

1    **Microscale Updrafts Within Northeast U.S. Coastal Snowstorms Using High-**  
2    **Resolution Cloud Radar Measurements**

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## ABSTRACT

Limited knowledge exists about  $\sim 100$  m scale precipitation processes within U.S. Northeast coastal snowstorms because of a lack of high-resolution observations. We investigate characteristics of microscale updraft regions within the cyclone comma head and their relationships with snowbands, wind shear, frontogenesis, and vertical mass flux using high-spatiotemporal resolution vertically-pointing Ka-band radar measurements, soundings, and reanalysis data for four snowstorms observed at Stony Brook, NY. Updraft regions are defined as contiguous time-height plotted areas with upward Doppler velocity without hydrometeor sedimentation that is equal to or greater than  $0.4 \text{ m s}^{-1}$ . Most updraft regions in the time-height data occur on a time scale of seconds ( $< 20$  s), which is equivalent to spatial scales  $< 500$  m. These small updraft regions within cloud echo occur more than 30% of the time for three of the four cases and 18% for the other case. They are found at all altitudes and can occur with or without frontogenesis and with or without snowbands. The updraft regions with relatively large Doppler spectrum width ( $> 0.4 \text{ m s}^{-1}$ ) occur more frequently within midlevels of the storms, where there are strong wind shear layers and moist shear instability layers. This suggests that the dominant forcing for the updrafts appears to be turbulence associated with the vertical shear instability. The updraft regions can be responsible for upward mass flux when they are closer together in space and time. The higher values of mean upward mass flux within updraft regions often occur during snowband periods.

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## SIGNIFICANCE STATEMENT

Small-scale ( $< 500$  m) upward motions within four snowstorms along the U.S. Northeast Coast are analyzed for the first time using high spatiotemporal resolution millimeter-wavelength cloud radar pointed vertically. The analysis reveals these updrafts appear in the storms regardless of whether snowbands are present or whether there is larger-scale forcing for ascent. The more turbulent and stronger updrafts frequently occur in midlevels of storms associated with instability from vertical shear and contribute to upward mass flux during snowband periods when they are closer together in space and time.

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45 **1. Introduction**

46

47 Snowbands in the comma head of winter storms are responsible for much of the heavy snowfall  
48 over the northeast United States. Because of the large societal impacts of these winter storms, they  
49 have been studied for decades using both numerical models and observations (e.g., Novak et al.  
50 2008; Stark et al. 2013; Plummer et al. 2014; McMurdie et al. 2022). These studies have shown a  
51 spectrum of snowbands, ranging from single bands, which are defined as a single reflectivity  
52 feature  $> 250$  km in length, 20–100 km in width, and with intensities of  $> 30$  dBZ maintained for  
53 at least 2 hours, to smaller-scale multibands that occur in groups (Ganetis et al. 2018). The  
54 mechanisms that result in the variations in snowband characteristics are not well known.

55 Previous studies on snowband formation focused on the mesoscale ingredients of lift along a  
56 mid-level trough, instability, and moisture (Novak et al. 2010). While primary bands were often  
57 associated with mid-level frontogenesis (e.g., Novak et al. 2004, 2008), multibands were also  
58 associated with mesoscale instabilities, such as boundary layer instability, moist/conditional  
59 symmetric instability (e.g., Byrd 1989; Schultz and Schumacher 1999), conditional instability  
60 (e.g., Reuter and Yau 1990; Trapp et al. 2001; Morales 2008) and inertial instability (e.g., Jurewicz  
61 and Evans 2004; Schultz and Knox 2007). Earlier modeling and theoretical studies of multibands  
62 focused on the superposition of a frontogenetical circulation and either conditional instability or  
63 conditional symmetric instability (Xu 1992). However, Ganetis et al. (2018) showed that these  
64 bands often develop in a region of little or no frontogenesis; thus, other mechanisms may be  
65 responsible, such as elevated convection, generating cells, shear instabilities, and gravity wave  
66 activity, as well as microphysics, which often occur at smaller scales than frontogenesis and  
67 mesoscale instabilities (Bosart and Sanders 1986; Zhang et al. 2001, 2003; Kumjian et al. 2014;  
68 Plummer et al. 2014, 2015; Rauber et al. 2014, 2017; Rosenow et al. 2014, 2018; Keeler et al.  
69 2016a, b, 2017; Kumjian and Lombardo 2017; Lackmann and Thompson 2019; McMurdie et al.  
70 2022).

71 Snowbands also have variable microphysics that can impact precipitation rates. Studies using  
72 in-situ surface measurements observed a variety of snowflake habits and degrees of riming within  
73 the cyclone comma head and storm evolution (e.g., Stark et al. 2013; Colle et al., 2014). In  
74 particular, the microphysical processes can change across the snowband from more rimed on the  
75 east (warmer) side to more dry snow on the west (colder) side. Those microphysical studies using

76 in-situ surface measurements have also revealed that upper-level cloud structures, dynamics, and  
77 microphysics (i.e., generating cells, turbulence) strongly impact the surface precipitation, while  
78 internal cloud processes can interact with each other producing complex microphysics. Field  
79 campaigns using high spatiotemporal resolution airborne radars (e.g., ProfiLing Of Winter Storms  
80 [PLOWS], e.g. Rauber et al. 2014) revealed microscale convective updrafts producing generating  
81 cells, which contributed to greater ice production by vapor diffusion, riming, and aggregation  
82 processes (Plummer et al. 2014). Kumjian and Lombardo (2017) observed planar crystal growth  
83 and precipitation-type transitions (snow/rain/ice) in snowbands using the dual-polarization  
84 capabilities from the WSR-88D radar network. They also found a signature of secondary ice  
85 production. Those detailed, complex microphysical evolution in winter storms might not be  
86 captured by regional models (e.g., the Rapid Refresh model reanalysis with 50 vertical levels and  
87 13-km horizontal grid spacing, Kumjian and Lombardo 2017) and even by cloud resolving models,  
88 likely due to deficiencies in many of the bulk microphysical schemes for winter storms (e.g.,  
89 Naeger et al. 2017;2020; Molthan et al. 2016). Motivated by those gaps in understanding  
90 microphysics and less representativeness in numerical models, the field campaign, the  
91 Investigation of Microphysics and Precipitation of Atlantic Coast-Threatening Snowstorms  
92 (IMPACTS, McMurdie et al. 2022) focuses on improving the understanding of snowfall processes,  
93 remote sensing of snow, and the prediction and evolution of banded structures. The campaign  
94 achieved multi-scale observations including in-situ and vertically-pointing and scanning remote  
95 sensing measurements from both airborne and ground-based platforms.

96 Previous observational studies (e.g., Plummer et al. 2015; Rauber et al. 2014; Rauber et al.  
97 2017) corroborated the importance of turbulence to generate ice particles and intensify snowfall.  
98 Turbulence often has been observed by fine-scale Doppler radar observations to occur in multiple  
99 layers within stratiform snow clouds with frequent periodic upward/downward patterns within  
100 stratiform precipitation clouds without terrain forcings (e.g., Rauber et al. 2017). The upward  
101 motions could contribute to supply water vapor and hence further ice formation and/or growth  
102 resulting in generating cells (Kumjian et al. 2014) and intensifications of snow fallstreaks  
103 (Plummer et al. 2015; Rauber et al. 2017), as shown by prominent radar reflectivity compared to  
104 the background. One of the common factors that generate instability and turbulence in stratiform  
105 clouds is wind shear (i.e. Kelvin–Helmholtz instability). Previous studies such as Boucher et al.  
106 (1965), Wexler et al. (1967), and Syrett et al. (1995) observed wind shear varying in time in winter

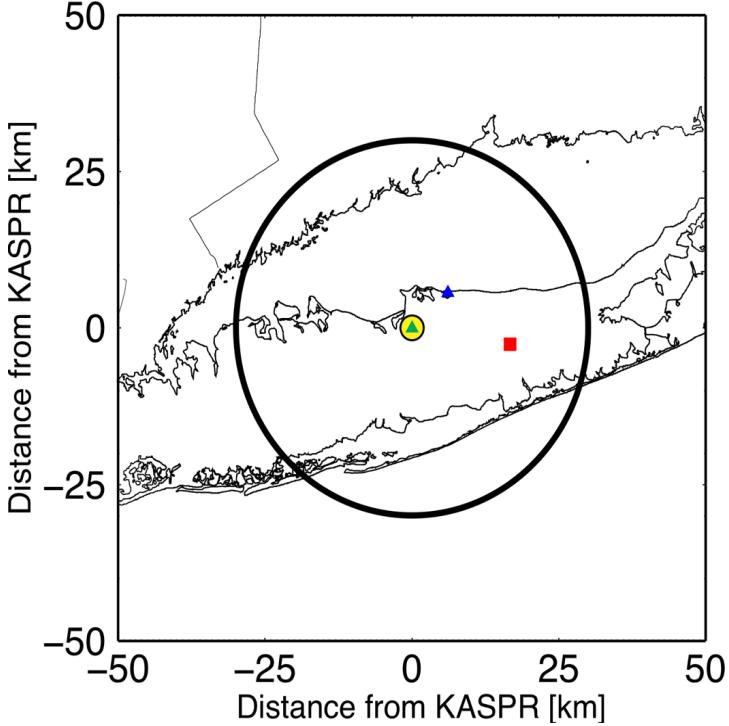
107 storms using Doppler radar measurements and suggested that the wind shear could play a role in  
108 the formation of generating cells. Rauber et al. (2014) revealed multiple sources of air masses with  
109 different wind directions and humidity associated with a comma head snowstorm. The multi air  
110 masses controlled cloud structure including cloud depth, instability, and vertical wind shear and  
111 produced cloud-top turbulence and generating cells. Despite the importance of generating cells, as  
112 well as turbulence that contributes to vertical air motion, for snowfall intensification, the formation  
113 process of the turbulent layers and the roles in snowfall formations are still unclear. Such  
114 turbulence structures are usually not resolved by either regional models or operational radars.

115 In this study, we hypothesize that 1) this upward motion associated with fine-scale turbulence  
116 contributes to snowband formation and 2) the turbulence producing the updrafts is generated by  
117 wind shear and/or thermodynamic instability in association with Kelvin–Helmholtz instability.  
118 Particularly we investigate i) frequency of upward motion component of fine-scale turbulence  
119 during the individual storms, ii) contribution of the updrafts to the upward mass transport, iii)  
120 relationships amongst the fine-scale upward motion, wind shear, and thermodynamic stability, and  
121 mesoscale forcing for ascent (e.g. frontogenesis).

122 The Ka-band Scanning Polarimetric Radar (KASPR) is an ideal remote sensing instrument to  
123 study the fine scale kinematic and microphysical characteristics of winter storms (e.g. Oue et al.  
124 2017; Kollias et al. 2020). KASPR has been part of the Stony Brook University and Brookhaven  
125 National Laboratory Radar Observatory (SBRO) since 2017 and is installed at the Stony Brook  
126 University site (Fig. 1). KASPR polarimetric and Doppler capabilities have revealed fine scale  
127 dynamical and microphysical features within winter storms (Kumjian et al. 2020; Lamer et al.  
128 2021; Oue et al. 2021).

129 This study uses data from four winter storms and focuses on characteristics of microscale (< 1  
130 km) updraft regions, their relationship with ambient conditions, and their role in vertical mass  
131 transport. Section 2 describes the datasets used in this study. Section 3 summarizes the  
132 meteorological context and evolution of the four events and the corresponding KASPR  
133 observations. The relationships among the observed updraft structures, precipitation features, and  
134 other storm parameters important for forcing for ascent (e.g., frontogenesis, Petterssen, 1956) and  
135 wind shear are discussed in section 4. Finally, summary and conclusions are presented in Section  
136 5.

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138  
 139  
 140 Figure 1: Locations of KASPR (the center of this display, yellow dot)  
 141 the nearest NWS sounding site  
 142 (OKX, red square), and two SBU mobile sounding sites (SBU by green triangle and Cedar Beach by blue  
 143 triangle). A large circle represents the KASPR's 30-km radius maximum observation range.

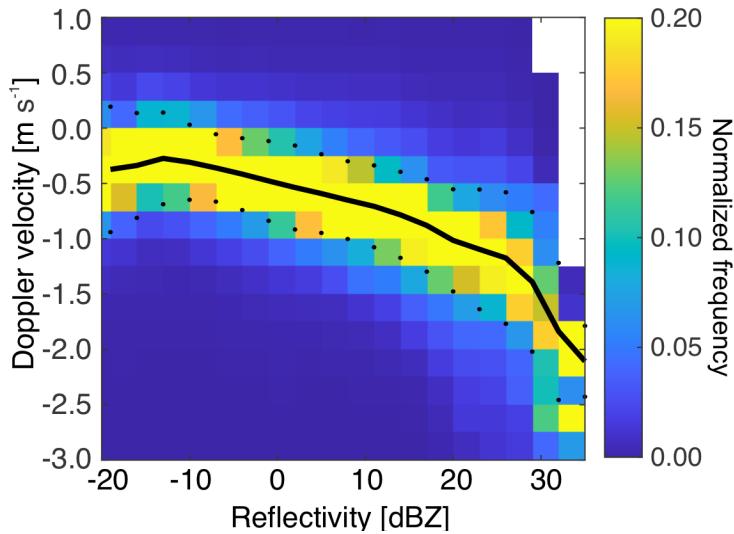
## 144 2. Data and Methods

### 145 a. *Ka-band Scanning Polarimetric Radar (KASPR)*

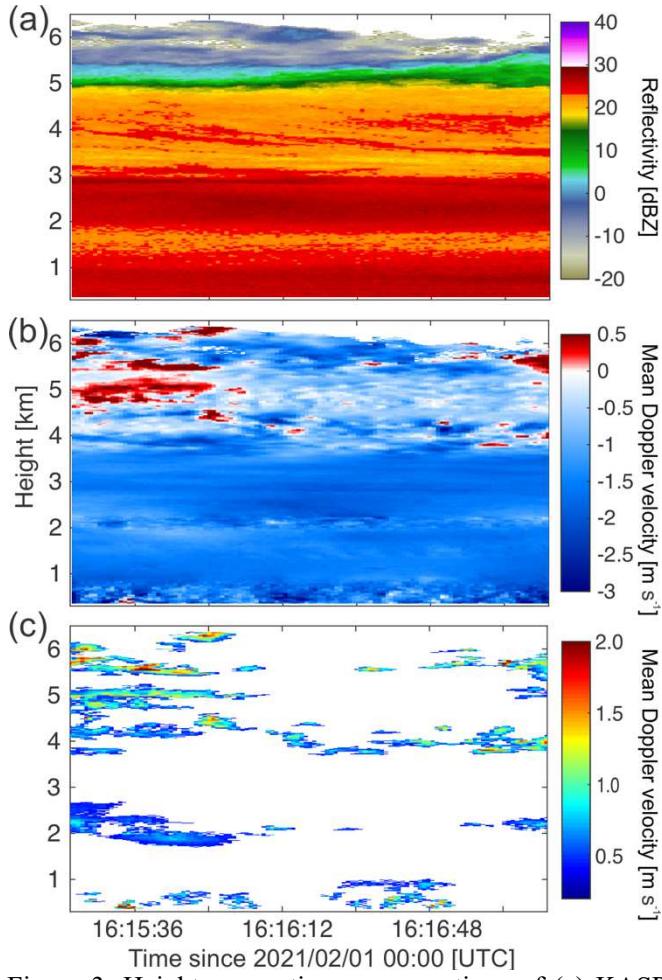
146 KASPR is a state-of-the-art 35-GHz cloud scanning radar with a beamwidth of  $0.32^\circ$ , capable  
 147 of collecting reflectivity, Doppler velocity, Doppler spectrum width, and the standard set of  
 148 polarimetric radar variables. The KASPR power measurements are calibrated using a corner  
 149 reflector technique. The detailed specification of KASPR is available in Kumjian et al. (2020) and  
 150 Kollias et al. (2020).

151 During winter storm observations, KASPR executed a scanning strategy that consisted of a  
 152 Plan Position Indicator (PPI) surveillance scan at  $15^\circ$  elevation angle (and  $20^\circ$  for the 2018 winter),  
 153 a zenith pointing ( $90^\circ$  elevation angle) PPI for calibration, horizon-to-horizon Range-Height  
 154 Indicator (RHI, Kollias et al., 2014) scans, and a vertically pointing mode (VPT). Following  
 155 Kumjian et al. (2020), this pattern was repeated and took approximately 13-15 minutes to  
 156 complete, providing the slant PPI surveillance scans every 7 minutes (15 minutes for the 2018  
 157 winter). RHI scans crossing and/or along the snowbands were performed in each cycle producing  
 158

160 RHI scans toward the same direction every 40 seconds to 3.5 minutes (40 seconds to 15 minutes  
 161 for the 2018 winter). The PPI and RHI scans were performed with a full polarimetry mode and  
 162 scan speeds of  $6^{\circ} \text{ s}^{-1}$  and  $2^{\circ} \text{ s}^{-1}$ , respectively, to collect data with a 30-m range-gate spacing,  $0.6^{\circ}$   
 163 PPI azimuthal spacing and  $0.3^{\circ}$  RHI elevation spacing. The VPT mode was executed with only  
 164 horizontally polarized waves transmitted and both horizontally and vertically polarized waves  
 165 received. The KASPR radar was in VPT mode for consecutive periods lasting 2 to 5 min with a  
 166 15-m range-gate spacing. Based on the beamwidth, the horizontal resolution at a 10 km range is  
 167 approximately 56 m. This study did not apply attenuation corrections for hydrometeors, and  
 168 carefully selected periods when ice precipitation dominates through the cloud to avoid significant  
 169 attenuation by liquid precipitation. If the column included liquid precipitation signatures, which  
 170 could be defined by the presence of melting layer(s) in the VPT measurements, the column is  
 171 excluded from the analysis.



172  
 173 Figure 2: Reflectivity versus mean Doppler velocity from all KASPR VPT data used in this study from the  
 174 selected four cases. Color shade represents frequency normalized every 3 dB. Solid and dotted lines represent  
 175 mean Doppler velocity every 3 dB and mean Doppler velocity  $\pm$  standard deviation, respectively. Negative  
 176 Doppler velocity values indicate downward motion. Doppler velocity is scaled for air density using the nearest  
 177 sounding data.



178  
179 Figure 3. Height-versus-time cross sections of (a) KASPR reflectivity and (b) mean Doppler velocity from  
180 vertically-pointing measurements on February 1, 2021 from 16:15:18 to 16:20:10 UTC, (c) the estimated upward  
181 vertical air motion  $> 0.4 \text{ m s}^{-1}$  (estimated after removing hydrometeor fall speed from the measured Doppler  
182 velocity). Very small updraft regions  $< 15$  range-bins are not included in the analysis.  
183

184 The VPT measurements were used to identify updraft regions (URs) in time-height plots based  
185 on mean Doppler velocity. First, Doppler velocity with reflectivity  $< -20 \text{ dBZ}$  was removed from  
186 the analysis since it was too noisy. Then the hydrometeor sedimentation component (i.e. fall speed)  
187 was removed from the measured Doppler velocity following Protat and Williams (2011) using  
188 reflectivity versus Doppler velocity (Z-V) relationships. In this study, the hydrometeor fall speed  
189 was estimated as the median Doppler velocity for every 3 dB from  $-20 \text{ dBZ}$  to  $34 \text{ dBZ}$ , assuming  
190 that the mean vertical air motion for a certain time period at a given height was  $\sim 0 \text{ m s}^{-1}$ . Whenever  
191  $> 100$  samples were available in a bin and median Doppler velocity was negative, we estimated the  
192 Z-V relationship every 500 m using hourly VPT data and used the median value of Doppler  
193 velocity as the hydrometeor fall speed at each Z bin at each height-hour window. For bins with few

194 observations (<100 samples) or a positive median Doppler velocity (ascent), we instead used the  
195 statistical value of the median Doppler velocity estimated from all VPT data shown in Fig. 2 (black  
196 line) as the hydrometeor fall-speed estimate. To avoid accounting for range gates not representing  
197 updrafts due to the variability of the Doppler velocity, we considered updraft regions to be range  
198 gates where the measured Doppler velocity minus the estimated hydrometeor fall speed was  
199 greater than equal to  $0.4 \text{ m s}^{-1}$ . The threshold of  $0.4 \text{ m s}^{-1}$  is the value of standard deviation of  
200 Doppler velocity in Fig. 2. For example, in Fig. 2, for -1 dBZ the estimated fall speed is  $-0.48 \text{ m s}^{-1}$   
201 so measured Doppler velocity values  $> -0.08 \text{ m s}^{-1}$  are considered to be updrafts. For 20 dBZ  
202 the estimated fall speed is  $-1.01 \text{ m s}^{-1}$  so measured Doppler velocity values  $> -0.61 \text{ m s}^{-1}$  are  
203 considered updrafts. We defined the updraft region (UR) as a region with at least 15 updraft range-  
204 gates connected in time or height. Small regions having less than 15 range-bins are removed from  
205 the analysis.

206 Figure 3 shows an example of 5-min VPT reflectivity (Fig. 3a), Doppler velocity (Fig. 3b), and  
207 the identified updrafts from 16:15:18 UTC for 5 minutes on 1 February 2021. We tested other  
208 Doppler velocity thresholds ( $0.0 \text{ m s}^{-1}$  and  $0.6 \text{ m s}^{-1}$ ) that were applied to the Doppler velocity after  
209 the sedimentation (i.e. fall speed) removal and confirmed that varying the threshold within this  
210 range had little impact on the results. For example, when the threshold is increased from  $0.4 \text{ m s}^{-1}$   
211 to  $0.6 \text{ m s}^{-1}$ , the total number of URs decreases by 28-49%; however, the shapes of the normalized  
212 size distribution (discussed in Section 4) and vertical distribution do not change. Because of the  
213 uncertainty of the detected numbers of URs that depend on the thresholds, the analysis in this study  
214 focuses on qualitative descriptions rather than quantitative discussion.

215 The duration of the detected URs is the time between the earliest and latest times of the  
216 appearance of the UR region. The vertical extent is the height between the lowest and highest  
217 range-bins of the appearance of the region. The altitude of individual updraft regions is estimated  
218 as a mean of altitudes of range gates in the updraft regions (Fig. 3c). Fine scale turbulence can  
219 contribute to Doppler spectrum width (SW, Appendix A). For the VPT measurements using the  
220 narrow beamwidth ( $0.3^\circ$ ) in this study, the wind shear component in the observed SW would be  
221 mostly dominated by horizontal/vertical gradients in vertical air motion within the radar range  
222 gate. We use  $SW > 0.4 \text{ m s}^{-1}$  from the VPT measurements to represent the fine-scale turbulence  
223 (Appendix A). The Doppler spectrum width (SW) from the VPT measurements is averaged in each  
224 UR.

225 To understand the role of URs on snowfall intensification, we estimate the upward mass flux  
 226 based on the detected URs. The bulk mass flux (MF) at each altitude can be estimated as:

227  $MF = \bar{W} \rho_d UF \quad [\text{kg m}^{-2} \text{ s}^{-1}] \quad (1)$

228 where  $\bar{W}$  is the mean updraft that is the positive Doppler velocity from the detected updraft regions  
 229 averaged over time at each altitude,  $\rho_d$  is the dry air density estimated from the nearest-time  
 230 sounding measurements, and  $UF$  is the updraft fraction estimated as the ratio of the time of updraft  
 231 (positive Doppler velocity in the detected updraft regions) to the total cloudy time for each VPT  
 232 file (bulk mass flux). The mass flux profile from each VPT file is then averaged  $> 1.2$  km above  
 233 ground level (AGL) for each VPT file (column mean mass flux).

234 Because we highlight the importance of shear-driven turbulence, we compute vertical wind  
 235 shear, which measures the likelihood of turbulence. The Kelvin–Helmholtz instability can appear  
 236 when the Richardson Number is small ( $< \sim 0.25$ ). Here we introduce the moist Richardson Number  
 237 ( $mR_i$ ) to represent the instability in this cloudy environment. First, we estimate wind direction and  
 238 speed from KASPR PPI measurements at an elevation angle of  $15^\circ$  (approximately every 7  
 239 minutes) using a velocity-azimuth display (VAD) technique (Browning and Wexler, 1968). Using  
 240 the VAD data up to maximum height of 7.8 km AGL, we estimated the vertical wind shear  
 241 ( $V_{shear}$ ):

242 
$$V_{shear} = \sqrt{(u_{z_2} - u_{z_1})^2 + (v_{z_2} - v_{z_1})^2} / (z_2 - z_1) \quad (2)$$

243 where  $z$  represents height and  $u_z$  and  $v_z$  represent horizontal wind components at  $z$ . We use a 100  
 244 m (124 m for the January 4 case) spacing to estimate  $V_{shear}$  at each VAD data point ( $z_2 - z_1 = 100$   
 245 m). Since only a single elevation angle is used for the VAD, the resulting horizontal wind profile  
 246 is based on an increasing diameter cone with increasing height ( $\sim 7.5$  km diameter for 1 km altitude  
 247 and  $\sim 59.7$  km for 8 km altitude). The moist Richardson Number ( $mR_i$ ) is calculated using the  
 248 following equation (Markowski and Richardson, 2010):

249 
$$mR_i = \frac{N_m^2}{\left(\frac{\Delta u}{\Delta z}\right)^2 + \left(\frac{\Delta v}{\Delta z}\right)^2} \quad (3)$$

250  $N_m$  is the Brunt–Väisälä frequency defined as:

251 
$$N_m = \sqrt{\frac{g}{\theta_e} \frac{\Gamma_m}{\Gamma_d} \frac{\Delta \theta_e}{\Delta z}} \quad (4)$$

252

253 where  $\theta_e$  is equivalent potential temperature,  $\Gamma_d$  and  $\Gamma_m$  are dry adiabatic lapse rate and pseudo  
254 adiabatic lapse rate, respectively, and  $g$  is Earth's gravity.  $\theta_e$  and  $\Gamma_m$  were estimated from the  
255 nearest soundings.  $u$  and  $v$  are horizontal wind components, which can be obtained from soundings  
256 or the KASPR VAD measurements. A  $\Delta z=100$  m (124 m for the January 4 case) was used in this  
257 study.  $\Gamma_m$  was estimated using the following equation (Markowski and Richardson, 2010):

$$258 \quad \Gamma_m = \frac{g + \frac{dL_v r_v}{dz}}{c_{pd} + r_v c_l} \quad (5)$$

259 where  $L_v$  is the latent heat of vaporization,  $r_v$  is the water vapor mixing ratio,  $c_{pd}$  is the specific  
260 heat of dry air at constant pressure, and  $c_l$  is the specific heat capacity of liquid water.

261

262 *b. Sounding and WSR-88D radar data*

263 Twice-daily radiosonde data at 0000 and 1200 UTC were used from the nearest NWS site  
264 (OKX; red square in Fig.1), which is ~22 km to the east of Stony Brook. For the events in 2020  
265 and 2021 additional soundings were launched every ~3 hours using the GRAW sounding system  
266 installed on a Stony Brook University (SBU) mobile radar truck. The SBU mobile radar truck was  
267 deployed in several locations in Long Island including Cedar Beach (40.965N, -73.030E; blue  
268 triangle in Fig. 1; for 18 January 2020) and Stony Brook University (40.897N, -73.127E; green  
269 triangle in Fig. 1; for 17 December 2020 and 1 February 2021).

270 To provide the regional precipitation context for these snow events, we used the radar  
271 reflectivity from the NEXRAD WSR-88D surveillance scans at the lowest two elevation angles  
272 (i.e., 0.5 and 0.8°) at the KOKX site.

273

274 *c. Reanalysis data*

275 The Rapid Refresh reanalysis (RAP; Benjamin et al. 2016) was used to examine the  
276 environment and frontogenesis during the snowstorm events. It uses hourly-updated data  
277 assimilation with 37 pressure levels and 13.5-km grid spacing.

278 The mean sea level pressure from the fifth generation of the European Centre for Medium-  
279 Range Weather Forecasts (ERA5; Hersbach et al. 2020) was used to track cyclone centers. It  
280 provides hourly data interpolated into 37 pressure levels at 0.25° grid spacing. The cyclone  
281 tracking was based on the mean sea level pressure value less than 1010 hPa, gradients of the mean

282 sea level pressure, and the distance from the previous cyclone center, following Crawford et al.  
283 (2021).

284

285 *d. Snow events*

286 We examined four snowstorm cases: 4 January 2018; 18 January 2020; 16 – 17 December  
287 2020; and 31 January to 1 February 2021. Three cases (4 January 2018; 16 – 17 December 2020;  
288 and 31 January to 1 February 2021) had multibands, while the other case (18 January 2020)  
289 included a primary band (Table 1). The track of the surface low pressure centers for the storms are  
290 shown in Fig. 4. The cases were chosen to provide a diverse set of band structures within the  
291 cyclone comma head. Table 1 summarizes the storm characteristics and sample size for each case.  
292 Brief descriptions of the cases and synoptic conditions will be shown in Section 3b.

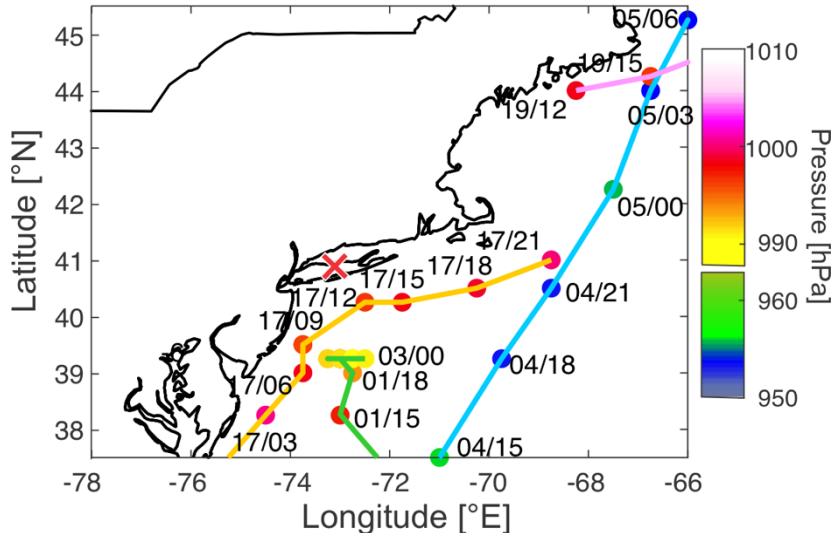
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	Storm Quadrant	Types of bands within storm	Trend in storm Z intensity as it passed over Stony Brook (based on NEXRAD)	Period	Total cloudy time analyzed by VPT [min]	No. of VPT files used	No. of VAD profiles (PPI data files) used	Sounding time (day/hour) [UTC]
4 Jan 2018	S	Multibands	Sustain Z >30 dBZ	10:09 – 23:55 UTC	278.2	56	138	04/12, 05/00
18 Jan 2020	NW~W	Single-band	Weaken from 30 dBZ to < 25 dBZ.	14:10 – 23:59 UTC	196.7	80	80	18/12, 18/18, 18/19, 18/21, 19/00
16 – 17 Dec 2020	N~NW	Multibands	Sustain Z >30 dBZ	18:22 UTC on Dec. 16 – 02:59 UTC on Dec. 17	184.1	75	76	16/12, 17/00, 17/03
31 Jan – 1 Feb 2021	N	Multibands	Weakening from 40 dBZ to 30 dBZ.	18:02 UTC on Jan. 31 – 23:51 UTC on Feb. 1	236.4	119	170	31/12, 01/00, 01/06, 01/08, 01/12, 01/15, 01/18, 01/21, 02/00

294

295 Table 1. Summary of the storm characteristics and sample sizes. Soundings at 00 and 12 UTC are from the  
296 NWS OKX soundings, and the others are from the SB mobile truck.

297  
298



299  
300 Figure 4: Tracks of the cyclone centers every three hours using ERA5 mean sea level pressure for the cases of  
301 4 Jan 2018 (blue line), 18 Jan 2020 (magenta line), 17 – 16 Dec 2020 (orange line), and 31 Jan–1 Feb 2021  
302 (green line). Color for each dot represents the surface central pressure (in hPa) every three hours. Cross mark  
303 represents the Stony Brook location.

304

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### 3. Results

308

#### 309 *a. Characteristics of detected updraft regions*

310 Table 2 lists the number of the URs detected within the cloud echo observed by the KASPR  
311 VPT measurements, and those normalized by the total cloudy time per 5 minutes (number per 5  
312 min), which is defined as an accumulated time where KASPR observed cloud echoes at any  
313 altitude during the analysis period. The warm frontal band case on 18 January 2020 has the lowest  
314 number of the updraft regions per 5 min among the four cases (~10 per 5 min). The other cases  
315 have 2-5 times more URs.

316 We identified the URs below ~1.2 km AGL for all cases. These are likely associated with the  
317 boundary layer turbulence. To focus on updrafts that may have strong relationships with storm  
318 evolution, wind shear, and shear-induced turbulence in clouds, we do not include the boundary  
319 layer updraft regions in the present analysis. The height of the planetary boundary layer (PBL) was  
320 determined as the bottom of the temperature inversion layer near the surface using the sounding

321 profiles. They are 0.5 km for 4 January, 0.9 km for 18 January, 1.1 km for 16-17 December, and  
322 0.7 km for 1 February 2021. The numbers of URs above the PBL are also listed in Table 2.

323

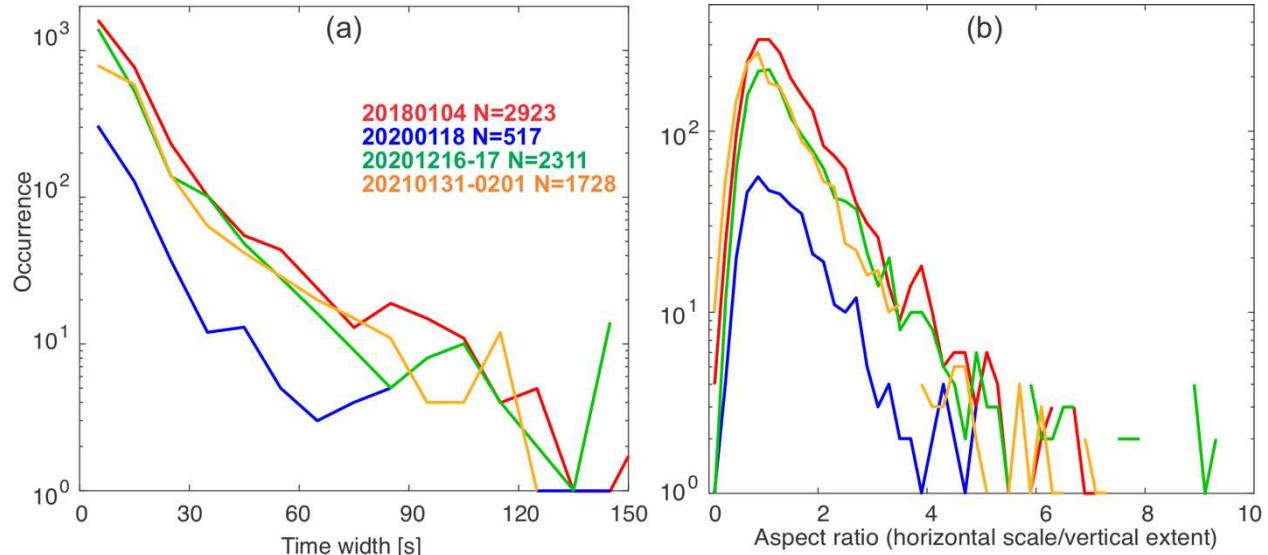
	Total number	Above PBL	Average Per 5 min	Normalized number per 5 min above PBL	Number with SW > 0.4 m s <sup>-1</sup> for height above PBL
Jan 4, 2018	3376	2923 (86.6%)	60.7	52.5	1142 (39.1%)
Jan 18, 2020	673	517 (76.8%)	17.2	13.2	21 (4.1%)
Dec 16 – 17, 2020	2774	2311 (83.3%)	75.3	62.8	667 (29.9%)
Jan 31 – Feb 1, 2021	2171	1728 (79.6%)	45.9	36.5	542 (31.4%)

324 Table 2. The number of detected updraft regions in total and normalized per 5 minutes (number per 5 min). The  
325 numbers of detected updraft regions counted for heights above the planetary boundary layer (PBL) are also  
326 listed. The percentages in the parenthesis in the right most column represents those of the total number of URs  
327 for height above the PBL.

328

329 Figure 5a shows frequency distributions of the duration of URs detected above the PBL within  
330 a 10-sec interval. Even though the number of detected URs varies by event, the qualitative  
331 characteristics of the duration time distribution is similar for all four cases. There is a peak at the  
332 smallest size bin for the four cases, and overall, the number exponentially decreases with size.  
333 Most of the updraft regions last for less than 20 sec (i.e., small size in the height-time plots),  
334 accounting for approximately 80% of the total. The maximum duration bin is 95 sec. This  
335 distribution shape and qualitative features do not change when different thresholds are used for  
336 mean Doppler velocity (Sect. 2a). The distribution curves from the three cases that had more than  
337 1700 updraft regions show similar exponential distribution. Considering that the horizontal extent  
338 of UR can be roughly estimated as duration  $\times$  horizontal wind speed estimated from the KASPR  
339 VAD, approximately 85% of URs have the horizontal scale  $< 500$  m. It is highly possible that  
340 portions of the URs passed through the radar site rather than the part of the maximum dimensions  
341 and the horizontal scale could be an underestimate. Although it is difficult to estimate URs using  
342 the tilted scans, the KASPR RHI measurements showed cell-like features with horizontal scale  $<$   
343 500 m. The mean vertical extent of URs from the VPT measurements is 261 m for all cases, and  
344 more than 81% of URs have vertical extents less than 330 m. The aspect ratio of the URs (defined

345 as the ratio of the horizontal scale to the vertical extent) for each case shown in Fig. 5b has a  
 346 lognormal frequency distribution with the frequency peak around 1 (i.e., circular).



347  
 348 Figure 5: Distributions of (a) time-width of the URs with the time bin size of 10 s for 4 Jan 2018 (red), 18 Jan  
 349 2020 (blue), 16 – 17 Dec 2020 (green), and 31 Jan – 1 Feb 2021 (orange) and (b) aspect ratio of the URs  
 350 defined as the ratio of horizontal scale to the vertical extent with the aspect ratio bin size of 0.2. The time width  
 351 for each updraft region is defined as the time between the time when the updraft region appeared first and the  
 352 time when it disappeared. The horizontal scale is estimated using the time width and the horizontal wind speed  
 353 estimated from the KASPR VAD measurements. The sample size for each case is shown in the right corner in  
 354 (a).  
 355

356  
 357 Table 3 summarizes the characteristics of the detected URs for each case. The URs have a  
 358 median vertical extent of 150 – 240 m. We classified the regions for smaller and larger SW from  
 359 the KASPR VPT measurements with a threshold of  $0.4 \text{ m s}^{-1}$ ; UR with  $\text{SW} < 0.4 \text{ m s}^{-1}$  named  
 360 URSWLO and UR with  $\text{SW} > 0.4 \text{ m s}^{-1}$  named URSWHI. Higher values of  $\text{SW} > 0.4 \text{ m s}^{-1}$  can be  
 361 a proxy for higher magnitudes of turbulence (Appendix A). The URSWHIs are thicker than  
 362 URSWLOs for all cases, but the difference varies between 0 m (16 – 17 December) and 90 m (18  
 363 January). Considering the range-gate spacing for the VPT measurements, the difference in the  
 364 number of range-gates for the vertical extent is less than 6. These qualitative characteristics do not  
 365 change significantly when the Doppler velocity threshold for the detection is changed to 0.2 and  
 366  $0.6 \text{ m s}^{-1}$ .  
 367  
 368

	Median duration (sec)		Median vertical extent (m)		Median maximum upward motion (m s <sup>-1</sup> )		Median reflectivity (dBZ)	
	URSWHI	URSWLO	URSWHI	URSWLO	URSWHI	URSWLO	URSWHI	URSWLO
4 Jan 2018	7.2	9.3	180	150	1.26	0.74	16.64	8.36
18 Jan 2020	6.2	8.3	165	150	1.15	0.78	19.95	5.78
16 – 17 Dec 2020	7.2	8.3	180	180	1.08	0.88	13.00	-3.36
31 Jan – 1 Feb 2021	9.3	11.4	165	135	1.29	0.77	7.77	4.32
All cases	7.2	9.3	165	150	1.18	0.79	14.45	4.24

369 Table 3. Median duration, vertical extent, maximum upward motion (maximum Doppler velocity without  
370 estimated fall speed in each UR), and reflectivity from the KASPR VPT moments for UR with SW>0.4 m s<sup>-1</sup>  
371 (URSWHI) and SW<0.4 m s<sup>-1</sup> (URSWLO) found at heights > 1.2 km.

372  
373

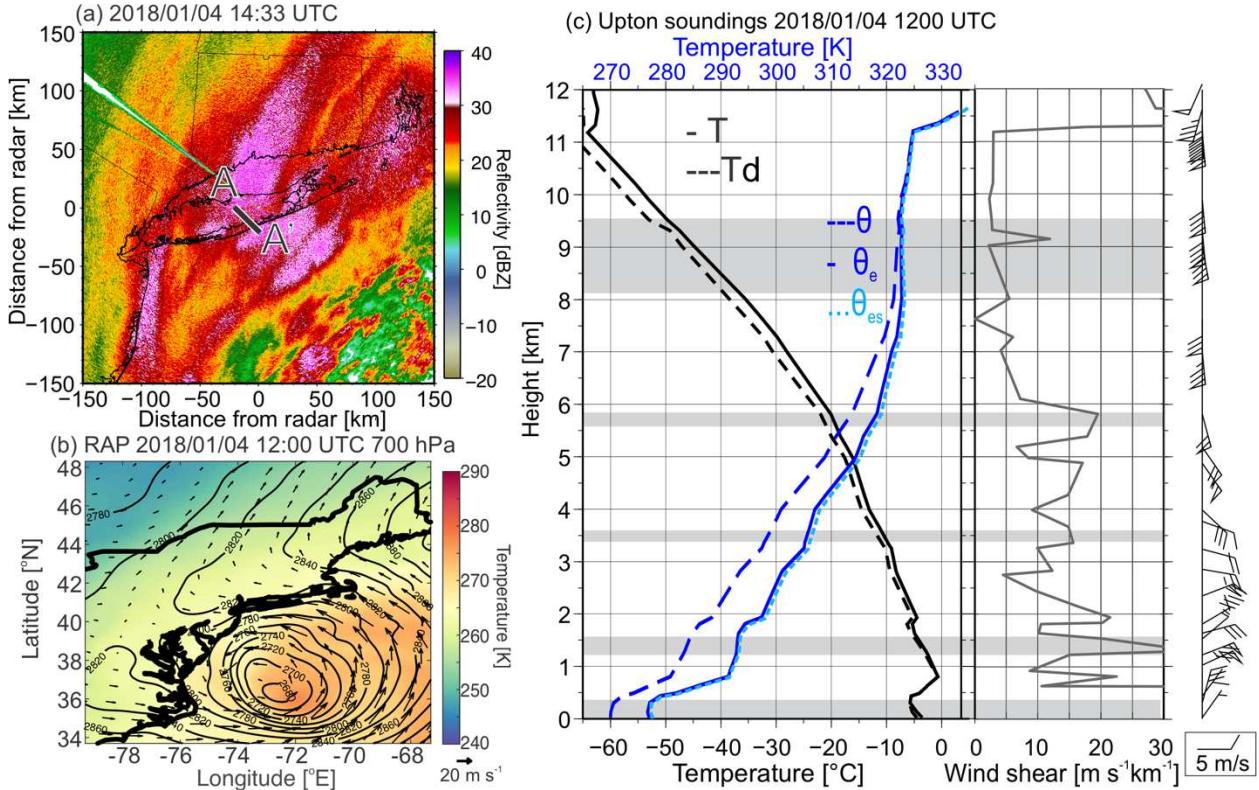
374 *b. Spatial distributions of updraft regions*

375 *1). 4 January 2018*

376 At 1200 UTC 4 January 2018, a deep surface cyclone (~968 hPa) was located a few hundred  
377 kilometers to the southeast of Long Island, NY. As the storm center moved from south of Long  
378 Island to the northeast (blue track in Fig. 4), the snowbands within the comma head produced  
379 heavy snow across Long Island from 1000 UTC 4 January to 0100 UTC 5 January 2018. Figure 6  
380 shows the horizontal distributions of the NEXRAD reflectivity at the time of snowbands, 700 hPa  
381 RAP, and the 1200 UTC sounding profile. From 0900-1730 UTC, narrow snowbands (to 25 dBZ)  
382 orientated north-northeast to south-southwest crossed SBU (Fig. 6a), about 300 km northwest of  
383 the surface cyclone (Fig. 4). The snowband pivoted and changed its motion from westward to  
384 eastward around 1730 UTC, as the cyclone center moved ~500 km southeast of Long Island. The  
385 snowbands passed over SBU from west to east around 2030-2100 UTC. After the passage of the  
386 snowband, several relatively weak reflectivity bands < 25 dBZ, oriented along north-to-south or  
387 north-northwest-to-south-southeast passed through SBU until 0200 UTC on 5 January 2018. At  
388 700 hPa (Fig. 6b), southeasterly wind with the closed low resulted in warm advection extending  
389 northward to Long Island, likely producing vertical wind shear.

390 The NWS sounding at 1200 UTC on 4 January shows a gradual wind veering from northeasterly  
391 below 2.5 km above mean sea level (ASL) to southerly at 6-km ASL (Fig. 6c), while there was a  
392 stable layer with a shallow frontal zone from 0.4-0.8 km ASL. From Eq. 1, there were relatively

393 large vertical wind shear layers ( $> 15 \text{ m s}^{-1} \text{ km}^{-1}$ ) below 6 km ASL (e.g., 0.8, 1.3, 1.9, 3.5, 4.6, and  
 394 5.7 km, Fig. 5c). They generally corresponded to heights of positive gradient of temperature (black  
 395 line) and  $\theta_e$  (blue solid line), and a few of them corresponded to the  $\text{mRi} < 0.25$  (indicated by gray  
 396 shades in Fig. 6c).

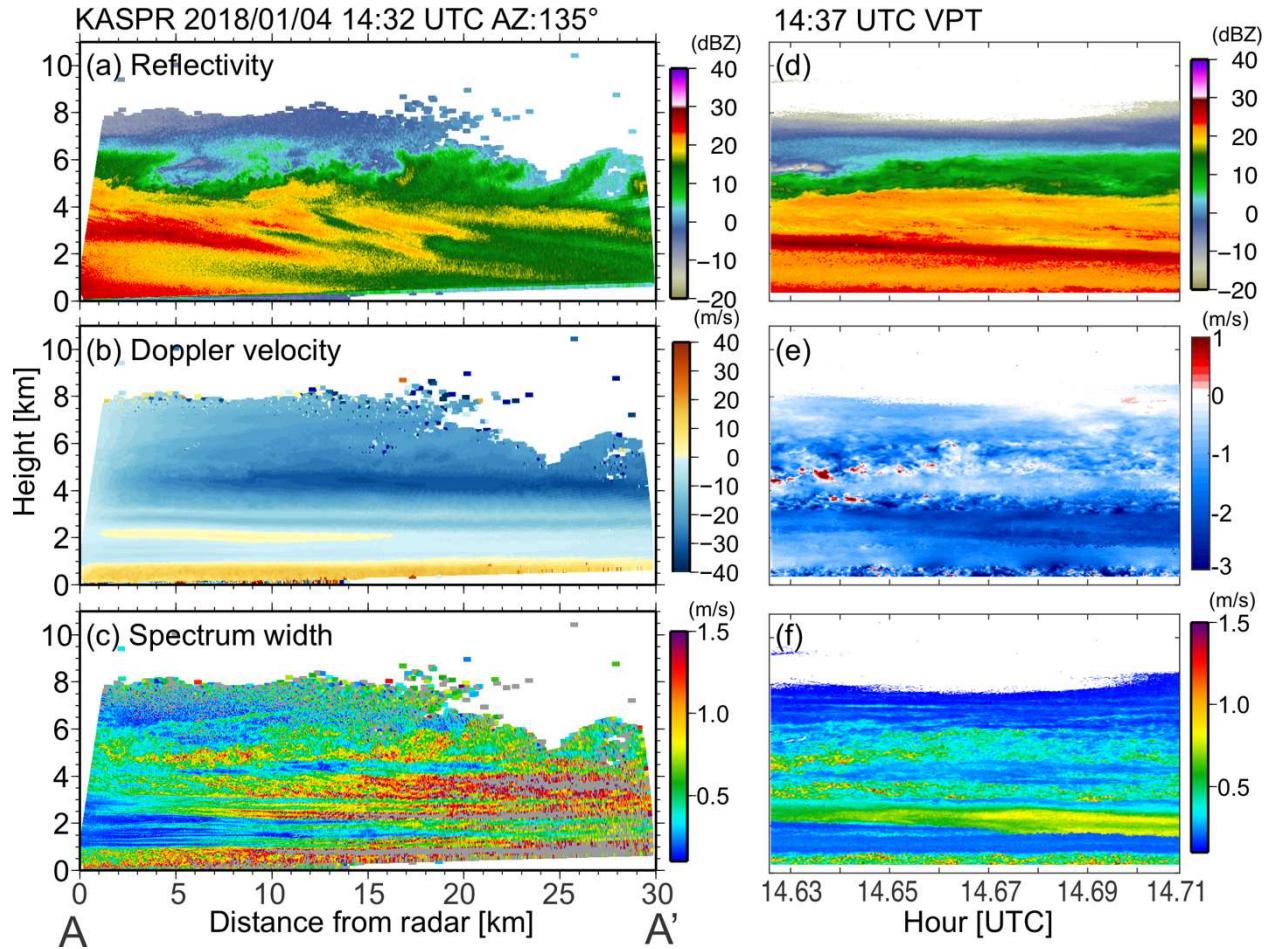


397  
 398 Figure 6: (a) Horizontal distribution of KOKX reflectivity from a PPI scan at an elevation angle of  $0.5^\circ$  at 1433  
 399 UTC on 4 Jan 2018, (b) 700-hPa temperature (color shade), geopotential height (contour) and horizontal wind  
 400 (arrows) from RAP reanalysis data at 12 UTC on 4 Jan 2018. (c) Sounding profiles of temperature (black solid  
 401 line), dew point temperature (black dashed line), potential temperature ( $\theta$  blue dashed line), equivalent  
 402 potential temperature ( $\theta_e$  blue solid line), saturated equivalent potential temperature ( $\theta_{es}$  light blue dashed  
 403 line), and vertical wind shear estimated from Eq. 1 (gray line in the right panel) at 1200 UTC from OKX. Gray  
 404 shaded layers in (c) represent layers of moist Richardson Number ( $\text{mRi} < 0.25$ ). The A – A' line in (a)  
 405 represents the KASPR RHI direction shown in Fig. 7a-c.  
 406

407 Figure 7 shows vertical cross sections of the snowbands from the KASPR RHI and height-  
 408 versus-time plots during the snowband from the VPT measurements. The KASPR RHI  
 409 measurements across the snowbands at 1432 UTC ( $135^\circ$  in azimuth, Fig. 7a-c) reveal individual  
 410 cells between 4-7 km altitudes and fallstreaks underneath. The KASPR reflectivity and Doppler  
 411 velocity from the VPT measurements (Fig. 7d,e) also reveal updrafts in the convective cell layer  
 412 and the fallstreaks attaining the 30-dBZ reflectivity. This convective cell layer at 4-7 km is  
 413 collocated with a relatively large SW of approximately  $0.4 \text{ m s}^{-1}$  in VPT (Fig. 7f). The SW layers

414 presented in the VPT measurements are consistent with the vertical wind shear shown at 5-6 km  
 415 ASL in the RHI Doppler velocity A-A' (Fig. 7b) and sounding profile (Fig. 6c). They suggest a  
 416 role of turbulence in forming convection and intensifying snow.

417



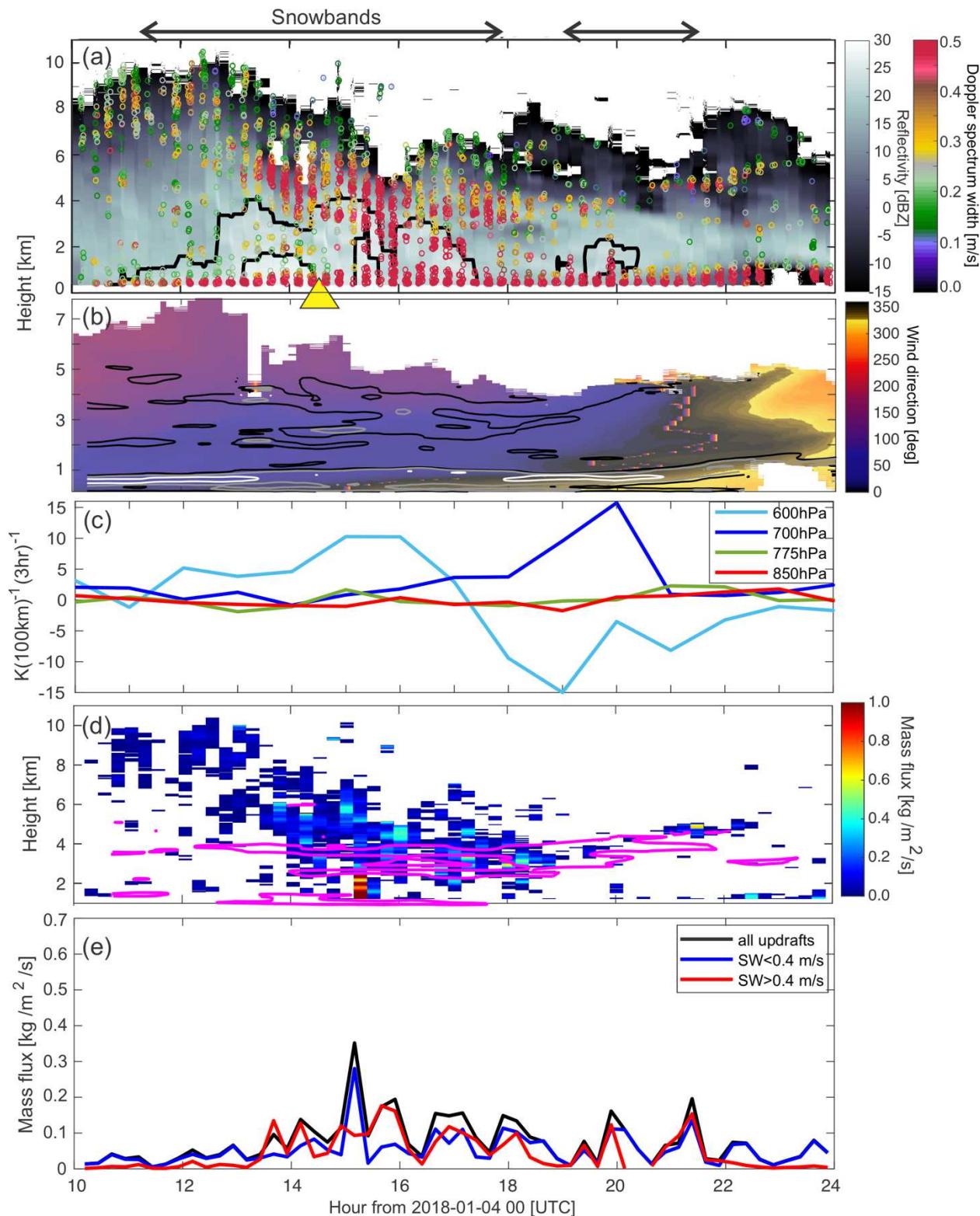
418

419 Figure 7: Vertical cross sections of KASPR (a) reflectivity (shaded in dBZ), (b) mean Doppler velocity in  $\text{m s}^{-1}$   
 420 , (c) Doppler spectrum width in  $\text{m s}^{-1}$  from an RHI scan along A-A' line in Fig. 6a at 1432 UTC, and height-  
 421 versus-cross sections of KASPR (d) reflectivity, (e) mean Doppler velocity, (f) Doppler spectrum width from  
 422 the VPT measurements.

423

424 To investigate the spatial characteristics of URs associated with wind shear, we show locations  
 425 of the detected updraft regions (dots) in the height-time cross section fields for individual cases in  
 426 Fig. 8a, with color shade representing SW from the KASPR VPT measurements and grayscale  
 427 representing KASPR reflectivity from VPT. We also show wind directions (color shade) and  
 428 vertical wind shear (contours) from the KASPR VAD analysis in Fig. 8b. To examine a relation  
 429 with a large scale forcing, we also show the time series of frontogenesis at 600, 700, 775, and 800

430 hPa (Bluestein, 1993) using RAP reanalysis data in Fig. 8c. The  $mRi < 0.25$  calculated from  
 431 sounding and KASPR VAD data is shown in Fig. 8d by magenta contours.



432

433 Figure 8: (a) Height-versus-time cross sections of KASPR reflectivity from VPT (gray scale) and locations of  
434 detected updraft regions (color dots) for 4 January 2018. Black contours represent KASPR reflectivity  $>20$  dBZ  
435 with 5 dB increment. Horizontal arrows on the top represent the period of snowband passages at SBU. Yellow  
436 triangle indicates the time of the snowbands near SBU shown in Fig. 4 and RHI measurements shown in Fig. 7.  
437 (b) Height-versus-time cross sections of wind direction (color shade) and vertical wind shear of 0.02, 0.03, and  
438  $0.05 \text{ s}^{-1}$  (black, gray, and white contours, respectively) estimated from KASPR VAD. (c) Time series of  
439 frontogenesis at four levels at four levels estimated using RAP reanalysis data for 4 January 2018. (d) Height-  
440 versus-time cross sections of the bulk mass flux (color shade) and mRi smaller than 0.25 (magenta contour). The  
441 mass flux is averaged every VPT file at each height. (e) Time series of column mean mass flux estimated from  
442 the detected updraft regions averaged over altitude. The mass flux is averaged every VPT file and then averaged  
443 over altitude. Black line represents the column mean mass flux from the contributions of all detected updraft  
444 regions, and blue and red lines represent that of updraft regions with  $\text{SW} > 0.4 \text{ m s}^{-1}$  and  $< 0.4 \text{ m s}^{-1}$ , respectively.  
445

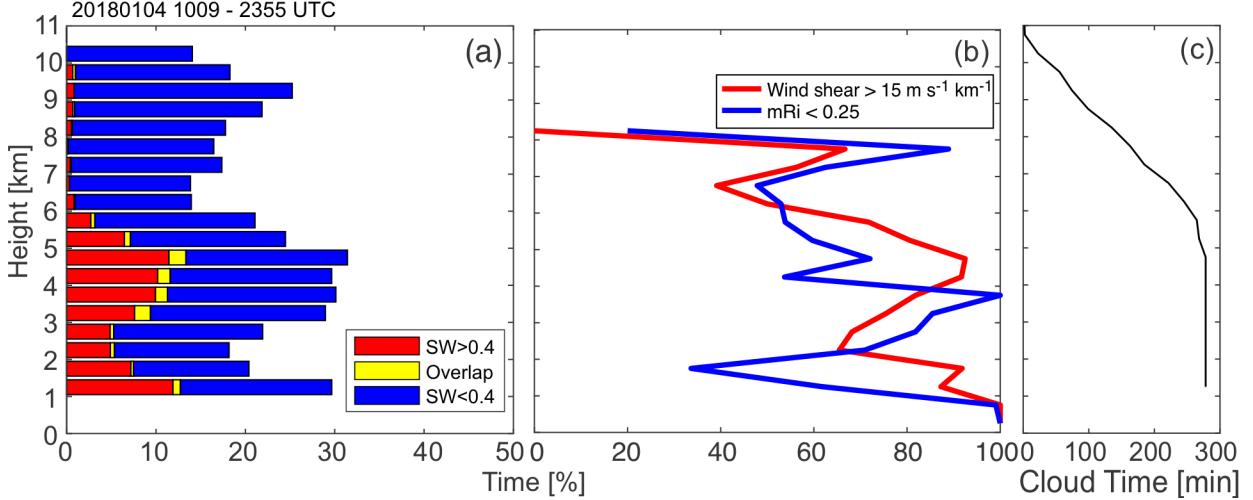
446 URs are more frequently observed during the first snowband period (1110–1800 UTC),  
447 suggesting a contribution of URs to the snowbands, as shown in the previous studies. The URs are  
448 also observed outside the snowbands. A difference between the URs during the first snowband and  
449 those outside the first snowband period is that the URs during the first more accompanied  
450 URSWHIs above 2 km ASL (Fig. 8a, represented by brown-red dots). The URSWHIs appear in  
451 the mid-level of the cloud, consistent with the earlier snowband period producing reflectivity  $>20$   
452 dBZ below 4 km altitude, which is 5-dB stronger than that of the latter snowband (Fig. 8a). The  
453 URSWHI height is 6 km at 1300 UTC and then descends to 3.5-4 km at 1800 UTC (Fig. 8a).  
454 URSWLOs (represented by yellow-green dots) are found at 2-km below the cloud top before the  
455 snowband period from 1000 UTC through the entire period of precipitation until 0000 UTC on 5<sup>th</sup>.  
456 From the KASPR VAD (Fig. 8b), the period (1300 - 1800 UTC) when URSWHIs dominate the  
457 mid-levels corresponds to the easterly wind associated with the cyclone below 5 km. This low-  
458 level wind produces a vertical wind shear with the upper-level southerly wind shown in the  
459 sounding (Fig. 6c). The low-level easterly flow corresponds to the veering wind around the storm  
460 center accompanying a warm air shown at 700 hPa. The URSWHIs are also collocated with  
461 vertical wind shear  $> 20 \text{ m s}^{-1} \text{ km}^{-1}$  in the KASPR VAD (Fig. 8b, black contour). This is consistent  
462 with the low mRi  $< 0.25$  (Fig. 8d, contour) suggesting shear instability. The number of the URs  
463 decreases after 1830 UTC close to the end of the first snowband passage and the frontal passage.  
464 The low mRi is still present at 4 km altitude, but the wind direction changes to north to  
465 northwesterly. The earlier snowband period (1110–1800 UTC) is well associated with 600-hPa  
466 frontogenesis, with a peak exceeding  $10 \text{ K (100 km)}^{-1} (3\text{hr})^{-1}$  (Fig. 8c, light blue), while there is  
467 700-hPa frontogenesis exceeding  $15 \text{ K (100 km)}^{-1} (3\text{hr})^{-1}$  for the latter snowband (1900 – 2110  
468 UTC) (Fig. 8c, blue).

469 The height-time cross sections of bulk mass flux and the time series of column mean mass flux  
470 are displayed in the time series plots in Figs. 8d and 8e, respectively. Figure 8d also show the mRi  
471 less than 0.25. Generally, the URs contributed to upward bulk mass flux when they are clustered  
472 in space and time. The URs during the snowbands above 2 km contributed relatively large mass  
473 flux (Fig. 8d), and the URSWHIs contribute to slightly larger mass flux compared to URSWLO  
474 (Fig. 8e) during the snowband periods, likely due to slightly stronger updrafts (Table 3). For the 4  
475 January case, the latter snowband accompanied less URs producing smaller bulk mass flux and  
476 smaller reflectivity (Figs. 8a and 8e), likely because the URs were not clustered in space in time.  
477

478 The upward bulk mass flux occurs within and above the moist shear instability ( $mRi < 0.25$ )  
479 layers (Fig. 8d) and the layers of large vertical wind shear  $> 30 \text{ m s}^{-1} \text{ km}^{-1}$  (Fig. 8b, gray contour).  
480 This suggests a role of shear instability for generating URs and hence upward mass transport.  
481 Those were common features for the three multiband cases, (4 January first band, 16 – 17  
482 December, and 1 February, shown later).

483 Figure 9a shows the vertical distributions of the URs. The total length of each bar is the  
484 accumulated time with updrafts divided by the accumulated time with cloud echo as a percent at  
485 each altitude bin (0.5 km). The UR accounts for more than 10% of cloud echo time at almost all  
486 levels with the maximum of 31 % at 5 km altitude. The URs are classified into URSWHIs and  
487 URSWLO, shown by color-coded subsets along each bar. For the 4 January case, URs collocated  
488 with  $SW > 0.4 \text{ m s}^{-1}$  (URSWHIs) account for approximately 30% of the total number of URs above  
489 the PBL height (Table 2) with the maximum at 5 km altitude. This was common for the three  
multiband cases.

490 Figure 9b shows profiles of the occurrences (time relative to the total cloudy time) at each  
491 height similar to Fig. 9a, but for wind shear  $> 15 \text{ m s}^{-1} \text{ km}^{-1}$  and  $mRi < 0.25$ . The midlevel UR and  
492 URSWHIs are well correlated with wind shear (red line, correlation of 0.76 between URSWHIs and  
493 wind shear). The wind shear is also correlated with the moist shear instability ( $mRi < 0.25$ ,  
494 correlation of 0.58). Note that the VAD-based wind data are limited to 7.8 km maximum altitude.  
495 Sounding in Fig. 6c shows the shear instability ( $mRi < 0.25$ ) at 8-9.5 km, consistent with local  
496 peak of the UR and URSWHIs occurrences.



497

498 Figure 9: (a) Histograms of occurrence (time relative to the total cloudy time observed by KASPR VPT) of  
 499 detected updraft regions shown as percentage of the entire cloud echo time at each altitude for the cases of 4 Jan  
 500 2018. The color bars represent different magnitudes of mean Doppler spectrum width (SW) with a SW threshold  
 501 (red:  $SW > 0.4 \text{ m s}^{-1}$  and blue:  $SW < 0.4 \text{ m s}^{-1}$ ). The SW from the KASPR VPT measurements was averaged  
 502 over each updraft region. The 15-m vertical resolution data were resampled every 0.5 km height. In a resampled  
 503 0.5 km height range, some URs located in different heights can be overlapped in time. Periods where UR having  
 504 SW > 0.4 m s<sup>-1</sup> and UR having SW < 0.4 m s<sup>-1</sup> are overlapped in time are represented by yellow bars. (b)  
 505 Occurrences of wind shear > 15 m s<sup>-1</sup> km<sup>-1</sup> (red), and mRi < 0.25 (blue) normalized by the entire cloud echo time  
 506 at each altitude for the cases of 4 Jan 2018. (c) Profiles of the time used for the analysis at each height bin. The  
 507 analysis was based on the VPT scans for 5 min duration every 15 minutes. The total analysis period is also shown  
 508 in Table 1.

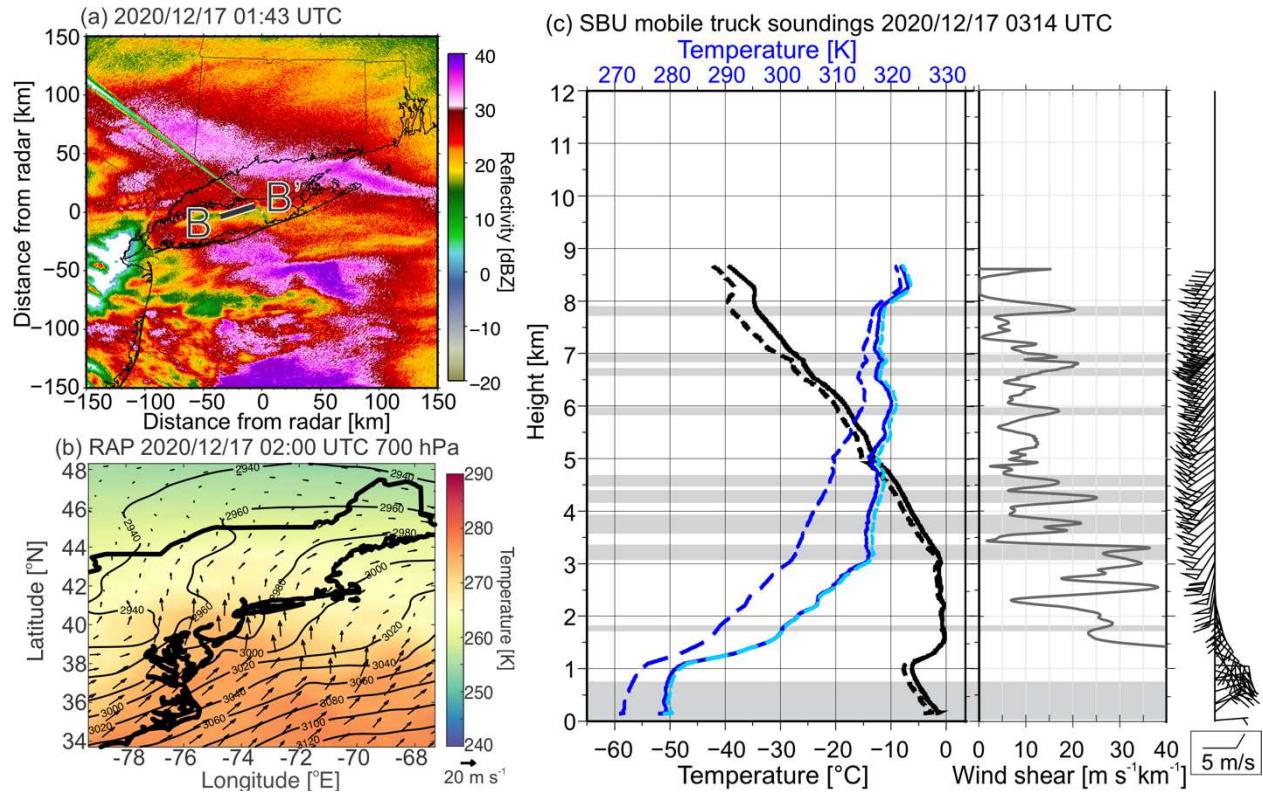
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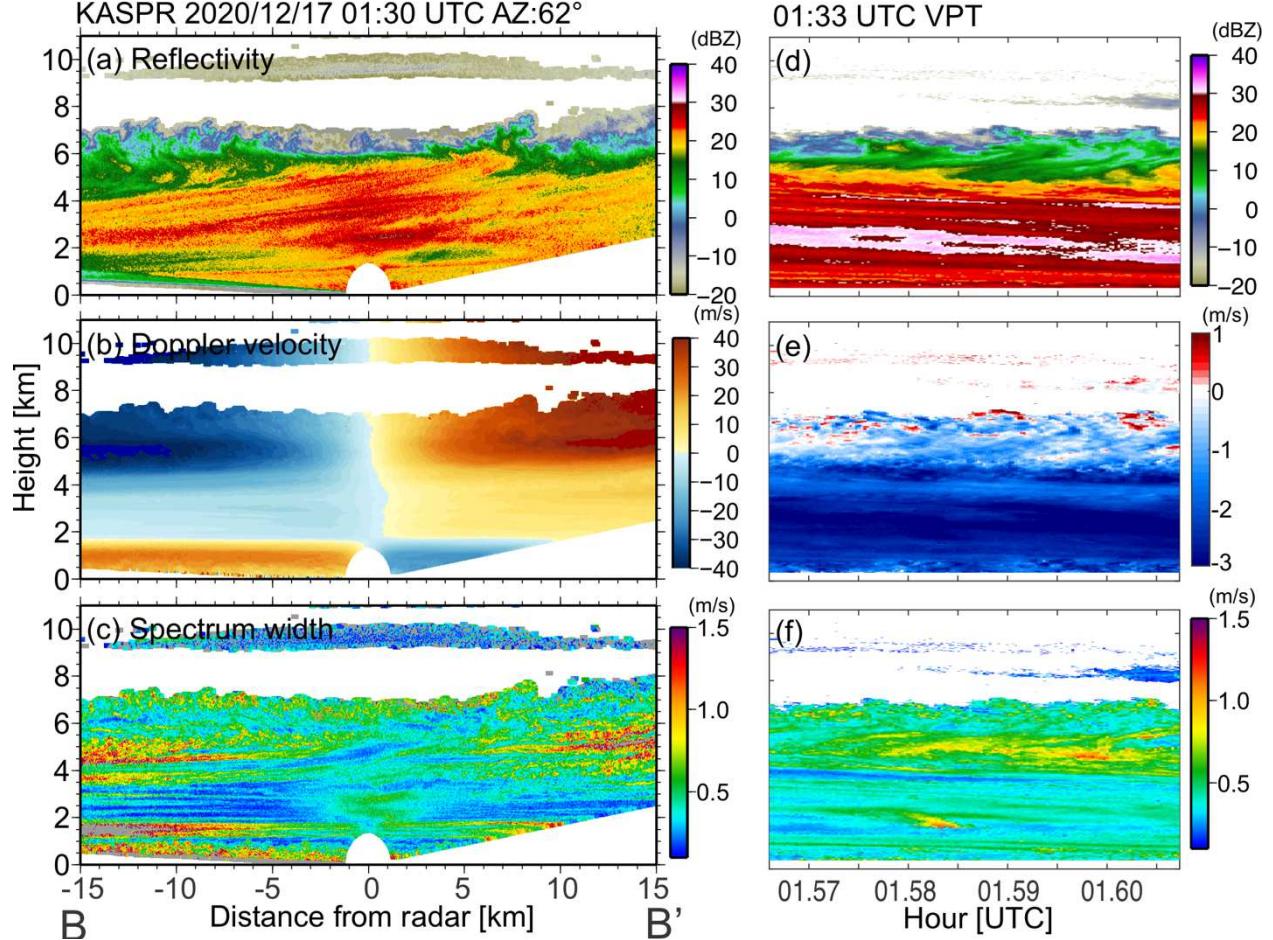
## 511 2). 16-17 December 2020

512 At 0200 UTC, surface low pressure (~1006 hPa) was along the mid-Atlantic coast and moved  
 513 northward offshore during the next 24 h (orange track in Fig. 4). Similar to the 4 January case, Fig.  
 514 10 shows the horizontal distributions of the NEXRAD reflectivity at the time of snowbands, 700  
 515 hPa RAP, and the 0314 UTC on 17<sup>th</sup> sounding profile. A well-defined warm front existed to the  
 516 north of the surface cyclone stretching from west to east south of Long Island, with warm advection  
 517 at 700 hPa over Long Island (Fig. 10b). The WNW-ESE oriented snowbands exceeding 30 dBZ  
 518 passed over SBU from south to north from 0000 to 0310 UTC 17 December (Fig. 10a). The SBU  
 519 surface temperature was below freezing until 0900 UTC 17 December, while soundings at 0000,  
 520 0606, and 0715 UTC 17 December at SBU show an inversion layer and a temperature > 0°C  
 521 centered around 1.5, 2.3, and 2.4 km ASL, respectively. KASPR RHI observed a melting layer  
 522 after 0350 UTC 17 December at 2.5 km ASL. We focus on the period prior to the melting signature  
 523 (Table 1).

524 Figure 11 shows the KASPR RHI and VPT scans near a snowband at around 01:30 UTC on  
 525 17<sup>th</sup>. The KASPR RHI scans (Fig. 11a-c) reveal generating cells at cloud top, with areas of  
 526 relatively large SW ( $> 0.75 \text{ m s}^{-1}$ ) from 6-7.5 km ASL, and fallstreaks extending downward to the  
 527 west below 6 km ASL where their directions are modified by the wind shear at around 5.4 km ASL  
 528 (Figs. 10c, 11b). The KASPR VPT measurements (Figs. 11d-f) also reveal updrafts with large SW  
 529 ( $> 0.4 \text{ m s}^{-1}$ ) in the generating cell layer. The cloud top generating cells are in conditionally  
 530 unstable layers with smaller regions of moist absolute instability layer (MAUL; Bryan and Fritsch,  
 531 2000) as  $\partial\theta_e/\partial z < 0$  and  $\partial\theta_{es}/\partial z < 0$  with  $\theta_e \approx \theta_{es}$  at 6 – 6.5 km and 6.8 – 7.0 km ASL (Fig.  
 532 10c). These layers correspond to the vertical wind shear layer observed by the RHI (Fig. 11b). The  
 533 sounding had several layers of vertical wind shear layers ( $> 20 \text{ m s}^{-1} \text{ km}^{-1}$ ), most of which above  
 534 3 km ASL were linked to shear instability ( $m\text{Ri} < 0.25$ ).



535  
 536 Figure 10: Same as Fig. 6 but at (a) 0143 UTC on 17 Dec 2020, (b) at 0200 UTC on 17 Dec 2020, and (c)  
 537 03:14 UTC on 17 Dec 2020 collected at Stony Brook (green triangle in Fig. 1). The B – B' line in (a)  
 538 represents the KASPR RHI direction shown in Fig. 11a-c.  
 539



540  
541 Figure 11: Same as Fig. 7 but (a-c) at the 0130 UTC RHI scan along B-B' line in Fig. 10a on 17 Dec 2020, and  
542 (d-f) VPT measurements on 17 Dec 2020.  
543

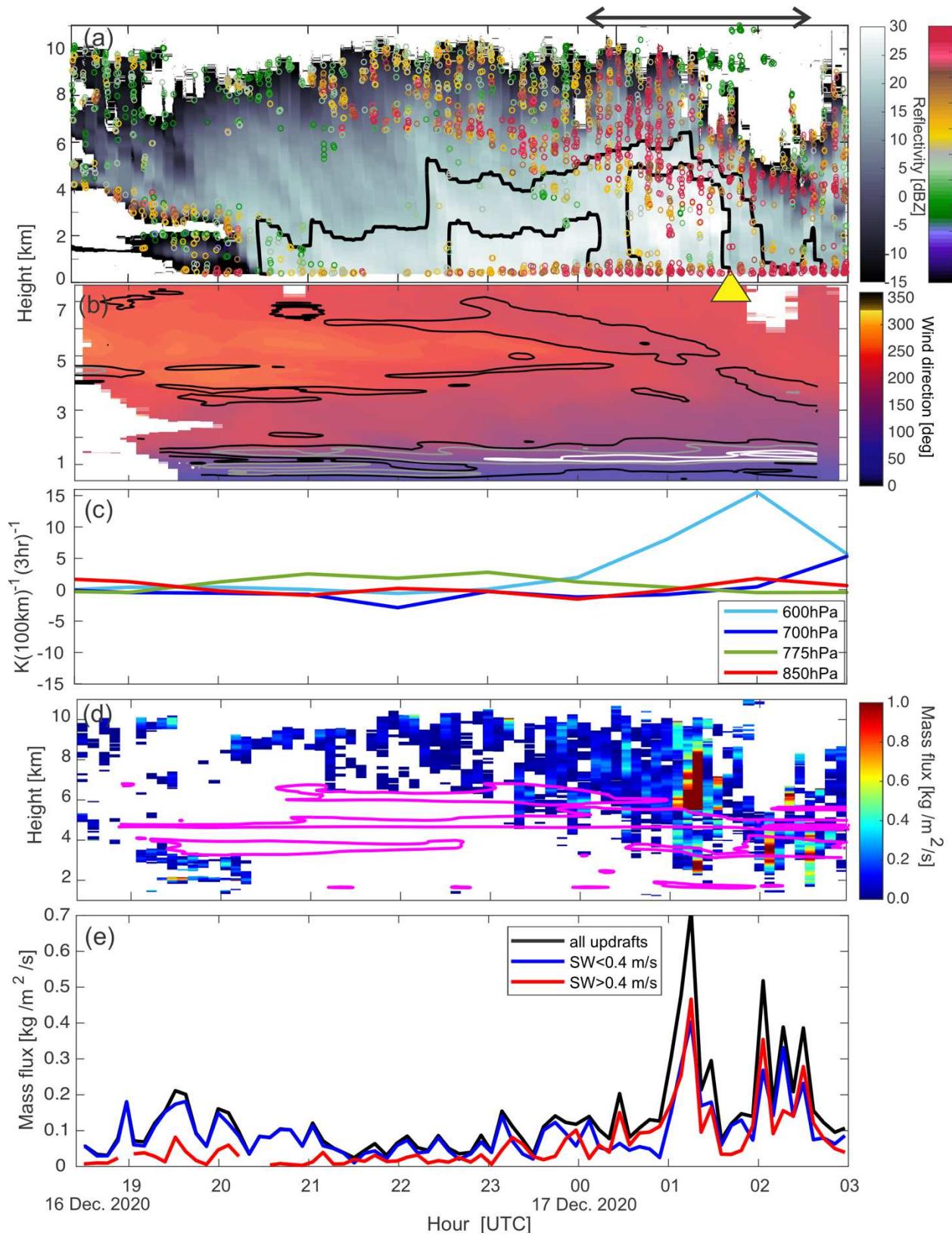
544 Figure 12 depicts the time series of the spatial distributions of URs, horizontal wind,  
545 frontogenesis, mRi, and mass flux for the case of 16-17 December 2020, similar to Fig. 8., The  
546 snowband period is also closely associated with the 600-hPa frontogenesis, with the peak  
547 exceeding  $15 \text{ K (100 km)}^{-1} (3\text{hr})^{-1}$  at 0200 UTC on 17 December (Fig. 12a,c). While the URs are  
548 observed throughout the event, during the snowband period URs dominate in the cloud depth  
549 above 2 km. URSWLOs are found within 2 km from the cloud top height starting before the  
550 snowband period from 1930 UTC 16 December to 0200 UTC 17 December (Fig. 12a). URSWHIs  
551 appear from 2100 UTC 16 December just below the URSWLO layer, corresponding to the period  
552 that includes larger reflectivity  $>25 \text{ dBZ}$  below 2 km (2230-0010 UTC) and below 4.5 km (0030 -  
553 0145 UTC, snowband period). During the snowband period, URSWHIs are more prominent in a  
554 deeper layer between 4 km to the cloud top. The KASPR VAD analysis (Fig. 12b) shows that the  
555 lower boundary of the URSWHI layer well corresponds to vertical wind shear  $> 20 \text{ m s}^{-1} \text{ km}^{-1}$  (Fig.

556 12b) starting at 6.2 km altitude at 2100 UTC until 3-4 km at 0230 UTC and low mRi < 0.25 (Fig.  
557 12d, contour).

558 In the occurrence profile in Fig. 13, similar to the 4 January case, the UR accounts for more  
559 than 10% of cloud echo time at almost all levels. The maximum occurrence of URs is found at 9  
560 km, but the maximum occurrence of URSWHIs is found at lower altitude around 6.5 km. The  
561 occurrence profile of the wind shear in Fig. 13b is somewhat noisy, but the shear of the wind  
562 direction (green line) is consistent with URSWHIs (black line, correlation of 0.33).

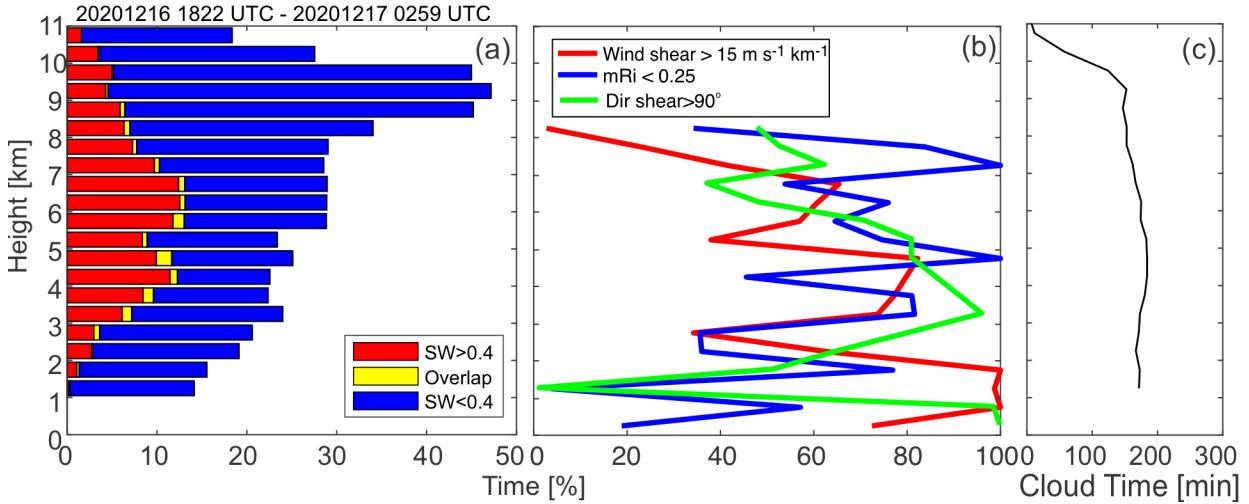
563 Similar to the 4 January case, the upward bulk mass flux occurs within the moist shear  
564 instability ( $mRi < 0.25$ ) layers (Fig. 12d). This suggests a role of shear instability for generating  
565 URs and hence upward mass transport. The mass flux shows a maximum during the snowband  
566 period where the URSWHIs are clustered, suggesting a contribution of URSWHIs to the upward  
567 mass transport.

568



569  
570  
571

Figure 12: Same as Fig. 8, but for 16 – 17 December 2020.



572

573 Figure 13: Same as Fig. 9, but for 16 – 17 December 2020. The occurrence of direction shear of horizontal  
 574 wind  $> 90^\circ \text{ km}^{-1}$  is also shown in (b) by a green line.

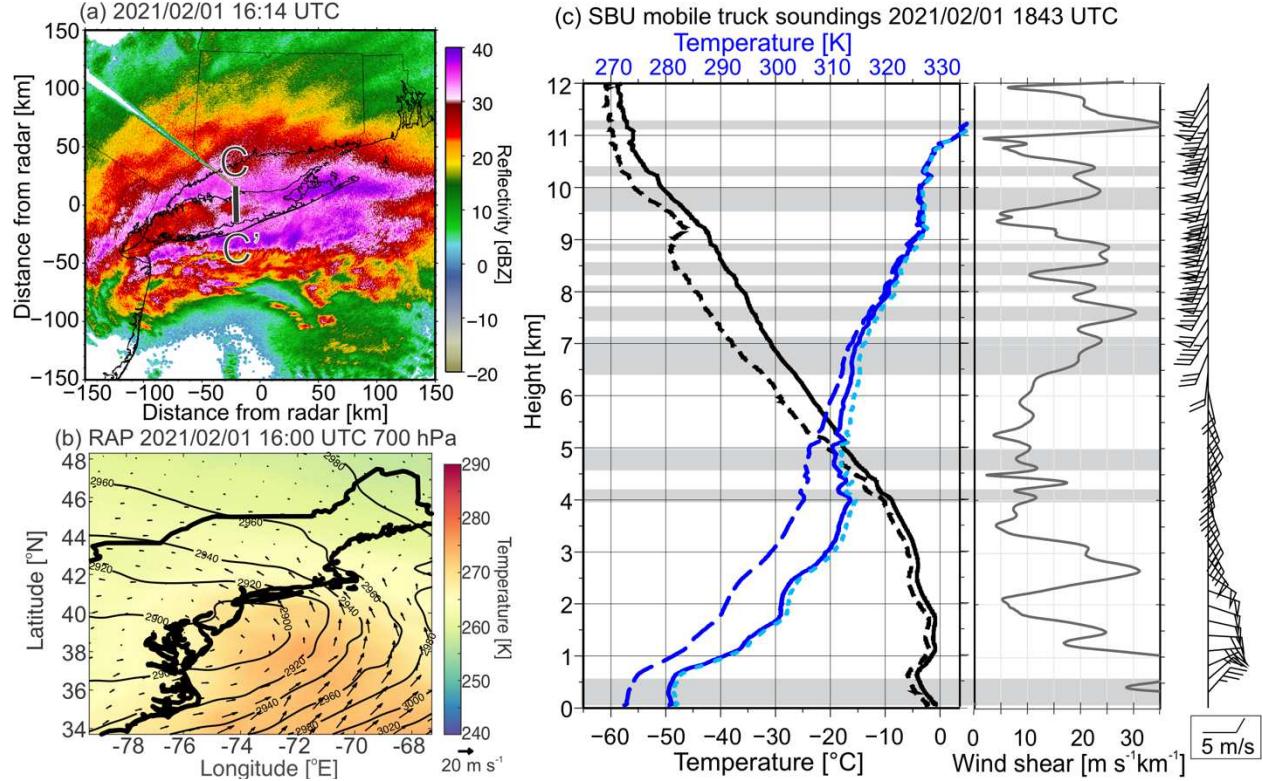
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576 3) 1 February 2021

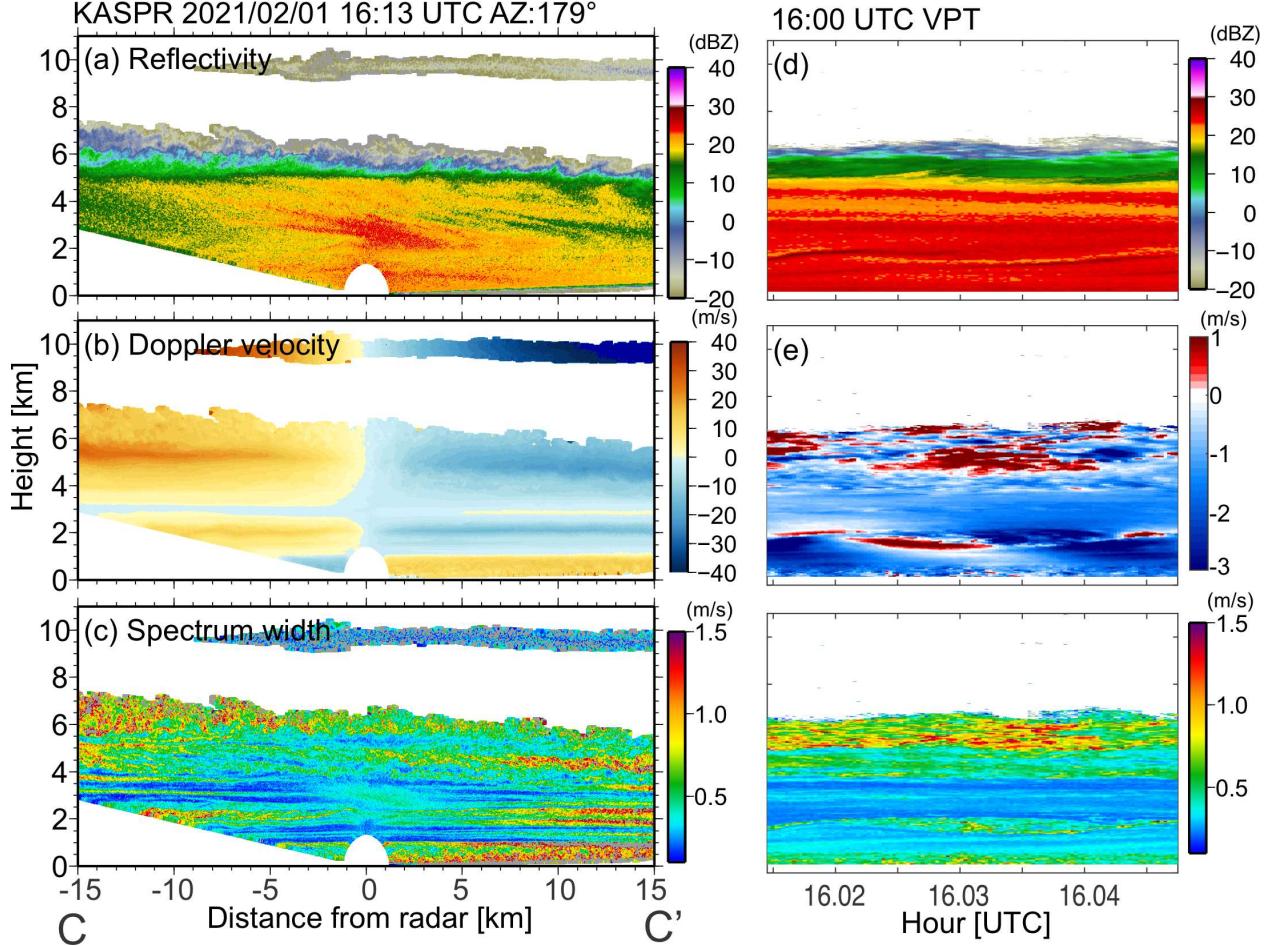
577 Similar to 16-17 December, the snow event was associated with a surface cyclone moving  
 578 northward to the south of Long Island (green track in Fig. 4). No snowbands were evident around  
 579 Long Island before 0210 UTC 1 February. After 0210 UTC 1 February, there were precipitation  
 580 areas with reflectivity  $> 30 \text{ dBZ}$  around SBU, but they were not banded until 1300 UTC 1  
 581 February. From 1400 UTC to 1830 UTC 1 February, two west-east oriented snowbands (30-40  
 582 dBZ) crossed SBU moving northward (Fig. 14). As shown in Fig. 14b, there was a surface warm  
 583 front just south of Long Island, with warm advection at 700 hPa over this region.

584 Figure 14c shows the SBU sounding at 1843 UTC 1 February near the snowband period. The  
 585 sounding profile shows that a moist unstable layer exists near cloud top (4-5.2 km ASL). The  
 586 KASPR RHI measurements in Fig. 15 reveal cloud-top generating cells at 5.2-7.5 km ASL near  
 587 the moist unstable layer and fallstreaks to 25 dBZ underneath the generating cell layer during the  
 588 snowband period (Figs. 15a,d). The Z difference between the WSR-88D and KASPR likely  
 589 represents a resonance scattering effect at Ka-band owing to the presence of larger snow particles  
 590 relative to the wavelength. The sounding for the generating cell layer includes moist instability,  
 591 and wind shear instability ( $mRi < 0.25$ , Fig. 14c). This layer also corresponds to supersaturation  
 592 with respect to ice (not shown). The VPT measurements shown in Fig. 15d-f show updrafts and  
 593 large SW, indicating that the generating cells accompanied turbulence. The cloud-top generating  
 594 cell layer is also well consistent with the Doppler velocity shear (Fig. 15b). Interestingly, there is

595 also a layer of updrafts collocated with relatively large SW ( $>0.4 \text{ m s}^{-1}$ , Fig. 15e,f) at 2 km in the  
 596 VPT, well consistent with vertical wind shear observed by the RHI (Fig. 15b) and the sounding  
 597 measurements (at 2.5 km, Fig. 14c).



598  
 599 Figure 14: Same as Fig. 6 but at (a) 1614 UTC on 1 Feb 2021, (b) at 1600 UTC on 1 Feb 2021, and (c) 18:43  
 600 UTC on 1 Feb, 2021 collected at Stony Brook (green triangle in Fig. 1). The C – C' line in (a) represents the  
 601 KASPR RHI direction shown in Fig. 15a-c.  
 602



603

604 Figure 15: Same as Fig. 7 but (a-c) at 1613 UTC RHIs along C-C' line in Fig. 14a on 1 Feb 2021, and (d-f) VPT  
 605 measurements on 1 Feb 2021.

606

607 As shown in the time series of this case shown in Fig. 16, the snowband period is associated  
 608 with a 700 hPa frontogenesis of  $7 \text{ K (100 km)}^{-1} (3\text{hr})^{-1}$  at 1800 UTC, and the 775 hPa frontogenesis  
 609 maximum of  $15 \text{ K (100 km)}^{-1} (3\text{hr})^{-1}$  at 2000 UTC (Fig. 16a,c). URSWLOs (represented by yellow-  
 610 green dots) are observed near the cloud top with weak reflectivity ( $\sim 0 \text{ dBZ}$ ) from the beginning  
 611 of the observation on 31 January (1800 UTC) until 1300 UTC on 1 February. URSWHIs occur  
 612 from 1200 UTC 1 February to the end of the analysis period (0000 UTC 2 February), which is  
 613 consistent with the snowband period. The KASPR VAD analysis (Fig. 16b) shows the vertical  
 614 wind shear produced by south-southeasterly wind in the low altitudes and southwesterly wind in  
 615 the upper altitudes. The shear line is found between 2 and 3 km altitudes until 1210 UTC on 1  
 616 February and then ascends to 6 km altitude by 1400 UTC. The low-level south-southeasterly wind  
 617 corresponds to the 700 hPa warm advection presented in the RAP data at 1400 UTC (Fig. 14b).

618 In the occurrence profile in Fig. 17, similar to the other multiband cases, the UR accounts for  
619 more than 10% of cloud echo time at almost all levels with the maximum of UR (and URSWLO)  
620 occurrence is found near the cloud top. The maximum occurrence of URSWHIs is found at lower  
621 altitude around 7 km, and the second peak is found at around 3 km. These peaks in the URSWHIs  
622 occurrence are well consistent with the peaks of the occurrence of large wind shear (red line,  
623 correlation of 0.24 with URSWHIs). The URSWHIs occurrence is also well correlated with low mRi  
624 (correlation of 0.63).

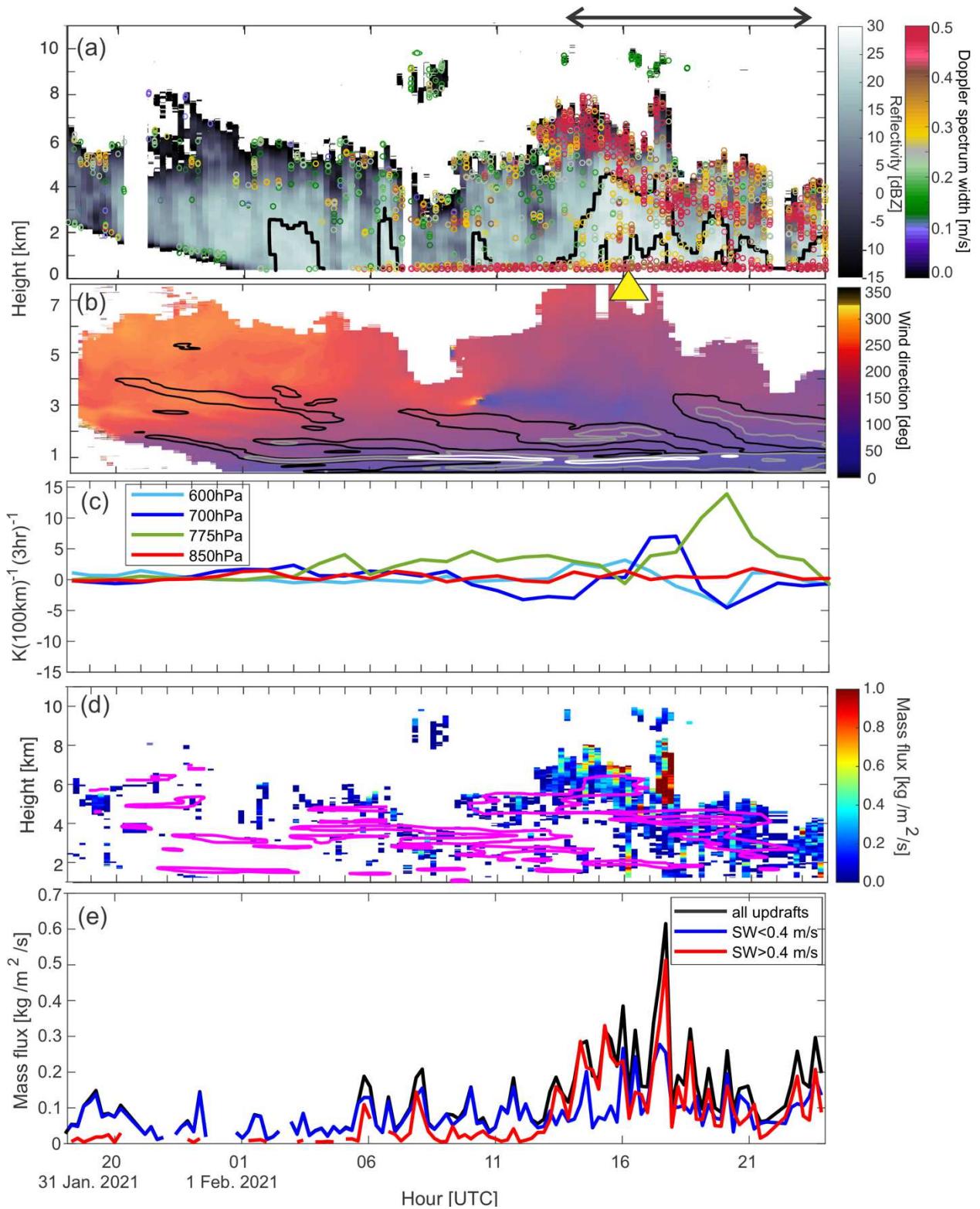
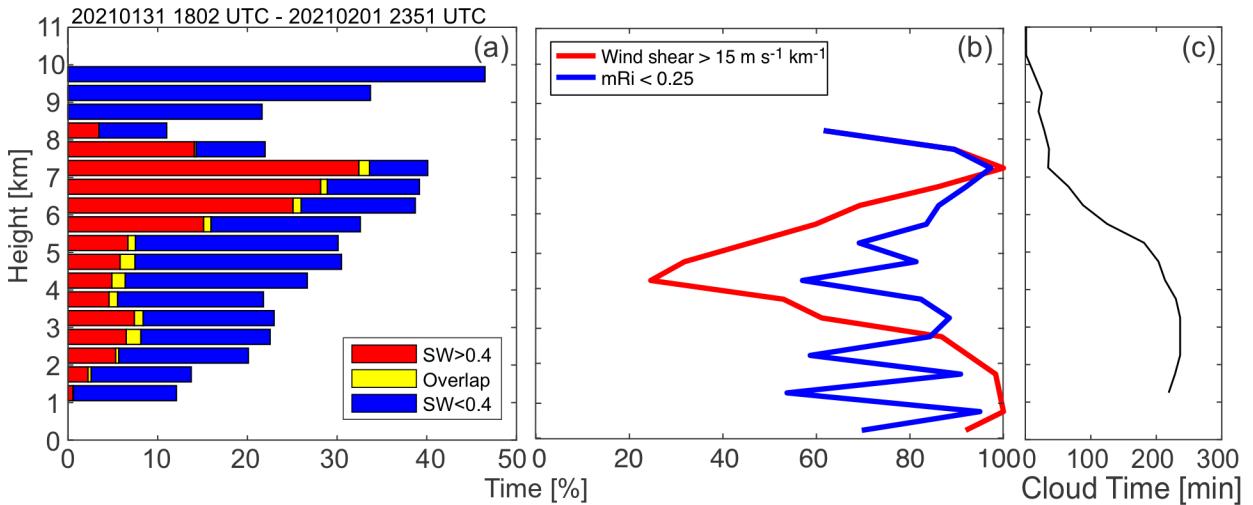


Figure 16: Same as Fig. 8, but for 31 January – 1 February 2020.

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626  
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628

629 Figure 17: Same as Fig. 9, but for the case of 1 Feb 2021.

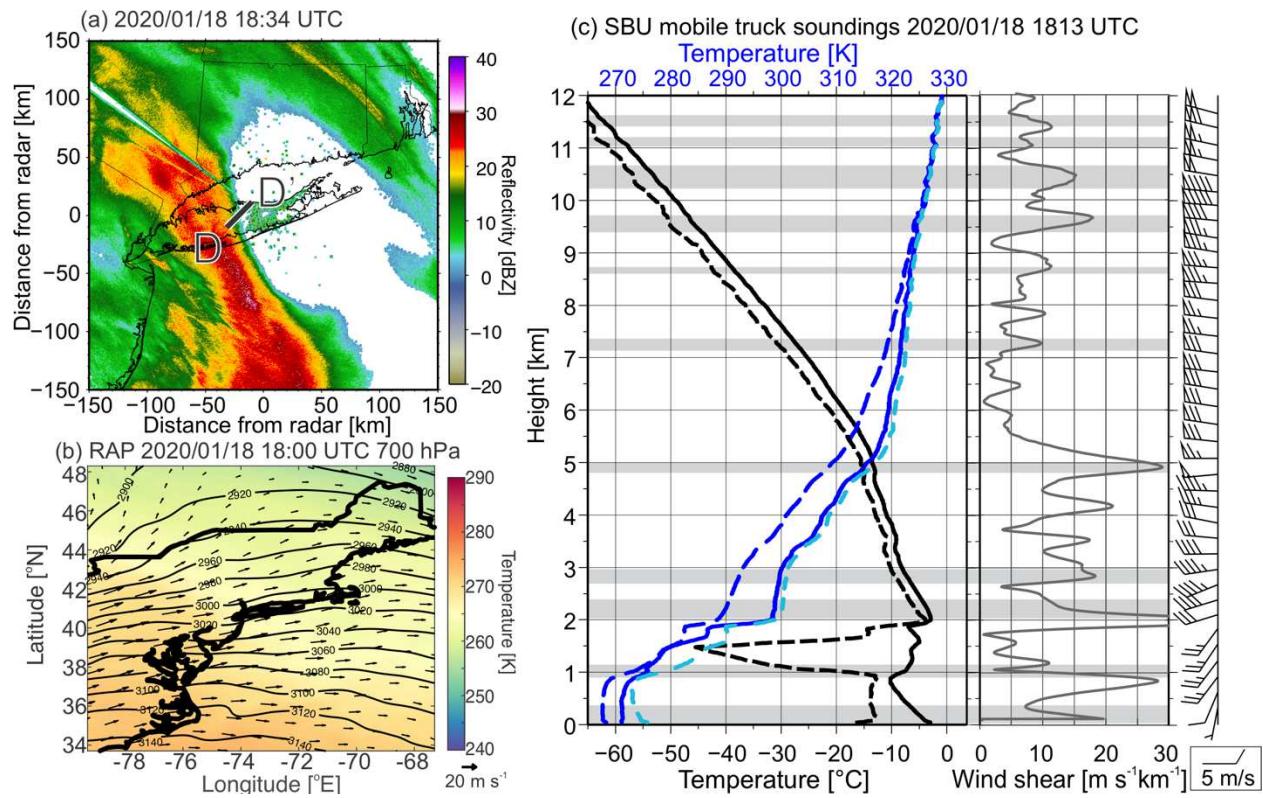
630

631 4) 18 January 2020

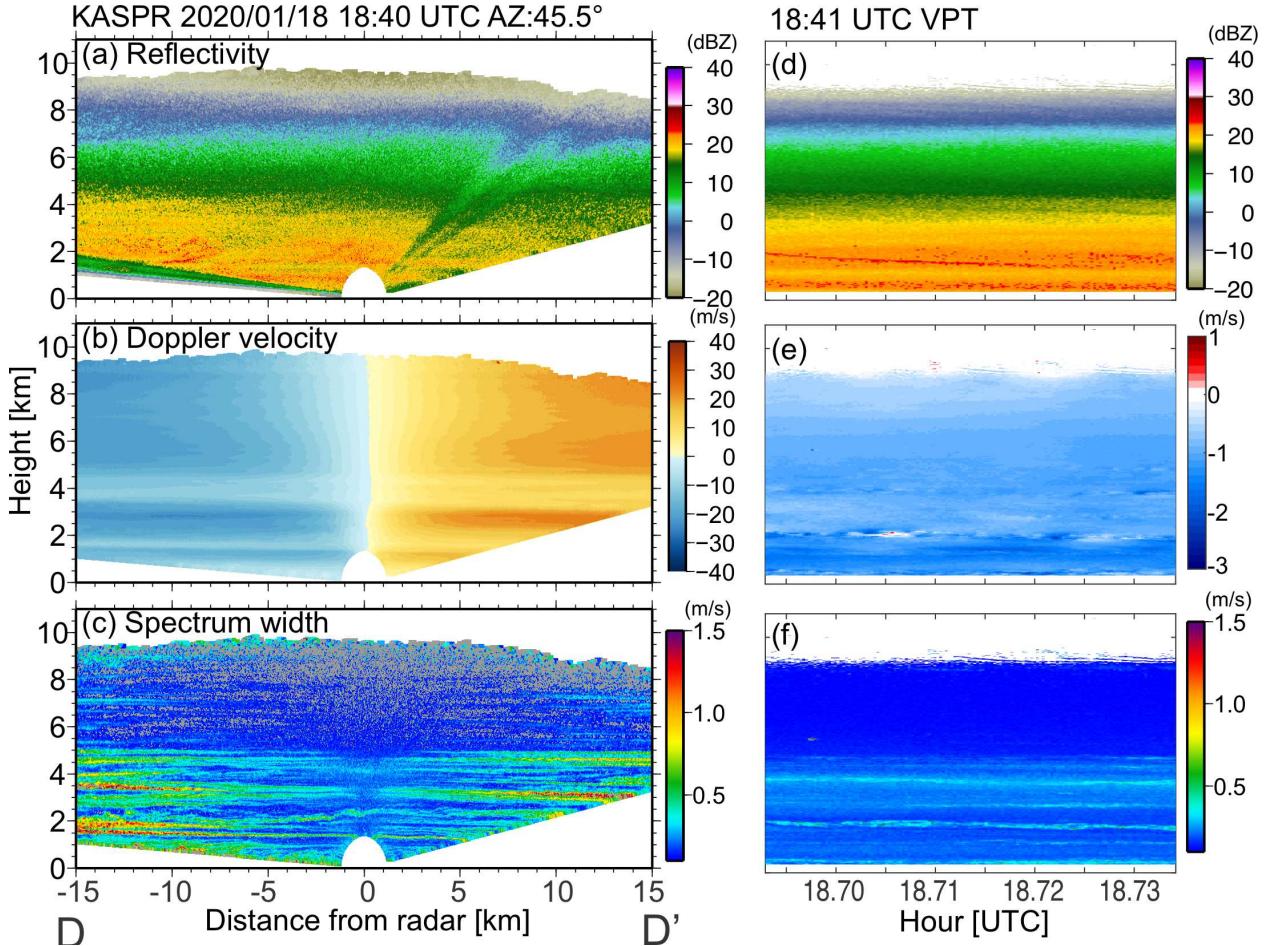
632 For the 18 January case, which accompanied a single snowband, by 0300 UTC 19 January, the  
 633 surface warm front associated with a surface cyclone (~1003 hPa) centered over northern Michigan  
 634 was southwest of Long Island along the mid-Atlantic coast as the storm center moved eastward  
 635 (magenta track in Fig. 4). The vertical wind shear seemed similar to the multiband cases, but the  
 636 spatial distributions of UR and the radar reflectivity showed some differences. Figure 18 shows  
 637 the horizontal distributions of the NEXRAD reflectivity at the time of snowbands, 700 hPa RAP,  
 638 and the 1813 UTC sounding profile. The warm-frontal snowband (> 20 dBZ) orientated from  
 639 northwest to southeast crossed SBU between 1830-2015 UTC 18 January (Fig. 18a) accompanying  
 640 a 700-hPa warm advection (Fig. 18b). The maximum WSR-88D reflectivity for this snowband  
 641 (~30 dBZ) was 5-10 dB weaker than the other cases. The surface precipitation transitioned to rain  
 642 after the warm frontal snowband passage. This study focuses on the warm frontal snowband to  
 643 avoid uncertain errors associated with hydrometeor attenuation in the rain layer.

644 The sounding from the SBU mobile truck deployed at Cedar Beach (blue triangle of Fig.1) at  
 645 1813 UTC 18 January, which was closest to the snowband passage, shows a warm-frontal stable  
 646 layer with veering winds from 1-2 km ASL and a near moist neutral layer from 2-3 km ASL (Fig.  
 647 18c). The sounding in Fig. 18c and KASPR measurements in Fig. 19 suggest that the cloud top  
 648 height reach 10 km ASL, where the vertical profile is less stable and near moist neutral (7-10 km  
 649 ASL). KASPR RHI scans oriented NE-SW (B-B' in Fig. 18a) and crossing the snowband (Figs.  
 650 19a-c) reveal a gradual increase of reflectivity toward the surface above ~3 km ASL and relatively

651 uniform reflectivity at a given altitude when the snowband passed. The VPT measurements (Figs.  
652 19d-f) do not show significant convection above 4 km, where the RHI Doppler velocity shows less  
653 vertical wind shear. There is a wave-like feature at the radar echo top (at ~10 km, Figs. 19a,b,c,e)  
654 with weak reflectivity, which could be a source of precipitation. The VPT measurements show  
655 layers of relatively large SW at 2 and 4 km. These layers also correspond to the vertical wind shear  
656 observed by the RHI and shear unstable layers observed by the sounding ( $mRi < 0.25$ , gray shades).  
657 The lower layer (2 km ASL) corresponds to a local intensification of reflectivity (~25 dB). These  
658 SW values, however, weaker (generally  $< 0.4 \text{ m s}^{-1}$ ) than the other cases (Fig. 19f).



659  
660 Figure 18: Same as Fig. 6 but at (a) 1834 UTC on 18 Jan 2020, (b) at 1800 UTC on 18 Jan 2020, and (c) 18:13  
661 UTC 18 Jan 2020 collected at Cedar Beach (blue triangle in Fig. 1). The D – D' line in (a) represents the  
662 KASPR RHI direction shown in Fig. 19a-c.  
663

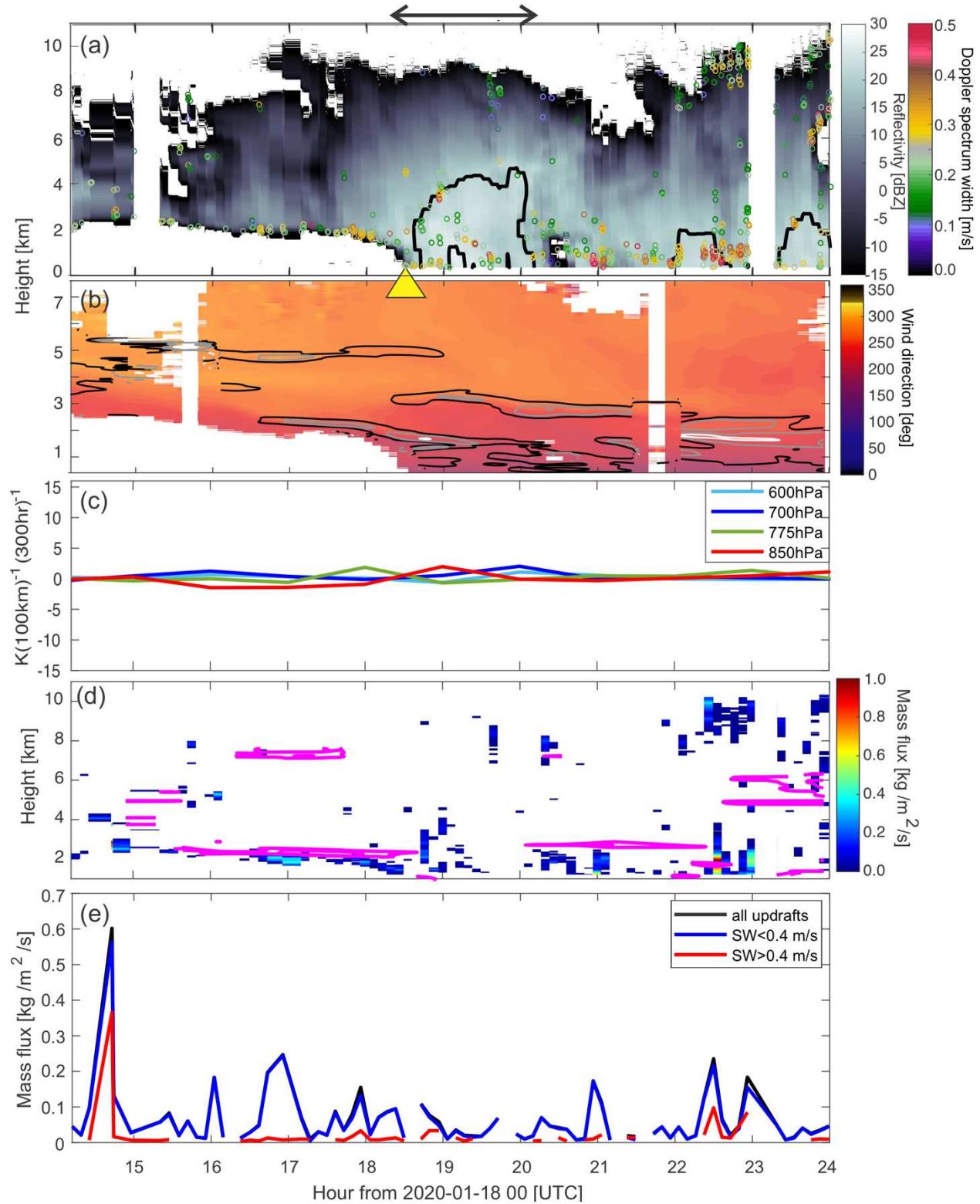


664  
665 Figure 19: Same as Fig.9 but (a-c) at the 1840 UTC RHI scan along D-D' line in Fig. 18a on 18 Jan 2020 and  
666 (d-f) VPT measurements on 18 Jan 2020.  
667

668 Figure 20 shows the set of time series for the 18 January case. Similar to the other three cases,  
669 URSWLOs (represented by yellow-green-blue dots) are observed near the cloud top with weak  
670 reflectivity ( $< 0$  dBZ), but the frontogenesis was very weak at all levels ( $< 2$  K  $(100\text{ km})^{-1}$   $(3\text{ hr})^{-1}$ )  
671 at 700, 775, and 800 hPa, Fig. 20c). Multiple layers of vertical wind shear  $> 30\text{ m s}^{-1}\text{ km}^{-1}$  (Fig.  
672 20b) are observed within the cloud, similar to the other three cases. In contrast to the three  
673 multiband cases, for the 18 January case, no significant increase in the mass flux is found (Fig.  
674 20d,e). Near the cloud top, where URSWLOs dominate, the mass flux varies in magnitude and is  
675 discontinuous in time.

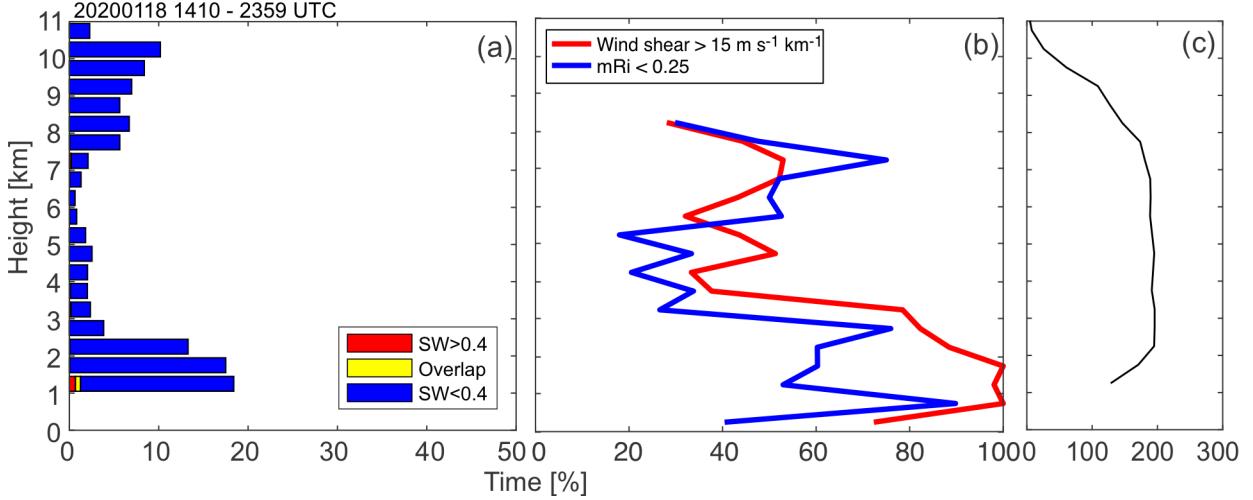
676 Figure 21 shows the occurrence profiles. The UR accounts for less than or equal to 10% of  
677 cloud echo time 2.5 km, and URSWHIs accounts for only 3% at all levels (Fig. 21a). Although the  
678 number of URSWHIs is small, the occurrences of UR and URSWHIs are correlated with the large  
679 wind shear (red line, correlation of 0.48 with URSWHIs, Fig. 20b). They are also weakly correlated

680 with shear instability; the correlation between UR (gray line) and low mRi (blue line) is 0.30, and  
681 that between large wind shear (red line) and low mRi (blue line) is 0.57. While the layers of updraft  
682 regions and wind shear were correlated, the strength of wind shear or shear instability might not  
683 be a factor of the strength and frequency of the updrafts. Further analysis is needed to evaluate this  
684 using high resolution model simulations.



685  
686

Figure 20: Same as Fig. 8, but for 18 January 2020.



687

688 Figure 21: Same as Fig. 9, but for the case of 18 January 2020.

689

690 *c. Summary*

691 The snowbands accompanied convective cells near cloud top and fallstreaks underneath, as  
 692 observed by the previous studies. The convective cells were well collocated with layers of updrafts  
 693 and large SW associated with vertical wind shear. They suggest that the near cloud top convective  
 694 cells (i.e., generating cells) contributed to the precipitation within the snowbands. The weak  
 695 snowband case (18 January) did not show significant updrafts and turbulence (large SW) near the  
 696 cloud top even within the cloud layer. The URs associated with wind shear was often multi-layered  
 697 in the precipitating clouds, as it was observed not only near the cloud top but also the middle of  
 698 the precipitating clouds. They were collocated with intensification of the precipitation underneath  
 699 as measured by reflectivity.

700 Although there is a case-to-case variability in the heights, the URs collocated with  $SW > 0.4 \text{ m s}^{-1}$  (URSWHIs) occur in more interior clouds of the multiband cases. URSWHIs tend to be shorter  
 701 lived compared to the URSWLOs for all cases (approximately 1 – 2 sec shorter, Table 3).  
 702 URSWHIs have slightly ( $0.20 – 0.52 \text{ m s}^{-1}$ ) stronger median upward motion and 3.5 – 16.4 dB  
 703 larger median reflectivity (Table 3). In contrast, URs collocated with  $SW < 0.4 \text{ m s}^{-1}$  (URSWLOs)  
 704 show higher frequencies at higher altitudes. They are found above 8 km altitude; some of them are  
 705 found very close to the echo tops, and most of them are found within 1-2 km distance from the  
 706 echo tops. The lower SW is likely due to a combination of less turbulence and a narrower particle  
 707 size distribution owing to weaker reflectivity. The peak of the URSWHIs occurrence is found at a  
 708 lower altitude than that of the URSWLOs.  
 709

710 Overall, URs are found throughout the snowstorms regardless of the snowband periods or  
711 frontogenesis for the four cases. The occurrences of URs and URSWHIs are well correlated with  
712 large wind shear and shear instability (Figs. 9, 13, 17, and 21). The URSWHIs are frequently found  
713 during the snowband periods for three of the cases.

714

715

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717

#### 718 4. Discussion

719 Novel high-vertical resolution, high-sensitivity vertically-pointing radar measurements in four  
720 winter storms along the US northeast coast showed that URs are ubiquitous features in the winter  
721 storms. They are correlated with layers of enhanced wind shear and are found frequently in the  
722 presence of snowbands, which suggests that URs could promote precipitating particle growth.  
723 They accompanied the snowbands, but they also occurred outside the snowbands. This suggests  
724 that the URs could promote precipitating particle growth. In particular, high-reflectivity bands  
725 from 4 January, 16-17 December, and 1 February accompanied URSWHIs. They suggest that URs  
726 associated with the wind shear contributed to the formation and /or intensification of snowbands.  
727 This is consistent with the previous studies (e.g., Plummer et al. 2014; Rauber et al. 2014). The  
728 present study revealed that the URs contributed to particularly significant upward mass flux when  
729 they were clustered in space and time especially during the snowband periods. RAP 700 hPa  
730 specific humidity  $> 5 \text{ g kg}^{-1}$  (not shown) was associated with periods with the larger mass flux in  
731 all four cases. It is suggested that the URs could play a role as a driver for supplying water vapor  
732 that could produce saturation with respect to liquid water and thus potentially the formation of  
733 supercooled liquid droplets. They could then contribute to forming/growing precipitation particles,  
734 as pointed out by Plummer et al. (2014). The URs were observed not only in the snowbands, but  
735 also outside the snowbands, although they were distributed sparsely compared to within the  
736 snowbands. This suggests a role of the URs on snow particle formation and/or intensification of  
737 snowfall outside the snowbands.

738 The URs, especially URSWHIs were well correlated with vertical wind shear and shear  
739 instability ( $mRi < 0.25$ , Figs. 8, 9, 12, 13, 16, 17, 20, and 21), suggesting that vertical wind shear  
740 caused turbulence, which included upward vertical velocity perturbations identified here as URs.  
741 The large mass flux is particularly associated with the moist shear instability, suggesting a role of

742 the turbulence induced URs on the vertical mass transport. The wind shears were multi-layered in  
743 the cloud systems. This is consistent with the former studies such as Boucher et al. (1965), Wexler  
744 et al. (1967), Sylett et al. (1995), and Rauber et al. (2014) that observed wind shears associated  
745 with frontal zones in winter storms using Doppler radar measurements. This suggests that the  
746 snowstorm systems in the present study were composed of several air masses originating from  
747 different sources producing frontal zones, as commonly observed by previous studies (e.g.  
748 Boucher et al. 1965; Rauber et al. 2014). The KASPR measurements presented in this study also  
749 showed wind shear layers with low mRi outside the frontal zones. The multi-layers of turbulence  
750 and updrafts associated with wind shear observed by the RHIs and soundings in this study suggest  
751 the presence of smaller scale air masses rather than synoptic scale air masses as presented in  
752 Rauber et al. (2014).

753 It is hard to determine the role of frontogenesis in the UR formation, because frontogenesis,  
754 larger (synoptic) scale forcing, associates larger scale ascending that generally has smaller ( $\text{cm s}^{-1}$   
755 scale) vertical air motion. The URs and updrafts observed by the vertically-pointing radar  
756 measurements in this study are finer scales having  $\sim 10$  s duration and  $> 1 \text{ m s}^{-1}$  upward velocity  
757 than those associated with frontogenesis.

758 It is possible that the updrafts at the cloud tops could be produced in part by instabilities caused  
759 by cloud-top radiative cooling. The cloud-top cooling could produce cloud-top destabilization  
760 leading to development of generating cells (e.g., Sylett et al. 1995; Keeler et al. 2016). That cooling  
761 might pair with heating associated with snow particle growth (depositional and riming growth)  
762 within the cloud, resulting in lapse rate maintaining convection (Kumjian et al. 2014).  
763 Alternatively, if there was advection of drier and/or cooler (i.e., lower  $\theta_e$ ) air aloft, it could play a  
764 role in producing the potential instability, as shown in the soundings from the 16-17 December  
765 and 1 February cases. Future studies should examine the relative roles of microscale updrafts as  
766 compared to mesoscale and synoptic scale updrafts.

767 The wind-shear instability (i.e. Kelvin–Helmholtz instability) can also produce downdrafts,  
768 which are not focused on in this study. The updrafts can contribute to supplying water vapor  
769 producing water saturation and supercooled water in the cells and hence enhancement of particle  
770 growth and snow intensification (Plummer et al. 2014; Kumjian et al. 2014), while the downdrafts  
771 can play a role in entrainment and sublimation (Rauber et al. 2014). The downdraft regions in this  
772 study were found between URs at midlevels and near the cloud tops (e.g., Figs. 11, 15). The spatial

773 distribution of the downdraft regions was very similar to that of URs, except near the cloud tops.  
774 This suggests that URs existed as a UR and downdraft region in the midlevels of the storms. The  
775 midlevels of the storms could be supersaturated, then precipitation particles did not have a chance  
776 of much sublimation within the downdraft regions. Near the cloud tops, detecting downdraft  
777 regions by the radar measurements might be difficult because hydrometeor particles could be  
778 sublimated. Considering the microphysics within the updraft/downdraft regions near the cloud tops  
779 and at midlevels should be important to analyze mass/water budgets and entrainment and  
780 understand the water cycle within the cloud systems. We will need lidar measurements to observe  
781 the clear-air updrafts and downdrafts and high-resolution simulations that resolve the microscale  
782 updrafts and downdrafts.

783 Limited cases in this study (four cases) result in some uncertainty in the relationships between  
784 URs and wind shear. Analysis including more cases using the high-resolution radar observations  
785 can help to better understand the formation mechanisms of URs and the roles of URs in  
786 microphysical processes in winter storms. Model simulations will help to further understand the  
787 mechanisms of UR generations and the microphysical impacts including the sources of multi layers  
788 that generate turbulence and ice particle formation, if the models can resolve the wind shear layers  
789 and use optimal turbulence kinematic energy parameterization. The model simulations are beyond  
790 the scope of this observational study.

791 The methodology to identify microscale updrafts described in this paper can be applied to other  
792 vertically-pointing Doppler radar datasets to examine the small scale velocity structures that are  
793 too fine scale to be resolved by other sensors such as the US National Weather Service PPI  
794 scanning radars. Further analysis of microscale updrafts and downdrafts can also shed light on the  
795 degree to which ice precipitation mass changes in a more episodic mode (smaller time and space  
796 scales) versus a more continuous mode (longer time and scale scales) in winter storms and in  
797 stratiform regions. The spatial/time scales of the updrafts in this study are larger than the radar  
798 sampling volume ( $\sim 15$  m in height and  $\sim 1$  s in time). There can be smaller scale turbulence (i.e.  
799 eddies) within the sampling volume, which can be observed by higher resolution radars/lidars  
800 (having a few cm scale resolution, Schmidt et al. 2012; Cooper and Chattopadhyay 2014).

801

802 **5. Summary**

803        Although microscale upward motions in the snowstorms could play an important role in snow  
804 precipitation, their characteristics relative to the snowbands and relationship with vertical wind  
805 shear instability from Kelvin–Helmholtz instability has not been shown yet. We investigated the  
806 relations among microscale updraft regions (URs) and snowbands, vertical wind shear, shear  
807 instability, and frontogenesis, and roles on vertical mass transport for four winter snowstorm cases  
808 (4 January 2018; 18 January 2020; 16–17 December 2020; 1 February 2021). Measurements were  
809 made as each storm passed over the Stony Brook and Brookhaven National Laboratory Radar  
810 Observatory (SBRO) in Stony Brook, NY and included periods with and without snowbands and  
811 in the northwest and northeast quadrants of the three storms (4 January, 16 – 17 December, and 1  
812 February) and the southwest quadrant of the 18 January storm. Three of the cases (4 January, 16 –  
813 17 December, and 1 February) included more than one snowband.

814        The mean Doppler velocity from the vertically-pointing Ka-band polarimetric radar  
815 measurements was used to identify updraft regions in this study. After the removal of hydrometeor  
816 sedimentation from the VPT measured mean Doppler velocity, the updraft region was simply  
817 defined as a region with the estimated vertical air motion (VPT measured mean Doppler velocity  
818 minus estimated particle fall speed) greater than or equal to  $0.4 \text{ m s}^{-1}$  (negative sign of Doppler  
819 velocity indicates a downward motion). Using vertically-pointing radar data does not provide  
820 information on the lifecycle of the updrafts, but the duration of the updraft overhead can be  
821 converted to an estimated spatial scale using horizontal wind estimated from a KASPR PPI VAD  
822 technique. We estimated mass flux based on the detected updraft regions and sounding data and  
823 investigated relationships among updraft regions, mass flux, and frontogenesis, and shear  
824 instability to discuss roles of mesoscale instability on generating the updrafts and vertical mass  
825 transport. We summarize the characteristics of the updraft regions and their associations with  
826 spectral width, wind shear and frontogenesis:

- 827        • In the vertical pointing radar data, the distribution of updraft durations is approximately  
828 logarithmic with most individual updrafts passing overhead in only a few seconds ( $<20$   
829 sec) (Fig. 5a). They are roughly equivalent to spatial scales  $< 500 \text{ m}$ . The aspect ratio  
830 of the updraft regions estimated from the duration and horizontal scale has a lognormal  
831 frequency distribution with a modal value near 1 (equivalent to circular, Fig. 5b).
- 832        • Overall, the snowbands occur in a region of low to mid-level frontogenesis, but the  
833 microscale updraft regions are found at all altitudes, occur both in regions with and

834 without snowbands. (Figs. 8a, 12a, 16a, and 20a). They suggest a role of the microscale  
835 updraft regions on snow particle formation /or intensification of snowfall not only in  
836 the snowbands but also outside the snowbands.

- 837 • The subset of updraft regions with larger VPT Doppler spectrum width ( $SW > 0.4 \text{ m s}^{-1}$ ), which imply more turbulent updrafts, frequently occur within mid-levels of the  
838 storms (several km below cloud echo top). They are associated with vertical wind shear  
839 and moist shear instability that are sometimes multi-layered in the precipitating clouds.  
840 They suggest that the updraft regions can be triggered in the shear unstable layers (Figs.  
841 9, 13, 17, and 21).
- 843 • Updraft regions that tend to be closer together in space and time yield more bulk mass  
844 flux. The duration of individual updraft regions is not the primary factor of the mass  
845 flux (Figs. 8e, 12e, 16e, and 20e).
- 846 • The calculated upward bulk mass flux occurs within and above the moist shear  
847 insurability layers. The highest mean mass flux is produced during snowband periods  
848 (Figs. 8d, 12d, 16d, and 20d).

849 The analysis implies that the dominant forcing of the microscale updrafts appears to be  
850 associated with the vertical shear instability. The updrafts are responsible for upward mass flux  
851 and then contribute to the precipitation mass growth regardless of whether snowbands are present  
852 or not. We also detected many microscale updrafts in the boundary layer during portions of each  
853 of the four storms but did not focus on them in this study. These low-level updrafts may further  
854 modify the microphysics and would be an interesting topic for future studies. Future studies with  
855 increasing cases should examine the relative roles of microscale updrafts, downdrafts, and  
856 turbulence as compared to mesoscale and synoptic scale updrafts and quantitatively estimate the  
857 contributions to the mass growth to better understand the precipitation production in winter storms,  
858 which in turn may aid in reducing uncertainties in snowfall accumulation forecasts.

## 861 Appendix A

862 There are six major Doppler spectral broadening mechanisms that contribute to the measured  
863 spectrum width  $\sigma_{tot}$  (Doviak and Zrnić, 2006): dispersion of hydrometeor fall speed  $\sigma_{psd}$ ,  
864 turbulence  $\sigma_{tur}$ , mean horizontal wind shear  $\sigma_{shr}$ , cross wind  $\sigma_{crs}$ , antenna motion  $\sigma_{ant}$ , and

865 parameters related to the particle orientation (including canting, horizontal orientation, wobbling,  
 866 and oscillation of shape, Dunnavan, 2021)  $\sigma_{ori}$ . Those contributions can be written as:

$$867 \quad \sigma_{tot} = \sqrt{\sigma_{psd}^2 + \sigma_{tur}^2 + \sigma_{shr}^2 + \sigma_{crs}^2 + \sigma_{ant}^2 + \sigma_{ori}^2} \quad (A1)$$

868 To simplify, the contributions of  $\sigma_{ant}$  and  $\sigma_{ori}$  are ignored, assuming that they are negligible  
 869 compared to the other terms. Moreover, for the VPT measurements,  $\sigma_{shr}$  and  $\sigma_{crs}$  can also be  
 870 negligible compared to  $\sigma_{psd}$  and  $\sigma_{tur}$ . The value of  $\sigma_{shr}$  in Eq. A1 differs from the original  
 871 surveillance-scan form proposed by Doviak and Zrnić (2006) because we neglect possible  
 872 contributions from vertical wind shear across the sampling volume; only the horizontal wind shear  
 873 component is considered here. For the VPT measurements in this study, we consider the  $\sigma_{psd}$  and  
 874  $\sigma_{tur}$  to evaluate the contributions to  $\sigma_{tot}$  as:

$$875 \quad \sigma_{tot} = \sqrt{\sigma_{psd}^2 + \sigma_{tur}^2} \quad (A2)$$

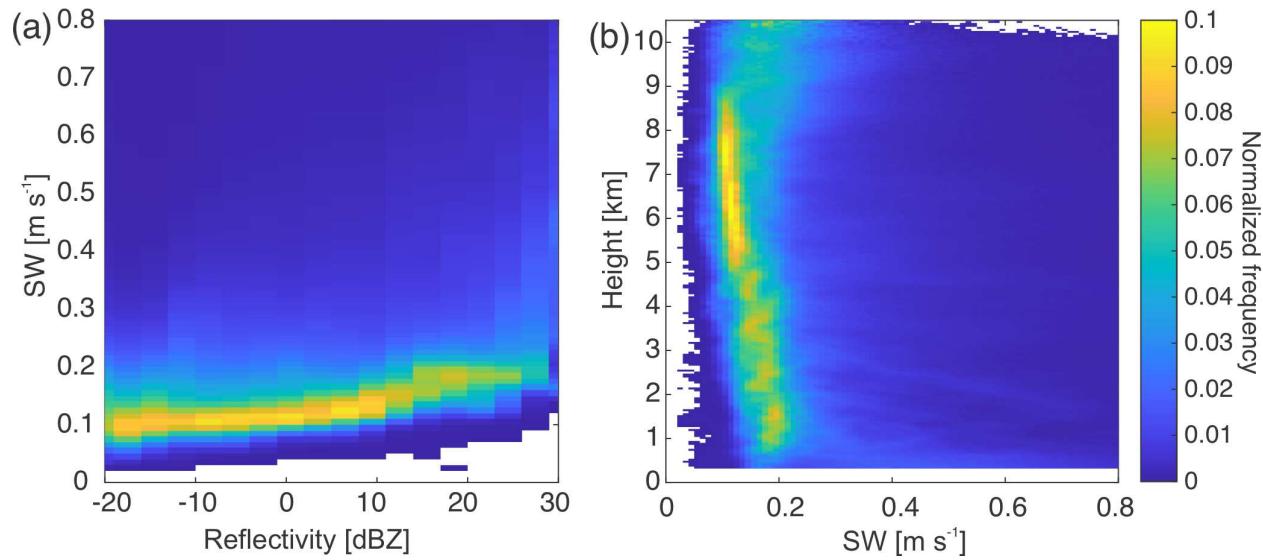
876 In the case of VPT measurements,  $\sigma_{tur}$  in Eq. (A2) can represent the dispersion of radial velocities  
 877 within the sampling volume owing to turbulent structures and/or horizontal gradients in coherent  
 878 vertical structures. To estimate general  $\sigma_{psd}$  for the present cases, Fig. A1a shows a frequency  
 879 distribution of SW versus reflectivity from the KASPR VPT measurements collected during the  
 880 selected four events in this study. In this figure, we assume that  $\sigma_{psd}$  can be correlated with  
 881 reflectivity statistically. The SW with the maximum frequency at each reflectivity bin increased  
 882 with reflectivity. This can represent statistical  $\sigma_{psd}$  associated with reflectivity. For the four cases,  
 883 which are stratiform snow events, radar reflectivity generally increases with decreasing height,  
 884 suggesting that hydrometeor particles generally grow downward. As the particle size distribution  
 885 broadens with increasing reflectivity,  $\sigma_{psd}$ , which is attributed to the spread of particle fall speeds  
 886 in the radar resolution volume can also increase. The SW with maximum frequency is generally  
 887 less than  $0.2 \text{ m s}^{-1}$ . Figure A1b shows vertical frequency distribution of the SW. The SW with  
 888 maximum frequency increases downward and is generally less than  $0.2 \text{ m s}^{-1}$ .

889 Figure A2 shows the relationship of Eq. A2 to represent the contributions of  $\sigma_{psd}$  and  $\sigma_{tur}$  to  
 890  $\sigma_{tot}$ . At  $\sigma_{tot}$  (observed SW) =  $0.3 \text{ m s}^{-1}$  with  $\sigma_{psd} = 0.2 \text{ m s}^{-1}$ ,  $\sigma_{tur}$  is approximately  $0.2 \text{ m s}^{-1}$ ,  
 891 comparable with  $\sigma_{psd}$ . For  $\sigma_{tot} > 0.35 \text{ m s}^{-1}$  with  $\sigma_{psd} = 0.2 \text{ m s}^{-1}$ ,  $\sigma_{tur} > \sigma_{psd}$ , indicating that  $\sigma_{tur}$   
 892 can be a primary contribution to the observed SW. We use  $\sigma_{tot} = 0.4 \text{ m s}^{-1}$  for the threshold of  
 893 observed SW to classify detected updraft regions related to turbulence. When we compared  
 894 statistics using a threshold of  $\sigma_{tot} = 0.4 \text{ m s}^{-1}$  and a threshold of  $0.5 \text{ m s}^{-1}$ , the medians of duration

895 and vertical extent for URSWHI and URSWLO did not change much (less than 1 sec and less than  
 896 15 m, respectively). The medians of reflectivity and upward motion slightly increased as we  
 897 expected; median reflectivity for URSWHI increased by  $\sim 0.5$  dB and that for URSWLO increased  
 898 by  $\sim 2$  dB, and median upward motions for both URSWHI and URSWLO increased by less than  
 899  $0.05 \text{ m s}^{-1}$ .

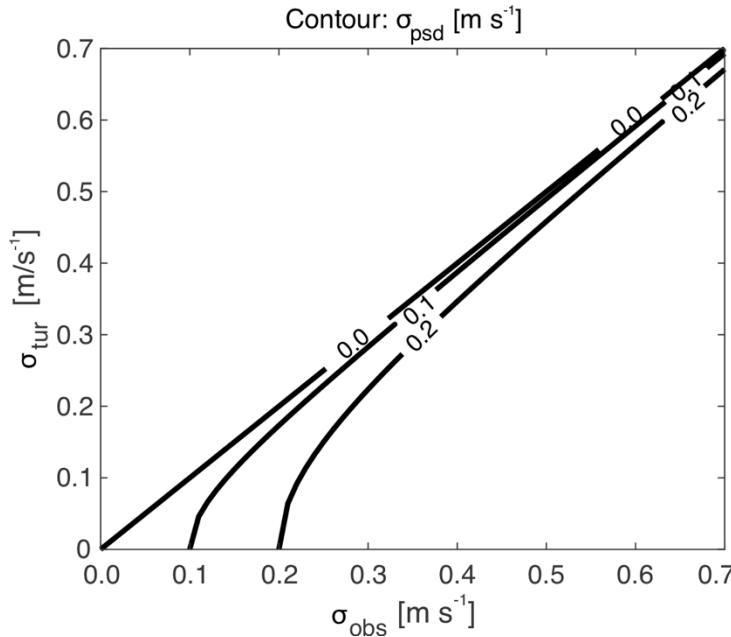
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901



902

903 Figure A1: (a) Frequency distribution of SW versus reflectivity from the KASPR VPT  
 904 measurements collected during the four events. Color shade represents the frequency normalized  
 905 every 2 dBZ from -20 dBZ. (a) Frequency by altitude distribution of SW. Color shade represents  
 906 the frequency normalized at each radar range-gate (every 15 m).  
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908

909 Figure A2:  $\sigma_{tur}$  versus  $\sigma_{tot}$  diagram with corresponding  $\sigma_{psd}$  based on Eq. A2.

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920

921 *Data Availability Statement.*

922 KASPR and sounding data used in this study are available at Stony Brook University Academic  
 923 Commons Repository (<https://commons.library.stonybrook.edu/somasdata/17>). Those data collected  
 924 under IMPACTS are also available at the NASA Global Hydrometeorology Resource Center (GHRC)  
 925 DAAC system (<http://dx.doi.org/10.5067/IMPACTS/RADAR/DATA101>,  
 926 <http://dx.doi.org/10.5067/IMPACTS/SOUNDING/DATA301>).

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