



1 Global climatology of vertical cloud motion observed by 2 EarthCARE/CPR: Cloud-type dependence with link to microphysics

3 Haruka Hotta^{1,2}, Kentaroh Suzuki², Maki Kikuchi¹, Shunsuke Aoki¹, Takuji Kubota¹

4 ¹Earth Observation Research Center, Japan Aerospace Exploration Agency, Tsukuba, 305-8505, Japan

5 ²Atmosphere and Ocean Research Institute, University of Tokyo, Kashiwa, 277-8564, Japan

6 *Correspondence to:* Haruka Hotta (hotta.haruka@jaxa.jp)

7 **Abstract.** Nadir-pointing cloud-radar Doppler velocity (V_D) quantifies the reflectivity-weighted vertical motion of
8 hydrometeors by combining sedimentation and vertical air motion. Thus, it provides an observational constraint on cloud
9 microphysical–dynamical processes. However, as spaceborne Doppler measurements have only recently become available, its
10 global climatology remains poorly documented. This study utilized EarthCARE Cloud Profiling Radar observations to
11 establish the global climatology of V_D and evaluate its association with temperature, reflectivity, and cloud type. Quality-
12 controlled V_D samples were analyzed after classifying cloud layers into five types based on the cloud-top temperature and
13 column-maximum reflectivity. Results indicate that V_D is predominantly downward, consistent with hydrometeor
14 sedimentation, and that the downward motion becomes faster at warmer temperatures. Below the melting layer, V_D varies
15 among cloud types, even at comparable reflectivities. Further, warm precipitating clouds exhibited faster downward V_D than
16 non-precipitating warm clouds, whereas precipitation originating from cold cloud layers demonstrated even faster downward
17 V_D . These cloud-type differences indicate that V_D describes precipitation pathways and vertical air motion, which are not
18 captured by reflectivity alone. Geographical patterns of mean V_D display faster downward velocities over tropical convective
19 regions dominated by precipitating clouds, but much weaker downward velocities over subtropical stratocumulus regions
20 dominated by non-precipitating warm clouds. Additionally, the fractional occurrences of substantially positive V_D highlight
21 upward particle motion signatures obscured in the mean fields, particularly in strongly precipitating cold clouds in the tropics.
22 Therefore, spaceborne Doppler measurements provide a benchmark for evaluating cloud microphysics, precipitation formation,
23 and convective transport in global models.

24 1 Introduction

25 The vertical motion of cloud hydrometeors links cloud microphysical processes to cloud dynamics. Within clouds, updrafts
26 and downdrafts transport hydrometeors, consequently modulating their microphysical evolution. For example, updrafts under
27 abundant water vapor sustain supersaturation, thereby promoting the activation, condensation, and depositional growth of



28 hydrometeors. Meanwhile, the entrainment of dry environmental air prevents supersaturation through turbulent mixing and
29 enhances evaporation/sublimation (Korolev and Mazin, 2003; Blyth, 1993; Chandrakar et al., 2021). Additionally,
30 hydrometeors sediment gravitationally at a terminal fall speed determined by their size, shape, and density (Pruppacher and
31 Klett, 2010; Mitchell, 1996; Heymsfield and Westbrook, 2010). Through these coupled dynamic and microphysical pathways,
32 the vertical motion of hydrometeors modulates cloud optical properties and persistence, and ultimately governs the Earth's
33 radiation budget (Hartmann et al., 1992; Albrecht, 1989; Stevens and Feingold, 2009). Therefore, characterizing vertical
34 motion within clouds on a global scale is important for accurately assessing the role of clouds in climate systems.

35 Cloud-radar Doppler velocity (V_D) is a powerful observable for evaluating vertical motions within clouds. In this study, we
36 adopt a sign convention in which upward motion is positive. For nadir- or zenith-pointing observations, V_D can be expressed
37 as:

$$38 \quad V_D = w - V_{t,z} \quad (1)$$

39 where w is the vertical air velocity and $V_{t,z} (> 0)$ is the reflectivity-weighted hydrometeor terminal fall speed (directed
40 downward) (Doviak and Zrnić, 1993). Therefore, V_D contains both microphysical information through $V_{t,z}$, which depends on
41 the particle type and size, and dynamic information through w . Since separating V_D into w and $V_{t,z}$ requires assumptions that
42 may vary across cloud regimes, we focus on the Doppler velocity itself as a directly observable quantity that can be compared
43 with forward-simulated radar measurements. When the mean vertical air motion is weak or cancels over sufficiently large
44 space–time averages, the mean V_D primarily reflects the sedimentation of the predominant hydrometeors (Mason et al., 2018;
45 Kalesse and Kollias, 2013). In contrast, the contributions from w can remain even in the mean V_D in cloud systems with active
46 convection or wave motion, or when statistics are computed on sufficiently fine scales to retain turbulent variability (Kollias
47 et al., 2003; Shupe et al., 2008). Furthermore, the broadening of the particle size distribution and variability in the vertical air
48 motion intensify the spread of the V_D distribution (Shupe et al., 2008; Zhu et al., 2021). Consequently, statistical
49 characterization of V_D distribution, such as its mean and spatiotemporal dispersion, offers useful information on cloud
50 microphysics–dynamics interactions.

51 Observational studies have actively used V_D from ground-based cloud radar and airborne radar platforms. These studies have
52 linked updraft–downdraft structures to raindrop size distributions in convective precipitation (Kollias et al., 2003; Yuter and
53 Houze 1995), documented weak sedimentation and gravity-wave signatures in the upper-tropospheric cirrus (Kalesse and
54 Kollias, 2013), and examined the roles of ice particle sedimentation and turbulent mixing in persistent Arctic mixed-phase
55 clouds (Shupe et al., 2004; Morrison et al., 2012). Despite these advances, previous investigations have been inherently
56 constrained by limited spatial and temporal sampling. Moreover, quantitative intercomparison across regions and cloud
57 regimes has been challenging because radar configurations and analysis procedures vary among studies, including operating
58 frequency, sensitivity, beamwidth, and processing steps such as noise removal and velocity de-aliasing (Kollias et al., 2007;
59 Doviak and Zrnić, 1993). Consequently, a globally consistent climatology for Doppler velocity statistics has not yet been
60 established (Illingworth et al., 2015; Kollias et al., 2023).



61 Importantly, such a climatology would provide an observational benchmark for evaluating model representations of
62 hydrometeor sedimentation, precipitation formation, and convective transport, thereby aiding in the diagnosis of systematic
63 model biases associated with cloud and precipitation processes. In the upper troposphere, climatological V_D statistics for ice
64 clouds can constrain the ice-particle fall speed formulations assumed in the models. In the lower troposphere, a shift in the V_D
65 distribution toward faster downward velocities from non-precipitating to precipitating warm-cloud regimes would provide a
66 process-oriented test of warm-rain microphysics. Similarly, the frequency of strongly upward V_D signatures in convectively
67 active clouds provides complementary information on convective dynamical activity. When compared with forward-simulated
68 radar observables, these diagnostics could inform constrain assumptions and tuning choices that differ substantially across
69 global models (Hourdin et al., 2017; Mauritsen et al., 2012; Mitchell et al., 2011; Stephens et al., 2010; Tsushima et al., 2020).
70 Launched in late May 2024, the Earth Cloud Aerosol and Radiation Explorer (EarthCARE) provides a unique opportunity to
71 establish homogeneous global Doppler velocity climatology and interpret it within a cloud-type-resolved framework (Wehr et
72 al., 2023; Kubota et al., 2026). Notably, the Cloud Profiling Radar (CPR) onboard EarthCARE is the first spaceborne cloud
73 radar with Doppler capability in W-band (94 GHz). The CPR provides global observations from a common instrument and
74 operational processing chain for climatological applications. Pre-launch algorithm studies and early post-launch assessments
75 in EarthCARE have begun to quantify the key error characteristics of the CPR Doppler product and validate its performance,
76 thereby supporting the feasibility of robust global statistics (Hagihara et al., 2023; Kollias et al., 2023; Kim et al., 2025).
77 Collectively, these capabilities offer a globally consistent observational basis for systematically characterizing and comparing
78 Doppler velocity signatures across diverse cloud regimes and geographical environments, including remote oceans and polar
79 regions.

80 Against this background, this study established the global climatology of V_D using EarthCARE/CPR observations. Our analysis
81 emphasizes the regional statistics of V_D , including its mean and an intense updraft frequency metric, which are defined as the
82 frequency with which V_D exceeds the specified thresholds. Agreement between observed and simulated means can arise from
83 compensating errors in cloud-type occurrence and type-conditioned V_D , and does not necessarily imply that the underlying
84 microphysical and dynamical processes are represented correctly. Accordingly, rather than treating the global field as a single-
85 regime quantity, we interpreted it as a superposition of distinct cloud populations. To achieve this, clouds were classified into
86 five types based on phase characteristics ("warm" versus "cold") and a reflectivity-based indicator of precipitation intensity.
87 For each cloud type, the regional occurrence of that type and V_D statistics conditioned on it were analyzed. Within this
88 framework, we investigate whether global Doppler velocity climatology can reveal physically interpretable signatures linked
89 to cloud microphysical regimes, precipitation pathways, and cloud dynamics, beyond those available from conventional radar-
90 reflectivity-based climatologies. We addressed this overarching issue using two specific questions: First, how do Doppler
91 velocity statistics vary among cloud regimes with different phase characteristics and precipitation states, and do these contrasts
92 persist at comparable radar reflectivities? Second, to what extent do cloud-type occurrences and cloud-type-conditioned
93 Doppler velocity statistics contribute to regional contrasts in the overall Doppler velocity climatology?



94 The remainder of this paper is organized as follows: Section 2 describes the EarthCARE/CPR data, quality control procedures,
95 and the cloud-type classification method. Section 3 examines the temperature dependence of V_D statistics and their contrasts
96 among cloud types. Thereafter, Section 4 documents the regional patterns in cap V sub dV_D and diagnoses the drivers of
97 regional contrasts using type-decomposition analysis, and Section 5 summarizes the main findings and discusses their
98 implications and challenges.

99 **2 Data and methods**

100 **2.1 Dataset**

101 We used observations from the Cloud Profiling Radar (CPR), an instrument onboard the EarthCARE satellite that measures
102 cloud structure using Doppler radar. EarthCARE follows a sun-synchronous near-polar orbit with an inclination of 97.05° ,
103 passes through the same local solar time, and crosses the equator at nominal local times of approximately 02:00 (ascending)
104 and 14:00 (descending) (Wehr et al., 2023). The W-band (94 GHz) CPR provides nadir-pointing profiles for the radar
105 reflectivity factor (Z_e) and Doppler radial velocity (V_D). The instantaneous beam-limited footprint was approximately 750 m
106 at the nadir, and profiles were generated at approximately 500 m spacing along the track in the native sampling. Further, the
107 intrinsic pulse-limited range resolution was 500 m, and the signal was oversampled on a 100 m vertical grid.

108 Furthermore, we used a cloud mask Z_e , V_D , and the air temperature from the standard JAXA CPR Level-2A one-sensor product
109 CPR_CLP Version Cb (JAXA, 2026b). All variables were provided on a 1 km horizontal grid and a 100 m vertical grid.
110 CPR_CLP screens second-trip echoes, following Battaglia (2021) (Aoki et al., 2026a). A cloud mask was derived using
111 established noise and echo-continuity tests, coupled with surface-echo information (Sato et al., 2025). Meteorological variables,
112 such as air temperature and pressure, were derived from the ECMWF forecast fields (Eisinger et al., 2024) and interpolated
113 onto a Level-2 grid. Throughout the study, we adopted the convention that positive (negative) V_D values denote upward
114 (downward) motion.

115 The CPR operates in three observation modes with profiling windows of 16, 18, and 20 km and the corresponding pulse
116 repetition frequencies (PRFs). Shorter profiling windows permit higher PRFs, consequently reducing the random Doppler
117 uncertainty and Nyquist folding likelihood; however, they limit the maximum observable altitude. Meanwhile, longer windows
118 allow observations at greater altitudes but result in larger Doppler uncertainty and a greater likelihood of Nyquist folding. The
119 latitude-dependent MIX-mode configuration was revised after launch. On July 8, 2025, the profiling window was changed
120 from 20 km to 18 km within 60°S – 60°N , whereas the 16-km mode remained in use at higher latitudes. Since November 26,
121 2025, the boundary between the 18- and 16-km modes has been 40° , with the 18-km mode used within 40°S – 40°N and the 16-
122 km mode at higher latitudes (Imura et al., 2026).

123 In this study, we restricted our analysis to the period between July 9, 2025, and July 8, 2026, after the profiling window was
124 set to 18 km or lower. Imura et al. (2026) demonstrated that the 16 km and 18 km modes provide more accurate Doppler



measurements than the 20 km mode, with comparable standard deviations between modes 16 and 18. Restricting the analysis to the 16- and 18-km modes reduces the temporal inhomogeneity in the Doppler velocity sampling. As cloud occurrence above 18 km is rare, excluding echoes above this altitude is expected to have a minimal impact on the resulting statistics (Imura et al., 2026).

2.2 Doppler velocity quality control

As presented in Eq. (1), the Doppler velocity (V_D) represents the combined vertical air velocity (w) and hydrometeor fall speed term ($V_{t,z}$). Although one term can be estimated by imposing assumptions, such as $w \approx 0$ after extensive averaging or prescribed $V_{t,z}$ -particle size relations, these assumptions introduce errors that may systematically vary across cloud regimes and regions. Therefore, this study established climatological statistics for V_D that can be directly compared with forward-simulated radar observables from the models. Accordingly, this section describes the preprocessing used to convert the gate-level CPR_CLP Doppler velocity into Doppler-velocity samples employed for climatological statistics.

CPR V_D is influenced by random error, antenna-pointing error, multiple scattering, non-uniform beam filling, and Nyquist folding (Tanelli et al., 2005; Kollias et al., 2014; Kollias et al. 2022; Hagihara et al., 2023; Sy et al. 2014). The CPR_CLP product corrects the antenna-pointing bias by assuming that the 100-km mean Doppler velocity near the surface is zero (JAXA, 2026a). Nonetheless, residual scene-dependent biases may remain, particularly over land. Therefore, regional signals should be interpreted with caution (JAXA, 2026a; Puigdomènech Treserras et al., 2025).

We restricted the analysis to cloud interior gates to reduce random uncertainty near weak echoes and cloud boundaries. A gate was considered cloud interior when the entire centered 3×3 neighborhood satisfied both the cloud-mask and reflectivity criteria (cloud mask ≥ 20 and $Z_e \geq -19$ dBZ). This screening preferentially excludes weak echoes and cloud edges whose V_D distributions may differ from those of the retained samples. Therefore, we interpret the reported statistics as conditional on these sampling criteria and emphasize the cloud-type- and reflectivity-conditioned contrasts.

The CPR_CLP product applies an operational velocity-unfolding procedure for warm gates ($T \geq 273.15$ K), wherein Doppler velocities exceeding 3 m s^{-1} on the ordinary branch are interpreted as fast-falling hydrometeors and shifted by one Nyquist interval (Hagihara et al., 2023; JAXA, 2026a). Since residual folding may remain, we additionally screened profiles for fold-like vertical discontinuities using the 3-km along-track averaged V_D and ray-wise Nyquist velocity. Additionally, distinct criteria were applied below and above 273.15 K to account for their different velocity branches. Profiles containing at least one fold-like discontinuity were excluded from the target samples used for the climatological statistics. The details of this procedure are provided in Sect. S1.2. Although excluding fold-like profiles is necessary to prevent aliased velocities from contaminating quantitative statistics, this screening may preferentially exclude some of the most dynamically active convective profiles where strong vertical motions are more likely to produce velocity folding (Galfione et al., 2025; Hotta et al., 2026).

For each retained target gate, V_D was locally averaged over a 3-km along-track $\times$ 200-m vertical window (3×2 gates) using reflectivity-weighted pulse-pair averaging. The average was retained only when all six gates in the window were valid. The



157 pulse-pair mean was calculated using the ray-wise Nyquist velocity, and its velocity branch was restored relative to the
158 CPR_CLP velocity at the target gate to preserve the operationally unfolded branch. The details of the averaging are provided
159 in Sect. S1.1. The cloud type was assigned according to the cloud layer containing the target gate (Section 2.3), whereas the
160 averaging window was not restricted to the gates of the same cloud type. Thus, the average V_D can include neighboring gates
161 belonging to differently classified cloud layers. Hereafter, V_D refers to this quality-controlled, locally averaged quantity,
162 whereas an overbar denotes the statistical average over a temperature bin, latitude–pressure bin, geographical region, or cloud-
163 type subset.

164 2.3 Cloud type classification

165 A vertically continuous sequence of analysis-valid cloudy gates, as defined in Section 2.2, is considered a cloud layer. Cloud
166 layers separated by one or more invalid gates were treated as distinct layers. Therefore, a single radar profile could contain
167 multiple cloud layers.

168 Each cloud layer was classified into one of five microphysically distinct types using (1) cloud-top temperature (CTT) and (2)
169 a radar-reflectivity-based precipitation indicator. In this study, the thermodynamic phase (liquid versus ice) and precipitation
170 intensity were the two fundamental axes representing the cloud microphysical characteristics. Such phase- and precipitation-
171 based categorizations are common in satellite studies (Jing et al. 2017; Matus and L'Ecuyer 2017). Accordingly, we developed
172 a classification scheme that separated distinct populations using only the aforementioned two characteristics.

173 (1) Warm/Cold (CTT-based phase proxy)

174 In this study, we defined air temperature at the cloud-layer top as the cloud-top temperature (CTT). Cloud layers with $CTT \geq$
175 273.15 K (0°C) are classified as Warm type, whereas those with $CTT < 273.15$ K are classified as Cold type. Using a CTT
176 threshold of 273.15 K is consistent with standard practice in satellite meteorology, where CTT helps to distinguish the cloud
177 phase (Gryspeerdt et al., 2014). Ice- and mixed-phase processes are important in the Cold-type layers. These include ice particle
178 formation and growth by vapor deposition, the Wegener–Bergeron–Findeisen mechanism, riming, and aggregation, which
179 affect cloud development in ways that differ from purely warm-liquid clouds (Pruppacher and Klett, 2010; Korolev et al.,
180 2017).

181 (2) No Precip/Weak Precip/Strong Precip (reflectivity-based proxy for precipitation formation)

182 Next, we distinguished precipitation intensity using radar reflectivity. Under the Rayleigh scattering approximation, the radar
183 reflectivity is strongly dependent on the particle size and is proportional to the sixth power of the particle diameter. The
184 appearance of precipitation-sized particles greatly increases reflectivity, so the maximum reflectivity within a cloud column is
185 typically used as a practical marker of the cloud-to-rain transition (Doviak and Zrnić, 1993; Beall et al., 2024; Mace, 2007).
186 For the Warm-type, we used the layer-maximum reflectivity Z_{e_max} , defined within each cloud layer. Accordingly, we define
187 Warm Precip when $Z_{e_max} \geq 0$ dBZ and Warm No Precip when $Z_{e_max} < 0$ dBZ. This choice follows common practice in
188 CloudSat-based warm-rain studies, where column-maximum reflectivity is often used to classify non-



precipitating/drizzling/raining regimes, with 0 dBZ serving as a practical threshold for “raining (certain rain)” conditions (Jing et al. 2017; Mülmenstädt et al., 2015; Beall et al., 2024). As our Doppler-velocity analysis is restricted to gates with $Z_e \geq -19$ dBZ, we did not further subdivide the warm population using the -15 dBZ drizzle threshold sometimes adopted in CloudSat analyses (Beall et al., 2024).

For the Cold-type, Z_{e_max} can be dominated by the melting layer and/or rain below it. To capture the precipitation formation associated with ice-phase particles (e.g., aggregation and riming), we used the maximum reflectivity within the subfreezing portion of the layer ($T \leq 273.15$ K), denoted $Z_{e_cold,max}$. We first defined Cold No Precip as $Z_{e_cold,max} < 0$ dBZ. For the remaining cold layers, we define Cold Strong Precip as $Z_{e_cold,max} \geq 10$ dBZ and Cold Weak Precip as $0 \leq Z_{e_cold,max} < 10$ dBZ. A 10 dBZ value is an indicator of a strong echo core in CloudSat-based structural metrics (Mülmenstädt et al., 2015; Beall et al., 2024; Luo et al. 2011).

Therefore, each cloud layer is classified into one of the following five types:

- Warm No Precip: $CTT \geq 273.15$ K and $Z_{e_max} < 0$ dBZ
- Warm Precip: $CTT \geq 273.15$ K and $Z_{e_max} \geq 0$ dBZ
- Cold No Precip: $CTT < 273.15$ K and $Z_{e_cold,max} < 0$ dBZ
- Cold Weak Precip: $CTT < 273.15$ K and $0 \leq Z_{e_cold,max} < 10$ dBZ
- Cold Strong Precip: $CTT < 273.15$ K and $Z_{e_cold,max} \geq 10$ dBZ

Since our cloud-type classification is a simple threshold-based proxy based on CTT and radar reflectivity, it faces several limitations. First, the cold type ($CTT < 0^\circ\text{C}$) can include clouds containing supercooled liquid water and, therefore, does not strictly separate the liquid phase from ice-phase clouds (Korolev et al., 2017). Second, for 94 GHz (W-band) observations, non-Rayleigh (Mie) scattering effects for large ice particles and melting hydrometeors, along with attenuation and multiple scattering, can modify the observed reflectivity. Consequently, the reflectivity thresholds used in this study (0 and 10 dBZ) cannot be uniquely mapped to specific particle sizes or precipitation intensities (Sassen et al., 2007; Heymsfield et al., 2008; Battaglia et al., 2010, 2011). Third, multilayer scenes can complicate the interpretation because attenuation or multiple scattering associated with other cloud layers in the same radar column may affect the observed reflectivity of the target layer. Considering these shortcomings, our classification cannot be interpreted as a definitive determination of the cloud microphysical state. Rather, this study offers a practical stratification framework for organizing globally comparable statistics.

3 Temperature and cloud-type dependence of Doppler velocity

3.1 Vertical profiles of reflectivity and Doppler velocity

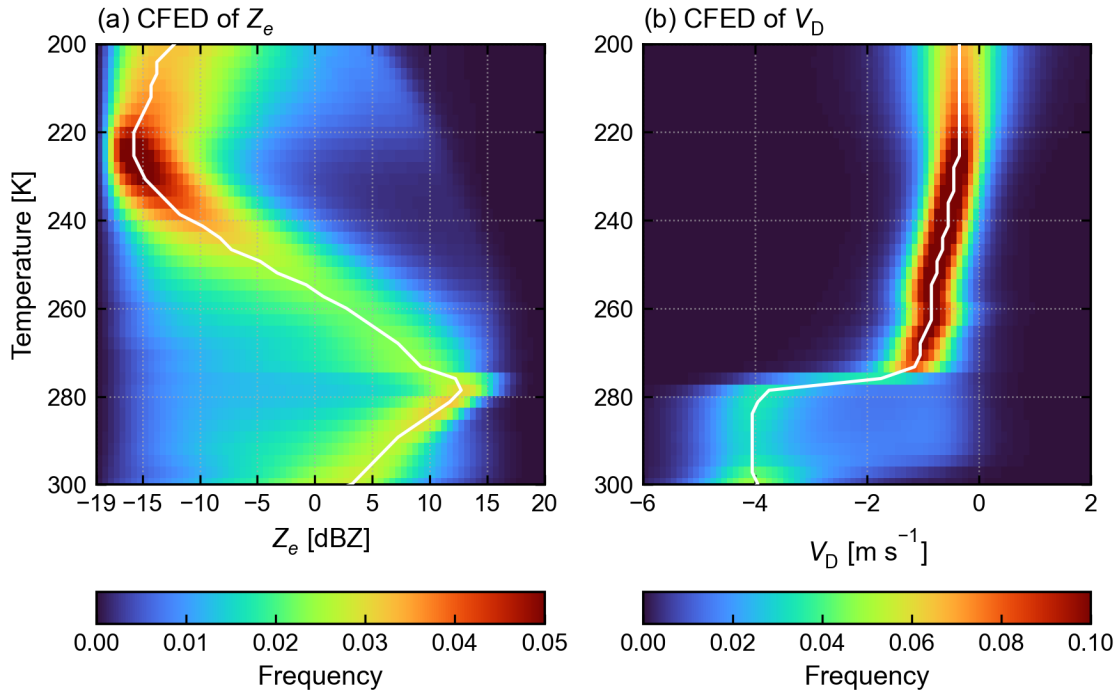
To characterize the overall temperature dependence of radar reflectivity (Z_e) and Doppler velocity (V_D) measured by EarthCARE/CPR, we constructed Contoured Frequency by tEmperature Diagrams (CFEDs; Hashino et al., 2013) for global clouds, as illustrated in Fig. 1. Since air temperature is closely linked to hydrometeor phases and growth pathways (Korolev



et al., 2017), we used temperature rather than geometric altitude on the ordinate. The modal value at each temperature is indicated by the white line.

Fig. 1a demonstrates that, at subfreezing levels, the reflectivity distribution transitions from lower to higher values as the temperature increases, which aligns with the results of Hashino et al. (2013). A distinct enhancement on the high-reflectivity side is evident near 273.15 K, where the mode reaches approximately 10–15 dBZ, thereby indicating the melting-layer bright band (Zhang et al., 2008; Sassen et al., 2007). At higher temperatures, the modal value shifts back toward lower reflectivities, which implies increased W-band attenuation in liquid hydrometeors along with changes in the hydrometeor population below the melting layer (Haynes et al., 2009; Mitrescu et al., 2010; Battaglia et al., 2020; Lebsock and L’Ecuyer, 2011; Xie et al., 2016).

Fig. 1b illustrates that the modal V_D indicates a downward motion across all temperatures, which is consistent with a substantial contribution from hydrometeor sedimentation (Shupe et al. 2008; Protat and Williams 2011). Between 200 and 270 K, the distribution is centered at approximately -1.5 to 0.5 m/s, with the mode shifting toward faster downward velocities as temperature increases. The weaker dependence near 260 K may suggest aggregation into low-density particles (Connolly et al., 2012), whose larger projected area and lower bulk density can limit increases in fall speed (Locatelli and Hobbs 1974; Heymsfield and Westbrook 2010; Korolev et al. 2017). Above approximately 273.15 K, the mode shifts to faster downward velocities of around -4 m s⁻¹, which aligns with substantially higher terminal velocities of raindrops (Gunn and Kinzer, 1949; Beard, 1976; Atlas et al., 1973). The broader distribution in the warm layer (from approximately -5 to 0 m s⁻¹) further suggests greater variability in raindrop-size distributions and stronger influences from turbulence and vertical air motion, including updrafts and downdrafts (Atlas et al., 1973; Doviak and Zrnić, 1993; Shupe et al., 2008).



240

241 **Figure 1. Contoured frequency-by-temperature diagrams (CFEDs) of (a) radar reflectivity factor (Z_e) and (b) Doppler velocity (V_D).**

242 **Shading indicates frequency normalized within each temperature bin, and the white curves show the mode at each temperature.**

243 Following the overall temperature dependence illustrated in Fig. 1, Fig. 2 compares the dependences of Z_e and V_D on the five
 244 cloud types. A cloud layer classified as Cold type (CTT < 273.15 K) extending across the melting level may include gates at
 245 temperatures ≥ 273.15 K. In Fig. 2a, the cloud types are separated according to the reflectivity-based classification. Both No
 246 Precip categories (Warm and Cold) were concentrated at lower dBZ values, whereas the precipitation categories were shifted
 247 toward higher reflectivities, as expected from their definitions. Among the three Cold categories, the median Z_e increased from
 248 Cold No Precip (lowest Z_e) to Cold Weak Precip to Cold Strong Precip (highest Z_e). Z_e in Cold-type layers increases as
 249 temperature approaches 273.15 K. Above 273.15 K, the median Z_e of the cold-origin precipitating categories decreases, most
 250 clearly for Cold Strong Precip, which is consistent with stronger W-band attenuation along the rain path. For the Warm Precip
 251 type, Z_e generally increases at higher temperatures. This downward increase aligns with previous warm-rain studies
 252 demonstrating that collision-coalescence and drizzle/rain growth shift the reflectivity maximum toward the cloud base (Wang
 253 and Geerts, 2003; Suzuki et al., 2010, 2015).

254 Fig. 2b indicates that the median V_D across all cloud types shifted toward faster downward velocities at warmer temperatures.
 255 For the three Cold types, although the medians were not well separated at subfreezing temperatures ($T < 273.15$ K), the
 256 precipitating types tended to exhibit faster downward Doppler velocities than the Cold No Precip. This was consistent with
 257 larger and/or more rimed particles having faster terminal velocities. Among the three Cold types, Cold Strong Precip displays
 258 a broad distribution spanning both upward and downward Doppler velocities under cold conditions. This aligns with greater



variability in particle fall speeds and vertical air motion in precipitating systems (Pruppacher and Klett, 2010; Heymsfield and Westbrook, 2010; Korolev et al., 2017; Kollias et al., 2003; Shupe et al., 2008). Furthermore, the differences among the cloud types became more pronounced above 273.15 K. The median Doppler motion becomes progressively faster downward from Cold No Precip to Cold Weak Precip and Cold Strong Precip. Meanwhile, Warm types typically exhibit a slower downward Doppler motion than Cold types. Within the Warm types, Warm Precip exhibits faster downward median Doppler velocities and a wider percentile spread, which is consistent with active collision-coalescence and a broader spectrum of drop sizes. In contrast, Warm No Precip is centered closer to zero, thereby aligning with the weak net downward motion typical of non-precipitating liquid clouds (Pruppacher and Klett, 2010; Beard, 1976). These results indicate that the all-cloud V_D distribution is not a single temperature-dependent microphysical sequence. Instead, they reflect a mixture of cloud populations with different phase characteristics and precipitation states. This motivated the joint Z_e – V_D analysis in the following section, which examined whether differences in Doppler velocity remained at comparable reflectivity.

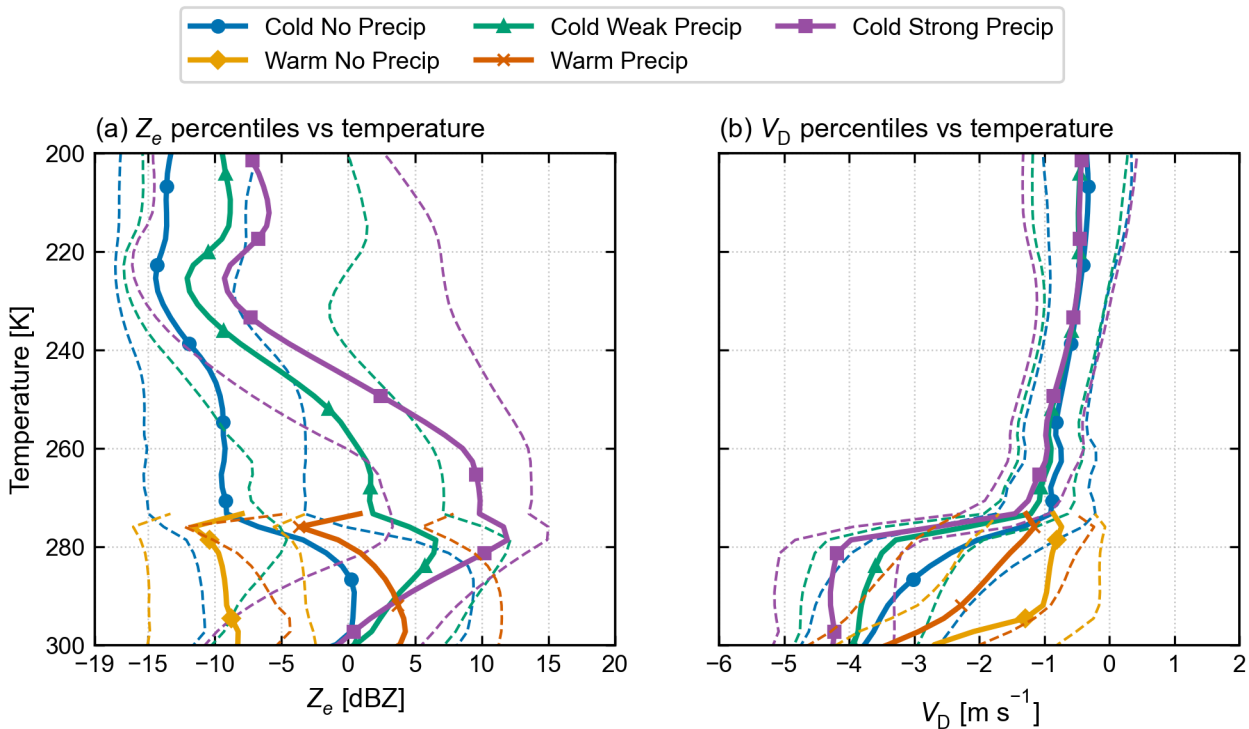


Figure 2. Temperature dependence of (a) radar reflectivity factor (Z_e) and (b) Doppler velocity (V_D) stratified by cloud type. Solid lines indicate medians, and dashed lines depict the 10th and 90th percentiles within each temperature bin. Cloud types are defined at the cloud-layer scale using cloud-top temperature and layer-maximum reflectivity metrics (Sect. 2.3).

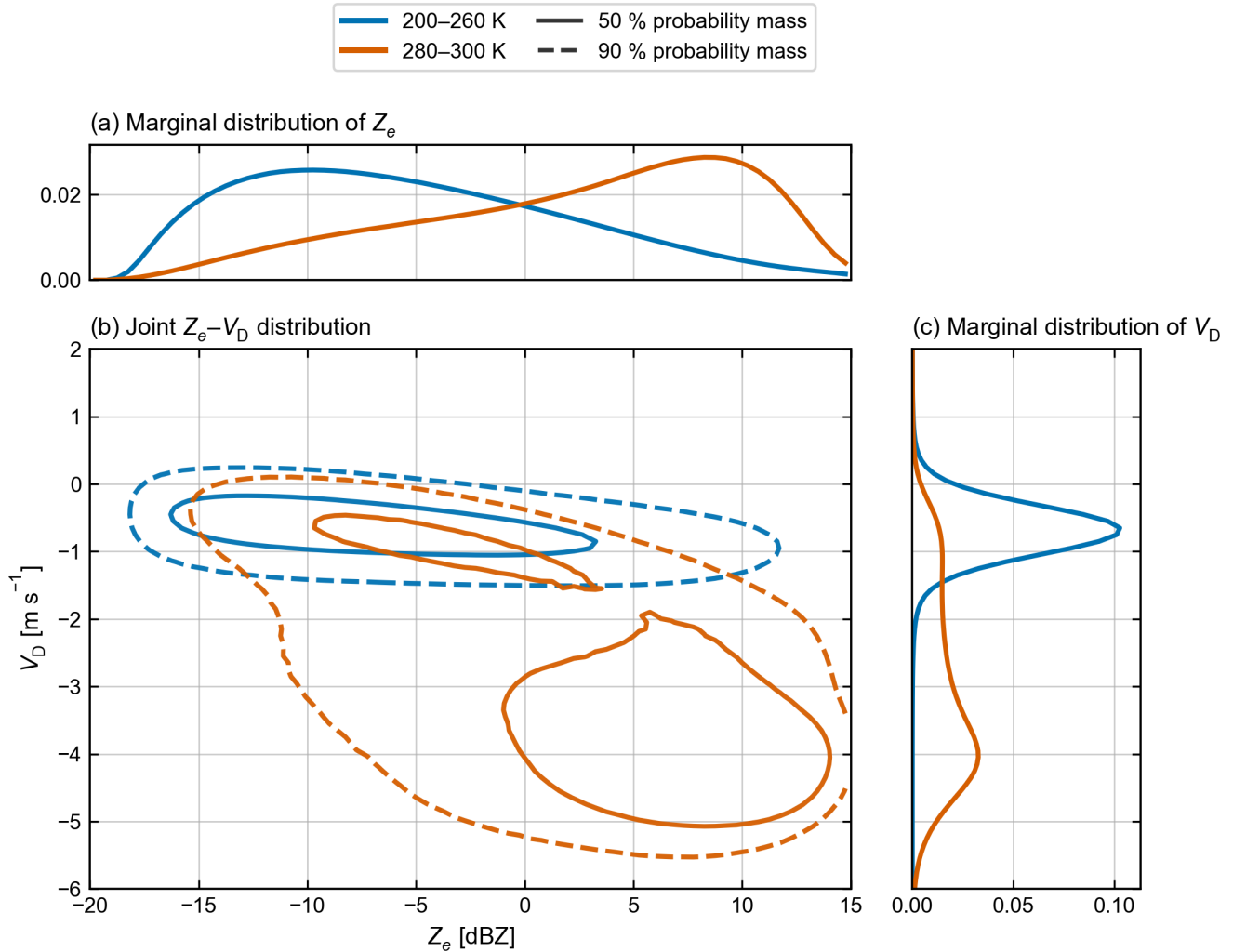


3.2 Joint reflectivity–Doppler velocity statistics of various cloud types

Figs. 1 and 2 illustrate that both Z_e and V_D vary systematically with temperature and cloud type. We examined the joint Z_e – V_D distributions in Fig. 3 for two representative temperature ranges: a cold regime (200–260 K) and a warm regime (280–300 K), while avoiding the melting-layer transition near 273.15 K, to determine whether V_D simply reflects changes in the observed Z_e or whether V_D contains additional information. Previous Doppler radar studies have employed similar joint-distribution diagnostics to interpret the combined effects of hydrometeor sedimentation and vertical air motion (Kollias et al., 2003; Shupe et al., 2008; Delanoë et al. 2007; Protat and Williams, 2011; Roh et al. 2026).

In the cold regime, Z_e spans a broad range, but V_D remains concentrated between -1.5 and 0.5 m s^{-1} , consequently producing nearly horizontal contours that indicate weak variation in V_D with Z_e (Seiki et al. 2025a, b). This weak dependence appears to arise primarily from the limited range of $V_{t,z}$ for ice and snow. Snow fall speeds are typically relatively low, and $V_{t,z}$ rarely reaches several-meter-per-second values, which is characteristic of rain, except where dense, strongly rimed particles dominate. Aggregation can substantially increase the particle dimensions and Z_e without a comparable increase in $V_{t,z}$ because it often produces open, low-density particles with a large projected area relative to the mass. Variations in particle habit, density, and riming further make the Z_e – $V_{t,z}$ relationship non-unique (Locatelli and Hobbs, 1974; Heymsfield and Westbrook, 2010; Kalesse et al., 2016; Mroz et al., 2023). Previous millimeter-wave Doppler radar studies have demonstrated that the reflectivity–fall–speed relationship varies with the dominant ice growth processes (Kalesse et al., 2013). Residual vertical air motion can add scatter; however, the relatively narrow V_D distribution suggests that this is not the primary cause of the weak climatological dependence.

In the warm regime, Z_e shifts toward higher values, while V_D spans a much wider range extending from near zero to approximately -5 to -6 m s^{-1} . Therefore, the joint contours exhibited a clearer shift toward a faster downward V_D with increasing Z_e than in the cold regime. The broader V_D distribution is attributable to the much wider range of $V_{t,z}$ among the liquid hydrometeors, although vertical air motion can also contribute to the spread. Compared to ice aggregates, whose bulk density generally decreases as they grow, liquid drops remain compact with a nearly constant material density. Therefore, their mass-to-projected-area ratio increases markedly with size, thereby allowing the terminal fall speeds to reach considerably larger values. Consequently, $V_{t,z}$ can reach much larger values in rainfall than in most ice particle populations. Since $V_{t,z}$ is reflectivity-weighted, the emergence of large, fast-falling drops through collision–coalescence further shifts $V_{t,z}$ toward higher values while simultaneously increasing Z_e (Gunn and Kinzer, 1949; Atlas et al., 1973; Beard, 1976; Heymsfield and Westbrook, 2010). Moreover, the warm regime combines warm cloud droplets and drizzle, warm rain, and precipitation originating from cold cloud layers, thereby further broadening the V_D distribution. This cloud type dependence is presented in Fig. 4. Collectively, these temperature-dependent interpretations are consistent with Fig. 1, where V_D spans a relatively narrow range at subfreezing temperatures but becomes much more broadly distributed in the warm layer.



307

308 **Figure 3. Joint and marginal distributions of radar reflectivity factor (Z_e) and Doppler velocity (V_D) for the cold (200–260 K; blue)**
 309 **and warm (280–300 K; red) regimes. (a) Marginal Z_e distributions, (b) joint $Z_e - V_D$ distributions, and (c) marginal V_D**
 310 **distributions. Solid and dashed contours in (b) enclose the highest-density regions containing 50% and 90% of the probability mass,**
 311 **respectively. Marginal distributions are normalized independently within each temperature regime.**

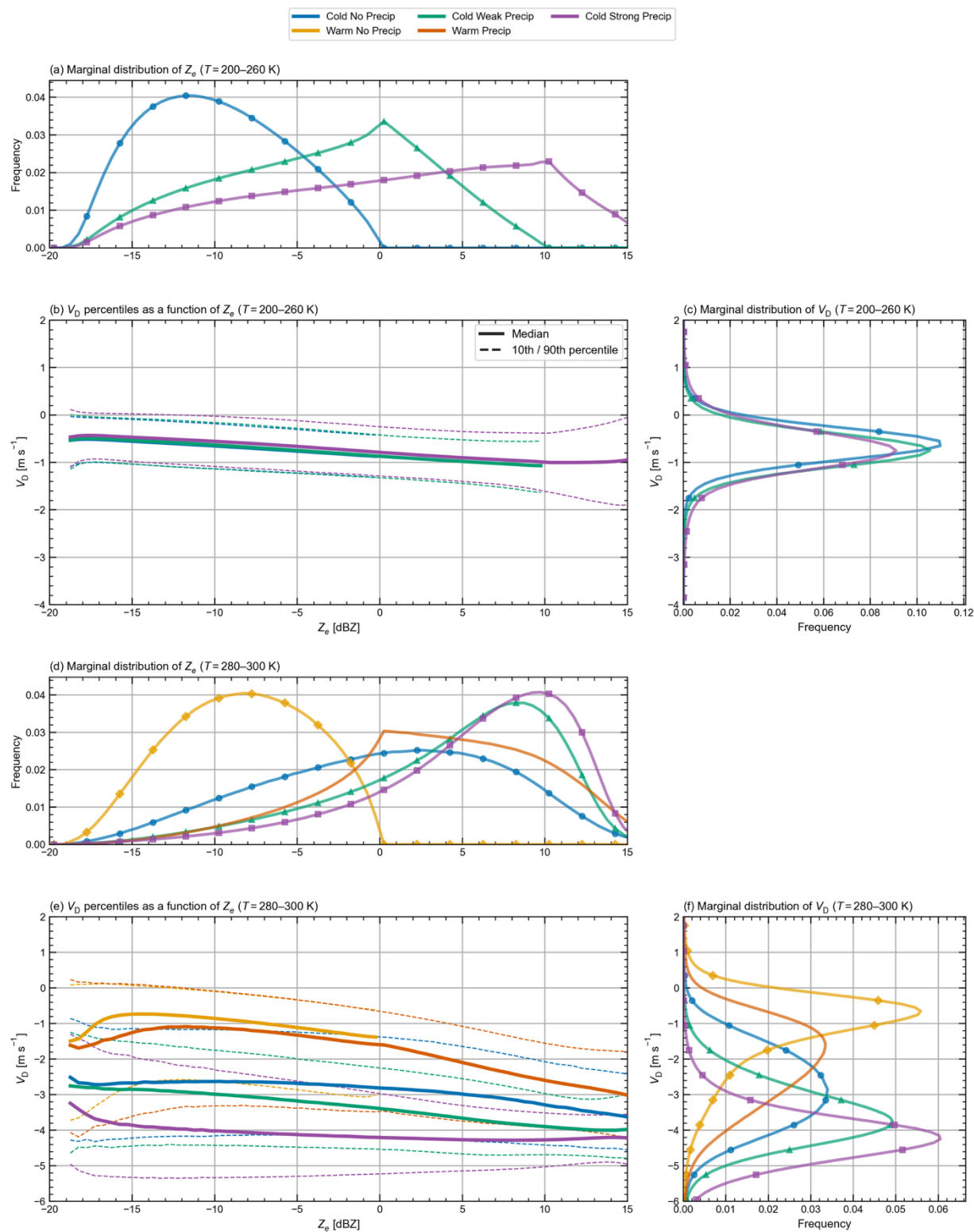
312 Fig. 4 further examines the $Z_e - V_D$ relationships in Fig. 3 by cloud type for the same two temperature ranges (200–260 K:
 313 cold regime; 280–300 K: warm regime). In the cold regime (Fig. 4a–c), the distributions comprise only the three Cold types.
 314 The marginal Z_e distribution (Fig. 4a) shifts toward higher reflectivity from Cold No Precip to Cold Weak Precip and Cold
 315 Strong Precip, which is expected from the reflectivity-based classification. Further, the conditional V_D statistics reveal
 316 differences that are not directly imposed by the reflectivity-based classification (Fig. 4 b). At a given observed Z_e , the median
 317 V_D differs only weakly among the cold categories. Meanwhile, Cold Strong Precip displayed the broadest upward and
 318 downward-velocity tails, which aligns with prior Doppler cloud-radar studies (Kollias et al., 2002; Shupe et al., 2008; Kalesse



319 et al., 2016; Zhu et al., 2021; Kneifel and Moisseev, 2020). This suggests that Cold Strong Precip clouds are linked to a wider
320 diversity of terminal fall speeds due to vigorous ice-particle growth or larger fluctuations in vertical air motion.

321 In the warm regime (Fig. 4d–f), the contrasts among the cloud types were especially pronounced. At a comparable Z_e , warm
322 precipitation exhibited a faster downward V_D than warm no precipitation, whereas cold-origin precipitating clouds generally
323 displayed a faster downward V_D , particularly cold strong precipitation (Fig. 4e). The separation between the warm precipitation
324 and cold-origin precipitation categories in Fig. 4e may indicate differences in the precipitation formation pathways.
325 Aggregation and riming aloft, followed by melting, can produce drop-size distributions that differ from those formed
326 predominantly through warm cloud collision–coalescence (Giangrande et al., 2016; Raut et al., 2021). At a given observed Z_e ,
327 a particle size distribution with a greater contribution from large particles may have a faster reflectivity-weighted terminal fall
328 speed $V_{t,z}$ than that dominated by smaller particles, because Z_e and $V_{t,z}$ depend differently on the particle size distribution
329 (Mroz et al., 2023). The cloud-type-dependent V_D separation at a given observed Z_e may also be enhanced by the differential
330 W-band attenuation (Sasikumar et al. 2026). If attenuation is stronger in deeper cold-origin precipitating columns, their
331 observed Z_e can be reduced while V_D is less directly affected; consequently, the same observed Z_e does not necessarily
332 represent comparable intrinsic hydrometeor populations across cloud types. Additionally, differences in the vertical air motion
333 could contribute to a lower w , whether due to stronger downdrafts or weaker updrafts, which would shift the observed Doppler
334 motion toward faster downward velocities. However, the present analysis does not independently separate w from $V_{t,z}$ and,
335 therefore, cannot determine the relative contributions of microphysical and dynamical differences to cloud-type separation, as
336 illustrated in Fig. 4e.

337 Cloud-type stratification reveals that the broad warm-regime distribution in Fig. 3 is primarily a population mixture signal (Fig.
338 S1) and not within a single cloud type. These populations range from warm non-precipitating clouds with low Z_e and weak
339 downward Doppler motion, warm precipitating clouds with high Z_e and faster downward motion, to cold-origin clouds with
340 high Z_e and yet faster downward motion. This broadening, attributed to the mixing of cloud types, directly reflects the key
341 contrast between regimes and is consistent with previous V_D analyses, wherein aggregating distinct microphysical–dynamical
342 populations led to broader joint and marginal statistics (Battaglia et al., 2020; Mroz et al., 2023).





344 **Figure 4.** Cloud-type-stratified Z_e and V_D statistics for the same temperature regimes as Fig. 3. The top row (a,b,c) illustrates the
345 cold regime (200–260 K), and the bottom (d,e,f) row presents the warm regime (280–300 K). Panels (a, d) indicate marginal Z_e
346 distributions, panels (b, e) present the 10th, 50th, and 90th percentiles of V_D as functions of Z_e , and panels (c, f) illustrate marginal
347 V_D distributions.

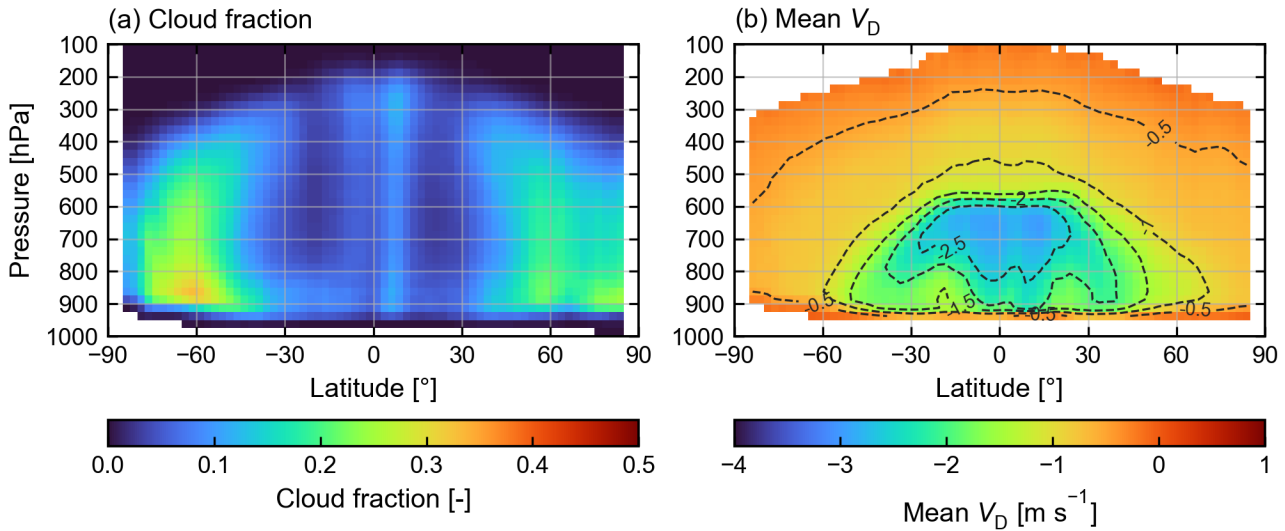
348

349 **4 Regional variability of Doppler velocity and cloud-type attribution**

350 **4.1 Zonal-mean picture of Doppler velocity**

351 Having established the temperature- and cloud-type-dependent characteristics of V_D , we examined how these signatures were
352 represented geographically. Fig. 5 illustrates the zonal-mean cloud fraction (left panel) and mean V_D (right panel) in latitude–
353 pressure coordinates. Here, the cloud fraction is defined as the frequency of the analysis-valid cloudy gates used in the V_D
354 statistics, and therefore, does not represent the total cloud occurrence. As it requires relatively strict conditions, the resulting
355 cloud fraction is expected to be smaller than that derived from W-band radar cloud masks (Marchand et al., 2008). Nevertheless,
356 the resulting latitude–pressure pattern is largely consistent with previous active-sensor cloud climatologies, with enhanced
357 low-level cloudiness at mid-to high-latitudes and deep cloud occurrences in the tropics (Stephens et al., 2008; Mace et al.,
358 2009; Chepfer et al., 2010; Naud et al., 2010; Wall et al., 2017).

359 The right panel elucidates that the zonal-mean Doppler velocity was downward throughout the latitude–pressure domain. In
360 the upper troposphere (approximately 100–400 hPa), the mean V_D indicated a relatively slow downward motion. Downward
361 motion progressively became faster toward lower levels, reaching approximately 600 hPa. The downward motion in the
362 tropical mid-to-lower troposphere (approximately 600–900 hPa) is particularly fast, with the mean V_D reaching approximately
363 -3 m s^{-1} near 700 hPa; below this level, the downward motion slowed and varied relatively little. This vertical structure was
364 broadly consistent with the temperature dependence illustrated in Figure 1b. Additionally, the downward Doppler motion at a
365 given lower-tropospheric pressure is generally slower in the subtropics than in the tropics or mid-to-high latitudes.



366

367 **Figure 5. Zonal-mean latitude–pressure cross-sections of (left) analysis-valid cloud fraction and (right) mean Doppler velocity (V_D).**

368 **Cloud fraction is defined as the frequency of analysis-valid cloudy range gates used in the V_D statistics (Sect. 2.2). Statistics were**
 369 **calculated in $2.5^\circ \times 25$ hPa latitude–pressure bins. Bins containing 10000 or fewer samples contributing to the respective statistics**
 370 **were masked.**

371 Next, these patterns were decomposed into cloud-type and cloud-type-conditioned mean V_D . Accordingly, the zonal mean
 372 V_D ($\overline{V_D}$) is represented as:

$$373 \quad \overline{V_D} = \sum_{k=1}^5 f_k \times \overline{V_D}^{(k)}, \quad (2)$$

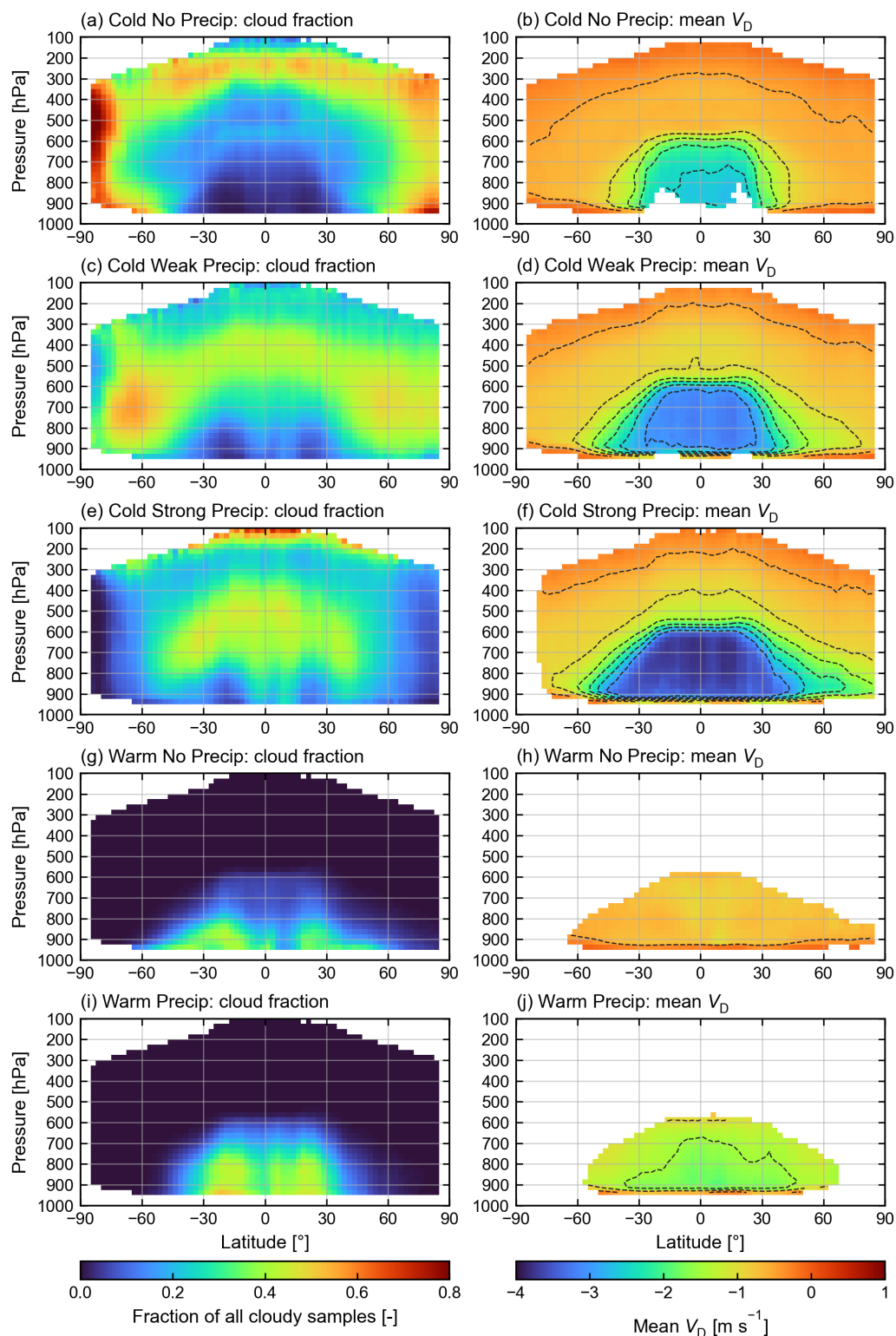
374 where f_k denotes the cloud-type fraction relative to the total cloud fraction and $\overline{V_D}^{(k)}$ represents the mean V_D conditional on
 375 cloud type k . This “frequency $\times$ conditional-mean” framework is broadly applied in satellite cloud-type or regime climatologies
 376 to attribute gridded mean structures to the relative prevalence of distinct cloud categories (Rossow and Schiffer, 1999; Mace
 377 et al., 2009; Jakob and Tselioudis, 2003).

378 Fig. 6 resolves the zonal mean V_D in Fig. 5 into the cloud-type fraction f_k (left column) and the corresponding cloud-type-
 379 conditioned mean Doppler velocity $\overline{V_D}^{(k)}$ (right column). By definition, $\sum_k f_k = 1$ in each latitude–pressure bin. The three Cold
 380 types dominated the upper and middle tropospheres and extended into the lower troposphere, whereas the warm types were
 381 confined mainly to low levels, which is consistent with the CTT-based classification.

382 The right column displays the common vertical dependence of $\overline{V_D}^{(k)}$ for the three Cold types. In the upper troposphere
 383 (approximately 200–400 hPa), all three types exhibited weak downward motion, generally between 0 and 1 m s^{−1}. The weak
 384 downward zonal mean V_D in the upper troposphere in Fig. 5 primarily reflects the common weakly downward $\overline{V_D}^{(k)}$ across the
 385 Cold types (Shupe et al., 2008; Protat and Williams, 2011; Heymsfield and Westbrook, 2010).



386 In the mid–lower troposphere, $\overline{V_D^{(k)}}$ becomes progressively faster downward for all Cold types, particularly in tropical and
387 subtropical regions (35°S–35°N). The fast downward zonal-mean $\overline{V_D}$ in the tropical mid–lower troposphere (Fig. 5) primarily
388 reflects this common vertical intensification of $\overline{V_D^{(k)}}$, rather than a regime shift to a specific precipitation type. Cloud-type
389 differences modulate the magnitude. Notably, faster downward motion for cold strong precipitation below approximately 600–
390 700 hPa, coupled with their frequent occurrence in the tropics, reinforced the tropical minimum in $\overline{V_D}$.
391 In contrast, the slower downward $\overline{V_D}$ in the lower troposphere, which was most evident in the subtropics, was primarily
392 explained by the cloud-type composition. At these levels, Warm clouds become frequent and display $\overline{V_D^{(k)}}$ values that are
393 much closer to zero than those of Cold clouds at similar pressures. Consequently, the zonal mean $\overline{V_D}$ shifts toward a weaker
394 downward motion near the surface, even though the Cold-type downward $\overline{V_D^{(k)}}$ itself remains nearly as fast as that in the layers
395 above. Meanwhile, among the Warm types, Warm No Precip exhibits a particularly weak downward $\overline{V_D^{(k)}}$. Its dominance in
396 the subtropical boundary layer shifts the $\overline{V_D}$ toward a weaker downward motion than that in the tropics.
397 Thus, Fig. 6 illustrates that the upper- and mid-tropospheric structures of the zonal mean $\overline{V_D}$ are governed chiefly by the
398 common vertical dependence within Cold clouds, whereas the subtropical lower-tropospheric weakening is primarily driven
399 by increasing contributions from Warm clouds.





401 **Figure 6. Decomposition of the zonal-mean $\overline{V_D}$ (Fig. 5) by cloud type. Left column: Cloud-type fraction f_k in each latitude–pressure**
402 **bin (summing to unity across types). Right column: Type-conditioned mean Doppler velocity $\overline{V_D}^{(k)}$ computed from gates assigned to**
403 **each type. Rows correspond to the five cloud types defined in Sect. 2.3. Statistics were calculated in $2.5^\circ \times 25$ hPa latitude–pressure**
404 **bins. or cloud-type fractions, bins with 10000 or fewer total cloudy samples were masked; for type-conditioned mean $\overline{V_D}$, bins with**
405 **10000 or fewer samples of the corresponding cloud type were masked.**

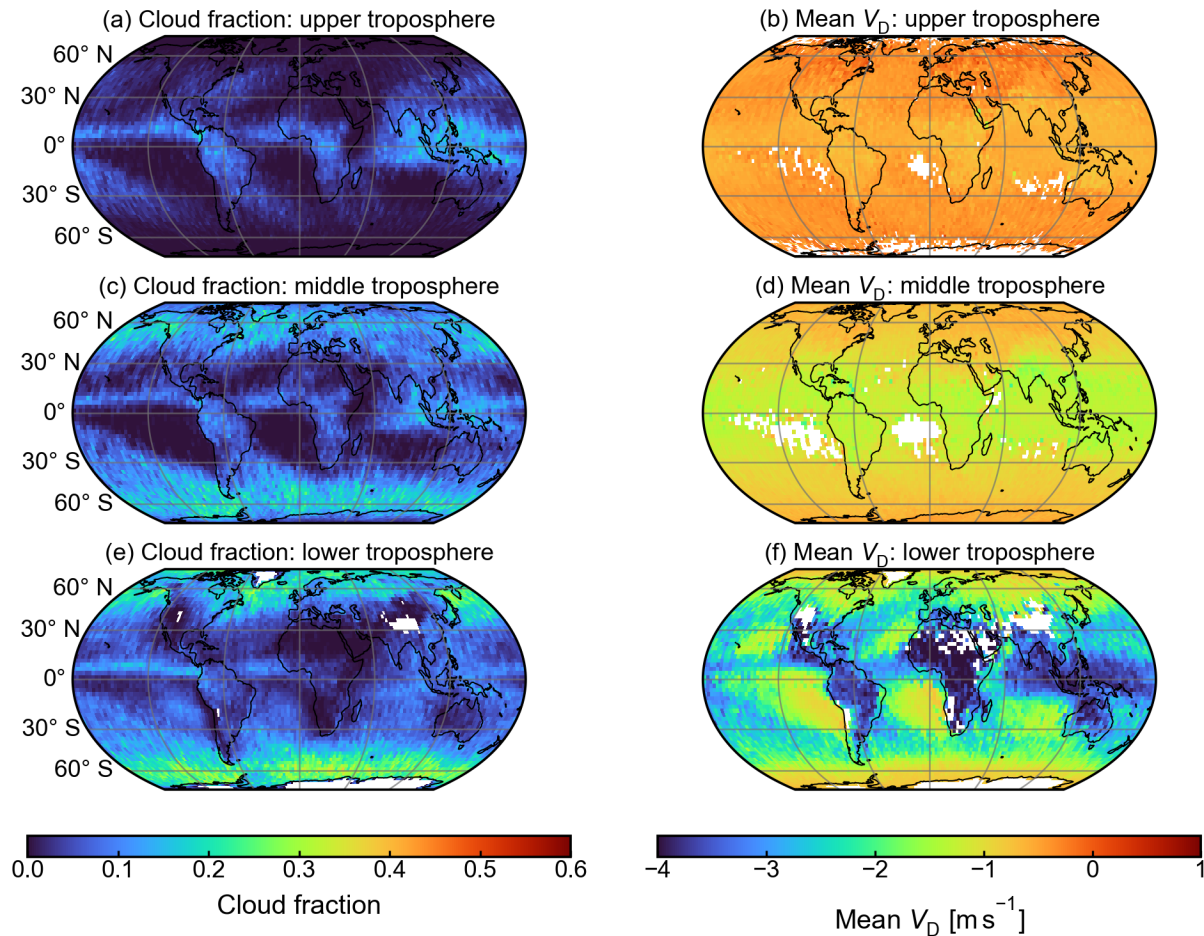
406 4.2 Geography of Doppler velocity at varying levels

407 To examine the geographical structure underlying the zonal means in Fig. 5, Fig. 7 illustrates the horizontal distributions of
408 the analysis-valid cloud fraction (left column) and mean $\overline{V_D}$ (right column) for the upper (200–300 hPa), middle (450–550 hPa),
409 and lower (800–900 hPa) troposphere. In this section, $\overline{V_D}$ represents the average Doppler velocity observed in each region and
410 pressure range.

411 In the upper troposphere (Fig. 7, top panels), the cloud fraction was elevated in the tropics (20°S–20°N), particularly in the
412 regions with active deep convection. These include the Indian Ocean–Maritime Continent–Western Pacific warm pool,
413 convergence zones over the tropical Atlantic and Pacific (ITCZ/SPCZ), and tropical continental convective centers such as the
414 Amazon and Congo basins (Schumacher and Houze, 2003; Stephens et al., 2008; Houze, 2014). The $\overline{V_D}$ indicates weak
415 downward motion (commonly between -1 and 0 m s⁻¹), with little geographical variation.

416 In the mid-tropospheric pressure range (Fig. 7, middle panels), the cloud fraction was relatively high over both tropical
417 convective regions and mid-latitudes (35°–70°), which aligns with previous satellite-based cloud climatologies (Mace et al.,
418 2009; Naud et al., 2010; Stephens et al., 2008; Houze, 2014). Relative to the upper-tropospheric pressure range, the downward
419 Doppler motion was faster in the tropics (20°S–20°N) but remained relatively weak in the midlatitudes.

420 In the lower tropospheric pressure range (Fig. 7, bottom panels), the cloud fraction exhibited pronounced regional variability.
421 High cloud occurrence was observed over subtropical eastern ocean stratocumulus regions, mid-latitude oceans, and tropical
422 convectively active regions. These distributions are consistent with those reported in previous studies (Klein and Hartmann,
423 1993; Stevens, 2005; Wood, 2012; Mace et al., 2009; Naud et al., 2010). $\overline{V_D}$ generally indicated the fastest downward motion
424 among the three pressure ranges and displayed the strongest regional contrast. Additionally, tropical convective regions
425 generally exhibit faster downward Doppler motion, whereas stratocumulus regions in the eastern subtropics and mid-latitude
426 low-cloud regimes exhibit weak downward motions close to zero.



427

428 **Figure 7. Horizontal distributions of (left) cloud fraction and (right) mean Doppler velocity (V_D) in the upper (200–300 hPa), middle**
 429 **(450–550 hPa), and lower (800–900 hPa) troposphere. Statistics were calculated in $2.5^\circ \times 2.5^\circ$ latitude–longitude grid boxes after**
 430 **aggregating samples within each pressure range. Grid boxes containing 100 or fewer samples contributing to the respective statistics**
 431 **were masked.**

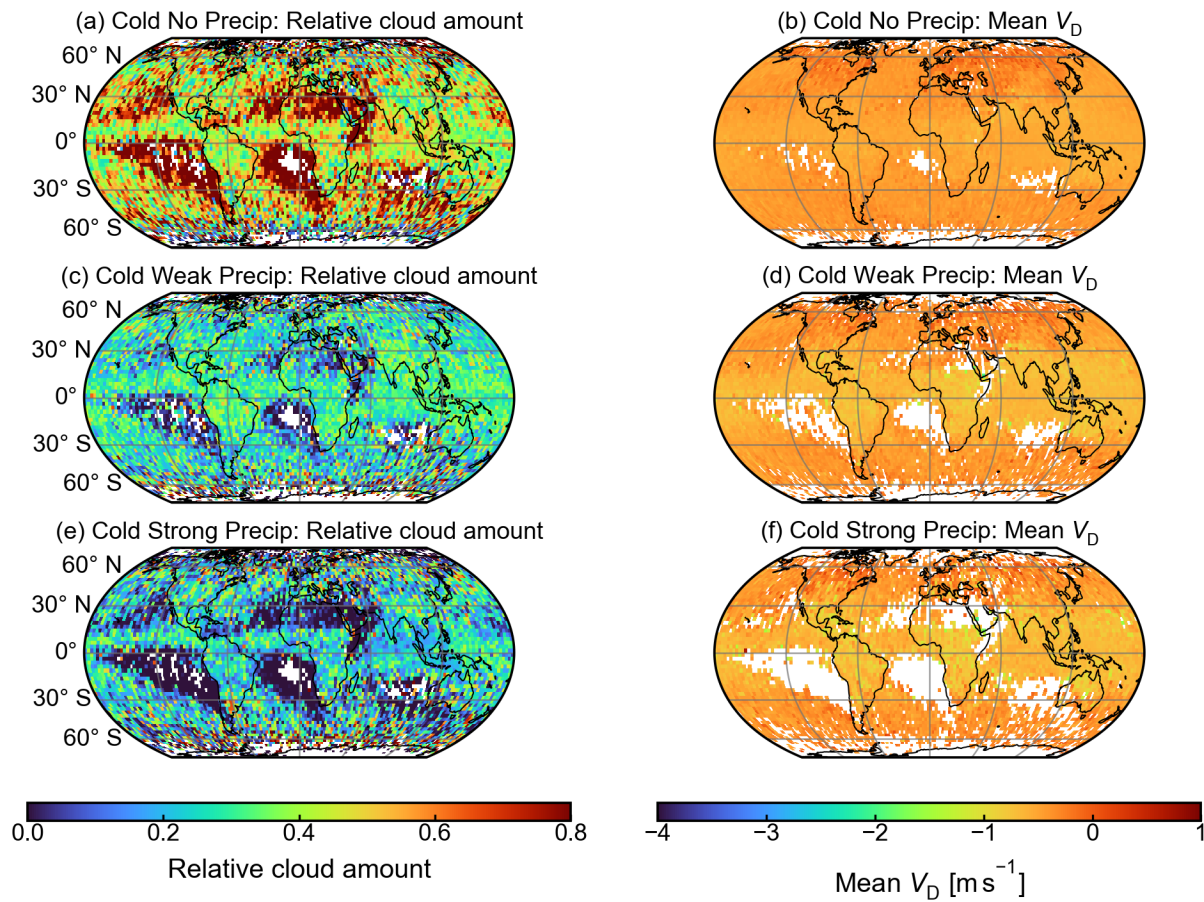
432 Figs. 8, S2, and 9 illustrate the cloud-type decomposition of the horizontal patterns of $\overline{V_D}$ depicted in Fig. 7 as described in Eq.
 433 (2). The left column and right columns of these figures present the relative cloud amount (f_k) and cloud-type-conditioned mean
 434 Doppler velocity ($\overline{V_D}^{(k)}$), respectively. This approach elucidates the spatial distributions of cloud types and which type-specific
 435 $\overline{V_D}^{(k)}$ contribute to the climatological map presented in Fig. 7.

436 Fig. 8 (upper level; 200–300 hPa) indicates that the upper-tropospheric cloud population consists almost exclusively of the
 437 three cold types. Among these, Cold No Precip dominated over extensive regions, although its relative contribution decreased
 438 in tropical convective regions. In contrast, strong cold precipitation exhibits a high relative cloud amount, specifically in these
 439 convective regions (Schumacher and Houze, 2003; Houze, 2004). Cold weak precursors are more broadly distributed and
 440 exhibit less spatial localization.



441 The $\overline{V_D}^{(k)}$ maps (Fig. 8 right column) indicate that all three cold categories generally exhibit weak downward $\overline{V_D}^{(k)}$ across
 442 extensive regions, generally ranging from -1 to 0 m s^{-1} . Across all three cold categories, $\overline{V_D}^{(k)}$ decreased slightly faster over
 443 tropical convective regions (20°S – 20°N), thereby implying contributions from larger or denser ice particles associated with
 444 slightly warmer temperatures. (Houze, 2014; Schumacher and Houze, 2003; Heymsfield and Westbrook, 2010). Overall, the
 445 upper-tropospheric V_D demonstrated limited regional variation with a modest shift toward faster downward motion in the
 446 tropics, which was consistent across all cold cloud types. Thus, the weak regional variability of the upper-tropospheric mean
 447 V_D illustrated in Fig. 7 primarily reflects the common V_D characteristics of cold clouds. The corresponding mid-tropospheric
 448 decompositions (Fig. S2) suggest that $\overline{V_D}^{(k)}$ remains broadly similar among the three cold types, thereby indicating that cloud-
 449 type dependence is still secondary to the structure of $\overline{V_D}$.

200–300 hPa (upper troposphere)



450

451 **Figure 8.** Upper-tropospheric (200–300 hPa) cloud-type decomposition of the mean V_D structure. Left column illustrates relative
 452 cloud amount (fraction of cloudy gates assigned to each type at this level), and right column presents type-conditioned mean V_D for



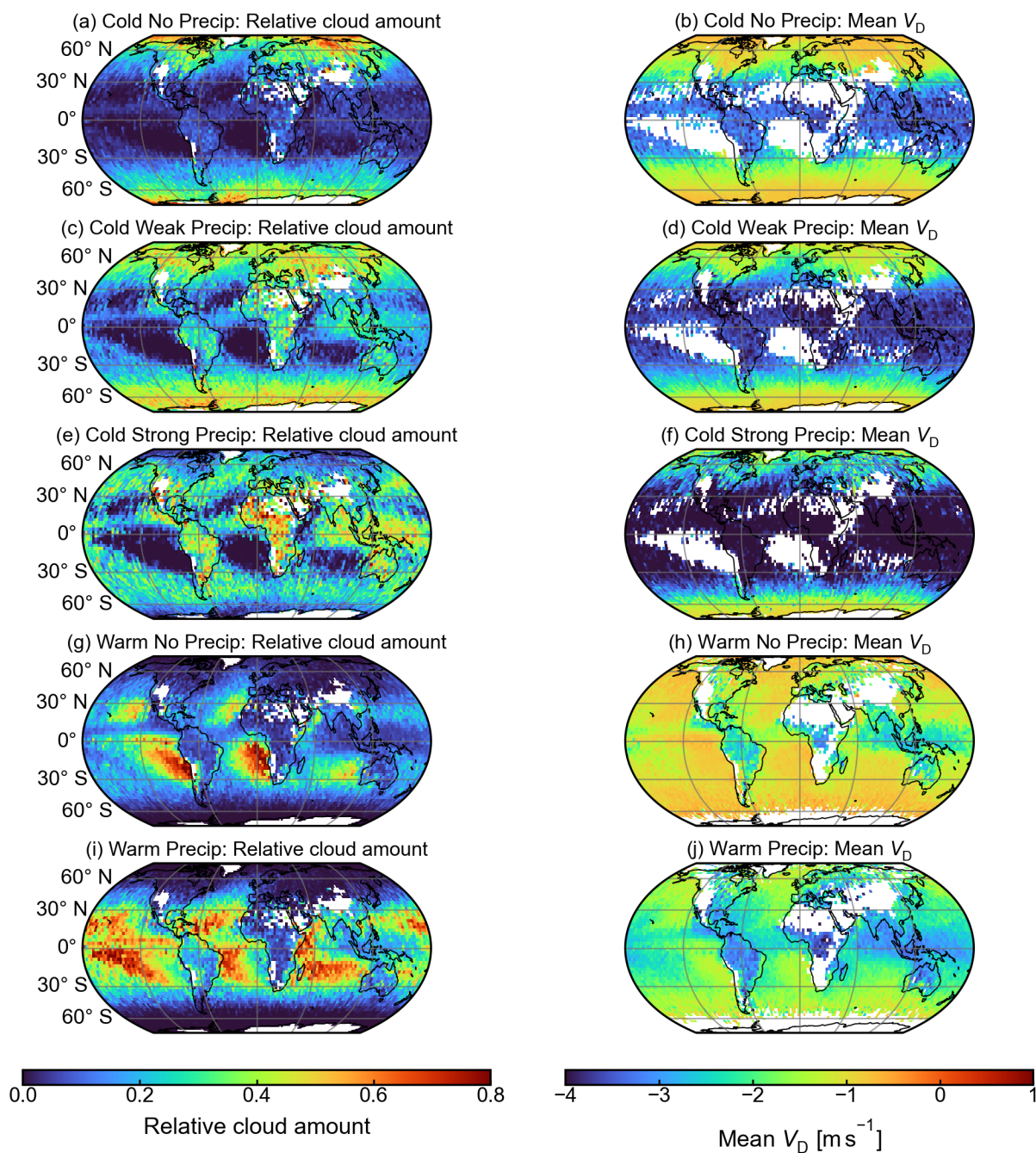
453 each type. Rows indicate the three Cold types. For relative cloud amount, grid boxes with 100 or fewer total cloudy samples were
454 masked; for type-conditioned mean V_D , grid boxes with 100 or fewer samples of the corresponding cloud type were masked.

455 Fig. 9, organized similarly to Figs. 8 and S2, presents values for the lower troposphere (800–900 hPa). The dominant type
456 differs markedly according to region. Warm No Precip exhibits substantial relative cloud amount over the subtropical eastern-
457 ocean stratocumulus regions. In contrast, Warm Precip is distributed along tropical–subtropical convergence zones, which is
458 consistent with the widespread occurrence of warm-rain and drizzle across these regions (Lebsock and L’Ecuyer, 2011; Kubar
459 et al., 2009). The Cold categories also contribute notably at these levels: Cold No Precip demonstrates increased relative cloud
460 amount in the high latitudes (poleward of 40°), while Cold Strong Precip is more prevalent in tropical (20°S–20°N) convective
461 regions.

462 Unlike the upper and middle troposphere, cloud-type differences in $\overline{V_D}^{(k)}$ were pronounced at lower levels, which is consistent
463 with the substantial inter-type variability in the warm regime (280–300 K) illustrated in Figs. 2 and 4. Cold types exhibited
464 faster downward V_D than Warm types, and the downward V_D became progressively faster for more strongly precipitating
465 categories. Further, Warm No Precip V_D remained close to zero, which indicated a weak net downward motion across broad
466 regions, whereas Warm Precip exhibited a faster downward V_D . Notably, within the Warm No Precip and Warm Precip types,
467 V_D over land is systematically more downward than over oceans, and considerable variability is evident among oceanic regions.
468 This residual regional dependence is not discussed in detail here, but possible contributors include differences in cloud depth,
469 dynamic environment, and aerosol-related warm rain processes (Takahashi et al. 2017; Albrecht, 1989; Rosenfeld et al., 2008).
470 These patterns explain much of the lower tropospheric structure illustrated in Figure 7. The rapid downward mean V_D in
471 tropical convective regions aligns primarily with the dominance of cold, strong, and warm precipitation. Meanwhile, regions
472 with a weak downward V_D over the subtropical eastern ocean stratocumulus regions correspond to the dominance of warm,
473 no-precipitation. Overall, these results indicate that the lower-tropospheric distribution of the mean V_D is governed by both
474 cloud-type and cloud-type-conditioned mean $\overline{V_D}^{(k)}$.



800–900 hPa (lower troposphere)



475

476 **Figure 9.** Same as Fig. 8 but for the lower-tropospheric pressure range (800–900 hPa).

477 The preceding discussion on the upper, middle, and lower atmospheric pressure ranges examined their region-specific
478 characteristics. Regional contrasts in mean V_D were categorized into differences in: (i) cloud-type occurrence and (ii) cloud-



type-conditioned V_D . In tropical convective regions, cold strong precipitation increases locally in the upper–mid troposphere, while warm and cold strong precipitation tends to dominate at low levels. The mean V_D is more likely to shift toward faster downward velocities, which is consistent with a larger $V_{t,z}$ in the precipitating categories (Schumacher and Houze, 2003; Stephens et al., 2008; Houze, 2014). In contrast, clouds over the subtropical eastern ocean stratocumulus decks are largely confined to the lower level, and Warm No Precip dominates. Consequently, mean V_D remains weak downward motion close to zero. In the mid-latitude storm-track regime, Cold No Precip and Cold Weak Precip prevail over broad areas, and mid- and low-level mean V_D tend to demonstrate intermediate downward values, which are slower than those in the tropics. Therefore, the analysis suggests that the climatology of mean V_D should be interpreted as a superposition of cloud-type occurrence and type-conditioned V_D rather than as a single-regime property at each altitude.

4.3 Occurrence of upward Doppler velocities

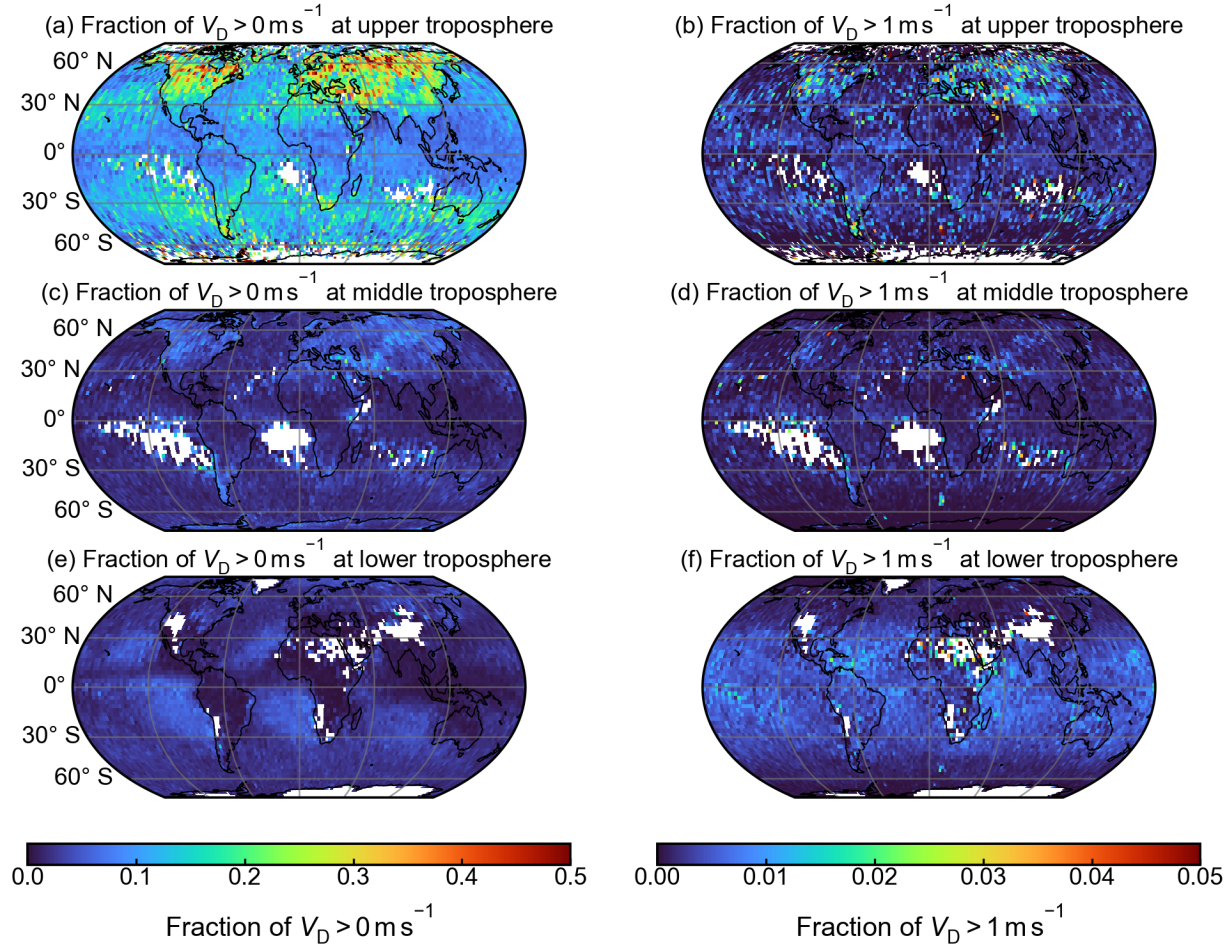
Figs. 5–9 present the regional distribution of mean Doppler velocity (V_D). However, regions with similar mean values may have markedly different upward tails in their V_D distributions. Even when the spatial or temporal mean V_D indicates downward motion, fluctuations in in-cloud vertical air motion can lead to a considerable number of positive V_D samples, which is closely linked to dynamically active updrafts (Kollias et al., 2003; Shupe et al., 2008; Protat and Williams, 2011; Kalesse and Kollias, 2013). We define two threshold-exceedance metrics to quantify the frequency of upward signatures: (i) $F_{V_D>0}$, the fraction of in-cloud samples with $V_D > 0 \text{ m s}^{-1}$, and (ii) $F_{V_D>1}$, the fraction with $V_D > 1 \text{ m s}^{-1}$. Given that $V_D = w - V_{t,z}$ (Eq. 1), the $F_{V_D>0}$ quantifies how often upward air motion exceeds the sedimentation contribution ($w > V_{t,z}$), and is sensitive to both the mean position and spread of the V_D distribution. Meanwhile, $F_{V_D>1}$ provides a more selective indicator of strongly upward Doppler-velocity signatures. Notably, V_D may still include measurement noise and residual, scene-dependent biases; therefore, some threshold exceedances, particularly over land, may reflect residual measurement biases rather than atmospheric dynamics (JAXA 2026a).

Fig. 10 illustrates these metrics for the upper, middle, and lower tropospheres. Some conspicuously high fractions over the Northern Hemisphere, particularly in the upper-tropospheric maps, may have resulted from measurement bias. For the $F_{V_D>0}$ (Fig. 10, left column), the upper troposphere displayed higher values in mid- to high-latitude regions and lower values in the tropics. This pattern corresponds to a slightly faster downward mean V_D in the tropics (Fig. 7). In the middle troposphere, the regional maxima were less pronounced, and the overall occurrence was suppressed. In the lower troposphere, the $F_{V_D>0}$ increased in regions where the mean V_D indicated only a weak downward motion (Fig. 7), particularly over the eastern subtropical ocean basins. These enhancements are not accompanied by similarly high $F_{V_D>1}$ (Fig. 10, right column), and thus, do not necessarily indicate strong updrafts but can result from modest variability around a weakly downward mean.



509 Further, $F_{V_D>1}$ is much less frequent than $F_{V_D>0}$ but is preferentially enhanced over tropical convective regions (Fig. 10, right
510 column). This pattern is qualitatively consistent with previous EarthCARE Doppler studies, which exhibited greater V_D and
511 vertical-air-motion variability in deep convective than in stratiform precipitation and frequent strong upward V_D signatures in
512 major tropical convective regions (Aoki et al., 2026b; Hotta et al., 2026). These enhancements were also evident in the tropical
513 lower troposphere despite the rapidly decreasing mean V_D (Fig. 7). This coexistence indicated broad V_D distributions with
514 intermittent and strong upward excursions.

515 Overall, Fig. 10 illustrates that $F_{V_D>0}$ and $F_{V_D>1}$ emphasize different aspects of the V_D distribution. Namely, $F_{V_D>0}$ primarily
516 reflects how readily the distribution crosses zero, whereas $F_{V_D>1}$ selectively identifies its strong upward tail, which is primarily
517 linked with tropical convective systems.



518
519 **Figure 10.** Horizontal distributions of the fractions of samples with $V_D > 0 \text{ ms}^{-1}$ ($F_{V_D>0}$; left) and $V_D > 1 \text{ ms}^{-1}$ ($F_{V_D>1}$; right). Rows
520 display the upper (200–300 hPa), middle (450–550 hPa), and lower (800–900 hPa) troposphere. Fractions were calculated in $2.5^\circ \times$
521 2.5° grid boxes after aggregating samples within each pressure range. Grid boxes containing 100 or fewer valid samples were masked.
522

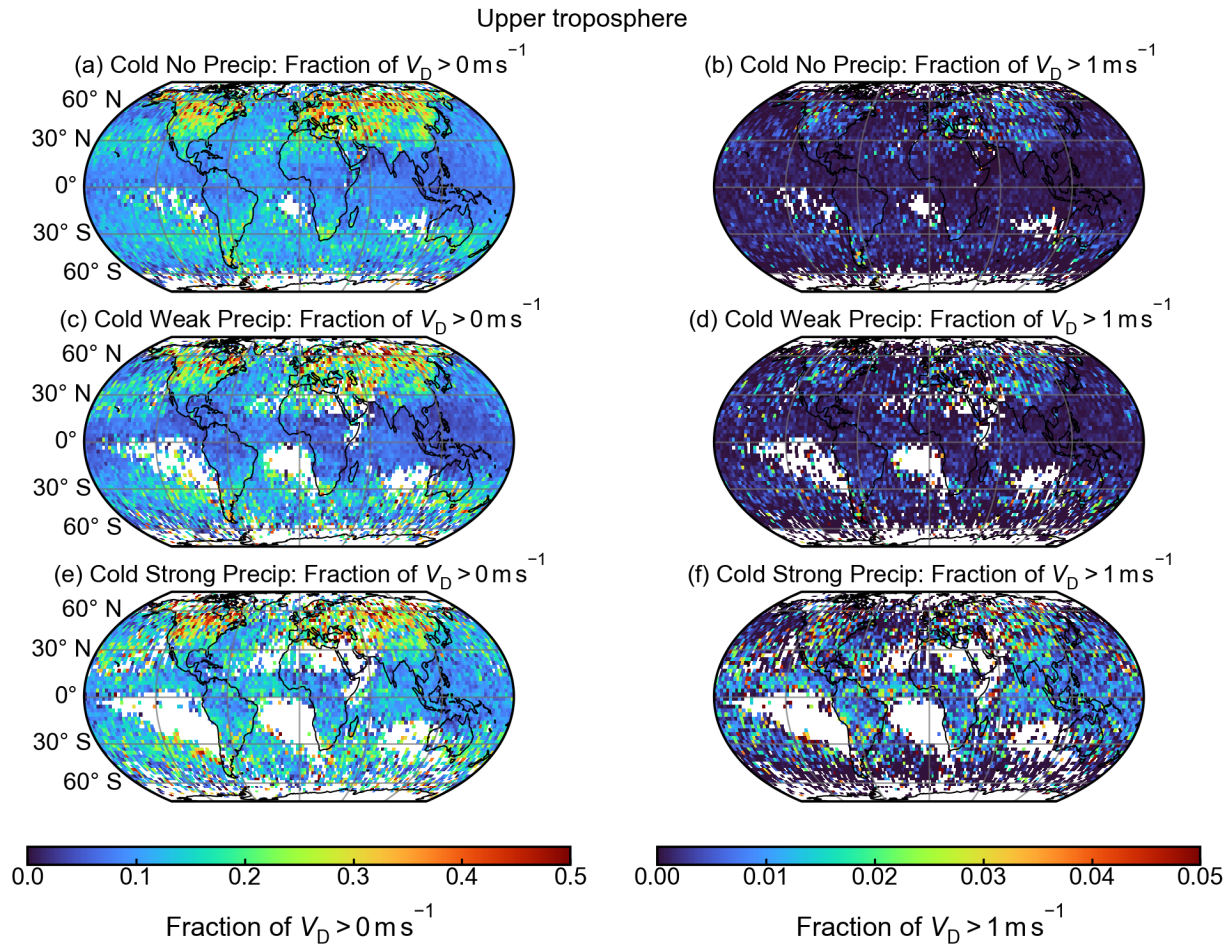


523 To clarify the cloud-type contributions to the spatial patterns in Fig. 10, Fig. 11 presents the cloud-type decompositions of
524 $F_{V_D>0}$ and $F_{V_D>1}$ in the upper troposphere (200–300 hPa). Since Warm-types are rare at these levels, only three Cold-types are
525 displayed. Additionally, the relative occurrence of the three Cold types illustrated in Fig. 8 should be considered when
526 interpreting their contributions to all-cloud frequencies, as provided in Fig. 10.

527 $F_{V_D>0}$ was non-negligible for all three Cold types. Elevated $F_{V_D>0}$ over Northern Hemisphere land is common to all three Cold
528 types and is likely to be influenced by residual V_D errors. Systematic contrasts among types are observed over tropical
529 convective regions (20°S–20°N), and the Cold Weak Precip typically exhibits a smaller $F_{V_D>0}$, whereas the Cold Strong Precip
530 exhibits a larger one. These cloud-type differences in $F_{V_D>0}$ can be interpreted in terms of the mean position and width of the
531 V_D distributions. The reduced $F_{V_D>0}$ in the Cold Weak Precip was consistent with a faster mean downward motion and limited
532 spread toward upward velocities, whereas the enhanced $F_{V_D>0}$ in the Cold Strong Precip reflected its broader upward tail, as
533 illustrated in Fig. 4.

534 Furthermore, $F_{V_D>1}$ is much smaller in absolute frequency but is preferentially enhanced for Cold Strong Precip in tropical
535 regions. As the mean V_D of the Cold Strong Precip is more downward in tropical regions than in other regions (Fig. 8), the
536 enhanced $F_{V_D>1}$ cannot be explained by an upward shift of the mean and instead reflects the broader upward tail of the V_D
537 distribution. Coupled with the enhanced tropical occurrence of Cold Strong Precip (Fig. 8), this largely explains the tropical
538 $F_{V_D>1}$ signal in Fig. 10. Similarly, mid-tropospheric maps (Fig. S3) indicate that $F_{V_D>1}$ was preferentially enhanced in the
539 tropical Cold Strong Precip.

540 The relatively broad tropical distribution of enhanced $F_{V_D>1}$ within the Cold Strong Precip may partly reflect the exclusion of
541 folded profiles from the present statistics. Hotta et al. (2026) demonstrated that folded profiles were concentrated over Central
542 Africa and the Amazon, thereby suggesting that the present $F_{V_D>1}$ statistic may omit a spatially localized subset of the most
543 intense convection.



544
545 **Figure 11. Cloud-type-resolved upward Doppler-velocity occurrence in the upper troposphere (200–300 hPa). Left column presents**
546 **the fractions of samples with $V_D > 0 \text{ m s}^{-1}$ ($F_{V_D>0}$) and right column exhibits the fractions of samples with $V_D > 1 \text{ m s}^{-1}$ ($F_{V_D>1}$);**
547 **rows correspond to cloud types. Maps were calculated on $2.5^\circ \times 2.5^\circ$ grid. For each cloud type, grid boxes containing 100 or fewer**
548 **samples of that type were masked.**

549
550 Fig. 12 illustrates the cloud-type-resolved $F_{V_D>0}$ and $F_{V_D>1}$ in the lower troposphere (800–900 hPa). Since cloud-type
551 occurrences vary strongly by region (Fig. 9), these type-conditioned frequencies should be interpreted together with the
552 corresponding relative cloud amounts.

553 $F_{V_D>0}$ is strongly suppressed for Warm Precip and all three Cold types, which is consistent with their fast downward mean V_D
554 under warm-temperature conditions in Figs. 2, 4, and 9. Therefore, their V_D distributions shifted well below the $V_D = 0$
555 threshold, consequently reducing the upward exceedances. In contrast, Warm No Precip exhibits a substantially higher $F_{V_D>0}$
556 because its mean V_D is closer to zero (Fig. 2, 4, and 9), and a modest variability in vertical air motion can shift the Doppler



557 velocity across the $V_D = 0$ threshold. Combined with its widespread occurrence over subtropical marine low-cloud regions
558 (Fig. 9), this largely explains the broad lower-tropospheric $F_{V_D>0}$ enhancement presented in Fig. 10.

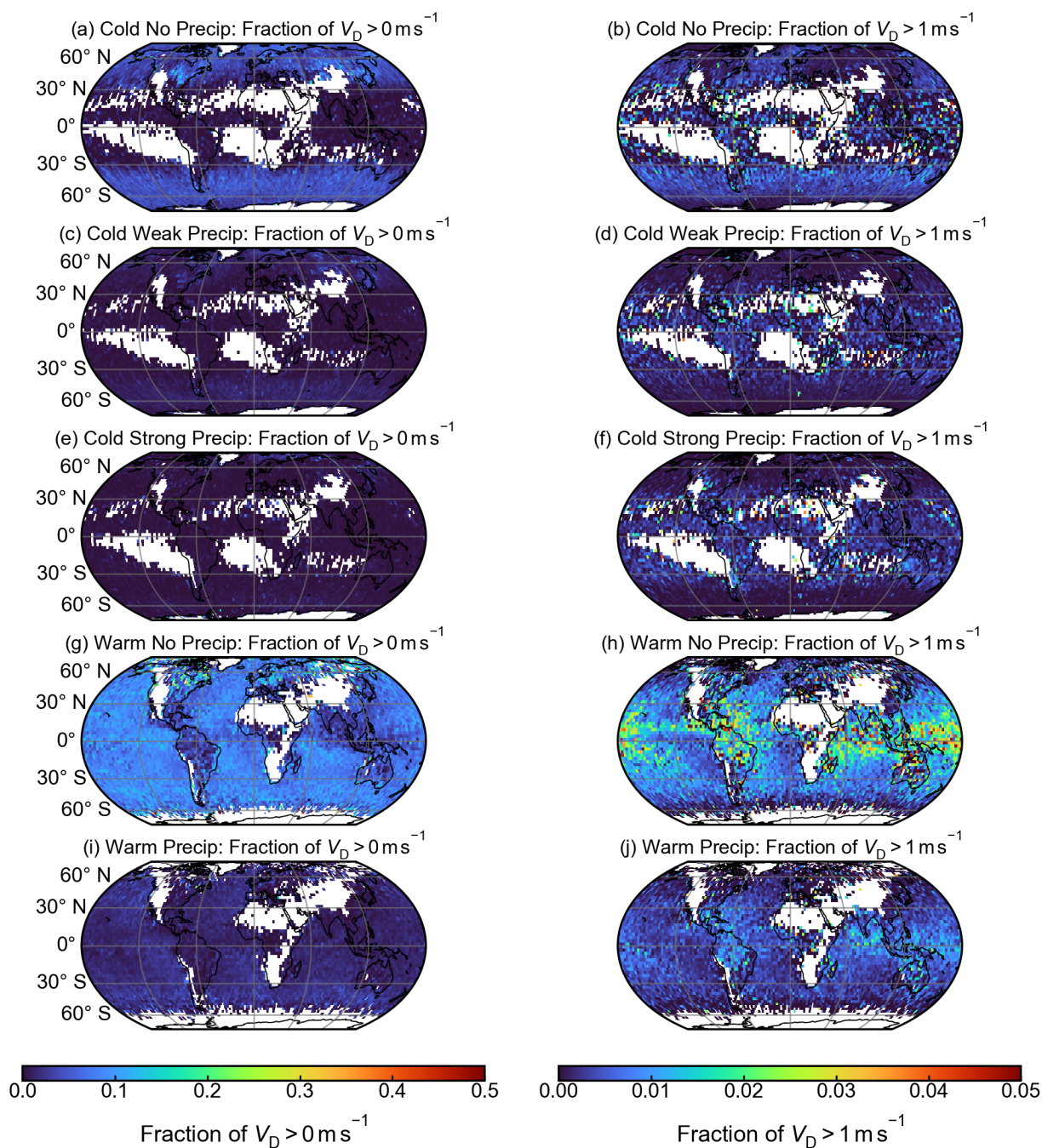
559 Moreover, $F_{V_D>1}$ was much less frequent than $F_{V_D>0}$ for all the cloud types. Nevertheless, Warm No Precip displayed the
560 highest $F_{V_D>1}$, but its spatial pattern is more localized than that of $F_{V_D>0}$, particularly near the tropical convergence zones.

561 This contrast indicates that the two thresholds emphasize different portions of the Warm No Precip V_D distribution. The broad
562 $F_{V_D>0}$ enhancement chiefly reflects clouds whose V_D values fluctuate around a mean close to zero, whereas the localized $F_{V_D>1}$
563 maxima selectively highlight the upward tail of the lower-tropospheric V_D distribution, thereby capturing rarer and stronger
564 upward excursions.

565



Lower troposphere



566

567 **Figure 12.** Same as Figs. 11 and S3 but for the lower-tropospheric pressure range (800–900 hPa).



5 Summary and implications

This study established a global climatology of cloud-radar Doppler velocity (V_D) using quality-controlled EarthCARE/CPR observations and examined its dependence on temperature, radar reflectivity (Z_e), cloud type, and geographical region. We considered V_D to be a directly observed radar quantity and interpreted it using five cloud types, defined according to cloud-top temperatures and reflectivity-based precipitation indicators. This framework allows the interpretation of all-cloud Doppler climatology as a mixture of cloud populations with different phases and precipitation characteristics.

V_D predominantly decreased over most temperature ranges and cloud regimes, which indicated a foremost contribution from hydrometeor sedimentation (Fig. 1, 3, 5, and 7). In the cold regime (200–260 K), V_D generally indicates a slow downward motion, which is consistent with the relatively low terminal fall speeds of most ice particles (Pruppacher and Klett, 2010; Heymsfield and Westbrook, 2010). V_D shifts toward a faster downward motion at warmer temperatures. In the warm regime (280–300 K), V_D spans a much wider range toward a faster downward motion. This behavior is consistent with the substantially higher terminal fall speeds of raindrops than those of most ice particles. Thus, the temperature dependence of V_D illustrates that cloud microphysical transitions are manifested in the Doppler velocity climatology.

Further, cloud-type conditioning revealed systematic V_D differences that were not explained by Z_e alone (Figs. 2–4). At a comparable observed Z_e in the warm regime, Warm Precip exhibited faster downward V_D than Warm No Precip, whereas cold-origin precipitating clouds still exhibited faster downward V_D , particularly Cold Strong Precip. This ordering aligns with contrasting precipitation pathways that can produce different drop size distributions below the melting level: Warm Precip is primarily linked to liquid-phase collision–coalescence, whereas cold-origin precipitation can undergo aggregation and riming aloft before melting into raindrops (Wang and Geerts, 2003; Suzuki et al., 2010; Mroz et al., 2023). Therefore, the broad all-cloud distribution in the warm regime largely reflects the superposition of cloud populations. In the cold regime, the median V_D differs only weakly among the three cold types at a given Z_e ; however, Cold Strong Precip had the broadest distribution, and extended toward both faster downward and upward motions. These broad tails are consistent with greater variability in ice particle fall speeds and vertical air motion. Collectively, these contrasts demonstrate that V_D provides information on the hydrometeor fall speed structure and precipitation pathways beyond those available from Z_e alone.

Regional mean V_D reflects both cloud-type occurrence and cloud-type-conditioned V_D (Figs. 5–9). Tropical convective regions exhibit a fast downward mean V_D because precipitation types are frequent and have large downward sedimentation contributions. Subtropical eastern-ocean stratocumulus regions exhibit a mean V_D close to zero because Warm No Precip dominates and has slower downward Doppler motion, whereas mid-latitude storm-track regions display intermediate values largely associated with Cold No Precip and Cold Weak Precip. Therefore, the regional all-cloud mean V_D reflected both the relative occurrence of different cloud types and their type-specific V_D .

The threshold exceedance metrics $F_{V_D>0}$ and $F_{V_D>1}$, defined as the fractions of samples with $V_D > 0 \text{ m s}^{-1}$ and $V_D > 1 \text{ m s}^{-1}$, respectively, provided complementary information regarding the upward tail of the V_D distribution (Figs. 10–12). $F_{V_D>0}$ is



strongly influenced by the position of the distribution relative to zero, and can therefore be enhanced when sedimentation is weak. In contrast, $F_{V_D>1}$ was less frequent and more selectively highlighted the strong upward particle motion signatures. Its enhancement in tropical convective regions, especially for Cold Strong Precip, occurs even where the mean V_D decreases faster than elsewhere, thereby indicating a broad distribution with intermittent upward excursions rather than an upward shift of the mean. Therefore, the mean V_D and $F_{V_D>1}$ characterize distinct aspects of cloud regimes: hydrometeor sedimentation characteristics and intermittent strong upward transport associated with convective dynamics (Protat and Williams, 2011; Kalesse and Kollias, 2013; Aoki et al., 2026b; Hotta et al., 2026). Importantly, these diagnostics provide a cloud-regime-resolved benchmark for the model evaluation. Slow downward V_D in upper-tropospheric cold clouds, the shift from Warm No Precip to faster downward Warm Precip, and the separation between warm- and cold-origin precipitation at comparable Z_e can validate model representations of ice and snow sedimentation, warm rain formation, and precipitation pathways, respectively. The broad V_D distributions in the Cold Strong Precip can additionally test the combined representation of precipitating ice microphysics and vertical-motion variability. Model comparisons should use forward-simulated radar observables that account for the CPR sensitivity, W-band attenuation, sampling, and reflectivity-weighted nature of V_D (Kollias et al. 2014; Sy et al. 2014; Kollias et al. 2022; Roh et al. 2026; Nakamura et al. 2026). Additionally, cloud-type conditioning is essential for separating biases in cloud regime occurrences from those in within-regime processes. Despite its strengths, this study also has several limitations that should be considered when interpreting the results. In the present analysis, V_D was treated as a directly observed quantity, and the contributions from vertical air motion w and hydrometeor sedimentation $V_{t,z}$ were not separated. Other limitations arise from W-band attenuation, multiple scattering, non-Rayleigh scattering, and residual Doppler velocity biases. Future studies should extend this analysis in several directions. First, a longer EarthCARE record and continued validation of the JAXA Level-2 products (JAXA, 2026a) against ground-based and airborne observations are expected to facilitate the explicit analysis of w and $V_{t,z}$, thereby enabling the quantification of dynamical and microphysical contributions to V_D . Second, applying similarly conditioned V_D diagnostics to the forward-simulated model can distinguish between biases in cloud occurrence, sedimentation, precipitation formation, and vertical-motion variability. Collectively, these efforts will strengthen spaceborne Doppler climatology as a process-oriented benchmark whose principal value lies in interpreting V_D jointly with Z_e and cloud types.

Data availability

The EarthCARE/CPR L2A CPR one-sensor Cloud Products (<https://doi.org/10.57746/EO.01jdvd2gqq34e6yz9p8kfe68x5>, JAXA,) used in this study can be downloaded from the JAXA G-Portal (<https://gportal.jaxa.jp/gpr/>).



631 **Author contributions**

632 HH conceptualized the study, performed the formal analysis, developed the methodology, and prepared the visualizations. KS
633 conducted the investigation and was responsible for funding acquisition. MK and TK contributed to the conceptualization of
634 the study. MA, SA, and TK contributed to the interpretation of the results. KS and TK supervised the study. HH wrote the
635 original draft of the manuscript. KS, MK, SA, and TK contributed to the review and editing of the manuscript.

636 **Competing interests**

637 At least one of the (co-)authors is a guest member of the editorial board of *Atmospheric Chemistry and Physics* for the special
638 issue “Early results from EarthCARE (AMT/ACP/GMD inter-journal SI)”. The authors have no other competing interests to
639 declare.

640 **Special issue statement**

641 This article is part of the special issue “Early results from EarthCARE (AMT/ACP/GMD inter-journal SI)”. It is not associated
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