
<td>7</td>
<td>Antenna gain</td>
<td>$\geq 35$ dB</td>
</tr>
<tr>
<td>8</td>
<td>Antenna types</td>
<td>Bistatic and parabolic antenna</td>
</tr>
<tr>
<td>9</td>
<td>Antenna diameter</td>
<td>1.6 m</td>
</tr>
<tr>
<td>10</td>
<td>Height coverage</td>
<td>0.1–24 km</td>
</tr>
<tr>
<td>11</td>
<td>Temporal resolution</td>
<td>2–3 s</td>
</tr>
<tr>
<td>12</td>
<td>Range resolution</td>
<td>15 m/30 m</td>
</tr>
<tr>
<td>13</td>
<td>Sensitivity</td>
<td>$\leq -20$ dBZ at 15 km</td>
</tr>
<tr>
<td>14</td>
<td>Velocity bin resolution</td>
<td>0.0895 m s$^{-1}$</td>
</tr>
<tr>
<td></td>
<td></td>
<td>Density spectra of reflectivity,</td>
</tr>
<tr>
<td></td>
<td></td>
<td>power density spectra, spectral</td>
</tr>
<tr>
<td></td>
<td></td>
<td>parameters</td>
</tr>
</tbody>
</table>

2.2. Snow events

The 6 snow events selected for this study were observed by the VPR-CFMCW during the winter of 2015–2016 in Shou County. Ground snow observations include
varying snow intensities. Snow periods and snowfall data are provided in Table 2. Note that before it began to snow on 24 November 2015, rainfall and sleet had occurred, so the snowfall data for this event also includes precipitation produced by light rain and sleet. The cumulative duration of the 6 snow events was 37 h and 39 min, and the total number of profiles was 26540. The snow events of 28 and 29 January 2015 had the longest duration (approximately 12 h). Ignoring the mixed snow event on 24 November 2015, the maximum snowfall was 6.9 mm per day on 28 January 2015.

The $Z$ and $V_r$ values detected during selected snowfalls on 27, 28, 29 January, 28 February, 24 November 2015, and 20 January 2016 are presented in Fig. 2. To exhibit the relatively complete evolution of snow clouds, the selected durations in the figure are slightly longer than those recorded by the ground snow observations. Among the 6 events, the echo tops reached 12 km or more on 28 February and 24 November 2015 (Figs. 2e and 2f), while clouds were mainly located below 8 km in the other events. Small erect turrets appeared at the edges of the cloud where the GCs occurred, with different distribution densities. The $V_r$ changed rapidly in the GCs, both updraft and downdraft were alternately observed. There were fibrous regions of large $Z$ below the GCs, with $Z$ values generally greater than 10 dBZ. These were FSs, corresponding to high $V_r$ above 1 m s$^{-1}$. The values of $Z$ and $V_r$ for the FSs showed a trend of gradual enhancement with decreasing height. FSs may be influenced by horizontal wind moving out of the beam or merging as the height decreases. Previous studies (e.g., Plummer et al., 2014; Keeler et al., 2016b) indicate that the existence of GCs is
almost ubiquitous during snowfall, serving as an initial nucleation and growth mechanism for ice crystals.

Table 2. The date, snow periods, duration and snowfall data for the 6 snow events.

<table>
<thead>
<tr>
<th>Events</th>
<th>Date</th>
<th>Snow periods</th>
<th>Duration (min)</th>
<th>Snowfall (mm)</th>
</tr>
</thead>
<tbody>
<tr>
<td>1</td>
<td>27 Jan 2015</td>
<td>0032UTC–0430UTC</td>
<td>238</td>
<td>1.9</td>
</tr>
<tr>
<td>2</td>
<td>28 Jan 2015</td>
<td>0000UTC–1200UTC</td>
<td>720</td>
<td>6.9</td>
</tr>
<tr>
<td>3</td>
<td>29 Jan 2015</td>
<td>0020UTC–1151UTC</td>
<td>691</td>
<td>4.3</td>
</tr>
<tr>
<td>4</td>
<td>28 Feb 2015</td>
<td>0000UTC–0335UTC</td>
<td>215</td>
<td>3.1</td>
</tr>
<tr>
<td>5</td>
<td>24 Nov 2015</td>
<td>0855UTC–1200UTC</td>
<td>185</td>
<td>27.3</td>
</tr>
<tr>
<td>6</td>
<td>20 Jan 2016</td>
<td>1230UTC–1600UTC</td>
<td>210</td>
<td>A little</td>
</tr>
</tbody>
</table>

The 6 snow events reveal that there are 2 types of location heights for GCs and 2 patterns of their associated FSs. The first type is represented by the 27, 28, 29 January 2015 and 20 January 2016 events (Figs. 2a, 2b, 2c and 2d). In these events, the GCs were located at the tops of relatively shallow clouds. The GC heights were consistently about 5 km in the 27 January 2015 and 20 January 2016 events. During the 28 and 29 January 2015 events, the GC heights fluctuated, ranging primarily from 5 km to 7 km. The vertical extent of the GCs was relatively small, approximately 0.5–1.5 km. The Z rapidly increased from about −20 dBZ to around 0 dBZ and V, fluctuated mainly between −1 and 1 m s\(^{-1}\) within the GCs. GC distribution was sometimes quite dense, and the fibrous FSs were traced from GCs that were densely distributed. The FSs with high Z generated by falling ice crystals gradually combined into bands as they got closer to the ground and were relatively uniform before
reaching the ground.
Fig. 2. Time–height images of $Z$ detected by the VPR-CFM CW during the 6 snow events. (a), (b), (c) and (d) are the shallow events, corresponding to 27, 28 and 29 Jan 2015 and 20 Jan 2016, respectively. (e) and (f) are deep events, corresponding to 28 Feb 2015 and 24 Nov 2015, respectively. The time spans within the black double arrows represent the snow periods recorded by ground observations. The typical time periods used in the following sections are marked by red double arrows. The black lines represent the boundaries between GC and St regions (the specific partition method is described in section 4.2).

The second type of snow event is represented by the 28 February 2015 and 24 November 2015 snowfalls (Figs. 2e and 2f). GCs were located at the tops of deep clouds, and the heights of the GC tops were above 10 km. The vertical extent of the GCs was relatively large, mainly around 2 km and the increase in $Z$ took place over a greater distance within GCs. The cloud structure was relatively uniform in height, while the fluctuations of $V_r$ within GCs were similar to those in the first type of snow event. The key similarity among the 6 events was that $Z$ values greater than 20 dBZ were only found below 5 km. The average $T$ at 5 km was about $-15^\circ$C.

3. Analysis of atmospheric environment

3.1. Synoptic background

Reanalysis data from the European Centre was used to analyze the weather conditions, with a resolution of 0.25° $\times$ 0.25°. The 20 January 2016 event was
selected to represent the first event type, and the 28 February 2015 event was selected
to represent the second event type.

For the 20 January 2016 event, weather conditions at 1200 UTC were analyzed. We
mainly focused on the wind, temperature, and specific humidity fields at 700 hPa, as
well as wind, height, and temperature fields at 850 hPa. As can be seen in Figs. 3(a)
and 3(b), the Shou County was located ahead of a southwest low-level jet, in advance
of a trough at 700 hPa. This low-level jet was a very important water vapor transport
channel, providing a source of water vapor for the snowfall. The Shou County was
also located in the upward motion area southeast of the shear line of horizontal wind
speed. Corresponding to the southwest jet at 700hPa, an inverted trough could be
observed at 850hPa. The Shou County was near the top of the inverted trough, which
was also the convergence zone formed by northerly airflow and easterly airflow,
leading to an area of upward air motion.

Fig. 3. Weather analysis chart at 1200 UTC on 20 Jan 2016. Blue solid lines represent
the height field or pressure field, red solid lines represent the temperature field, green
solid lines represent the specific humidity field, and the black pentagram shows the
location of Shou County.
As can be seen from the 500 hPa plot in Fig. 4(a), corresponding to the 28 February 2015 event at 0000 UTC, the temperature trough to the north of the Shou County lagged behind the height trough. The cold advection following the trough was favorable for the trough’s development. The Shou County was still located ahead of the southwesterly jet, as shown in Fig. 4(b), where the warm, wet southwesterly jet converged with the dry, cold northwesterly airflow. The significant wind shear above the Shou County made the airflow converge and rise. The frontal system tilted to the northwest with height, and the Shou County lay behind the surface cold front. During this snow process, cold air in mid- to high latitudes advected southward and influenced the Shou County. The system was deeper than that of the 20 January 2016 event.

Fig. 4. Weather analysis chart at 0000 UTC on 28 Feb 2015. SLP in Fig. 4(d) is sea
level pressure. Blue solid lines, red solid lines, green solid lines and the black pentagram represent the same parameters as in Fig. 3.

3.2. Vertical stratification

Due to a lack of sounding data in Shou County, atmospheric stratification was analyzed at the Fuyang meteorological station using the Skew-T diagrams and the profiles of relative humidity with respect to water (RH) and supersaturation with respect to ice ($S_i$). The same events as in section 3.1 were selected for analysis, as shown in Fig. 5.

The RH was directly obtained by air sounding data. The detected RH can be transformed into relative humidity with respect to ice ($\text{RH}_i$) using WMO formulations about calculating saturation vapor pressure over a surface of ice and liquid water below 0°C (WMO, 2008).

The saturation vapor pressure over liquid water below 0°C ($e_w$) was calculated by Eq. (1):

$$e_w = 6.112e^{(17.62T/(243.12+T))}, \quad (1)$$

with $T$ in °C, $e_w$ in hPa.

The saturation vapor pressure over ice ($e_i$) was calculated by Eq. (2):

$$e_i = 6.112e^{(22.46T/(272.62+T))}, \quad (2)$$

with $T$ in °C, $e_i$ in hPa.

The actual vapor pressure ($e$) was obtained by multiplying $e_w$ by RH ($e = e_w\text{RH}$), and $\text{RH}_i$ was obtained by dividing $e$ by $e_i$ ($\text{RH}_i = e/e_i$). The $S_i$ ($S_i = \text{RH}_i - 100\%$) was
also calculated.

In Figs. 5a and 5b, the thermal profiles showed that the $T$ below 1000 hPa were all lower than 0°C. Thus, snowflakes did not undergo any melting or phase transformation while falling. There was a $T$ inversion layer between 850–700 hPa and the $T$-dew point difference was very small, indicating that stratiform clouds with increasing $T$ and humidity existed in this layer.

As shown in Fig. 5a (deep event on 28 Feb 2015), the $T$-dew point difference below 300 hPa was very small, corresponding to large RH values shown in Fig. 5c. The $S_i$ values below 13 km were greater than 0, indicating that the entire region was saturated with respect to ice. The $S_i$ reached a maximum of 40% at 12 km, where the echo top located.

As shown in Fig. 5b (shallow event on 20 Jan 2016), the dew point profile above the saturated layer sloped steeply towards low $T$, resulting in a sharp decrease in ambient humidity and the formation of a dry layer. Thus, there was no cloud above ~6 km, corresponding to the similar height of echo top seen in Fig. 2d. There was a near-saturated thin layer above the dry layer located near 300hPa. However, maybe due to its short duration and thin thickness, the echo of cloud was not observed at the corresponding height. The region above the dry layer had no effect on the snowfall process, which was been ignored in the following analysis.

In the Fig. 5d, similar to the analysis results of the Skew-T diagram, RH values stayed above 90% under 4 km, and then decreased slightly with height until they sharply decreased to ~20% above 6 km. The $S_i$ values were greater than 0 between 1
and 6 km, indicating the region inside cloud was saturated with respect to ice. Ignoring the region above the dry layer, the $S_i$ reached a maximum of 40% near 6 km, where the echo top located.

![Fig. 5. The Skew-T figures on (a) 0000 UTC 28 Feb 2015, and (b) 1200 UTC 20 Jan 2016, along with RH and $S_i$ profiles on (c) 0000 UTC 28 Feb 2015, and (d) 1200 UTC 20 Jan 2016 respectively at Fuyang radiosonde station. In the Skew-T figures, the light red lines represent the dry adiabatic line ($^\circ$C), constant saturation specific humidity lines (g/kg) are purple, moist adiabate lines ($^\circ$C) are cyan, the gray and white lines are isotherms ($^\circ$C), the black curve is the temperature sounding curve, the blue curve is the dew point temperature sounding curve, the red dotted line is the state curve.](image-url)
4. Analysis of vertical structure

In this section, 6 typical time periods marked by red double arrows in Fig. 2 were selected for specific analyses firstly, and then the partition method for GC and St region was determined. Finally, the characteristics of GC regions in the 6 events were analyzed.

4.1. Properties of the vertical structure

4.1.1. Vertical structure

The contoured frequency by altitude diagrams (CFADs; Yuter and Houze, 1995) of $Z$, $V_r$, the gradients of $Z$ (d$Z$/d$h$) and $V_r$ (d$V_r$/d$h$) were exhibited in Fig. 6 (4 shallow events) and Fig. 7 (2 deep events) to study vertical evolution properties. To prevent noise near the echo top from interfering with the analysis, following the method of estimating cloud-top altitude described by Plummer et al. (2014), the threshold of $V_r$ variance still adopted 0.8 m$^2$ s$^{-2}$. However, according to the sensitivity of VPR-CFMCW near 6 km and 12 km (the height of echo top for the 2 types of snow events), the threshold of $Z$ was changed to $-28$ dBZ and $-22$ dBZ respectively. Since the data reliability of the first 4 range gates was not high, the CFADs began with the fifth range gate (150 m).

The d$Z$/d$h$ and d$V_r$/d$h$ between each 2 adjacent range gates (30 m) were obtained to bolster the understanding of vertical variation about $Z$ and $V_r$. They were vertical gradients in the downward direction calculated with respect to the direction toward the ground. Positive values of d$Z$/d$h$ and d$V_r$/d$h$ indicate that $Z$ and $V_r$ increase with decreasing height. Negative values imply the opposite.
As height decreased, $Z$ values increased initially and then decreased slightly in the 6 snow events. The maximum $Z$ values ($Z_{\text{max}}$) were approximately 25 dBZ near 2.5 km above the ground. $Z$ values were usually between 10 dBZ–20 dBZ when snowflakes reached the ground. Compared with the distribution range of $Z$, that of $V_r$ was relatively concentrated, mainly from 0–2 m s$^{-1}$. Negative $V_r$ appeared in the upper regions of the clouds, indicating the existence of upward air motions.

In Fig. 6 and Fig. 7, within 1–2 km below the cloud-top, the distribution range of $dZ/dh$ and $dV_r/dh$ widened. The distribution width of $dV_r/dh$ reached ~1 m s$^{-1}$, which was significantly higher than that in the rest of cloud (mainly within 0.4 m s$^{-1}$). The fluctuation of $dV_r/dh$ indicated that the $V_r$ changed rapidly, maybe influenced by air motion. Within ~1 km above the ground, $Z$ values showed a decreasing trend, possibly due to the effect of decreasing humidity, as shown in Fig. 5.

4.1.2. Differences of shallow and deep events

Figure 6 illustrates that the $Z$ values increased most quickly in the regions approximately 1–2 km below the cloud-top, and the maximum values of $dZ/dh$ were encountered at these depths. As for shallow events, the maximum values reached 2 dB/30 m, even above 5 dB/30 m. Deeper in these regions, the growth rate of $Z$ values got slower. The range of $dV_r/dh$ was also largest in those same regions. The ranges of $dZ/dh$ and $dV_r/dh$ values were relatively narrow beneath the regions near cloud-top, indicating that the lower regions were relatively uniform and stable.
Fig. 6. CFADs of the Z, \(V_r\), \(dZ/dh\) and \(dV_r/dh\) values for the 4 events of the shallow events. (a1)–(a4) corresponds to 0230–0330 UTC 27 Jan 2015, (b1)–(b4) 1100–1200 UTC 28 Jan 2015, (c1)–(c4) 0900–1000 UTC 29 Jan 2015, and (d1)–(d4) 1230–1330 UTC 20 Jan 2016. The CFADs begin with the fifth range gate (150 m).

The CFADs shown in Fig. 7 correspond to the 2 deep events. Z values increased a little faster in the regions around 2 km beneath the cloud-top than in the regions below. The range of \(dZ/dh\) values in the regions near cloud-top was narrower than that of
shallow events, indicating that the rate and amplitude for the variation of $Z$ values were relatively small in the deep events. The differences in $dZ/dh$ between regions near cloud-top and the rest were also smaller than shallow events. The range of $dV_r/dh$ values in the upper regions was similar to that of shallow events, while that of $dZ/dh$ values was a little narrower in the deep events. Note that due to the limitation of radar sensitivity, some weak signals may not be captured near the top of the deep events, so that the calculated $dZ/dh$ may be smaller than the actual values.

Fig. 7. CFADs of the $Z$, $V_r$, $dZ/dh$ and $dV_r/dh$ values for the deep events. (a1)–(a4) and (b1)–(b4) correspond to 0020–0120 UTC 28 Feb 2015, and 0900–1000 UTC 24 Nov 2015 respectively. The CFADs begin with the fifth range gate (150 m).

4.2. Characteristics Analysis

The characteristics of $Z$ and $V_r$ in the regions near cloud-top were similar to those of the GC regions described in the literature (e.g., Plummer et al. 2014, Rosenow et al.
2014). As shown in Figs. 2, 6, and 7, the uniformity and character of the echoes are distinct between the GC regions near cloud-top and the rest of clouds termed St regions. The 2 regions could be divided by the differences in $dZ/dh$ and $dV_r/dh$.

After analyzing the full datasets, it was determined that $dZ/dh > 0.6$ dB/30 m and $|dV_r/dh| > 0.08$ m s$^{-1}$/30 m could be used as GC indicators for the shallow events, while $dZ/dh > 0.2$ dB/30 m and $|dV_r/dh| > 0.08$ m s$^{-1}$/30 m could be used as GC indicators for the deep events. If 3 continuous range gates designated as $x_1$, $x_2$, and $x_3$ did not simultaneously meet the indicator criteria, the corresponding height of $x_1$ (the range gate at the top of the 3) was considered to be the height of the bottom of the GC. Note that, the differences between GC and St regions are gradual, so it is hard to determine a clear boundary. The partition method presented here extracts the regions with obvious GC features from a statistical perspective, and regards the rest as relatively stable St regions. The purpose is to quantitatively describe the characteristics of GC regions for further analyses. The black lines in Fig. 2 represent the boundaries between the 2 regions obtained by this method, and the periods with obvious GC features account for about 90% of all the periods during the 6 events.

The characteristics of GC regions in the 6 events are shown in Table 3 (The statistics only refer to the snow periods recorded by ground observations), including the maximum times of GC appearance within 1 h, depth, the ranges of $T$ and $RH$, the average proportions with respect to the whole cloud depth and the average contribution of GC region to the growth of $Z$ value. Among them, the depth is obtained by subtracting the height of the bottom of GC from the calculated cloud-top...
height. The average proportion with respect to the whole cloud depth is the average ratio of the GC region depth to the whole cloud depth. The difference of $Z$ values ($\Delta Z$) between the cloud-top height and the height of the bottom of GC is defined as the $\Delta Z$ within GC region, and the difference between $Z$ value at the cloud-top and $Z_{\text{max}}$ is defined as the total $\Delta Z$ in the cloud. The average ratio of $\Delta Z$ within GC region to total $\Delta Z$ is defined as the average contribution of GC region to the growth of $Z$ value. The depths of GC regions are in good accord with previous studies (e.g., Plummer et al. 2014, Rosenow et al. 2014).

Table 3. The characteristics of GC regions for the shallow and deep snow events.

<table>
<thead>
<tr>
<th>Event type</th>
<th>Date</th>
<th>Appearance times (1h)</th>
<th>Depth (m)</th>
<th>Range of $T$ (°C)</th>
<th>Range of RH$_i$ (%)</th>
<th>Avg. proportion (%)</th>
<th>Avg. contribution (%)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Shallow</td>
<td>27 Jan 2015</td>
<td>18</td>
<td>420–800</td>
<td>−21 to −17</td>
<td>137–148</td>
<td>11.5</td>
<td>42.1</td>
</tr>
<tr>
<td></td>
<td>29 Jan 2015</td>
<td>19</td>
<td>500–1500</td>
<td>−25 to −18</td>
<td>127–142</td>
<td>13.3</td>
<td>40.5</td>
</tr>
<tr>
<td></td>
<td>20 Jan 2016</td>
<td>15</td>
<td>600–840</td>
<td>−20 to −16</td>
<td>130–140</td>
<td>12.6</td>
<td>41.3</td>
</tr>
<tr>
<td>Deep</td>
<td>28 Feb 2015</td>
<td>7</td>
<td>1400–2700</td>
<td>−62 to −46</td>
<td>125–140</td>
<td>16.7</td>
<td>32.6</td>
</tr>
<tr>
<td></td>
<td>24 Nov 2015</td>
<td>8</td>
<td>1300–2500</td>
<td>−59 to −47</td>
<td>128–141</td>
<td>16.4</td>
<td>33.2</td>
</tr>
</tbody>
</table>

These statistics indicate that for the shallow events, the $T$ measurements at the height of GCs were mainly around −20° C. The GC regions typically have a depth of 500–1200 m, with $\Delta Z$ mainly distributed between 23 and 27 dB. The depths of GC regions accounted for about 11–14% of the whole cloud. As for shallow events, the
average contributions of GC regions reached 42%. GCs appeared up to 23 times within 1 h, resulting in dense fibrous FSs below.

There were some differences in the statistical characteristics between the 2 types of snow events. For the deep events, the depths of GC regions were greater, reaching over 2 km. The times of GC appearance within 1 h were about one third to one half of those in the shallow events, leading to the relatively sparse FSs and uniform St regions below. The $T$ measurements at the height of GCs were low, mainly distributed around $-50^\circ$C. Due to the high altitude, the RH values were also lower than those found in the shallow events, but the RH$_i$ values at the height of GCs were greater than 120% for all 6 events, indicating that the environment was supersaturated with respect to ice. The depths of GC regions accounted for about 16% of the whole cloud, while the average contributions to the growth of $Z$ values reached 33% in these regions. Many previous studies also presented that the majority of ice growth typically occurred below the GC level (e.g., Matejka et al. 1980; Houze et al. 1981; Plummer et al. 2015), while the contributions of GC regions to $Z$ value growth were larger than those to ice mass growth described in the above articles. Overall, the results revealed that in the deep events, the proportions of GC regions were slightly larger, but the average contributions to the growth of $Z$ values were lower than those in the shallow events.

During the shallow events, due to the higher temperatures in the GC regions, supercooled water may still exist (Plummer et al. 2014). Thus the ice particles may grow through deposition and riming. Previous studies (e.g., Connolly et al. 2012)
reported that the aggregation efficiency is relatively high between \(-20\) and \(-10^\circ\text{C}\) and reaches its maximum at \(-15^\circ\text{C}\). Ice particles in the GC regions may also grow rapidly by aggregation.

During the deep events, the GC regions featured low \(T\) and water vapor content, and riming was not able to exist without supercooled water \((T < -40^\circ\text{C})\), which led to the relatively slow growth of ice particles. However, the GC region was located at great heights with low \(T\) and high RH, an environment that was conducive to the deposition nucleation of ice crystals (Welti et al. 2014), which might contribute to the increase in particle number concentration. The ice particles in the GC regions may feature high number concentration and slow growth rate, while \(Z\) is more sensitive to the changes in particle size. Therefore, we consider that in the GC regions of deep events, the \(Z\) values may not increase as fast as they do in the GC regions of shallow events, so the calculation error of \(dZ/dh\) may not be significant. We will provide evidence for the inference in the next article about microphysical properties.

According to the apparent differences between the surrounding regions and the inside of GCs and FSs exhibited in time–height contours of \(Z\), the \(dZ/dh\) values between the inside and outside of GCs and FSs were presumed to be different. To facilitate statistical analyses, the definition of the inside of GC presented by Plummer et al. (2014) was accepted, and we also use the threshold of 4 dB relative maxima in the \(Z\) time series measurements. To identify FSs, the method described in Plummer et al. (2015) was followed and the 45-s window was adopted. Since FSs were tilted, the calculation of \(dZ/dh\) within FSs was additionally explained here. The calculation of
dZ/dh inside FSs was based on the measurements classified as FSs at each height. We believed that particles affected by the same horizontal wind moved at the same speed in the horizontal direction, so the earliest falling particles also reached a same lower height firstly. Between 2 adjacent range gates, the dZ/dh values inside FSs were obtained by Z values classified as FSs at a lower range gate minus those at a adjacent higher range gate successively (i.e., the first Z value classified as FS at a lower range gate minus that at a adjacent higher range gate and so on). The statistical results are summarized in Table 4.

Table 4. The statistics on the average dZ/dh of GC and St regions along with the averages inside and outside GCs and FSs for the shallow and deep snow events.

<table>
<thead>
<tr>
<th>Event type</th>
<th>Date</th>
<th>Avg. dZ/dh in GC region</th>
<th>Avg. dZ/dh in St region</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td></td>
<td>(dB/30 m)</td>
<td>(dB/30 m)</td>
</tr>
<tr>
<td>Shallow</td>
<td>27 Jan 2015</td>
<td>1.02 1.49 0.51 0.17 0.24 0.08</td>
<td></td>
</tr>
<tr>
<td></td>
<td>28 Jan 2015</td>
<td>0.71 1.05 0.37 0.13 0.21 0.07</td>
<td></td>
</tr>
<tr>
<td></td>
<td>29 Jan 2015</td>
<td>0.73 1.03 0.36 0.12 0.19 0.08</td>
<td></td>
</tr>
<tr>
<td></td>
<td>20 Jan 2016</td>
<td>0.98 1.48 0.50 0.15 0.22 0.10</td>
<td></td>
</tr>
<tr>
<td>Deep</td>
<td>28 Feb 2015</td>
<td>0.29 0.41 0.15 0.10 0.13 0.06</td>
<td></td>
</tr>
</tbody>
</table>
As illustrated in Table 4, between the 2 types of events, the differences of average $dZ/dh$ in GC regions were significant, while the differences in St regions were relatively small. In the shallow events, the differences of average $dZ/dh$ between the 2 regions were greater than those in the deep events. The 2 types of events also had similarities: the average $dZ/dh$ values were usually 2–3 times (average 2.6) larger inside GCs compared to outside, and the differences between the inside and outside of the FSs were of similar magnitude.

### 5. Analysis of dynamical properties

5.1. Retrieval of dynamical properties

Vertical air motion is an important process in cloud dynamics, and is crucial for the formation and maintenance of GCs (e.g., Heymsfield, 1975; Hogan et al., 2002; Lothon et al., 2005). It is necessary to acquire $W_a$ values in the cloud in order to analyze the dynamical processes during the nucleation and growth of particles. The retrieval method of $W_a$ is introduced below.

5.1.1. Algorithm

The $V_r$ measured by vertically pointing radar is the sum of $V_z$ and $W_a$ (Matrosov et al. 2002). According to Rogers (1964), in order to obtain $W_a$, the relationship between $Z$ and $V_z$ first needs to be established, followed by estimating $V_z$ via the $Z$ measured by radar, correcting this estimation with air density, and finally calculating $W_a$ by subtracting the corrected $V_z$ from $V_r$. There are 2 main methods for acquiring the $Z$–$V_z$
relationship using a single radar. The first uses particle size distribution and the
relationship between $V_z$ and particle size ($D$) to deduce the $Z$–$V_z$ relationship (e.g.,
Rogers, 1964; Hauser and Amayenc, 1981; Ulbrich and Chilson, 1994). The second
consists of separating $V_z$ and $W$ by means of statistical methods along with the
addition of some hypothetical conditions (e.g., Orr and Kropfli, 1999; Matrosov et al.,
2002; Protat et al., 2003; Plana-Fattori et al., 2010). The $Z$–$V_z$ relationship for liquid
particles has been expressed clearly (e.g., Marks and Houze, 1987; Black et al., 1996;
Joss and Waldvogel, 1970), whereas the $V_z$ of solid particles is related to their habits
and densities, which increases the complexity of obtaining the $Z$–$V_z$ relationship.

As mentioned in section 2.1, VPR-CFMCW can output the precise power spectra of
the return signal. Since the minimum detection capability of VPR-CFMCW reaches
$-170$ dBm, the weak signals returned by small particles can be captured. Since small
particles at upper levels fall slowly, basically moving with the air motion, they can be
regarded as tracers of the mean vertical air motion (e.g., Gossard et al. 1997; Babb et
al. 1999; Kollias et al. 2001). When the vertical air motion is upward, these small
particles are driven to move upward, forming low spectral peaks in the negative
velocity region (downward $V_r$ values are positive) of the Doppler velocity spectra.
The upward small particles in the negative velocity region and the falling particles in
the positive velocity region form a distinctly bimodal spectrum. When the vertical air
motion is downward, bimodal spectra sometimes occur, depending upon the breadth
of the particle size distribution. In previous studies (e.g., Shupe et al. 2008; Zheng et
al. 2017), the identification and separation of bimodal Doppler velocity spectra were
often used to extract the $W_a$ in clouds. In our study, $W_a$ and $V_z$ were retrieved, and the $Z-V_z$ relationship established, using spectral data with distinctly bimodal distributions.

Figures 8a and 8c present the typical detection results of the bimodal spectra in the GC and St regions during 20 January 2016 event, respectively. Figure 8b is selected from the single modal spectra region between the 2 bimodal spectra regions. To clearly show the bimodal phenomenon, the ordinates in the figures are represented by normalized power and all Doppler spectra are smoothed using a three-point boxcar averaging window. The heights of the spectra are marked in the figures. Inspection of the spectra showed that the 1 peak in the GC region occurred in the positive velocity range and the other in the negative velocity range, as seen in Fig. 8a. In the St region, both peaks were primarily in the positive velocity range, as seen in Fig. 8c.

In the GC region, the spectral peak first appeared at negative (ascending) velocities, and also appeared at positive velocities with the reduction of height. Initially, the right spectral peak was small and the left peak was dominant. As the height decreased, the right spectral peak increased rapidly, eventually exceeding the left one, which gradually weakened and disappeared. The left peak was associated with small particles, could be interpreted as reflecting the vertical air motion. As the particles grew, those with fall speeds greater than the updraft velocities began to descend, leading to the formation of the spectral peak at positive velocities (the peak corresponds to the large particles, hereinafter referred as LP peak). As the height decreased, the sizes and masses of the particles gradually increased, leaving fewer small particles. This might explain the gradual disappearance of the left spectral peak.
and the dominance of the right peak. This result is consistent with the study by Shupe et al. (2004).

The spectral peak for $W_a$ in the St region appeared near 0 m s$^{-1}$ and was always located to the left of the LP peak. The $W_a$ peak became slightly enhanced with decreasing height, but was always weaker than the LP peak. The weaker peaks mostly appeared near $-8$ to $-3^\circ$C. Some relevant studies (e.g., Zawadzki et al. 2001; Oue et al. 2018) have also shown that bimodal spectra in snow were often observed near $-8$ to $-3^\circ$C, which corresponds to secondary ice generation via Hallett–Mossop process (Hallett and Mossop 1974), indicating the existence of supercooled droplets as well. The typical cloud droplets with negligible terminal fall speed ($V_t$) were regarded as tracers of vertical air motions by many studies (e.g., Shupe et al. 2004; Shupe et al. 2008). However, $Z$ of the weaker spectral peak in the St region was sometimes caused...
by cloud droplets and the small ice crystals produced by secondary ice generation, rather than just by the cloud droplets. In this case, the velocity corresponding to the weaker peak was the sum of downward air velocity and \( V_t \) of small secondary ice crystals, leading to the uncertainty in estimation of \( W_a \) and \( V_z \). The error estimation is discussed in section 5.1.2.

Since there are evident differences between the main spectral characteristics of the GC and St regions, in order to retrieve \( W_a \) and \( V_z \) more accurately, the \( W_a \) and LP peaks in the bimodal spectra of the 2 regions were identified respectively according to the following spectral peak-picking algorithm described in Shupe et al. (2004). Note that the criteria was adjusted because the cases, radar wavelengths and sampling parameters were different in our study. The criteria for picking secondary peaks was adjusted to at least 2 standard deviations of the noise greater than the spectral noise level (the original threshold was 2.5). Additionally, the width of spectral peaks was adjusted to at least 0.4475 m s\(^{-1}\) (or 4 continuous velocity bins) above the noise level (the original threshold was 7 continuous velocity bins).

When a \( W_a \) peak was identified in a given Doppler spectrum, the \( W_a \) value was then estimated by the offset of the spectral peak from 0 m s\(^{-1}\). The influence of \( W_a \) was then removed from the detected \( V_r \) in order to obtain a more accurate \( V_z \). Using the calculated \( V_z \) (in m s\(^{-1}\)) and the corresponding \( Z \) values (in mm\(^6\) m\(^{-3}\)), the \( Z-V_z \) power-law relationship with the generic form of \( V_z = aZ^b \) can be established by fitting (Note that the \( V_z \) values were referenced to ground level). The data points obtained using the bimodal spectra in the GC regions of shallow and deep events are...
shown in Figs. 9a and 9b, respectively. Equations (3) and (4) are the relationships that
were established for the GC regions in the shallow and deep events respectively (the
subscripts 's' and 'd' represent shallow and deep events respectively and the subscript
'1' represents GC region), represented by the blue solid lines in Figs. 9a and 9b:

\[ V_{Z1s} = 0.46Z^{0.08}, \quad (3) \]

\[ V_{Z1d} = 0.55Z^{0.11}. \quad (4) \]

The quality of fit for Eq. (3) was indicated by an R-square of 0.9172, with a
root-mean-square error (RMSE) of 0.093, which represented a reasonably good fit. As
for Eq. (4), the R-square was 0.9025 and RMSE was 0.090.

The data points obtained from the bimodal spectra in the St regions of shallow and
depth events are shown in Figs. 9c and 9d, respectively. The Z values of those weaker
peaks were below −20 dBZ. To reduce the retrieval error, the data obtained from
weaker peaks with Z lower than −25 dBZ were selected to establish the Z–Vc
relationships. Equations (5) and (6) are the relationships for the St regions in the
shallow and deep events respectively (the subscript '2' represents St region),
represented by the blue solid lines in Figs. 9c and 9d:

\[ V_{Z2s} = 0.78Z^{0.09}, \quad (5) \]

\[ V_{Z2d} = 0.80Z^{0.09}. \quad (6) \]

The quality of fit for Eq. (5) was indicated by an R-square of 0.9128, with a
root-mean-square error of 0.091. As for Eq. (6), the R-square was 0.8986 and RMSE
was 0.089.
Fig. 9. Fitting curve of the $Z-V_z$ relationships and comparison results with the relationships of other solid particles with different habits for the (a) GC region of shallow events, (b) GC region of deep events, (c) St region of shallow events, and (d) St region of deep events, DP represents the data points, FIT represents the fitting curve, COL represents hexagonal columns, ROS represents bullet rosettes, AGG represents aggregates, and PLA represents hexagonal plates. Among them, the $Z-V_z$ relationships of COL, ROS, PLA and AGG were presented by Protat and Williams (2011), and the relationship of snow was presented by Heymsfield et al. (2010).

Orr and Kropfli (1999) reported that, a single power-law equation cannot be used to produce profiles of $V_z$ throughout the depth of the cloud. There was usually a height
range without double peaks between the 2 regions with bimodal spectra. To obtain the continuous changes of $W_a$ and $V_z$ in the whole cloud, the height weighted algorithm presented by Heymsfield et al. (2010) was used to calculate $W_a$ and $V_z$ for the region without bimodal spectra. The corresponding height at top of the region was designated as $h_1$, and that at bottom of the region was designated as $h_2$. When a certain height ($h$) is between $h_1$ and $h_2$, the weights for Eqs. (3), (4) and Eqs. (5), (6) are given by Eq. (7) respectively:

$$w_1(h) = \frac{(h_1 - h)^2}{(h_1 - h_2)^2},$$

$$w_2(h) = 1.0 - w_1(h).$$

The $V_z$ in this region can be derived with the Eq. (8):

$$V_z(h) = w_1(h)V_{z1}(h) + w_2(h)V_{z2}(h),$$

where $V_{z1}(h)$, $V_{z1}(h)$ and $V_{z2}(h)$ are $V_z$, $V_{z1}$ and $V_{z2}$ at height $h$, respectively. The $Z$ values measured by the VPR-CFMCW during the 6 snow events in our study were substituted into Eqs. (3)–(6) to calculate the $V_z$ of the solid particles. Following the correction for air density, the $W_a$ in the GC and St regions can be retrieved with the following equations, respectively:

$$W_{a1}(h) = V_z(h) - V_{z1}(h)\left(\frac{\rho_0}{\rho(h)}\right)^{0.4},$$

$$W_{a2}(h) = V_z(h) - V_{z2}(h)\left(\frac{\rho_0}{\rho(h)}\right)^{0.4}.$$

where $h$ is height, $W_a(h)$, $V_z(h)$ and $V_{z}(h)$ are $W_a$, $V_z$ and $V_{z}$ at height $h$, respectively, the subscripts '1' and '2' represent GC and St region respectively, $\rho_0$ is the air density at the ground, and $\rho(h)$ is the air density at height $h$. Additionally, the $W_a$ in the regions without bimodal spectra can be retrieved with Eq. (11):
\[ W_0(h) = V_f(h) - V_z(h)\left(\rho_0 / \rho(h)^{0.4}\right). \]  \hspace{1cm} (11)

5.1.2. Test of the retrieval results

Equations (3)–(6) were compared with the \(Z-V_z\) relationships of other solid particles with different habits. Hong (2007) introduced 5 particle habit assumptions: hexagonal columns (COL), bullet rosettes (ROS), aggregates (AGG), hexagonal plates (PLA), and droxtals (DRO). Protat and Williams (2011) deduced 5 relationships for the above 5 particle habits using the maximum particle dimension of Heymsfield and Iaquinta (2000) and the radar backscattering coefficients of Hong (2007) to calculate the radar backscattering cross-section from the assumed ice particle size distribution. They also found that the use of these 5 habits was sufficient to represent the natural variability of solid particles. The \(a\) and \(b\) coefficients of the \(Z-V_z\) relationships corresponding to COL, PLA, and ROS were: \(a = 0.65, b = 0.10\); \(a = 0.44, b = 0.09\); and \(a = 0.52, b = 0.10\), respectively. The results of the comparison among Eqs. (3), (4) and the 3 relationships for COL, PLA, and ROS are shown in Figs. 9a and 9b. As can be seen in the figure, Eq. (3) was closest to the relationship corresponding to PLA and Eq. (4) was closest to the relationship corresponding to ROS. These results indicated that PLA might be the predominant particle type in the GC region of shallow events, while ROS might be the predominant particle type in the GC region of deep events. The inferences about particle habit are consistent with the observations presented by Plummer et al. (2014).

The \(a\) and \(b\) coefficients of \(Z-V_z\) relationship corresponding to rimed AGG were \(a = \ldots\)
0.78, b = 0.11 (Protat and Williams 2011). The a and b coefficients of the Z–Vz relationship corresponding to unrimed snow were a = 0.817, b = 0.063 (Heymsfield et al. 2010). The results of the comparison among Eqs. (5), (6) and the 2 relationships for AGG and snow are shown in Figs. 9c and 9d. As can be seen in the figure, the calculated Vz values from Eq. (5) basically fall between the Vz from the other 2 equations of AGG and snow. In cases with the same Z value, the maximum difference between Eq. (5) and the other 2 equations was nearly 0.1 m s⁻¹. The comparison result of Eq. (6) was similar to that of Eq. (5), but the Vz calculated by Eq. (6) was slightly larger for a same Z value. This result indicated that either unrimed snow or rimed AGG might be the predominant particle type in the St region.

As described in section 5.1.1, using Z–Vz relationships to estimate Wa and Vz in the St regions may produce errors, and the error estimation is the calculation of the Vt of small secondary ice crystals. Mossop (1976) have reported that on average, one ice splinter is thrown off for every 250 droplets of diameter ≥ 24 μm accreted. Since the droplets smaller than approximately 13 μm are also necessary for the Hallett–Mossop process (Mossop 1978), it is assumed that the average diameter of droplets in the cloud is 20 μm with backscattering cross-section about 1.99×10⁻¹⁵ mm². Many studies have shown that the secondary ice crystals are mainly needle and columnar crystals (e.g., Heymsfield and Willis 2014; Oue et al. 2018) with a mean particle size of 120 μm (Zawadzki et al. 2001). Taking long columnar ice crystal as an example, the backscattering cross-section is about 9.77 × 10⁻¹⁴ mm² calculated by the discrete dipole
approximation (DDA) method. The number concentration ratio of cloud droplets to ice crystals is supposed to be 250:1, so the Z ratio is calculated to be 5.1:1. The maximum Z value of the selected weaker spectral peak is −25 dBZ, and the Z of small ice crystals is estimated according to the Z ratio. Using Z of small ice crystals, backscattering cross-section calculated by DDA method, assumed ice particle size distribution (Hong 2007) and the relationship between \( V_I \) and maximum particle dimension for columns (Heymsfield and Iaquinta 2000), it can be estimated that the \( V_I \) of the secondary columns is about 0.16 m s\(^{-1}\). Since the Z values of selected weaker peaks in the St regions were below −25 dBZ, 0.16 m s\(^{-1}\) is the maximum error caused by ignoring the \( V_I \) of small secondary particles.

When the \( V_I \) of small ice crystals is not considered, the downward air velocity in the lower part of St region with bimodal spectra may be overestimated, resulting in the underestimation of \( V_Z \). In the upper part of St region without bimodal spectra, the underestimated \( V_Z \) calculated by the Z–\( V_Z \) relationship would lead to the underestimation of absolute value of upward air velocity (negative value).

The dynamical properties of the 2 cloud regions are analyzed in the following sections. Two typical events and time periods are selected to represent the shallow and deep events respectively.

5.2. Case study

5.2.1. Shallow event

Figure 10 shows the time–height contours of Z, \( W_a \) and \( V_z \) and contoured frequency...
by altitude diagrams (CFADs; Yuter and Houze, 1995) of \( W_a \) and \( V_z \) during
1230–1330 UTC of the 20 January 2016 event (beginning with the fifth range gate). As shown in Figs. 10b and 10d, \( V_z \) evidently accelerated within GCs, with speeds rapidly increasing from 0.2 m s\(^{-1}\) to over 0.6 m s\(^{-1}\). The \( V_z \) in the St region, which was mainly over 0.8 m s\(^{-1}\), was faster than that found in the GC region. The proportion for the velocity range of 0.8–1.0 m s\(^{-1}\) was largest, reaching more than 30%. As the height decreased, the proportion increased. FSs with enhanced Z values were shown clearly in Fig. 10a. The \( V_z \) within FSs was higher than that between them, with a rate above 1 m s\(^{-1}\).

As exhibited in Figs. 10c and 10e, GC locations usually correlated with strong upward air motions, having speeds up to 1.5 m s\(^{-1}\), while downward air motions often appeared in the regions between GCs. The convective activities were obvious in GC region. The high Z values in GC regions usually correspond to the large values of \( V_z \) and upward air motions. Upward motions play a crucial role in cold clouds, since they affect the supply rate of water vapor and the activation rate of ice nuclei (Lin et al. 1998). Stronger upward motions bring more water vapor and activate more ice nuclei. Hence, more ice crystals are formed by nucleation with faster growth, leading to higher Z values. Meanwhile, upward motions increased the probability of collisions among particles, which might promote the aggregation growth of particles and contribute to the increase of Z values. For shallow events, RH obtained by air sounding was usually close to or even more than 100% in GC regions, so there could also be a possibility of riming.
The upper part of the St region (~3.5–4.5 km) contained mainly upward motions, with speeds between 0.2 and 0.6 m s\(^{-1}\), corresponding to the magnitude of synoptic-scale rising motion, as analyzed in section 3.1. Below ~3.5 km, the weak downward air motions were dominated, with speeds slower than 0.5 m s\(^{-1}\). There was no apparent difference in \(W_a\) inside and outside the FSs, which was consistent with the finding of Plummer et al. (2015).

![Image](image_url)

**Fig. 10.** Analysis of the dynamical properties from 1230 to 1330 UTC on 20 Jan 2016; (a) \(Z\), (b) \(V_z\) (positive is downward), and (c) \(W_a\) (positive is downward); CFADs of (d) \(V_z\) and (e) \(W_a\). If the speeds are between −0.1 and 0.1 m s\(^{-1}\), the vertical air motions were considered to be negligible (grey areas in the Fig. 10c).

5.2.2. Deep event

Figure 11 shows the time–height contours of \(Z\), \(W_a\) and \(V_z\) and CFADs of \(W_a\) and \(V_z\) during 0020–0120 UTC of the 28 February 2015 event (beginning with the fifth range gate). The distribution characteristics of \(W_a\) and \(V_z\) in this event were similar to those
in the shallow event. To avoid repeating above description, the following discussion focused on the differences between the 2 types of events.

In the GC region, the maximum velocity of the downdraft was slower than that during the shallow event. In the St region, $V_z$ increased from 0.7 m s$^{-1}$ to 1.6 m s$^{-1}$, with the velocity range of 1.2–1.4 m s$^{-1}$ accounting for the largest proportion, over 40%. The upper part of the St region featured weak upward air motions (~5–8km), with speeds between 0.2 and 0.6 m s$^{-1}$, also corresponding to the magnitude of synoptic-scale rising motion, as analyzed in section 3.1. The weak downward air motions dominated the region below 5 km, with speeds slower than 0.5 m s$^{-1}$.

Fig. 11. Analysis of the dynamical properties from 0020 to 0120 UTC on 28 Feb 2015; (a) $Z$, (b) $V_z$ (positive is downward), and (c) $W_a$ (positive is downward); CFADs of (d) $V_z$ and (e) $W_a$. If the speeds are between −0.1 and 0.1 m s$^{-1}$, the vertical air motions were considered to be negligible (grey areas in the Fig. 11c).

In the St regions, there were few double peaks in the weak updraft layers (i.e.,
upper parts of the St regions). The vertical air velocities in these layers were obtained by subtracting $V_z$ calculated by the $Z$-$V_z$ relationships from the observed $V_r$. The radar observations showed that the $V_r$ values in the upper updraft layers were smaller than those in the lower downdraft layers, so the upward air velocities (negative values) occurred in the upper parts of St regions after retrieval. The $Z$ and calculated $V_z$ in the upper updraft layers were also smaller, so the negative air velocities were mainly caused by the observed slow $V_r$. Therefore, we believed that the existence of updraft layers was reasonable.

5.3. Characteristics analysis

In this section, all 6 events were statistically analyzed to discuss the similarities and differences for the characteristics of $W_a$ and $V_z$ in 2 regions of the 2 event types and to compare the differences between the inside and outside of the GCs along with FSs. The statistical characteristics of $W_a$ are summarized in Table 5. Since there was no apparent difference in $W_a$ inside and outside the FSs, while the upper and lower parts of St regions seemed to feature different $W_a$, the statistics on the averages and standard deviations (SDs) of $W_a$ in the upper and lower parts of St regions replaced the statistics on those inside and outside the FSs. The negative values represented the upward motions.

In the GC regions, the distribution range of $W_a$ was mainly between $-1.6$ and $1.5$ m s$^{-1}$ for the 2 types of events, which was similar to previous studies (e.g., Rosenow et al. 2014; Rauber et al. 2015). The upward air velocities within GCs for the 2 types
of events were similar, and the average speeds were mostly distributed around 1.2 m s\(^{-1}\). There was few difference in the downward air velocities between GCs for the 2 types of events, and the average speeds were mostly distributed around 1.0 m s\(^{-1}\). The standard deviations (SDs) inside GCs were larger than outside, with an average value of about 0.3 m s\(^{-1}\), while the SDs outside GCs were about 0.25 m s\(^{-1}\) on average, indicating that the distribution of \(W_a\) inside GCs was more discrete than outside.

Evident difference in SDs did not exist between the 2 types of events.

Table 5. The statistics on characteristics of \(W_a\) in GC and St regions for the 6 events.

<table>
<thead>
<tr>
<th>Event type</th>
<th>Date</th>
<th>Range</th>
<th>(W_a) in GC region</th>
<th>(W_a) in St region</th>
<th>Upper Range</th>
<th>Lower Range</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td></td>
<td>Avg. (SD)</td>
<td>Avg. (SD)</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td></td>
<td>inside GCs</td>
<td>outside GCs</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td></td>
<td>Avg. (SD)</td>
<td>Avg. (SD)</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Deep</td>
<td>27 Jan 2015</td>
<td>1.7 to 1.5</td>
<td>-1.2 (0.27)</td>
<td>1.1 (0.22)</td>
<td>-0.4 to 0.4</td>
<td>-0.2 (0.08)</td>
</tr>
<tr>
<td></td>
<td>28 Jan 2015</td>
<td>1.9 to 1.8</td>
<td>-1.3 (0.32)</td>
<td>1.2 (0.31)</td>
<td>-0.6 to 0.4</td>
<td>-0.4 (0.12)</td>
</tr>
<tr>
<td></td>
<td>29 Jan 2015</td>
<td>1.8 to 1.6</td>
<td>-1.1 (0.35)</td>
<td>1.0 (0.24)</td>
<td>-0.5 to 0.4</td>
<td>-0.3 (0.10)</td>
</tr>
<tr>
<td></td>
<td>20 Jan 2016</td>
<td>1.6 to 1.6</td>
<td>-1.2 (0.21)</td>
<td>1.2 (0.20)</td>
<td>-0.6 to 0.5</td>
<td>-0.3 (0.09)</td>
</tr>
<tr>
<td></td>
<td>28 Feb 2015</td>
<td>1.9 to 1.6</td>
<td>-1.2 (0.36)</td>
<td>1.0 (0.27)</td>
<td>-0.6 to 0.5</td>
<td>-0.4 (0.11)</td>
</tr>
<tr>
<td></td>
<td>24 Nov 2015</td>
<td>1.7 to 1.5</td>
<td>-1.1 (0.31)</td>
<td>0.9 (0.29)</td>
<td>-0.5 to 0.4</td>
<td>-0.3 (0.09)</td>
</tr>
</tbody>
</table>
In the St regions, the distribution ranges of $W_a$ were similar for the 2 types of events, mainly between $\pm 0.4$ m s$^{-1}$. The vertical air motions were much weaker in the St regions compared to the GC regions. The upper parts of the St regions consisted primarily of weak upward motions with the average speeds mostly distributing around 0.3 m s$^{-1}$, which was consistent with previous studies (e.g., Rauber et al. 2014b; Plummer et al. 2015), while weak downward motions dominated the lower parts with the average speeds mostly distributing around 0.2 m s$^{-1}$. The SDs of $W_a$ were smaller in the St regions compared to the GC regions, indicating that the distribution of $W_a$ in the St regions was more concentrated. The SDs of $W_a$ were similar in the upper and lower parts of St regions during the 2 types of events.

The statistical characteristics of $V_z$ were summarized in Table 6. To ensure retrieval accuracy, the minimal effective value for the $V_z$ retrieval result was set to 0.2 m s$^{-1}$. In the GC regions, the average speeds inside GCs were around 0.6 m s$^{-1}$, while those outside GCs were around 0.4 m s$^{-1}$, with the differences between 0.1 and 0.3 m s$^{-1}$. The SDs inside GCs were larger than outside, indicating that the distribution of $V_z$ inside GCs was more discrete than outside. The SDs inside and outside GCs were slightly larger during the shallow events compared to the deep events, indicating that the distribution of $V_z$ in the shallow events was more discrete.

In the St regions, $V_z$ values were generally larger than GC regions, mainly distributed above 0.8 m s$^{-1}$. The average speeds inside FSs were around 1.3 m s$^{-1}$, while those outside FSs were around 1.0 m s$^{-1}$, with the differences between 0.2 and 0.4 m s$^{-1}$. As reported by Locatelli and Hobbs (1974), the fall speeds of solid particles
are related to such as the maximum dimensions, mass, density, habits, and degree of riming of particles. The specific causes of the enhanced $V_z$ within GCs and FSs will be further analyzed in the next article on microphysical properties. At present, it can be concluded that the enhanced $V_z$ within GCs and FSs implied the more conducive conditions for particle growth existed in GCs and FSs. The characteristics of SDs are similar to those in GC regions, with larger SDs inside FSs and in shallow events.

Table 6. The statistics on the range of $V_z$ in GC and St regions along with the averages and SDs of $V_z$ inside and outside GCs and FSs for the shallow and deep snow events.

<table>
<thead>
<tr>
<th>Event type</th>
<th>Date</th>
<th>Range</th>
<th>Avg. (SD)</th>
<th>Avg. (SD)</th>
<th>Avg. (SD)</th>
<th>Avg. (SD)</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td></td>
<td>inside GCs</td>
<td>outside GCs</td>
<td>inside FSs</td>
<td>outside FSs</td>
<td></td>
</tr>
<tr>
<td>Shallow</td>
<td>27 Jan 2015</td>
<td>0.2–0.9</td>
<td>0.7 (0.19)</td>
<td>0.4 (0.08)</td>
<td>0.7–1.6</td>
<td>1.3 (0.14)</td>
</tr>
<tr>
<td>Shallow</td>
<td>28 Jan 2015</td>
<td>0.2–1.0</td>
<td>0.6 (0.23)</td>
<td>0.4 (0.09)</td>
<td>0.9–1.8</td>
<td>1.4 (0.17)</td>
</tr>
<tr>
<td>Shallow</td>
<td>29 Jan 2015</td>
<td>0.2–0.8</td>
<td>0.5 (0.16)</td>
<td>0.3 (0.06)</td>
<td>0.8–1.7</td>
<td>1.3 (0.15)</td>
</tr>
<tr>
<td>Shallow</td>
<td>20 Jan 2016</td>
<td>0.2–1.0</td>
<td>0.6 (0.21)</td>
<td>0.4 (0.08)</td>
<td>0.8–1.5</td>
<td>1.2 (0.13)</td>
</tr>
<tr>
<td>Deep</td>
<td>28 Feb 2015</td>
<td>0.2–0.8</td>
<td>0.5 (0.15)</td>
<td>0.3 (0.04)</td>
<td>0.8–1.7</td>
<td>1.3 (0.10)</td>
</tr>
<tr>
<td>Deep</td>
<td>24 Nov 2015</td>
<td>0.2–0.7</td>
<td>0.5 (0.13)</td>
<td>0.4 (0.05)</td>
<td>0.7–1.6</td>
<td>1.2 (0.09)</td>
</tr>
</tbody>
</table>

6. Conclusions
The continuous observation data of 6 snow events was obtained by VPR-CFM CW with extremely high resolution during the winter of 2015–2016 in middle latitudes of China. The GCs described in previous research had been found near the cloud tops in every event. Four of the 6 events were shallow processes, with the echo top mainly below 8 km. The other 2 were deep processes, with the echo top mostly above 10 km. The vertical fine structure and dynamical properties inside the 2 types of snow clouds were analyzed. The main conclusions of this study may be summarized as follows:

1) According to the characteristics of echo distribution and the changes of $dZ/dh$ and $dV_r/dh$, snow clouds could be divided into GC regions and St regions. For the shallow events, statistical analyses of the GC regions indicated that they typically have a depth of 500–1200 m. The depths of GC regions accounted for about 11–14% of the whole cloud, while the average contributions to the growth of $Z$ values reached 42% in these regions. GCs occurred up to 23 times within 1 h, resulting in dense fibrous FSs below. As for the deep events, the depths of GC regions were greater, reaching over 2 km. The times of GC appearance within 1 h were about one third to one half of those in the shallow events, leading to the relatively sparse FSs and uniform St regions below. The depths of GC regions accounted for about 16% of the whole cloud, while the average contributions to the growth of $Z$ values reached 33% in these regions. The contributions of GC regions to $Z$ value growth were larger than those to ice mass growth described in some previous studies (e.g., Matejka et al. 1980; Houze et al. 1981; Plummer et al. 2015). In the deep events, the proportions of GC regions were slightly larger, but the average contributions to the growth of $Z$ values
were lower than those in the shallow events. The depths of GC regions are in good accord with previous studies (e.g., Plummer et al. 2014, Rosenow et al. 2014).

2) Between the 2 types of events, the differences of average $dZ/dh$ in GC regions were significant, while the differences in St regions were relatively small. In the shallow events, the differences of average $dZ/dh$ between the 2 regions were greater than those in the deep events. The 2 types of events also had similarities: the average $dZ/dh$ values were usually 2–3 times (average 2.6) larger inside GCs compared to outside, and the differences between the inside and outside of the FSs were of similar magnitude.

3) Both updraft and downdraft were alternately observed in GC regions. The distribution range of $W_a$ was mainly between $-1.6$ and $1.5 \, \text{m s}^{-1}$ for the 2 types of events, which was similar to previous studies (e.g., Rosenow et al. 2014, Rauber et al. 2015). GC locations usually correlated with strong upward air motions, while downward air motions often appeared in the regions between GCs. The air velocities in GC region for the 2 types of events were similar, with average velocities mostly distributed around $-1.2$ and $1.0 \, \text{m s}^{-1}$ respectively. The standard deviations (SDs) of $W_a$ inside GCs were larger than outside, indicating that the distribution of $W_a$ inside GCs was more discrete than outside. Evident difference in SDs of $W_a$ did not exist between the 2 types of events.

In the St regions, the speeds of $W_a$ were mainly within $0.5 \, \text{m s}^{-1}$. There was no apparent difference in $W_a$ inside and outside the FSs, which was consistent with the finding of Plummer et al. (2015). The upper parts of the St regions consisted primarily
of weak upward motions, which was in good accord with previous studies (e.g.,
Rauber et al. 2014b; Plummer et al. 2015), while weak downward motions dominated
the lower parts. The SDs of $W_a$ were smaller in the St regions compared to the GC
regions, indicating that the distribution of $W_a$ in the St regions was more concentrated.
The SDs of $W_a$ were similar in the upper and lower parts of St regions during the 2
types of events.

4) In the GC regions, the SDs of $V_z$ inside GCs were larger than outside, indicating
that the distribution of $V_z$ inside GCs was more discrete than outside. The SDs of $V_z$
inside and outside GCs were slightly larger during the shallow events compared to the
deep events, indicating that the distribution of $V_z$ in the shallow events was more
discrete. In the St regions, the characteristics of SDs of $V_z$ are similar to those in GC
regions, with larger SDs of $V_z$ inside FSs and in shallow events. The average speeds
were slightly faster inside GCs and FSs compared to outside, with the differences of
0.1–0.3 m s$^{-1}$ and 0.2–0.4 m s$^{-1}$ respectively. The enhanced $V_z$ within GCs and FSs
implied the more conducive conditions for particle growth existed in GCs and FSs.

Using $Z$–$V_z$ relationships to estimate $W_a$ and $V_z$ in the St regions may produce errors,
and 0.16 m s$^{-1}$ is the maximum error caused by ignoring the $V_t$ of small secondary ice
crystals related to the weaker spectral peaks.

The results of this study have clarified findings of the past half-century. The $W_a$ and
$V_z$ values were retrieved more precisely using bimodal spectra from VPR-CFMCW.
Moreover, characteristics of GC regions, as well as the average reflectivity gradients
and dynamical properties inside and outside GCs and FSs, were quantified. The
results in this research may improve the understanding of the differences between the inside and outside of GCs and FSs during the 2 types of snow events in terms of vertical reflectivity gradients and dynamical properties. The vertical structure and continuous evolution of $W_a$ and $V_z$ in the snow clouds were also exhibited more clearly. The microphysical properties and physical mechanisms of the snow processes induced by GCs will be further analyzed using in situ observation data and simulation models based on this study.

Acknowledgment

This work has been supported by the National Key Research and Development Program of China under Grant 2017YFC1501703, National Natural Science Foundation of China (41675029, 41975046), research project of State Key Laboratory of Severe Weather (LaSW). We would like to thank LetPub (www.letpub.com) for providing linguistic assistance during the preparation of this manuscript. The assistance of Jisong Sun (a researcher at State Key Laboratory of Severe Weather, Chinese Academy of Meteorological Sciences) in analyzing the environmental background is greatly appreciated.

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