

1 **Characterizing precipitation and improving rainfall estimates over the Southern**
2 **Ocean using ship-borne disdrometer and dual-polarimetric C-band radar**

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15 **Key Points:**

- 16 • Different synoptic types across the Southern Ocean exhibit distinctive polarimetric
17 signatures and surface precipitation properties.
- 18 • Small raindrops of less than one millimeter contribute up to 47% of total accumulation
19 during the Austral warm seasons over the region.
- 20 • A new formulation for radar rainfall estimates that reflects the large numbers of small
21 drops over the Southern Ocean is proposed.

22 **Abstract**

23 Large satellite discrepancies and model biases in representing precipitation over the Southern
24 Ocean (SO) are related directly to the region's limited surface observations of precipitation. To
25 help address this knowledge gap, the study investigated the precipitation characteristics and rain
26 rate retrievals over the remote SO using ship-borne data of the Ocean Rainfall And Ice-phase
27 precipitation measurement Network disdrometer (OceanRAIN) and dual-polarimetric C-band
28 radar (OceanPOL) aboard the Research Vessel (RV) Investigator in the Austral warm seasons of
29 2016 to 2018. Seven distinct synoptic types over the SO were analyzed based on their radar
30 polarimetric signatures, surface precipitation phase, and rain microphysical properties.
31 OceanRAIN observations revealed that the SO precipitation was dominated by drizzle and light
32 rain, with small-sized raindrops (diameter < 1 mm) constituting up to 47 % of total
33 accumulation. Precipitation occurred most frequently over the warm sector of extratropical
34 cyclones, while concentrations of large-sized raindrops (diameter > 3 mm) were prominent over
35 synoptic types with colder and more convectively unstable environments. OceanPOL
36 observations complement and extend the surface precipitation properties sampled by
37 OceanRAIN, providing unique information to help characterize the variety of potential
38 precipitation types and associated mechanisms under different synoptic conditions. Raindrop size
39 distributions (DSD) measured with OceanRAIN over the SO were better characterized by
40 analytical DSD forms with two-shape parameters than single-shape parameters currently
41 implemented in satellite retrieval algorithms. This study also revised a rainfall retrieval algorithm
42 for C-band radars to reflect the large amount of small drops and provide improved radar rainfall
43 estimates over the SO.

44 **Plain Language Summary**

45 Precipitation is a major component of the hydrologic cycle in high-latitude regions including the
46 remote Southern Ocean (SO). However, large differences continue to exist among current
47 precipitation products in the region, owing in part to the absence of high-quality surface
48 observational records suitable for evaluation across a range of temporal and spatial scales. This
49 work uses two instruments aboard the RV Investigator over the Australian sector of the SO in the
50 Austral warm seasons of 2016 to 2018: the OceanRAIN disdrometer and OceanPOL radar. We
51 focused our analysis on seven distinct synoptic conditions over the SO and found the variability
52 in their radar features and surface precipitation properties. This work also discussed two
53 important findings related to remote sensing retrievals of SO rain. First, we demonstrated why
54 the rainfall retrieval assumptions in satellite algorithms may need to be refined to account for the
55 unique rainfall properties in the SO. Second, we formulated a new set of equations suitable for
56 shipborne C-band radars in improving rain rate estimates over the region. This work leads
57 toward more accurate, high-resolution estimates of precipitation over the measurement-sparse
58 SO to better understand a range of climatological and meteorological processes in the region.

59 **1 Introduction**

60 Clouds and precipitation over the Southern Ocean (SO) play a critical role in influencing
61 freshwater fluxes, air-sea fluxes, and radiative properties of the region (Wood, 2012; Caldeira &
62 Duffy, 2000; Siems et al., 2022; Pauling et al. 2016). The SO is characterized by fewer land
63 masses and anthropogenic aerosol sources than the Northern Hemisphere, creating a more
64 pristine environment and a distinct mix of cloud and precipitation processes. Climate models
65 continue to have large uncertainties in the cloud forcing over the SO, including their inability to

66 reproduce the correct cloud phase, supercooled liquid cloud opacity, and cold cloud processes in
67 the region (Cesana et al., 2022). These large uncertainties and biases have limited the ability of
68 the models to represent important local climate features and their teleconnections such as surface
69 warming, storm activity, and precipitation patterns (Ceppi et al., 2014; McFarquhar et al., 2021;
70 Vergara-Temprado et al., 2018).

71 Our current knowledge of precipitation over the SO is primarily derived from surface
72 measurements from island sites, satellite remote sensing observations, and reanalysis products
73 (Siems et al., 2022). However, these precipitation products have notable limitations. Historical
74 precipitation records across the SO are rare due to the sparsity of island sites. Orographic effects
75 may have also strongly influenced these records (Lewis et al., 2018; Manton et al., 2020; Siems
76 et al., 2022), limiting their ability to represent precipitation characteristics over the vast open
77 oceans. Indirect measurements of cloud and precipitation from satellite-based products lack
78 calibration for the Southern Hemisphere, which contributes to the large discrepancies among
79 satellite precipitation estimates over the region (Skofronick-Jackson et al., 2017; McFarquhar et
80 al., 2021). Precipitation estimates from reanalysis products strongly depend on model
81 parameterizations and are at scales that do not resolve key processes and, therefore, potentially
82 inherit the climate model biases over the region (Lang et al., 2018; Naud et al., 2014). Further,
83 the reanalysis and satellite estimates do not agree with each other and have large observed errors
84 (Montoya Duque et al., 2023).

85 In recent years, several observational programs have taken place to address the
86 longstanding knowledge gaps in the nature and variability of precipitation over the SO, including
87 their interaction with other climate components (e.g., McFarquhar et al. 2021). Ship-borne field
88 campaigns, in particular, have provided better spatiotemporal sampling necessary for evaluating
89 satellite-based and model precipitation products over the region. From 2016 to 2018, the
90 Australian Research Vessel (RV) Investigator conducted multiple scientific voyages over the
91 Australian Sector of the SO, enabling comprehensive surface and remote sensing measurements.
92 Routine observations from the RV Investigator along with coordinated field campaigns have led
93 to the recent understanding of the distinct microphysical characteristics of SO precipitation
94 compared with other latitudes (Protat et al., 2019a, 2019b), their variability across different
95 synoptic environments (Lang et al., 2021; Montoya Duque et al., 2022), and case studies of
96 shallow convection and frontal systems over the high-latitude SO (Mace et al., 2023).
97 Observations and analysis are still needed to further understand how the precipitation properties,
98 including rain rate, vary across synoptic types in the SO.

99 The RV Investigator carries a dual-polarization C-band (5.5 GHz) weather radar called
100 OceanPOL, one of only three ship-borne dual-polarization weather radars in the world and the
101 only one operating over the SO (Protat et al., 2022). OceanPOL provides high-resolution 3-D
102 measurements of precipitation at multiple elevation angles, and its dual-polarization capability
103 enables improved retrievals of hydrometeor species and spatial distributions. Its volumetric scans
104 allow wide coverage of precipitation-size particles that can subsequently reach the surface and
105 are complementary to the profile measurements from the vertically-pointing W-band (95 GHz)
106 cloud radar that was also deployed on the RV Investigator for some cruises (e.g., Lang et al.,
107 2021; Mace & Protat, 2018a & 2018b; Montoya Duque et al., 2022). The RV Investigator also
108 carries the Ocean Rainfall And Ice-phase precipitation measurement Network (OceanRAIN), an
109 optical disdrometer that samples the particle size distribution of precipitation along the ship track

110 (Klepp, 2015; Klepp et al., 2018) and can be used to improve the rainfall estimates of
111 OceanPOL.

112 This study aims to investigate the nature of precipitation and associated properties under
113 various synoptic conditions over the SO. We aim to address the following research questions
114 using the OceanRAIN and OceanPOL data from seven field cruises of the RV Investigator:

115 (1) What are the key precipitation characteristics over the SO, and how do they vary
116 under different synoptic conditions?

117 (2) Are the commonly applied analytical forms of rain drop size distribution (DSD) able
118 to accurately represent the observed DSD over the SO?

119 (3) Can the rainfall properties simulated from OceanRAIN observations be used to
120 improve the rainfall estimates of OceanPOL?

121 The remainder of the paper is structured as follows: Section 2 provides information about
122 the two instruments and the synoptic type classification. Section 3 provides a sample case of a
123 precipitation event associated with an extratropical cyclone, the bulk analysis of precipitation
124 characteristics, and optimization of the rain rate retrieval algorithms. Finally, Section 4 provides
125 discussion and conclusions.

126

127 **2 Materials and Methods**

128 2.1. OceanRAIN

129 OceanRAIN's primary instrument is the ODM470 disdrometer, which counts and sorts
130 precipitation particles into 128 logarithmically distributed size bins from 0.04 to 22.28 mm at
131 one-minute resolution (Klepp, 2015; Klepp et al., 2018). The design of the disdrometer
132 minimizes the impact of artificial small droplets due to splashing, while its algorithm resolves
133 edge effects, coincidence effects from overlapping particles, and precipitation fall velocities. As
134 an initial quality control, the OceanRAIN algorithm automatically removes data from size bins
135 below 0.39 mm, since these smaller droplets are often contaminated with artificial signals from
136 gusty winds and ship propulsion. OceanRAIN identifies the thermodynamic phase of
137 precipitation (liquid, solid, or mixed) following Burdanowitz et al. (2016), which is reported to
138 be more reliable in detecting rain than mixed-phase precipitation and rain-snow transitions at
139 ambient temperatures of -3 and 6 °C.

140 2.1.1. Pre-processing and quality control of OceanRAIN data

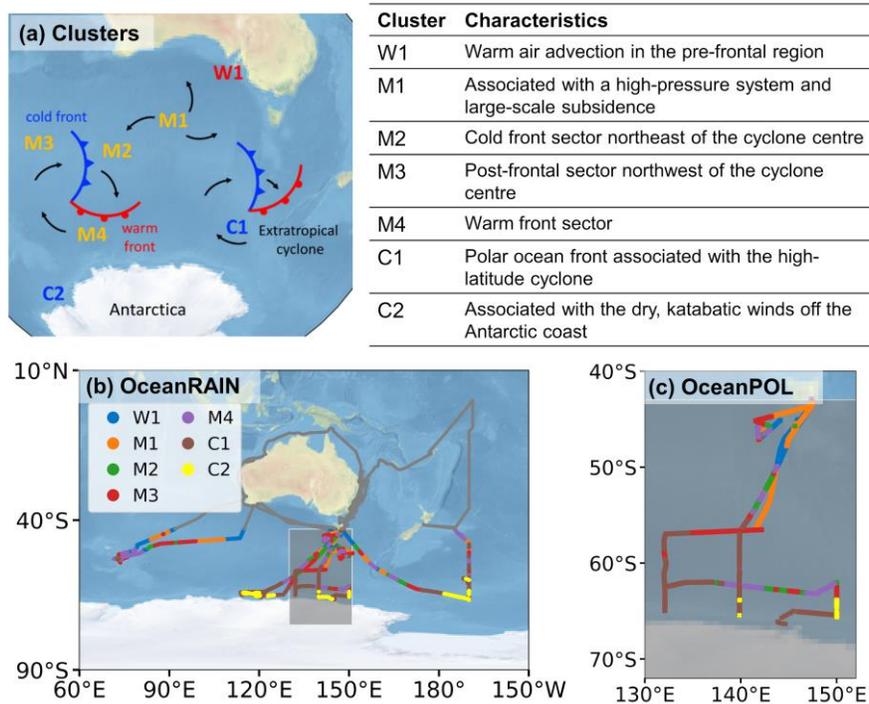
141 We used OceanRAIN data from seven voyages of the RV Investigator south of 43 °S in
142 the Austral warm seasons of 2016 to 2018 (Figure 1b and Table S1). Rain, mixed-phase, and ice
143 precipitation samples were used to analyze the surface precipitation frequency and
144 thermodynamic phase under different synoptic conditions. Rain samples were used to examine
145 the observed DSD and improve the rain rate retrieval algorithm of OceanPOL over the SO.

146 Recent research using disdrometer observations over Macquarie Island (54.5°S, 158.9°E)
147 showed that small raindrops, less than 1 mm, were significant and contributed ~10 % of the total
148 annual precipitation over the island (Tansey et al., 2022). This finding opens up questions on
149 whether and to what extent the small-sized raindrops vary under different synoptic conditions

150 over the broader SO, and if the widely-used analytical forms of DSD can reasonably capture the
 151 variability in the observed DSD.

152 A quality control procedure for rain samples was implemented for this analysis. First, we
 153 remove rain samples with diameters >8 mm since these samples are likely ice-contaminated or
 154 have misclassified precipitation phase, considering that the maximum size of a raindrop is
 155 typically around 8 mm (Blanchard & Spencer, 1970; Hobbs & Rangno, 2004). We retained the
 156 samples with rain rates of $0.01\text{--}100$ mm h^{-1} and have at least 20 droplets distributed into a
 157 minimum of 5 size bins to produce a valid analytical DSD fit (Jaffrain & Berne, 2011; Tokay et
 158 al., 2013; Protat et al., 2019a). Altogether, the quality control procedure discarded 33.9 % of total
 159 minutes of rain observation south of 43°S , with most of these being very light rain and
 160 comprising only up to 1% of total rainfall accumulation. The number concentrations for the
 161 different diameter bins ($N(D)$; $\text{m}^{-3} \text{mm}^{-1}$) were then used to calculate the following rain
 162 microphysical variables: liquid water content (LWC; g m^{-3}), rain rate (R , mm h^{-1}), total number
 163 concentrations (N_t , m^{-3}), mass-weighted mean diameter (D_m , mm), and the generalized intercept
 164 parameter (N_w , $\text{m}^{-3} \text{mm}^{-1}$).3 Data, or a descriptive heading about data.

165



166

167 Figure 1. (a) Conceptual illustration of the seven synoptic types over the SO adopted from
 168 Truong et al. (2020) and Montoya Duque et al. (2022). Clustered samples of (b) OceanRAIN and
 169 (c) OceanPOL from the seven voyages of the RV Investigator in the Austral warm seasons of
 170 2016 to 2018 (Table S1). The gray circles in (b) show the OceanRAIN measurements north of 43°S
 171 that were discarded from the analysis.

172

173 2.1.2. Dual-polarimetric radar variables simulated from OceanRAIN data

174 Dual-polarimetric radar variables were calculated from the observed DSD of OceanRAIN
 175 using the open-source Python library ‘PyTMatrix’ (Leinonen, 2014), which is based on the T-
 176 matrix scattering method (Mishchenko et al., 1996). Previous studies with the micro-rain radar
 177 (24 GHz, MRR-PRO) and cloud radar (94 GHz, BASTA) showed good agreement between
 178 reflectivity measurements and estimated radar variables from OceanRAIN (Delanoë et al., 2016;
 179 Protat et al., 2019a).

180 The following assumptions in the T-matrix calculations were used for the C-band
 181 properties (Protat et al., 2019a, 2019b): (1) the drop shape–size relation from Thurai et al.
 182 (2007), (2) drop temperature of 10 °C, and (3) canting angles that follow a Gaussian distribution
 183 of 0° mean and 10° standard deviation. The following radar variables were then calculated for
 184 comparison with the OceanPOL variables to be discussed in the next section: horizontal
 185 reflectivity (Z_H ; dBz), differential reflectivity (Z_{DR} ; dB), and specific differential phase (K_{DP} ; °
 186 km^{-1}).

187 2.2. OceanPOL radar

188 OceanPOL has a beamwidth of 1.3°, a range sampling of 125 m (pulse length of 1
 189 microsecond), and a maximum radial distance of 150 km. It typically scans about 14 elevation
 190 angles from 0.7° to 32° at 1° azimuth intervals every 6 minutes (Protat et al., 2022), but the
 191 numbers of elevation angles and sampling intervals vary between cruises. The antenna control
 192 system of OceanPOL is used to stabilize the antenna for the radar to operate on a ship.

193 Two OceanPOL data sets have been made available by the Australian Bureau of
 194 Meteorology: (1) the Plan Position Indicator (PPI) volume data, and (2) an interpolated and
 195 gridded data set using a Barnes (1964) analysis. Here, we use the PPI data to preserve the pixel
 196 values of radar observables and avoid smearing of reflectivity features due to interpolation. The
 197 following variables from the PPI data were then extracted: Z_H , Z_{DR} , K_{DP} , cross-correlation
 198 coefficient (ρ_{HV}), signal-to-noise ratio (SNR), and the hydrometeor classification based on
 199 Thompson et al. (2014). The PPI data came from three voyages of the RV Investigator south of
 200 43° S with collocated OceanRAIN measurements (Figure 1c and Table S1).

201 The OceanPOL calibration follows the framework applied to operational radars in
 202 Australia (Warren et al., 2018; Protat et al., 2022). However, we implemented an additional
 203 quality control step to the PPI data to remove, to the extent possible, non-meteorological signals
 204 (e.g., sea clutter signals), as well as a Z_{DR} calibration adjustment of -0.4 dB. We then calculated
 205 each radar pixel’s refractivity-corrected altitude, distance from the ship, and coordinates using
 206 the Python library ‘Wradlib’ v1.20 (Heistermann et al., 2013). Finally, we retain only radar
 207 pixels with the following properties: (1) Z_{DR} between -4 and 4 dB, (2) $\rho_{HV} > 0.85$, and (3) SNR $>$
 208 10 dB (Figures S1–S3). We also limit the analysis to ranges of 10–50 km to minimize the impact
 209 of beam broadening (Ryzhkov, 2007). The quality control procedure was necessary and
 210 effectively removed the sea clutter signals from the PPI data, but we note that it will have
 211 removed some weak meteorological signals, which will be discussed in detail in Sections 3.4 and
 212 3.5.

213

214 2.3. Synoptic type classification using ERA5

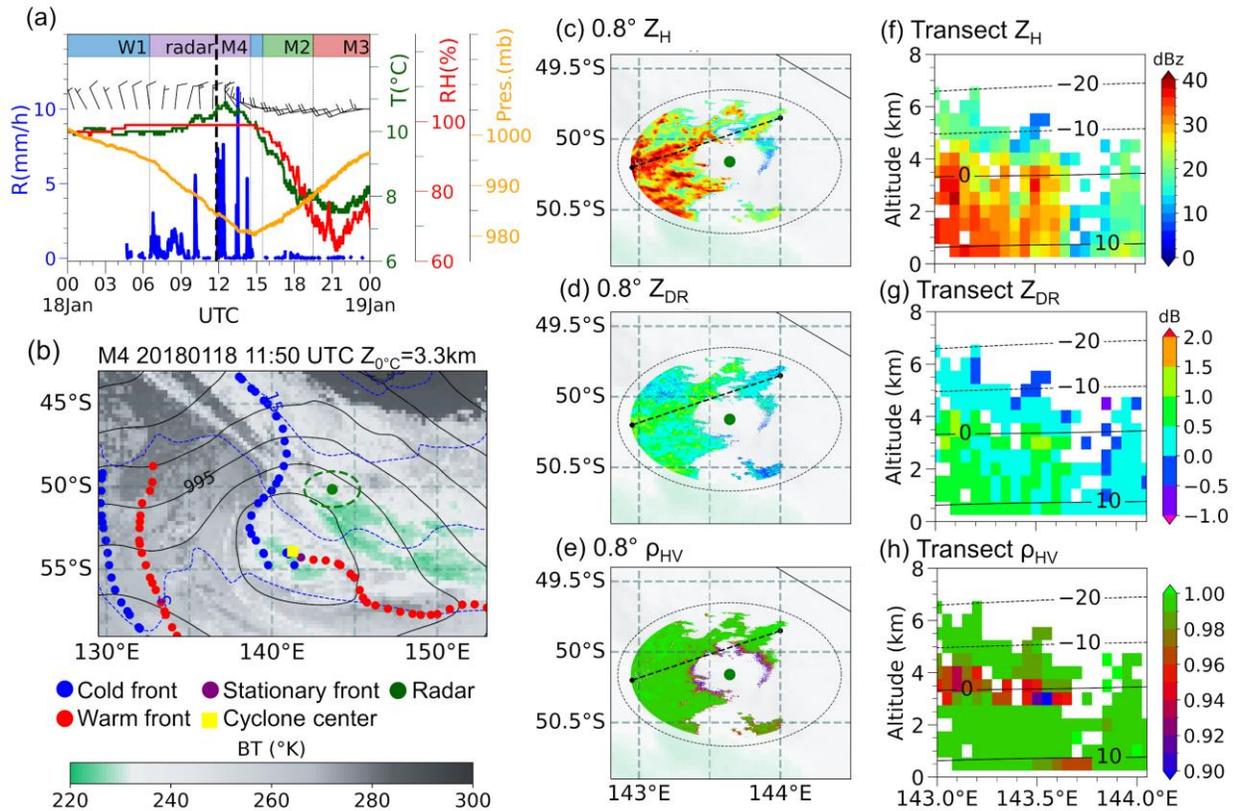
215 The diverse cloud and precipitation properties over the SO are strongly influenced by the
216 synoptic meteorology and thermodynamical environments over this region (Lang et al., 2018;
217 McFarquhar et al., 2021; Truong et al., 2020; Montoya Duque et al., 2023). Truong et al. (2020)
218 identified seven distinct synoptic types over the SO (Figure 1a) from a K-means cluster analysis
219 using upper air soundings from Macquarie Island and recent shipborne and aircraft campaigns
220 over the region (Figure 1a). These synoptic conditions extend the established cyclone and front
221 compositing methods over the SO by identifying two synoptic types that are unique over the
222 high-latitude SO. The K-means centroids of these seven clusters were used to identify the
223 synoptic types sampled by OceanRAIN (Figure 1b) and OceanPOL (Figure 1c). The synoptic
224 types near the ship location were identified following Montoya Duque et al. (2023) using hourly
225 data from the European Centre for Medium-Range Weather Forecast 5th generation climate
226 reanalysis product (ERA5; Hersbach, et al., 2020) (Figure S4).

227 **3 Results**

228 3.1. Precipitation over the warm sector of an extratropical cyclone: a sample case

229 The passage of an extratropical cyclone southwest of Tasmania on 18 January 2018 was
230 sampled by the RV Investigator (Figure 2 and Animation S1 in the supplementary material). The
231 cyclone was initially located 715 km west of the ship location (143.8 °E and 49.9 °S) and was
232 moving east-southeastward, allowing the ship to record information during several synoptic
233 phases during its passage (Figure 2a).

234 The event started with pre-frontal warm air advection (W1) at the ship location, with
235 north-northwesterly winds, surface temperatures around 10 °C, and precipitation developing
236 towards the transition to the warm sector (M4) between 06:30 UTC and 14:30 UTC. The cyclone
237 was closest to the ship (450 km southwest) at 11:50 UTC (Figure 2b), with a surface pressure
238 drop, northerly winds, and increasing precipitation (vertical dashed line in Figure 2a).
239 Precipitation during this period came from the trailing edge of cold optically thick clouds
240 indicated by the low brightness temperatures (<230 K) from the Himawari-8 (Figure 2b) and 0.8°
241 PPI scan of OceanPOL (Figures 2c–2e). A marked increase in Z_H and Z_{DR} and a decrease in ρ_{HV}
242 around the 3.3 km freezing level height (Figures 2f–2h) was detected from the radar vertical
243 cross-sections near the ship at 143–143.7 °E. These signals indicate stratiform precipitation with
244 a bright band signature, consistent with steady rain rates below 10 mm h⁻¹ detected by
245 OceanRAIN (Figure 2a). A brief period of pre-frontal warm air advection was detected at 14:50–
246 15:30 UTC as the ship location moved into the cold sector of the cyclone.



247

248 Figure 2. (a) Surface conditions sampled by OceanRAIN on 18 January 2018 as a cyclone
 249 traversed east-southeastward of the ship. The evolution of synoptic conditions at the ship
 250 location is shown at the top of the panel. The black vertical dashed line denotes the 11:50 UTC
 251 timestamp highlighted in the next panels. (b) Synoptic condition around the RV Investigator at
 252 11:50 UTC classified as an M4 cluster. Shown in the panel are the Himawari-8 Channel 13
 253 Brightness temperature (BT); cyclone center and associated fronts from objective identification
 254 methods (Murray & Simmonds, 1991; Berry et al., 2011); mean sea level pressure contours
 255 (solid black lines), surface temperature contours (dashed blue lines), and freezing level height at
 256 the ship location ($z_{0^\circ\text{C}}$ at the title) from the ERA5 data. The green-bordered circle denotes the
 257 150 km radius of OceanPOL. PPI scans of (c) Z_H (d) Z_{DR} , and (e) ρ_{HV} with 0.8° elevation at
 258 11:50 UTC. The black dashed circles denote the 1 km refractivity-corrected altitudes. Vertical
 259 profiles of (f) Z_H , (g) Z_{DR} , and (h) ρ_{HV} along the transect line near the ship, denoted by the black
 260 dashed diagonal line in (c)–(e). The ERA5 isotherms are also shown. We used each dataset’s
 261 nearest time offset to 11:50 UTC for (b)–(h) considering their different temporal resolutions.

262

263 Cold-frontal conditions (M2) were seen from 15:30–19:30 UTC, with westerlies and
 264 colder and drier air than in the M4 condition (Figure 2a). Finally, post-frontal conditions (M3)
 265 were encountered after 19:30 UTC, as the ship emerged from the cold sector around 670 km
 266 northwest of the cyclone center. Light rain ($< 1 \text{ mm h}^{-1}$) from multiple open cellular convective
 267 clouds was present in the M2 and M3 periods, characterized by widespread patchy shallow (< 2
 268 km) radar returns with $Z_H < 20 \text{ dBz}$ (Animation S1). These radar signatures are consistent with

269 previous observations of open mesoscale cellular convection (MCC) in the cold and post-frontal
270 sectors of SO cyclones (Huang et al. 2021; Lang et al., 2022).

271 The precipitation event presented above illustrated the different cloud organization,
272 polarimetric signatures, and surface variable characteristics including precipitation for the
273 different sectors of the extratropical cyclone. Individual PPI scans also revealed the temporal
274 consistency of precipitation macrostructures across various synoptic conditions.

275 3.2. Bulk statistics from OceanPOL and OceanRAIN

276 To examine further how the precipitation properties vary among synoptic types based on
277 OceanPOL and OceanRAIN observations, we present the bulk statistics of the polarimetric
278 signatures as well as the surface precipitation frequencies, thermodynamic phase, and rain
279 intensities. Contour Frequency by Temperature Diagrams (CFTD; Huang et al., 2015) were used
280 to illustrate the general structure and statistical properties of Z_H and Z_{DR} as a function of
281 temperature (Figure 3). The CFTD is a modified version of the Contour Frequency by Altitude
282 Diagram (Figures S5–S6; Yuter and Houze, 1995). The temperature associated with each
283 precipitation pixel was estimated using linear interpolation to the temperature field of ERA5 at
284 the nearest hour and 3D grid points. We also plot the fractional area of precipitation pixels
285 relative to the PPI scan area at 0.8° elevation. The following temperature regions were
286 highlighted to provide qualitative insights into the precipitation types and microphysical
287 processes aloft: (1) the freezing layer or 0°C line; (2) the Hallett-Mossop temperature range
288 between -8 and -3°C layer (Hallett & Mossop, 1974), which is often associated with mixed-
289 phase clouds and enhanced ice particle production; and (3) the -20 and -10°C layer, where
290 dendritic ice and hexagonal plate growth commonly develops within cold clouds (Bailey &
291 Hallett, 2009; Kennedy & Rutledge, 2011; Williams et al. 2015). The CFTDs were then related
292 to the bulk statistics of surface precipitation from OceanRAIN observation (Figure 4).

293 The OceanPOL and OceanRain data used here was collected over approximately 218
294 days, with precipitation observed at the ship approximately 20 % of the time. The precipitation
295 coverage of the warm front (M4) cluster had the largest areal fraction compared to other synoptic
296 types (first column of Figure 3 and Table 1). This result indicates the widespread precipitation in
297 the warm front sector consistent with the stratiform regime shown in the sample case (Figure 2),
298 and the smaller horizontal scales of precipitation in other synoptic types (e.g., Animation S1).
299 The high variability in precipitation coverage of the M4 cluster is related to the movement of the
300 warm front/sector from the ship location.

301 The vertical depth of precipitation also varies among synoptic types. Precipitation echoes
302 were often detected up to 7 km during the W1 and M4 cluster periods (Table 1 and Figure S5).
303 Lower precipitation echo tops were found in other synoptic types including high-pressure
304 conditions (M1), cold fronts (M2), post-frontal sectors (M3), polar ocean fronts (C1), and the dry
305 coastal Antarctic (C2).

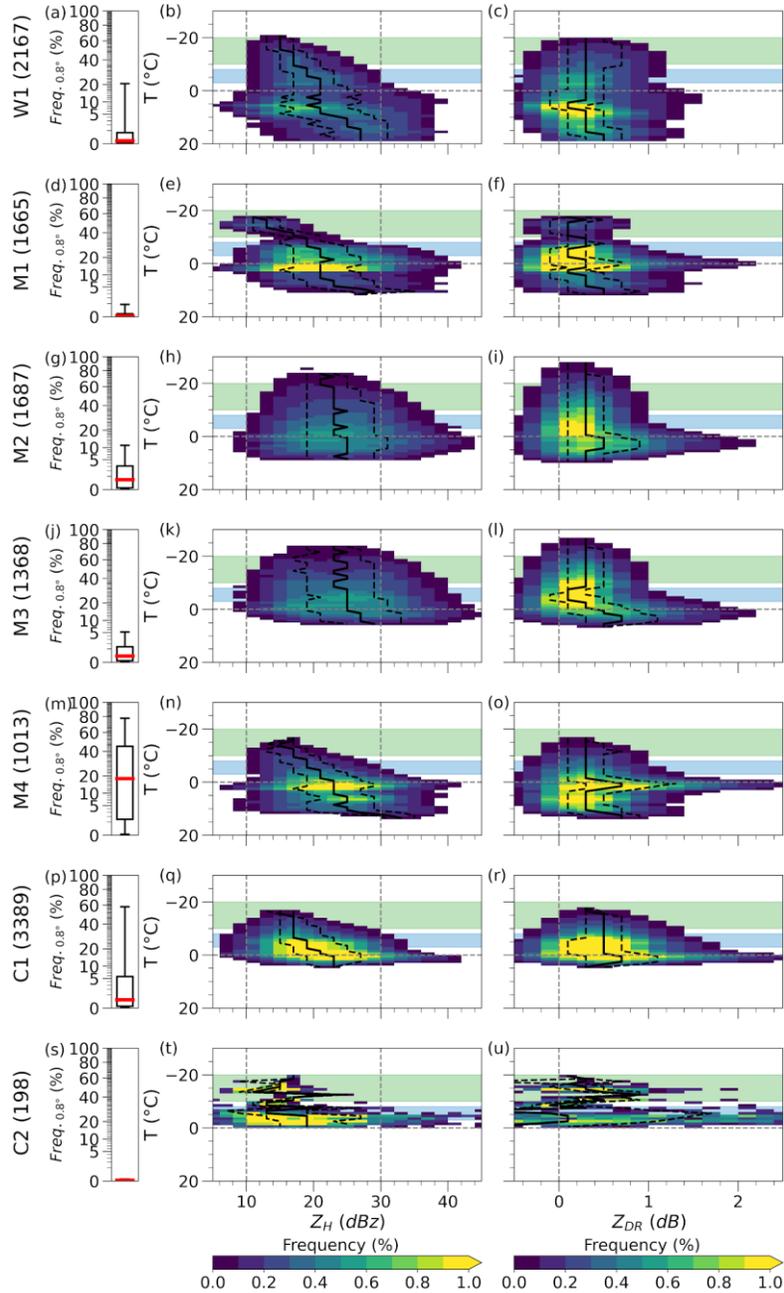
306 The CFTDs for Z_H and Z_{DR} provided insights into the polarimetric signatures and
307 possible microphysical processes related to precipitation particle growth. Low to moderate Z_H
308 (<30 dBz) and Z_{DR} values (<1 dB) were evident at temperatures between -20 to -10°C in the
309 CFTDs of all synoptic types (second and third columns of Figure 3 and Table 1). This radar
310 signature suggests the possible presence of quasi-isotropic ice particles that grow preferentially

311 in water-saturated environments (Giangrande et al., 2016; Griffin et al., 2018; Williams et al.,
312 2015; Wolde & Vali, 2001).

313 Looking at the polarimetric signatures above the freezing level for the different synoptic
314 types, the W1, M1, and M4 clusters had increasing median Z_H and uniform small median Z_{DR}
315 from -20 to -10 °C. These radar properties suggest the possible presence of active aggregation
316 and/or riming that could dilute the anisotropy and shape diversity of ice particles (Kumjian et al.,
317 2022; Ryzhkov et al., 2016; Williams et al., 2015; Wolde & Vali, 2001). The steady increase in
318 median Z_H values from the sub-freezing temperatures towards 0 °C also indicates the less
319 convective nature of the W1, M1, and M4 clusters. On the other hand, the M2, M3, and C1
320 clusters had broader Z_H distributions and increased presence of $Z_{DR} > 1$ dB extending towards
321 the Hallett-Mossop temperature range of -8 to -3 °C. These radar properties suggest diversity in
322 precipitation types and shapes (Giangrande et al., 2016; Ryzhkov et al., 2016; Keat &
323 Westbrook, 2017), and possibly mixed-phase precipitation associated with the convective nature
324 of the three synoptic types. Such a result is seemingly consistent with the limited in-situ and
325 remote data analysis that has shown the Hallett-Mossop ice multiplication process being active in
326 the M3 and C1 clusters (Huang et al., 2017, 2021; Montoya Duque et al., 2022; Mace et al.,
327 2023).

328 Finally, the largest spread to higher Z_H and Z_{DR} values occurred around 0 °C, but was less
329 pronounced in the W1 cluster and stronger in the colder clusters (from M1 to C1 clusters). This
330 radar feature suggests the melting of large ice particles (e.g., aggregates and rimed particles)
331 created in colder thermodynamic environments and is a typical bright-band signature.

332



333

334 Figure 3. (first column) Boxplots denoting the fraction (%) of precipitation coverage at 0.8° PPI
 335 elevation. The number of PPI data for each synoptic type is shown in parenthesis at each row
 336 label. Contour Frequency by Temperature Diagram (CFTD) of Z_H (second column) and Z_{DR}
 337 (third column) for frequencies above 0.05%. The dashed, solid, and dashed black lines along the
 338 abscissa show the 25th, 50th, and 75th percentiles. The shaded regions indicate possible dendritic
 339 growth layer (DGL) commonly occurring at -20 to -10 °C (green), and the Hallett-Mossop (H-
 340 M) temperature range at -8 to -3 °C (blue) often associated with mixed-phase clouds and
 341 enhanced ice particle production.

342

343 Table 1. Precipitation information from OceanPOL in terms of the median and 95th percentile
 344 values of precipitation coverage at 0.8° elevation (%), precipitation echo top (km), and ranges of
 345 median Z_H (dBz) and Z_{DR} (dB) values for the following temperature regions: dendritic growth
 346 layer (DGL; -20 to -10 °C), Hallett-Mossop (H-M; -8 to -3 °C), and above-freezing temperatures
 347 (>0 °C).

Synoptic type	Median (95th percentile) areal cover (%)	Echo top (km)	Median Z _H (dBz)			Median Z _{DR} (dB)	
			DGL	H-M	>0 °C	DGL & H-M	>0 °C
W1	0.01 (16.5)	7	15–17	19–21	19–27	0.3	0.1–0.5
M1	0 (0.5)	5.5	11–17	19–21	21–29	0.1–0.3	0.1–0.5
M2	0.4 (10)	5.5	21–23	23–25	23–25	0.3	0.3–0.5
M3	0.09 (2.5)	5.5	23–25	25	25–29	0.1–0.3	0.5–0.7
M4	16 (78)	7	15–17	19–21	23–33	0.3	0.3–0.7
C1	0.1 (53)	5.5	15–17	17–19	23	0.3–0.5	0.3–0.7
C2	0 (<0.1)	3.5	11–23	13–21	-	-1.1-0.9	-

348 Note: Numerical values found in this table are also shown graphically in Figures 3 and S5.

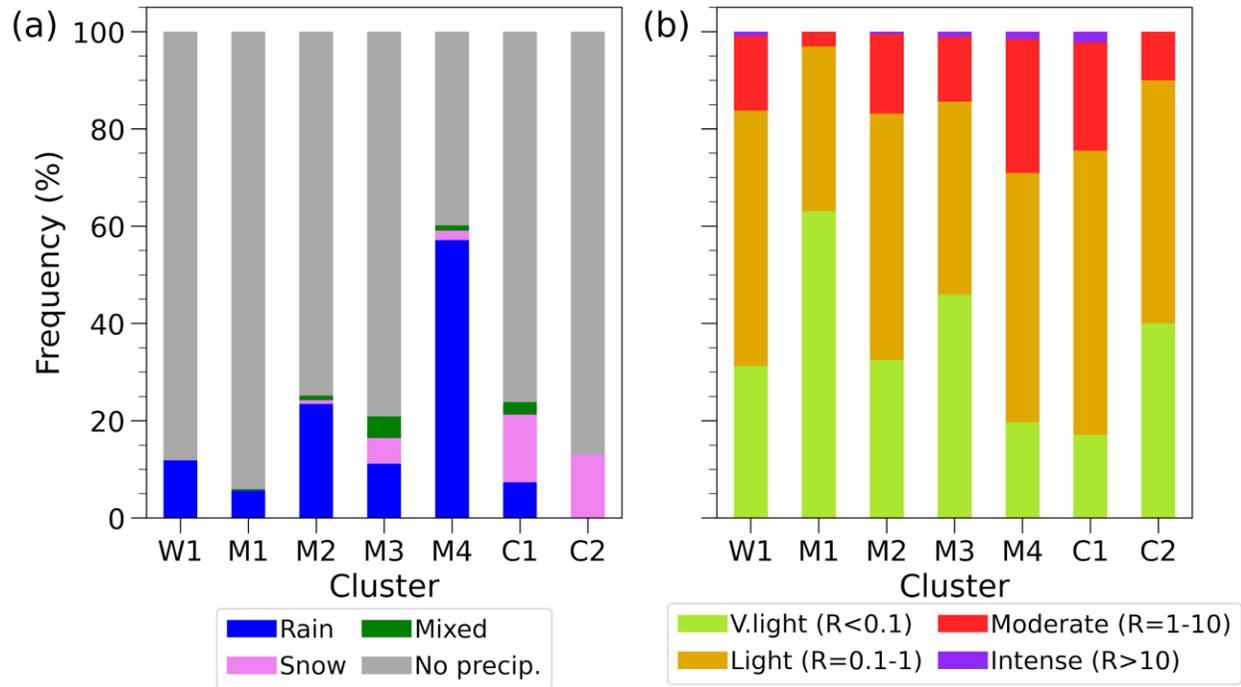
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350 At the surface, OceanRAIN sampled mainly rain in most synoptic types (Figure 4a), with
 351 71–97 % of the time being light rain rates (Figure 4b). The M4 cluster had the most precipitation
 352 occurrences, the M3 and C1 clusters had relatively higher fractions of mixed and snow
 353 precipitation, and the C2 cluster only had snow. Tansey et al. (2022) found similar results for the
 354 precipitation phase over Macquarie Island relative to the cyclone locations during summer, but
 355 our result expands this to higher latitudes and a broader area of the SO. We also examined
 356 whether the lowest 1 km radar returns from the OceanPOL data can be used to infer qualitatively
 357 the surface precipitation phase sampled by OceanRAIN using the ERA5 temperature values
 358 assigned to OceanPOL precipitation pixels (Figure S6). Results showed that the majority of
 359 precipitation pixels for most synoptic types were above 0 °C. A narrower temperature range near
 360 0 °C was found in the precipitation pixels of the M3 and C1 clusters, while the C2 cluster had all
 361 precipitation pixels occurring at sub-freezing temperatures. This highlights the general
 362 consistency in the precipitation characteristics detected by OceanRAIN and OceanPOL despite
 363 their very different sampling strategies.

364 In summary, the OceanPOL radar features and OceanRAIN surface observations provide
 365 useful information to characterize key precipitation properties and potential microphysical
 366 processes associated with the seven synoptic types over the SO. The M4 cluster had the largest
 367 precipitation coverage and the most frequent surface precipitation. Synoptic types with relatively
 368 warmer and less convectively unstable thermodynamic environments (W1, M1, and M4 clusters;
 369 Truong et al., 2020) showed clearer polarimetric signatures of potential aggregation/riming
 370 processes at sub-freezing temperatures. On the other hand, synoptic types with colder and more
 371 convectively unstable environments (M2, M3, and C1 clusters) showed higher variability in

372 polarimetric signatures, suggesting a wide diversity of precipitation types and shapes that are
 373 possibly associated with mixed-phase precipitation. There is also a general consistency in the
 374 surface thermodynamic phase of precipitation between OceanRAIN and OceanPOL.

375



376

377 Figure 4. (a) OceanRAIN frequency of precipitation and thermodynamic phase and (b) frequency
 378 of very light ($R < 0.1$ mm h^{-1}), light ($0.1-1$ mm h^{-1}), moderate ($1-10$ mm h^{-1}), and intense ($R > 10$
 379 mm h^{-1}) rain rates per synoptic type.

380

381 3.3. Rain microphysical properties

382 3.3.1. Observed Drop size distribution (DSD)

383 Knowledge of the DSDs is central in calculating the bulk rainfall properties and radar
 384 variables used for developing the rainfall estimators. Here, we examine the observed DSD
 385 obtained by OceanRAIN, and how the contributions of different raindrop sizes to rainfall
 386 accumulation varied among synoptic types (Figure 5). We have excluded the C2 cluster because
 387 of its very few rain samples.

388 The median values of the number concentrations $N(D)$ for each synoptic type were
 389 generally within the interquartile ranges of the total samples (Figure 5a). To examine whether
 390 this result is dependent on rain rates, we reduced the DSD variability by scaling the individual
 391 $N(D)$ per minute by their respective mass-weighted mean diameter (D_m) and generalized
 392 intercept parameter ($\log_{10}N_w$) (Testud et al., 2001; Protat et al., 2019a). The mean scaled $N(D)$ of
 393 all synoptic types generally converges into a single scaled $N(D)$ line (not shown), indicating that
 394 the median DSD shape found in Figure 5a is within the range of variability of the observed DSD
 395 across the SO.

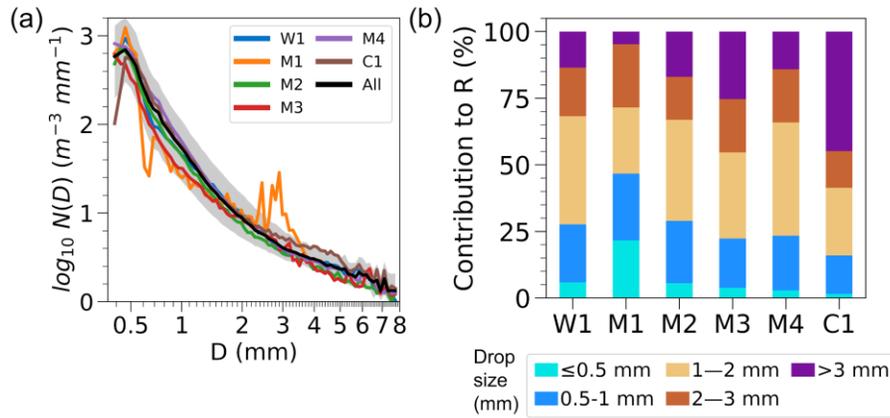
396 The contributions of the different raindrop sizes to total accumulation were also
397 examined (Figure 5b and Table 2). The contribution of small-sized drops to rainfall accumulation
398 across synoptic types (16–47 %) is higher than what was previously reported over Macquarie
399 Island in summer (5%; See Table 2 of Tansey et al., 2022). Data processing and instrument
400 differences may have contributed to this discrepancy rather than the fundamental differences in
401 rainfall properties alone. In particular, the higher detection rate of OceanRAIN to small-sized
402 raindrops can be due to its intended design for high sea-state measurements (Klepp 2015). On the
403 other hand, the Parsivel disdrometer used over Macquarie Island has been documented to
404 undercount small-sized droplets (Löffler-Mang & Joss, 2000; Tokay et al., 2013), which was
405 also validated in Tansey et al (2022).

406 Looking at the individual clusters, large-size raindrops had higher contributions to
407 rainfall accumulation in the M3 and C1 clusters. These raindrops possibly came from mixed-
408 phase precipitation aloft (e.g., frozen drops and rimed particles), produced by the convective
409 nature of the said clusters (Truong et al., 2020). These particles likely retained their large sizes
410 upon reaching the surface because the fall distance from the melting level to the surface was
411 small limiting breakup. The CFTDs of the M3 and C1 clusters support this interpretation,
412 showing broad Z_H and Z_{DR} distributions (Figure 3) and precipitation pixels occurring near 0 °C at
413 the lowest 1 km (Figure S6). We note that the M2 cluster, being associated with cold fronts,
414 features lower concentrations of large-size raindrops. This is likely due to the common presence
415 of multi-layer clouds in this cluster (Truong et al. 2022), which are not efficient in developing
416 heavy precipitation.

417 In contrast, large-size raindrops made a smaller contribution to rainfall accumulation in
418 the W1, M1, and M4 clusters. The three synoptic types have a less convective nature (Figure S4;
419 Truong et al., 2020) and thus limited collision-coalescence processes that are typically more
420 active in a convective and turbulent environment. These synoptic types also have higher freezing
421 level heights, which likely allowed break-up processes of large-sized ice particles created aloft.
422 The high contribution of mid-size raindrops to rainfall accumulation in the W1 and M4 clusters
423 may be explained by raindrop growth by coalescence below the freezing layer. These
424 interpretations are particularly consistent with the M4 cluster's CFTD (Figure 3) and sample
425 cases (not shown) that displayed a bright band signature and an increase in Z_{DR} at warmer
426 temperatures, although such polarimetric signatures are less apparent in the W1 and M1 clusters.

427

428



429

430 Figure 5. (a) median values of number concentrations across rain drop size spectra $N(D)$ for each
 431 synoptic type and all samples. The shaded region denotes the interquartile ranges from the
 432 overall median $N(D)$. (b) Contributions to rainfall accumulation of small-sized (< 1 mm; blue
 433 bars), mid-sized (1–3 mm; brown bars), and large raindrops (> 3 mm; violet bar).

434

435 Table 2. Contributions of raindrop sizes to rainfall accumulation for each synoptic type from
 436 OceanRAIN.

Synoptic type	Contribution to rainfall accumulation (%)		
	Small (< 1 mm)	Mid-sized (1–3 mm)	Large (> 3 mm)
W1	27.6	58.8	13.5
M1	46.7	48.6	4.8
M2	29.0	54.1	17.0
M3	22.3	52.4	25.3
M4	23.4	62.5	14.1
C1	16.0	39.2	44.9

437 Note: Numerical values found in this table are also shown graphically in Figure 5b for more raindrop size groups.

438

439 3.3.2. Analytical DSD

440 This section examines how well the commonly used analytical DSD forms capture the
 441 observed DSD and rain rates over the SO, given that analytical DSD forms are commonly used
 442 in remote sensing precipitation retrievals. Two analytical DSD formulations were evaluated,
 443 extending the analysis in Protat et al. (2019a) for different synoptic conditions. The first
 444 analytical form (Equation 1) is the Normalized Gamma distribution (Testud et al., 2001; Bringi
 445 et al., 2003; referred to as Normalized Gamma fit), which is a 3-parameter function used in the

446 DSD retrievals of the Global Precipitation Measurement (GPM) satellite products (Liao and
447 Meneghini, 2022). Its analytical $N(D)$ is given as

$$448 \quad N(D) = N_w \frac{\Gamma(4)(3.67 + \mu)^{4+\mu}}{3.67^4 \Gamma(4 + \mu)} \left(\frac{D}{D_m}\right)^\mu \exp\left[-(3.67 + \mu)\frac{D}{D_m}\right]$$

449 (1)

450 where Γ is the gamma function, μ the shape parameter, N_w the generalized intercept parameter,
451 and D_m is the mass-weighted mean diameter of the DSD.

452 The second analytical form (Equation 2) is the double-moment Normalized gamma
453 distribution by Delanoë et al. (2014; referred to as Delanoë fit), which has two shape parameters
454 (α and β). It also uses the N_w and D_m as input parameters, and its analytical form is given as:

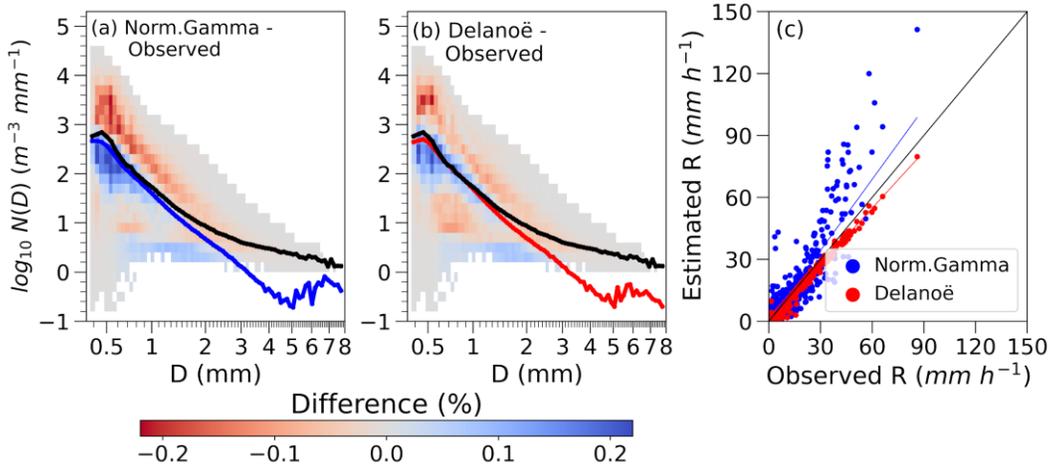
$$455 \quad N(D) = N_w \beta \frac{\Gamma(4) \left[\Gamma\left(\frac{\alpha + 5}{\beta}\right)\right]^{(4+\alpha)}}{4^4 \left[\Gamma\left(\frac{\alpha + 4}{\beta}\right)\right]^{(5+\alpha)}} \left(\frac{D}{D_m}\right)^\alpha \exp\left[-\left(\frac{\Gamma\left(\frac{\alpha + 5}{\beta}\right)}{\Gamma\left(\frac{\alpha + 4}{\beta}\right)}\right)^\beta \left(\frac{D}{D_m}\right)^\beta\right]$$

456 (2)

457 The analytical $N(D)$ from the Normalized Gamma and Delanoë fits were calculated by
458 fitting Equations (1) and (2) and their required inputs to individual observed $N(D)$ of
459 OceanRAIN every minute. These values were then used to estimate rain rates that were then
460 compared with OceanRAIN observations (Figure 6). The Delanoë curves fitted the observed
461 DSD (Figure 6b) better than the Normalized Gamma fit (Figure 6a) with lower $N(D)$ biases for
462 small-sized particles. This result is particularly important given the greater significance of small-
463 sized particles in SO rainfall (Figure 5b). The estimated rain rates from the Delanoë fit correlated
464 better with OceanRAIN observation and with less spread compared to the Normalized Gamma
465 fit results, although the Delanoë fit was slightly biased low (Figure 6c).

466 Satellites such as GPM use the Normalized Gamma fit with a constant shape parameter of
467 $\mu=3$ (Dual-frequency Precipitation Radar; Seto et al., 2013) and $\mu=2$ (Combined radar-
468 radiometer; Grecu et al., 2016). However, we found that these constant shape assumptions were
469 higher than the peak shape parameter values of -2 to 1 in all synoptic types (not shown),
470 consistent with what was reported in Protat et al. (2019a). Therefore, the shape parameter
471 assumptions may also contribute to the biases on rainfall retrievals of the GPM satellite products,
472 aside from the abovementioned limitation of the Normalized Gamma fit in retrieving the small-
473 sized particles over the SO.

474



475

476 Figure 6. Differences in the joint frequency distributions of analytical DSD using (a) the
 477 Normalized Gamma fit and (b) Delanoë fit relative to the observed DSD from all OceanRAIN
 478 samples. The lines denote the median $N(D)$ for the observation (black), Normalized Gamma fit
 479 (blue in (a)), and Delanoë fit (red in (b)). (c) Scatterplots of estimated rain rates using the
 480 Normalized Gamma fit (blue circles) and Delanoë fit (red circles) against OceanRAIN
 481 observation. Regression lines for the two gamma fits were also shown.

482

483 3.3.3. Rain microphysical parameters

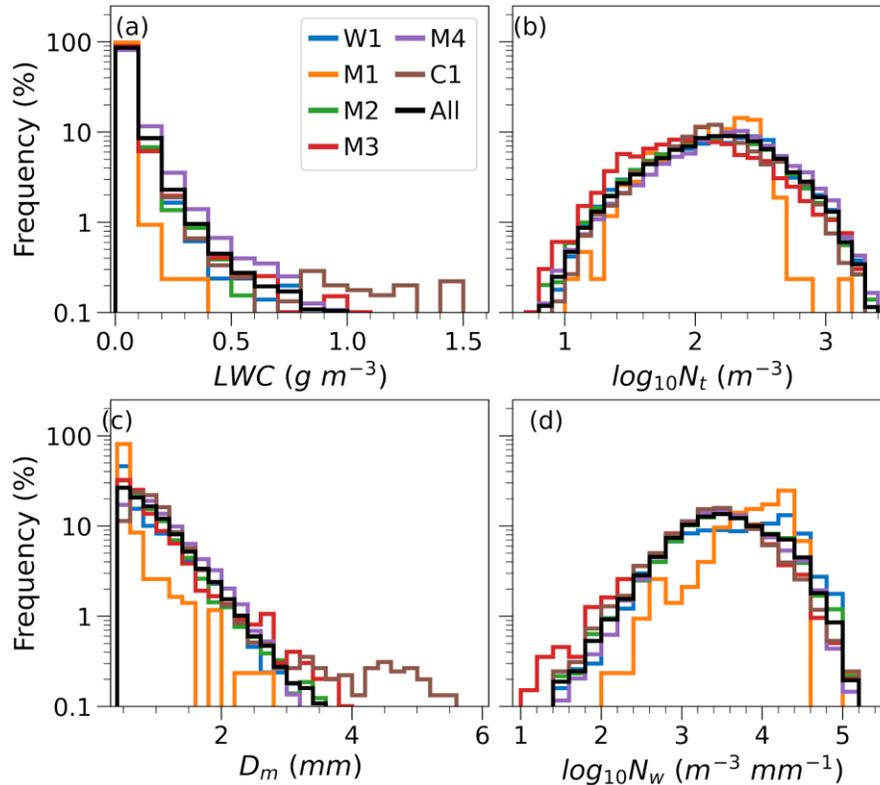
484 The frequency distributions of other rain microphysical variables such as the LWC,
 485 $\log_{10} N_t$, D_m , and $\log_{10} N_w$ were also examined (Figure 7). Some of these variables are related to
 486 lower DSD moments compared to rain rates and reflectivity, and therefore, are more
 487 significantly affected by small-sized raindrops (Raupach et al., 2019).

488 Results showed that the LWC values below 0.1 g m^{-3} occurred over 80 % of the time
 489 (Figure 7a), consistent with the dominance of drizzle and light rain across synoptic types (Figure
 490 4b). The overall $\log_{10} N_t$ distribution had a spread of 0.8–3.5 with minimal deviation (Figure 7b),
 491 likewise emphasizing the dominant number concentrations of drizzle and light rain in the data.
 492 More variability was seen in size-dependent variables such as D_m (Figure 7c) and $\log_{10} N_w$
 493 (Figure 7d), consistent with the different fractional contributions of raindrop sizes to total
 494 accumulation (Figure 5b). The M2, M3, and C1 clusters had lower fractions of $D_m < 1 \text{ mm}$
 495 compared with the M1, W1, and M4 clusters (Figure 7c). The joint frequencies of D_m and
 496 $\log_{10} N_w$ (not shown) further revealed that these convective clusters had more frequent samples of
 497 low $\log_{10} N_w < 3$ and high $D_m > 3 \text{ mm}$, highlighting the significant contributions of large-sized
 498 raindrops to their observed DSD. The overall $\log_{10} N_w$ peaked around $\log_{10} N_w = 3.6$, which is
 499 lower than what is typically found in the tropics (e.g., Protat et al., 2019a), and has a spread of
 500 1.4–5.2 for most synoptic types (Figure 7d). The W1 and M1 clusters had higher $\log_{10} N_w$ peaks
 501 at $\log_{10} N_w = 4.4$ due to their lower D_m compared with other synoptic types (Figure 7c).

502 In summary, Section 3.3 examined the rain microphysical properties from OceanRAIN
 503 measurements and their relation to OceanPOL polarimetric signatures, thermodynamic profiles,
 504 and potential microphysical processes for different synoptic environments. Small-sized raindrops
 505 contributed up to 47% of total accumulation across synoptic types. Large-size raindrops, on the

506 other hand, had more contribution to total accumulation in convective clusters (M3 and C1)
 507 compared with less convective clusters (W1, M1, and M4). The dominance of drizzle and light
 508 rain over the SO are manifested in other rain microphysical variables, also highlighting the
 509 importance of small-sized raindrops in the observed DSD over the SO. Given these
 510 characteristics, the analytical form by the Delanoë fit based on two shape parameters can better
 511 estimate the observed DSD and rain rates, as compared to the Normalized Gamma fit currently
 512 implemented in the DSD retrievals of GPM satellite products.

513



514

515 Figure 7. Frequency distributions of (a) liquid water content (LWC), (b) total concentration
 516 ($\log_{10}N_t$), (c) mass-weighted mean diameter (D_m), and (d) generalized number concentration
 517 ($\log_{10}N_w$) for the synoptic types and all samples. These variables were calculated from the DSD
 518 observations of OceanRAIN.

519

520 3.4. DSD-simulated radar variables and updated rainfall estimators from OceanRAIN

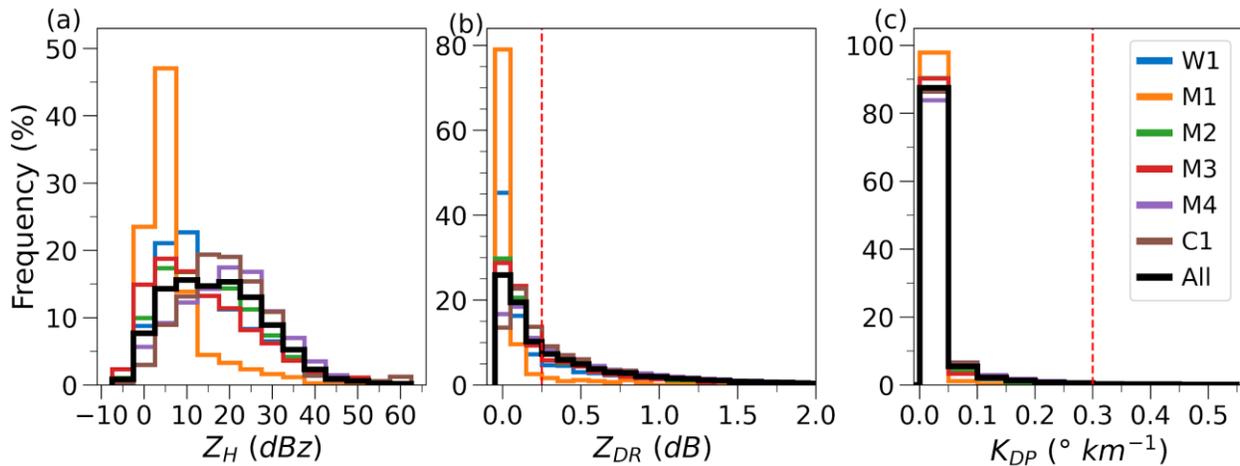
521 3.4.1. Z_H , Z_{DR} , and K_{DP} simulations

522 The observed DSD from OceanRAIN enables simulations of Z_H , Z_{DR} , and K_{DP} (Bringi et
 523 al., 2009; Cifelli et al., 2011; Thompson et al., 2018). Note that Z_H is proportional to the sixth
 524 power of raindrop sizes for Rayleigh scatter, Z_{DR} is related to the average particle oblateness, and
 525 K_{DP} to the number concentrations of non-spherical particles within a sampling volume (Bringi &
 526 Chandrasekar, 2001; Kumjian et al., 2022). Therefore, the DSD-simulated radar variables from

527 OceanRAIN observations provide important “ground-truth” to examine the quantitative rainfall
 528 estimates from OceanPOL for the remote SO.

529 Figure 8 presents the frequency distributions of Z_H , Z_{DR} , and K_{DP} values simulated from
 530 the OceanRAIN DSD. The Z_H distributions of most synoptic types were skewed to low values of
 531 $Z_H < 20$ dBz (Figure 8a). $Z_{DR} > 0.25$ dB occurred only 44 % of the time (Figure 8b), which is
 532 lower than what was found in the tropics (57 %) reflecting the smaller D_m values. $K_{DP} > 0.3$ °
 533 km^{-1} was virtually absent over the SO (Figure 8c), while it was relatively common in the tropics
 534 (11 %; Thompson et al., 2018). These results illustrate that an optimized set of radar-based
 535 rainfall estimators will better capture SO rainfall.

536



537

538 Figure 8. Frequency distributions of OceanRAIN DSD-simulated (a) Z_H , (b) Z_{DR} , and (c) K_{DP}
 539 values for the synoptic types and all OceanRAIN data using T-matrix calculations for C-band
 540 properties. The red vertical lines in (b) and (c) denote the threshold values of $Z_{DR} = 0.25$ dB and
 541 $K_{DP} = 0.3$ ° km^{-1} employed for rainfall retrieval equations.

542

543 3.4.2. Updated rainfall estimators for the SO (SO23)

544 The current rainfall retrieval algorithm used for the OceanPOL data sets is based on
 545 Thompson et al. (2018; hereafter TH18). TH18 has four rainfall estimators with different
 546 combinations of radar variables based on K_{DP} and Z_{DR} thresholds (second column of Table 3).
 547 The coefficients of these equations were derived from the DSD over the tropical ocean, and we
 548 have updated these to reflect the DSD characteristics observed by OceanRAIN over the SO
 549 (hereafter SO23; third column of Table 3). The K_{DP} and Z_{DR} thresholds were retained, since
 550 these values are associated with statistical uncertainty rather than detailed microphysics
 551 (Thompson et al., 2018). The $R(Z_H)$ and $R(Z_H, \zeta_{DR})$ are used mainly to estimate very light to
 552 moderate rain rates, and $R(K_{DP})$ and $R(K_{DP}, \zeta_{DR})$ to heavier rain (Cifelli et al., 2011; Thompson
 553 et al., 2018). We also performed a k-fold cross-validation (Kohavi, 1995) using k=10 iterative
 554 folds for training and validation of OceanRAIN data to confirm the robustness of SO23 against
 555 potential coefficient overfitting.

556 Table 3. Radar rainfall estimators for C-band properties based on Thompson et al. (2018; TH18)
 557 developed over the tropical oceans and OceanRAIN data over the SO derived in this study
 558 (SO23).

Criteria	TH18	SO23
$K_{DP} \leq 0.3$ and $Z_{DR} \leq 0.25$	$R(Z_H) = 0.021 z^{0.72}$	$R(Z_H) = 0.016 z^{0.846}$
$K_{DP} \leq 0.3$ and $Z_{DR} > 0.25$	$R(Z_H, \zeta_{DR}) = 0.0086 z^{0.91} \zeta_{DR}^{-4.21}$	$R(Z_H, \zeta_{DR}) = 0.011 z^{0.825} \zeta_{DR}^{-3.055}$
$K_{DP} > 0.3$ and $Z_{DR} \leq 0.25$	$R(K_{DP}) = 30.62 K_{DP}^{0.78}$	$R(K_{DP}) = 16.171 K_{DP}^{0.742}$
$K_{DP} > 0.3$ and $Z_{DR} > 0.25$	$R(K_{DP}, \zeta_{DR}) = 45.70 K_{DP}^{0.88} \zeta_{DR}^{-1.67}$	$R(K_{DP}, \zeta_{DR}) = 24.199 K_{DP}^{0.827} \zeta_{DR}^{-0.488}$

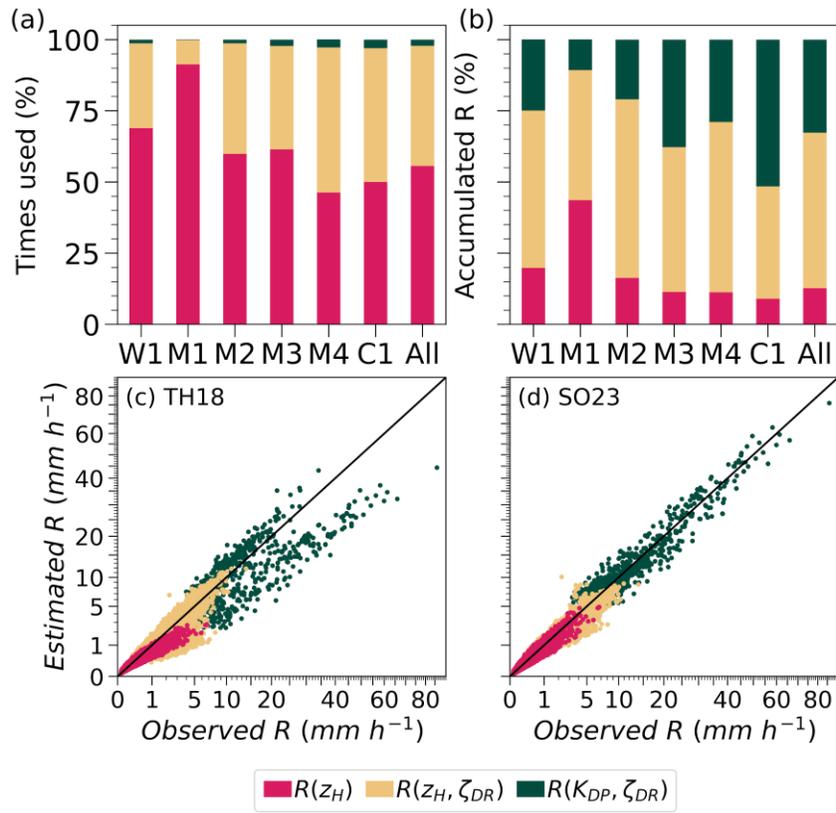
559 Note: The z_H and ζ_{DR} are the linear versions of Z_H and Z_{DR} , given by $10^{0.1 Z_H}$ and $10^{0.1 Z_{DR}}$, respectively.

560

561 The observed rain rates were first categorized into different estimators depending on their
 562 simulated Z_{DR} and K_{DP} values. Then, we examined how frequently the different estimators were
 563 used (Figure 9a) and their contributions to rainfall accumulation (Figure 9b). The $R(Z_H)$ was used
 564 about 56 % of the time for the SO rainfall (Figure 9a). On the other hand, moderate rain rates
 565 associated with $R(Z_H, \zeta_{DR})$ contributed most of the total accumulation (55 %; Figure 9b). These
 566 frequencies are 1.3 and 2.1 times higher than those in the tropics, signifying how the lower rain
 567 rates over the SO made these two rainfall estimators more important compared with the case
 568 over the tropics. The contributions of $R(K_{DP}, \zeta_{DR})$ to total accumulation in the M3 and C1
 569 clusters were higher (up to a factor of 5 higher than in other synoptic types; Figure 9b),
 570 signifying how the more frequent large-size raindrops in these clusters required the utility of K_{DP}
 571 and Z_{DR} values. The $R(K_{DP})$ was not used since there were no OceanRAIN samples with $K_{DP} >$
 572 $0.3 \text{ } ^\circ \text{ km}^{-1}$ and $Z_{DR} < 0.25 \text{ dB}$. Nonetheless, for completeness, we still derived the $R(K_{DP})$ using
 573 the samples with $K_{DP} > 0 \text{ } ^\circ \text{ km}^{-1}$ for the analysis with OceanPOL (Section 3.5).

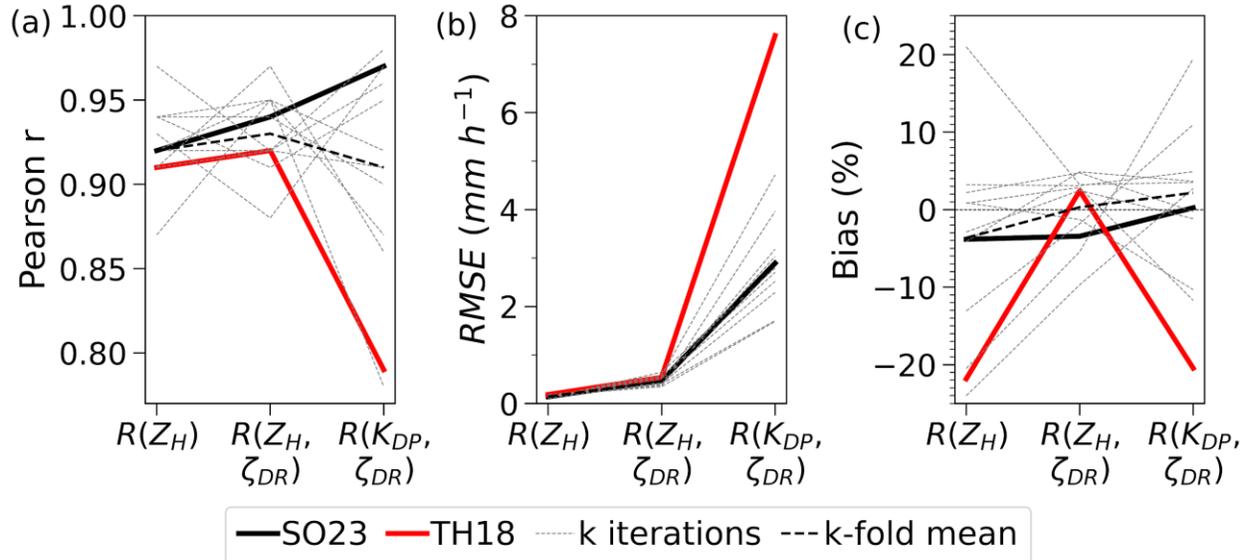
574 The observed rain rates were then compared against the OceanRAIN radar simulation-
 575 estimated rain rates of TH18 (Figure 9c) and SO23 (Figure 9d). The $R(Z_H)$ estimator of TH18
 576 tends to underestimate OceanRAIN observation (Figure 9c). This result demonstrates that Z_H is
 577 higher in the tropics than in SO for a given rain rate due to higher concentrations of large drops
 578 in tropical rain. There is also more spread in estimated rain rates using $R(Z_H, \zeta_{DR})$ and $R(K_{DP},$
 579 $\zeta_{DR})$ in TH18, which were notably improved in SO23 (Figure 9d). Estimated rain rates using
 580 SO23 correlate better with OceanRAIN observations (Figure 10a), and had lower root-mean-
 581 square error (RMSE; Figure 10b) and total accumulation bias (Figure 10c) compared with TH18.
 582 Results from k-fold cross-validation (black dashed line) were also more skillful than that of
 583 TH18, confirming the robustness of SO23 coefficients in accounting the variability within the
 584 OceanRAIN data.

585



586

587 Figure 9. (a) Frequency of times used and (b) contribution to total rainfall accumulation of
 588 different rainfall estimators using the OceanRAIN DSD-simulated Z_H , Z_{DR} , and K_{DP} values.
 589 Estimated rain rates of (c) TH18 and (d) SO23 retrieval equations (Table 3) relative to
 590 OceanRAIN observation. Note that the x- and y-axes were scaled to show lower rain rates.
 591



592

593 Figure 10. (a) Pearson correlation coefficient (r), (b) root mean squared error (RMSE), and (c)
 594 percent bias to total accumulation of OceanRAIN radar simulation-estimated rain rates using
 595 TH18 (red line) and SO23 (black line) for $R(Z_H)$, $R(Z_H, \zeta_{DR})$, and $R(K_{DP}, \zeta_{DR})$ relative to
 596 OceanRAIN observation. The figure also shows the k-fold cross-validation results for SO23 with
 597 $k = 10$ models (thin black dashed lines) and their mean values (thick black dashed lines) for the
 598 different metrics across three rainfall estimators. No OceanRAIN samples satisfied the $R(K_{DP})$
 599 criteria of $K_{DP} > 0.3 \text{ } ^\circ \text{ km}^{-1}$ and $Z_{DR} \leq 0.25 \text{ dB}$.

600

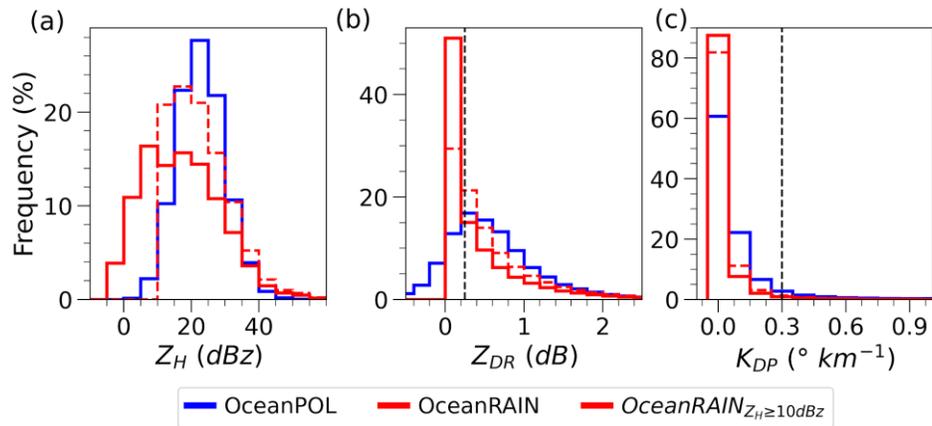
601 3.5. Comparison between OceanRAIN and OceanPOL radar variables

602 The DSD-simulated radar variables from OceanRAIN were compared against the quality-
 603 controlled radar observables of OceanPOL below 1 km (Figure 11). Only the OceanPOL
 604 precipitation pixels classified as rain in its hydrometeor classification product were included in
 605 this analysis. Such a comparison allows for a qualitative assessment of the consistency between
 606 the two datasets, despite the inherent differences in their instrumentation and sampling
 607 procedures. This method helps ensure the applicability of the SO23 rainfall retrieval algorithm to
 608 the OceanPOL radar observables.

609 About 31 % of OceanRAIN-simulated Z_H values were below 10 dBz (Figure 11a). This
 610 low Z_H value is outside the reliable measurements of OceanPOL (Section 2.2 and Figures S1–
 611 S3). Only the M4 and C1 clusters, which had heavier rain rates, had similar Z_H distributions in
 612 OceanPOL and OceanRAIN data (Figure S7). The OceanPOL's limitation to $Z_H \sim 10 \text{ dBz}$ also
 613 resulted in higher Z_{DR} (Figure 11b) and K_{DP} (Figure 11c) distributions compared to OceanRAIN-
 614 simulated radar values as samples with small drops and low Z are preferentially removed. The
 615 discrepancies between OceanRAIN and OceanPOL generally reduced after removing the subset
 616 of OceanRAIN data with $Z_H < 10 \text{ dBz}$ (thin red line in Figure 11). This result means that the
 617 OceanPOL data is comparable to OceanRAIN-simulated radar values excluding low Z_H , which
 618 gives confidence in using the SO23 algorithm to improve OceanPOL rainfall estimates.

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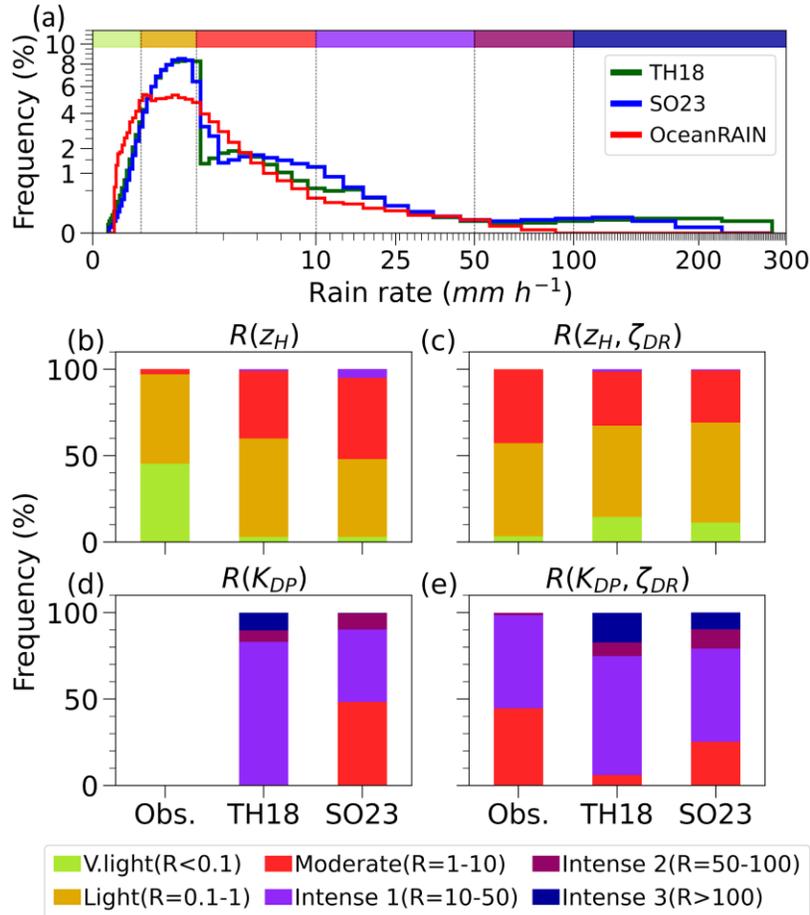
621

622 Figure 11. Frequency distributions of (a) Z_H , (b) Z_{DR} , and (c) K_{DP} from all data of OceanPOL
 623 (blue line) and OceanRAIN (thick red solid line), and the subset of OceanRAIN data with $Z_H \geq$
 624 10 dBZ (thin red dashed line). The OceanRAIN values were simulated from the surface DSD
 625 information using the T-matrix calculation (Section 2.1.2), while the OceanPOL values
 626 comprised the quality-controlled rain pixels within 10–50 km at the lowest 1 km altitude (Section
 627 2.2).

628

629 Figure 12 compares the frequency distributions of OceanPOL rainfall estimates using
 630 TH18 and SO23 relative to OceanRAIN observations. Note that a direct validation of OceanPOL
 631 estimates with OceanRAIN observations is not possible because the OceanRAIN was located in
 632 the “blind zone” of the OceanPOL. The OceanPOL rain rate estimates using SO23 showed better
 633 agreement with observation than the previous algorithm, particularly at the right tail (Figure
 634 12a). This result is highlighted in $R(K_{DP})$ (Figure 12d), where the OceanPOL estimates from
 635 SO23 had fewer intense rate rates, and in $R(K_{DP}, \zeta_{DR})$ (Figure 12e), where the OceanPOL
 636 estimates from SO23 were closer to observation. OceanPOL estimates for $R(Z_H)$ (Figure 12b)
 637 and $R(Z_H, \zeta_{DR})$ (Figure 12c) are generally comparable to OceanRAIN observations, except for
 638 the very light rain rates that were not present in OceanPOL due to its limitation to weak signals.

639



640

641 Figure 12. (a) Frequency distributions of rain rates from OceanRAIN observation and OceanPOL
 642 estimates using TH18 and SO23 retrieval equations (Table 3). The bars at the top of the panel
 643 denote the ranges of categorized rain rates. Note that the x- and y-axes were scaled to highlight
 644 lower rain rate and frequency values. (b–e) Frequencies of categorized rain rates from
 645 OceanRAIN observation and OceanPOL estimates using TH18 and SO23. There were no
 646 OceanRAIN observations that used the $R(K_{DP})$ in (d).

647

648 4 Discussion and Conclusions

649 This study used the OceanRAIN disdrometer and OceanPOL C-band polarimetric radar
 650 to characterize precipitation and improve radar rainfall estimates over the Southern Ocean (SO).
 651 Quality-controlled OceanRAIN and OceanPOL data from seven voyages of the RV Investigator
 652 in the Austral warm seasons of 2016 to 2018 were analyzed. The data was divided into seven
 653 distinct synoptic types. Key results include:

- 654 1. Precipitation over the broad SO during the Austral warm season is dominated by drizzle
 655 and rain rates less than 1 mm h^{-1} . Small-sized raindrops with diameters less than 1 mm
 656 contributed 16–47 % of total accumulation across all synoptic types.

- 657 2. Precipitation was most frequent in the warm sector (M4) of an extratropical cyclone,
 658 while least frequent in high-pressure conditions (M1) and coastal Antarctic-associated
 659 (C2) clusters.
- 660 3. Larger mass-weighted mean drop diameters were found in synoptic types with colder
 661 thermodynamic profiles and more convectively unstable environments such as the cold
 662 front sector (M2), post-frontal sector (M3), and ocean polar front at the sub-Antarctic
 663 region (C1), as compared to synoptic types with warmer thermodynamic environments,
 664 such as the warm-air advection (W1), M1, and M4 clusters.
- 665 4. Polarimetric signatures from OceanPOL provided information on the possible presence of
 666 quasi-isotropic ice particles within water-saturated environments, more active
 667 aggregation/riming processes in less convective clusters (W1, M1, and M4), and a wider
 668 variety of precipitation types and microphysical processes in more convective clusters
 669 (M2, M3, and C1).
- 670 5. The analytical form of raindrop size distribution (DSD) by Delanoë et al. (2014), which
 671 uses a double-moment normalization with two shape parameters better captures the
 672 observed DSD and rain rates over the SO compared with the Normalized Gamma
 673 distribution currently implemented in GPM satellite retrievals.
- 674 6. Radar rainfall estimators developed specifically for the SO using observed DSD from
 675 OceanRAIN outperformed the tropics-based retrieval equations (Thompson et al., 2014)
 676 currently used by OceanPOL. The stability of the coefficients of the new retrieval
 677 equations was also confirmed.

678 The quality control procedure applied in OceanPOL data, including the $\rho_{HV} > 0.85$ and
 679 $SNR > 10$ dB, can be configured depending on the synoptic type that will be examined in future
 680 case studies. On the other hand, the Z_{DR} offset of -0.4 dB will also change with future data of
 681 OceanPOL, given the ongoing efforts in updating OceanPOL data with improved calibration,
 682 K_{DP} estimation, and quality control. We also note the current limitation of OceanPOL in
 683 differentiating meteorological signals from noise and sea clutter at $Z_H < 10$ dBZ, which
 684 highlights the existing challenges in retrieving the bulk properties of drizzle dominant over the
 685 SO.

686 Direct in-situ measurements are essential in validating the polarimetric signatures from
 687 OceanPOL. For instance, future studies that incorporate multi-frequency radars collocated on the
 688 ship, and combined Doppler spectral analysis with radar polarimetry (e.g., Oue et al., 2018; Keat
 689 & Westbrook, 2017) would help in better understanding the variety of mixed and ice
 690 precipitation and processes involved in the region. Additionally, the prevalence of mixed
 691 precipitation and snow over the high-latitude SO necessitates the retrievals of their bulk
 692 properties (e.g., Mace et al., 2023).

693 Finally, the use of the Normalized gamma distribution (Testud et al., 2001; Bringi et al.,
 694 2003) may contribute to the biases of GPM satellite products in retrieving DSD information over
 695 the SO. The observed shape parameter over the SO is more likely to decrease and deviate further
 696 from the GPM assumptions if the reconstructed DSDs at drizzle mode (Thurai et al., 2018;
 697 Raupach et al., 2019) are considered to resolve small raindrops (< 0.4 mm) at OceanRAIN's
 698 truncation limit. This suggests the potential need for GPM retrievals to refine the shape
 699 parameter assumptions or integrate a new analytical DSD form, such as the double moment

700 normalization by Delanoë et al. (2014), for better retrievals of the drizzle-dominant rainfall
701 regime commonly observed over the high-latitude oceans including the SO.

702

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715

716 **Data Availability Statement**

717 The OceanRAIN version 2 data from the RV Investigator is available upon request to
718 Australia's Bureau of Meteorology through Dr. Alain Protat (alain.protat@bom.gov.au). The
719 OceanPOL PPI data are publicly available at <https://www.openradar.io/oceanpol> (doi:
720 10.25914/5fc4975c7dda8). The GADI server of Australia's National Computational
721 Infrastructure (<https://nci.org.au/our-systems/hpc-systems>) enabled access to Himawari-8, ERA5,
722 and OceanPOL data; user registration is needed. The ship tracks of the RV Investigator where
723 OceanRAIN and OceanPOL operated can be accessed at
724 http://www.marine.csiro.au/data/trawler/survey_list.cfm?source_id=309.

725

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