

1 **Measured ice nucleating particle concentrations**
 2 **improve the simulation of mid-level mixed-phase clouds**
 3 **over the high-latitude Southern Ocean**

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22 **Key Points:**

- 23 • WRF simulations of mixed-phase clouds over the high-latitude Southern Ocean
 24 are evaluated with remotely-sensed data from the MARCUS ;
 25 • Accounting for the low concentration of ice nucleating particles is critical to sim-
 26 ulate thin supercooled liquid water layer at cloud top;
 27 • Further parameterization developments targeting the convection at cloud top are
 28 needed to reproduce the turbulence-microphysics interplay.

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29 **Abstract**

30 Climate models exhibit major radiative biases over the Southern Ocean owing to a poor
 31 representation of mixed-phase clouds. This study uses the remote-sensing dataset from
 32 the Measurements of Aerosols, Radiation and Clouds over the Southern Ocean (MAR-
 33 CUS) campaign to assess the ability of the Weather Research and Forecasting (WRF)
 34 model to reproduce frontal clouds off Antarctica. It focuses on the modeling of thin mid-
 35 level supercooled liquid water layers which precipitate ice. The standard version of WRF
 36 produces almost fully glaciated clouds and cannot reproduce cloud top turbulence. Our
 37 work demonstrates the importance of adapting the ice nucleation parameterization to
 38 the pristine austral atmosphere to reproduce the supercooled liquid layers. Once sim-
 39 ulated, droplets significantly impact the cloud radiative effect by increasing downwelling
 40 longwave fluxes and decreasing downwelling shortwave fluxes at the surface. The net ra-
 41 diative effect is a warming of snow and ice covered surfaces and a cooling of the ocean.
 42 Despite improvements in our simulations, the local circulation related to cloud-top ra-
 43 diative cooling is not properly reproduced, advocating for the need to develop a param-
 44 eterization for top-down convection to capture the turbulence-microphysics interplay at
 45 cloud top.

46 **Plain Language Summary**

47 Among the major shortcomings of climate models is a poor representation of clouds
 48 over the Southern Ocean. Thanks to new measurements from the Measurements of Aerosols,
 49 Radiation and Clouds over the Southern Ocean campaign that took place aboard the
 50 Aurora Australis ice breaker, we can now better assess the ability of models to repre-
 51 sent clouds off Antarctica. In particular, we focus here on clouds that are mostly com-
 52 posed of ice crystals but that are topped by a thin layer of so-called ‘supercooled’ liq-
 53 uid droplets that form at temperatures below zero Celsius. While the standard version
 54 of the model produces almost fully-glaciated clouds (clouds composed only of ice), we
 55 show that by adapting the formulation of ice crystal formation to the very pristine at-
 56 mospheric conditions peculiar to the Southern Ocean it is possible to successfully repro-
 57 duce thin layers of supercooled liquid droplets observed in mixed-phase clouds. The lat-
 58 ter significantly changes how much sunlight these clouds reflect to space, which is crit-
 59 ical to understanding the climate. Compared to ice crystals, liquid droplets tend to re-

60 reflect more solar energy towards space and at the same time, they enhance the cloud in-
61 frared emission towards the surface of the Antarctic ice sheet.

62 **1 Introduction**

63 The Southern Ocean is a region where radiative biases in models involved in the
64 5th Coupled Model Intercomparison Project (CMIP) are amongst the largest globally
65 (Flato & coauthors, 2013; Hyder et al., 2018). Such biases have been attributed to a poor
66 representation of clouds that cover more than 80 % of the total Southern Ocean surface
67 on average (Mace, 2010) and that are mostly of mixed-phase composition, i.e. contain-
68 ing both ice crystals and supercooled liquid water (SLW). Low-level mixed-phase clouds
69 are the primary source of those biases (Bodas-Salcedo et al., 2014) but mid-level clouds
70 associated with the passage of warm fronts are also partly responsible (Mason et al., 2014).
71 While the climate sensitivity in some of recent climate models highly depends on South-
72 ern Ocean clouds (Gettelman et al., 2019; Zelinka et al., 2020), substantial shortcom-
73 ings regarding the simulation of mixed-phase clouds persist (e.g., Lenaerts, Van Tricht,
74 Lhermitte, & L'Ecuyer, 2017; Kawai et al., 2019).

75 The SLW amount in austral mixed-phase clouds is particularly high in summer,
76 at low altitude and over ice-free surfaces (Listowski et al., 2019). Highly reflective SLW
77 droplets substantially enhance the cloud albedo and therefore the amount of shortwave
78 radiation reflected towards space (Kay et al., 2016; Protat et al., 2017). By significantly
79 increasing the cloud optical depth, the amount of SLW in clouds is also critical for their
80 radiative forcing in the infrared spectrum.

81 Atmospheric models generally struggle to reproduce the albedo (Bodas-Salcedo et
82 al., 2014, 2016; Varma et al., 2020) and the surface longwave radiative flux associated
83 with frontal clouds over the Southern Ocean, that can be further advected over the Antarc-
84 tic ice sheet (King et al., 2015; Listowski & Lachlan-Cope, 2017; Vignon et al., 2018; Hines
85 et al., 2019; Ricaud et al., 2020). This is highly problematic for reproducing the net cloud
86 radiative forcing at the ice sheet surface and for predicting melting events associated with
87 oceanic intrusions of warm, moist and cloudy air masses (Nicolas et al., 2017; Wille et
88 al., 2019; Silber et al., 2019; Gilbert et al., 2020). Along the Antarctic edge, SLW is also
89 a key ingredient for precipitation generation and growth, through riming of snowflakes

90 (Grazioli et al., 2017; Vignon, Besic, et al., 2019) and through secondary ice production
 91 processes (Young et al., 2019; Sotiropoulou et al., 2020).

92 In mixed-phase clouds, SLW is thermodynamically unstable and depletes through
 93 transfer of water vapor towards ice crystals by the Wegener-Bergeron-Findeisen (WBF)
 94 process. The presence of SLW in mixed-phase clouds for more than a few hours is thus
 95 explained by a complex interplay between radiative exchanges, turbulent mixing and mi-
 96 crophysics (Morrison et al., 2012; A. V. Korolev & Mazin, 2003). A body of literature
 97 has documented this *a priori* surprising resilience of SLW in cold clouds, especially in
 98 boundary-layer clouds in the Arctic (see reviews in A. Korolev et al., 2017 and Andronache
 99 & coauthors, 2017). In particular, for typical mixed-phase stratocumulus and altocumu-
 100 lus found at mid- or high latitudes (Hogan et al., 2003; P. A. Barrett et al., 2020), the
 101 SLW resilience results from the following mechanism. At cloud top, the radiative cool-
 102 ing of the air - and to a second extent the sublimation and evaporation of hydromete-
 103 ors - drive a top down turbulent mixing that in turn generates compensating updrafts.
 104 If the updrafts are intense enough (A. V. Korolev & Mazin, 2003), the relative humid-
 105 ity can exceed saturation with respect to liquid through air adiabatic cooling during as-
 106 cent. Cloud droplets thus form and are advected upward, thereby forming a thin - a few
 107 hundred meter deep - layer of SLW at cloud top, below which ice crystals grow through
 108 the WBF process and possibly other mechanisms like riming and then sediment. SLW
 109 formation is further favored in conditions of high concentrations of cloud condensation
 110 nuclei (CCN) and low concentrations of ice nucleating particles (INPs).

111 The difficulty for atmospheric models to simulate SLW in austral mixed-phase clouds
 112 - be they either low-level stratocumulus or mid-level clouds - mostly lies in: *i*) their too
 113 coarse vertical resolution since SLW layers are a few tens or hundreds meters deep, i.e.,
 114 often thinner than model layers in common atmospheric models (A. I. Barrett et al., 2017b);
 115 *ii*) in a deficient representation of the turbulent mixing at the sharp cloud top bound-
 116 ary (Sotiropoulou et al., 2016) and *iii*) inadequate parameterizations or tuning of cold
 117 microphysical processes for the typical conditions encountered at high latitude (Forbes
 118 & Ahlgrim, 2014; A. I. Barrett et al., 2017a; Furtado et al., 2016; Listowski & Lachlan-
 119 Cope, 2017). The atmosphere above the Southern Ocean being particularly pristine, with
 120 INPs in the boundary layer that mostly originate from sea spray aerosols only (DeMott
 121 et al., 2016; McCluskey et al., 2018; Uetake et al., 2020), many current model formula-
 122 tions for ice nucleation may be inadequate. Such formulations were indeed mostly de-

123 developed for mid-latitude conditions where much higher concentrations of INPs are typ-
 124 ically present. They can potentially lead to substantial underestimation of SLW droplets
 125 in clouds and hence major radiative biases in models (Vergara-Temprado et al., 2018).
 126 In addition, previously underappreciated processes like secondary ice production through
 127 ice particle break-up also seem particularly critical to explain the concentration of ice
 128 crystals in clouds over the Antarctic coast (Young et al., 2019; Sotiropoulou et al., 2020).

129 During the austral summer 2017-2018, the Measurement of Aerosols, Radiation and
 130 Clouds over the Southern Ocean (MARCUS) campaign was conducted aboard the Aus-
 131 tralian ice-breaker Aurora Australis as the ship made three return crossings of the South-
 132 ern Ocean from Hobart to East Antarctica in order to resupply the three Australian Antarc-
 133 tic stations.

134 The MARCUS campaign offers a unique dataset to evaluate the ability of atmo-
 135 spheric models to represent frontal mixed-phase clouds adjacent to the Antarctic coast
 136 and to foster the development, evaluation and tuning of adequate microphysics and tur-
 137 bulence parameterizations in models.

138 In this study, we make use of those data to evaluate and improve the representa-
 139 tion of austral mixed-phase clouds in the Weather Research and Forecasting (WRF) model.
 140 We focus on clouds associated with the passage of a warm front above Mawson station
 141 (67.6°S, 62.9°E, green dot in Figure 1) between the 14 and the 16 February 2018. We
 142 pay particular attention to the challenging representation of SLW layers at the top of
 143 mid-altitude clouds preceding and following the front. Beyond the WRF evaluation, the
 144 aim of the paper is to identify priorities and propose pathways for parameterization de-
 145 velopment and tuning which can assist cloud modeling over the Southern Ocean.

146 **2 Meteorological setting, observations and simulations**

147 **2.1 Remotely-sensed and in situ observations from the Aurora Australis**

148 A comprehensive suite of instrumentation from the second Atmospheric Radiation
 149 Measurement (ARM) Mobile Facility (McFarquhar et al. 2020, submitted to BAMS) was
 150 deployed aboard the ship. A vertically-pointing W-band (95 GHz) Doppler cloud radar
 151 (MWACR) sampling every 2 s and set-up on a stabilizing platform provided vertical pro-
 152 files of reflectivity, Doppler velocity and spectral width. The reflectivity measurements
 153 were calibrated following Kollias et al. (2019). During the study case period, the ship

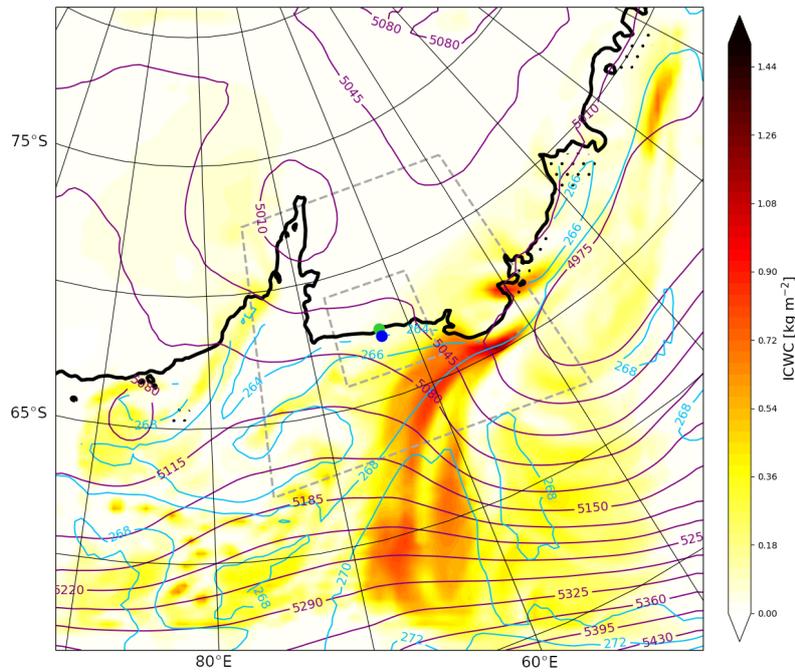


Figure 1. Map of synoptic conditions around Mawson station at 00 UTC, 15 February 2018, from the ctrl Polar WRF simulation (27-km resolution domain). The purple (resp. cyan) contours show the 500 hPa geopotential height in m (resp. the 900 hPa temperature above the Ocean in K). The color shading shows the vertically integrated condensed water content (ICWC, sum of cloud liquid droplets, cloud ice crystals, snow, rain and graupel species). Dashed grey lines delimit the 9-km and 3-km resolution domains. Regions where the sea ice concentration is greater than 0.5 are marked with small black dots. The green circle locates Mawson station while the blue circle indicates the position of the Aurora Australis at 00 UTC, 15 February 2018.

154 was in very calm waters thanks to offshore ice that damped sea swells. Subsequently, the
155 radar Doppler velocity uncertainty due to ship's heave is very low (the standard devi-
156 ation of the heave velocity during the three days of interest is lower than 0.01 m s^{-1}).
157 From the processing of Doppler velocity time series, it is possible to estimate the dissi-
158 pation rate ϵ of turbulent kinetic energy (TKE) within the cloud (see Sect. 1 of the sup-
159 porting information). A micro-pulse lidar (MPL) with a polarization sensitive system
160 and a 5-min temporal resolution allowed for the identification of SLW layers following
161 Alexander and Protat (2018). Radiosondes were launched every six hours - 0530, 1130,
162 1730, 2330 UTC (Sato, Inoue, Alexander, McFarquhar, & Yamazaki, 2018) - and stan-
163 dard meteorological variables were also measured on the ship, including downward short-
164 wave and longwave radiative fluxes. The liquid water path (LWP) was estimated from
165 microwave radiometer data following Marchand et al. (2003).

166 Ice nucleating particles were also measured from aerosol filter collections, as in prior
167 ship campaigns (McCluskey et al., 2018). Cumulative temperature spectra of the num-
168 ber concentration of INPs active via immersion freezing were derived from data collected
169 on the freezing of dilute (purified) water droplet suspensions of collected aerosols using
170 the Colorado State University ice spectrometer instrument system (McCluskey et al., 2018).
171 Details of the instrument methods, clean protocols, calculation of cumulative INPs per
172 volume of suspension, conversion of these to numbers per liter of sampled air versus tem-
173 perature, and calculation of confidence intervals (95%) are discussed in DeMott et al.
174 (2018). Filter samples were 24 or 48 hour collections, representing approximately 21 or
175 42 m^3 of air, respectively. Temperature spectra (six represented) of the INP concentra-
176 tions measured close to Mawson station during MARCUS are plotted in Figure 2d.

177 2.2 WRF simulations

178 This work is based on the version 4.1.1 of the WRF model. The simulation con-
179 figuration follows that used by Vignon, Besic, et al. (2019). The model has been run with
180 a downscaling method where a 27-km resolution parent domain contains a 9-km reso-
181 lution domain which itself contains a $102 \times 102 \text{ km}^2$ nest at a 3-km resolution (see Fig-
182 ure 1). Note that achieving a 3-km resolution is needed to correctly capture the dynam-
183 ics of Antarctic katabatic winds and in particular their coastal transition (Vignon, Traullé,
184 & Berne, 2019; Vignon et al., 2020). All WRF domains have been built with the same
185 polar stereographic projection and they are centered over Mawson station. The nesting

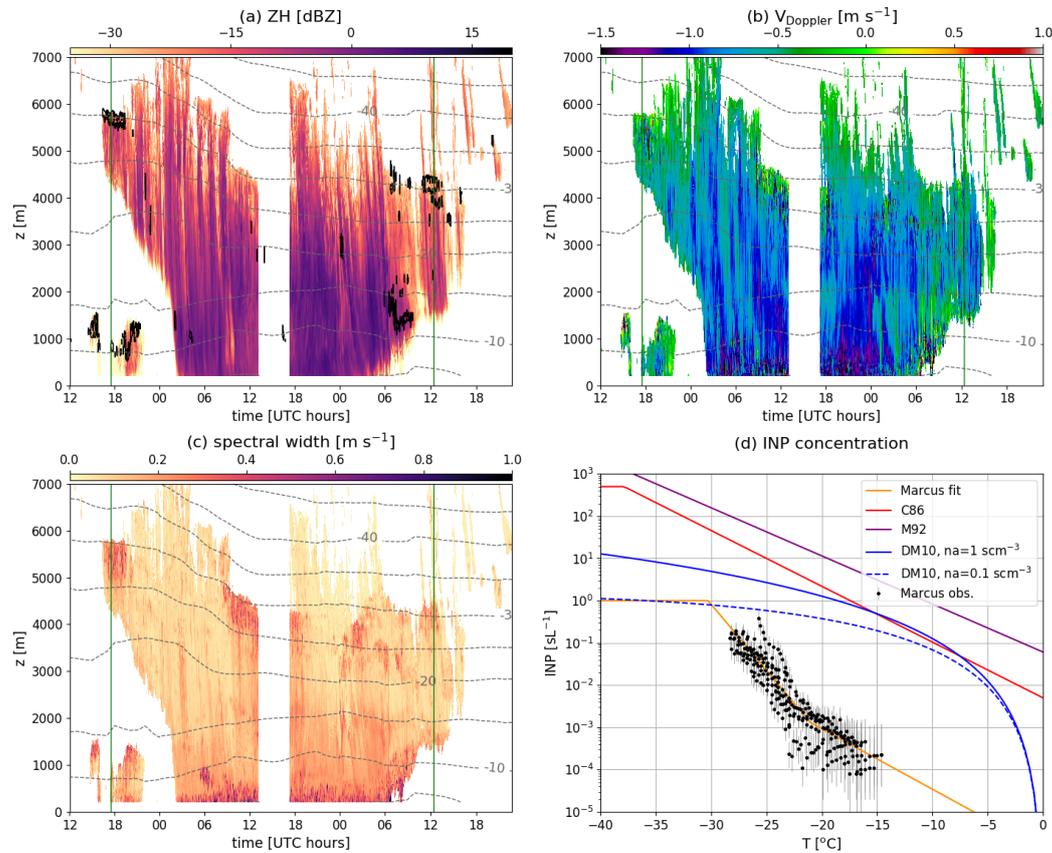


Figure 2. Panel a, b, c: Time height plots of the radar reflectivity ZH, Doppler velocity V_{Doppler} and Doppler spectral width respectively, measured by the MWACR above the ship between the 14 and 16 February 2018. Grey contours indicate the air temperature (5°C intervals) from the ERA5 reanalyses. Vertical green lines indicate the two specific times analyzed in Figure 6. In panel a, black outlines locate regions where the MPL detects SLW. Panel d: Temperature spectrum of the INP concentration. Black dots show measurements off Mawson station during the present MARCUS case study. Errorbars represent the 95% confidence intervals calculated in the same manner as in McCluskey et al. (2018). The orange line shows a fit on the data (see eq. 1). The red line shows the relationship from Cooper (1986) (C86). The purple line shows eq. 2.6 in Meyers et al. (1992) (M92). Blue lines show the DeMott et al. (2010)’s relationship for two extreme values of the concentration of aerosols larger than $0.5 \mu\text{m}$ (na) which commonly ranges between 0.1 and 1 scm^{-3}

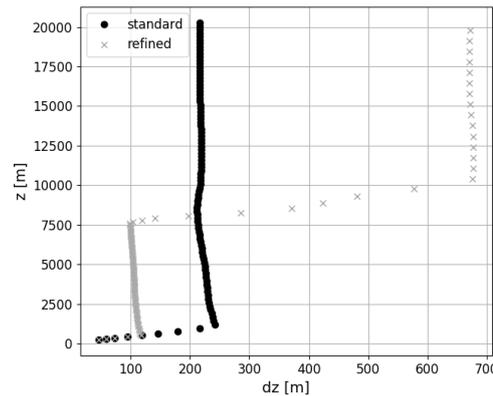


Figure 3. Mean altitude of WRF η levels (z) plotted versus the corresponding layer thickness (dz). Black circles refer to the ‘standard’ 96-level grid. Grey crosses refer to the ‘refined’ 96-level grid with thinner layers in the low- and mid-troposphere.

186 is one way i.e. no information is passed in return from one domain to its parent. Lat-
 187 eral forcings, sea ice concentration, sea surface temperature and initial conditions are from
 188 the ERA5 reanalysis (Hersbach et al., 2020). The topography is from the 1-km resolu-
 189 tion Reference Elevation Model of Antarctica dataset (Howat et al., 2019). The model
 190 is run with 96 vertical levels up to 50 hPa. The so-called ‘standard’ grid (black circles
 191 in Figure 3) is automatically generated by WRF after setting the vertical level number.
 192 It shows layer thicknesses between 200 and 250 m in the mid-troposphere. Using 1D sim-
 193 ulations of mixed-phase altocumulus, A. I. Barrett et al. (2017b) stress that a resolution
 194 of at least 100 m is needed to sustain a SLW layer at cloud top. A so-called ‘refined’ grid
 195 has thus been set-up to refine the vertical resolution in the mid-troposphere to about 100
 196 m at the expense of the representation of the stratosphere (grey crosses in Figure 3).

197 Simulations start on February, 14 2018 00 UTC corresponding to a 17 h spin-up
 198 time before the arrival of the first frontal clouds above the ship location. To allow for
 199 a concomitant comparison between *in situ* observations and simulations and to ensure
 200 a realistic synoptic dynamics in the model, the 27-km resolution domain has been nudged
 201 above the boundary layer towards ERA5 reanalysis for zonal and meridional wind speed,
 202 with a relaxation time scale of 6 h. The nudging only helps provide the best lateral forc-
 203 ing for the free 9-km and 3-km resolution domains. The physics options employed through-
 204 out the study include the new version of the Rapid Radiative Transfer Model for Gen-

205 eral Circulation Models radiation scheme for longwave and shortwave spectra, the Noah
 206 land surface model with adaptations by Hines and Bromwich (2008) and the Mellor-Yamada-
 207 Nakanishi-Niino (MYNN) planetary boundary layer scheme coupled with its associated
 208 surface layer scheme. For the domains with a resolution greater than or equal to 9 km,
 209 the Kain-Fritsch cumulus scheme has been activated. For a proper comparison with MWACR
 210 data, W-band radar reflectivity from WRF outputs has been calculated by means of the
 211 Cloud Resolving Model Radar Simulator (CR-SIM, Oue et al., 2020) version 3.1. CR-
 212 SIM uses the T-matrix method for computing the scattering properties of cloud water,
 213 cloud ice, rain, snow, graupel, and hail hydrometeors. In this study, CR-SIM has been
 214 configured as a virtual MWACR vertically profiling radar - with a frequency of 94 GHz
 215 (close to the 95 GHz frequency of the real instrument) and similar radar beamwidth and
 216 range resolution - that follows the track of the Aurora Australis.

217 *2.2.1 Microphysical scheme setting*

218 We employ the microphysical parameterization from Morrison et al. (2005) which
 219 was shown to produce more realistic amounts of liquid water in Antarctic clouds com-
 220 pared to less advanced WRF parameterizations and also produces realistic precipitation
 221 in coastal Adélie Land (Listowski & Lachlan-Cope, 2017; Hines et al., 2019; Vignon, Besic,
 222 et al., 2019). The scheme has a single-moment treatment of cloud droplets and a double-
 223 moment treatment of cloud ice, rain drops, snow and graupel particles. The activation
 224 of cloud droplets on CCN is not parameterized in the Morrison scheme (except when cou-
 225 pling WRF with its chemical module) and the droplet number concentration is a con-
 226 stant number. We set it to 100 cm^{-3} , a value that reasonably concurs with other stud-
 227 ies over the Antarctic coast and with CCN measurements collected aboard the Aurora
 228 Australis during MARCUS (see Sect. 2 of the supporting information).

229 Regarding primary ice production, tendencies of ice number and mass concentra-
 230 tions associated with homogeneous freezing of droplets (at temperatures $\leq 233.15 \text{ K}$)
 231 and three heterogeneous ice nucleation mechanisms are parameterized. In our control
 232 simulations, immersion freezing of cloud droplets and raindrops is taken into account fol-
 233 lowing the stochastic approach of Bigg (1953). Contact freezing is parameterized as a
 234 flux of contact INP to cloud droplets and the number of contact nuclei is given by Meyers
 235 et al. (1992) (M92). Deposition/condensation freezing nucleation is parameterized as a
 236 nudging term towards an INP concentration predicted as a function of temperature fol-

237 lowing Cooper (1986) (C86). Although our control (ctrl) simulation has been run with
 238 this configuration, the heterogeneous nucleation schemes are questionable for our study
 239 case. First, Bigg (1953)'s scheme based on laboratory data does not specifically account
 240 for ice nuclei and it was shown to be poorly reliable for polar conditions (e.g., de Boer,
 241 Hashino, Tripoli, & Eloranta, 2013; Paukert & Hoose, 2014). Second, except at temper-
 242 atures warmer than about -10°C where contact freezing dominates, the ice production
 243 in the ctrl WRF simulation during MARCUS is dominated by the deposition/condensation
 244 freezing nucleation scheme, but especially at temperatures lower than -15°C (see Fig-
 245 ure S2). Immersion freezing nucleation is thought to be the dominant nucleation mode
 246 in most mixed-phase clouds (Andronache & coauthors, 2017). It is likely that this mode
 247 is represented in the mixed-phase cloud observations from C86 that are parameterized
 248 as deposition/condensation freezing in WRF, but the number concentrations are repre-
 249 sentative of the mid-latitude, continental regions where the observations were primar-
 250 ily collected. Indeed, the INP concentration prescribed in the C86's deposition nucle-
 251 ation scheme is much higher than the measured INP concentration in the immersion freez-
 252 ing mode for the Mawson region at the time of this case study (Figure 2d). This excess
 253 of INP also impedes the generation of SLW and of all subsequent freezing processes.

254 As underlined by O'Shea et al. (2017), C86 and M72 parameterizations were de-
 255 veloped for continental conditions in which the INP concentrations are several orders of
 256 magnitude higher than in the pristine atmosphere above the Southern Ocean (DeMott
 257 et al., 2016; Kanji et al., 2017). DeMott et al. (2010) further developed an INP param-
 258 eterization using not only the temperature but also the concentration of aerosols. This
 259 parameterization better predicts the ice crystal number concentration present in clouds
 260 over the Antarctic Peninsula than C86 or M92 (Listowski & Lachlan-Cope, 2017). How-
 261 ever it overestimates the INP concentration off Mawson station (Figure 2d) and using
 262 it instead of C86's formulation only - as in Young et al. (2019) - decreases the ice nu-
 263 cleation rate but maintains ice formation at temperatures lower than -20°C (see Figure
 264 S2).

265 We thus replaced all the heterogeneous nucleation parameterizations in the Mor-
 266 rison microphysical scheme with a unique empirical one - reflecting immersion freezing
 267 - in the manner of Paukert and Hoose (2014). Note that the Bigg's parameterization is
 268 nonetheless kept active for the freezing of big rain drops. INP measurements during MAR-
 269 CUS have first been fitted with the following equation (see orange line in Figure 2d):

$$\log(N_{INP}) = \begin{cases} -0.14(T - T_2) - 2.88, & \text{if } T > T_1 \\ -0.31(T - T_1) - 2.88, & \text{if } T_2 \leq T \leq T_1 \\ 0.0 & \text{if } T < T_2 \end{cases} \quad (1)$$

270 with N_{INP} the INP number concentration in sL^{-1} , T the temperature in $^{\circ}\text{C}$, $T_1 =$
 271 -21.06 $^{\circ}\text{C}$ and $T_2 = -30.35$ $^{\circ}\text{C}$. INP measurements were performed at $T > -28$ $^{\circ}\text{C}$
 272 questioning extrapolation of the curve at very low temperatures. Here, we taper the ex-
 273 ponential increase with decreasing temperature and constrain N_{INP} not to exceed 1 sL^{-1} ,
 274 a value close to the prediction from the DeMott et al. (2010)’s parameterization for
 275 low aerosol concentrations (Figure 2d). Setting such a threshold is motivated by recent
 276 measurements during the CAPRICORN campaign over the Southern Ocean in McCluskey
 277 et al. (2018). The authors revealed that the INP concentration in the immersion mode
 278 no longer increases with decreasing temperature - staying below 1 sL^{-1} when temper-
 279 ature is lower than about -28°C . Similar behavior has been observed for other geograph-
 280 ical contexts (Kanji et al., 2017).

Then, the ice crystal production term follows the equation:

$$\left. \frac{dN_i}{dt} \right|_{nucleation} = \begin{cases} \frac{N_{INP} - (N_i + N_s + N_g)}{\Delta t}, & \text{if } N_{INP} > N_i + N_s + N_g \\ 0.0 & \text{otherwise} \end{cases} \quad (2)$$

281 where Δt the model timestep and N_i , N_s and N_g the number concentration of ice
 282 crystals, snowflakes and graupel particles respectively. As this empirical parameteriza-
 283 tion reflects immersion freezing, the produced mass of cloud ice is removed from cloud
 284 liquid water. It is worth noting that this new ice nucleation parameterization is based
 285 on INP measurements in the boundary-layer off Mawson station. 5-day back-trajectories
 286 revealed that the air parcels arriving in the mid-troposphere above the ship during the
 287 study case mostly originate from the north and west of the station and has been lifted
 288 from the marine boundary-layer in the vicinity of the station (see Figure S3). The present
 289 nucleation scheme should therefore be reasonably valid in both boundary-layer and mid-
 290 level frontal clouds.

291 Furthermore, the Morrison scheme accounts for secondary ice production through
 292 the rime-splintering process (Hallett-Mossop) in the $[-8^{\circ}\text{C}, -3^{\circ}\text{C}]$ temperature range.
 293 However, Young et al. (2019) show that this process should be artificially enhanced by

294 a factor of 10 to reproduce the observed ice crystal concentrations over the Weddell Sea.
 295 Sotiropoulou et al. (2020) suggest that it may be due to the absence of parametrization
 296 for the secondary ice production through ice particle break-up after hydrometeor colli-
 297 sion. By default in our simulations we do not activate a parameterization of collisional
 298 break-up but complementary sensitivity experiments have been carried out.

299 **2.2.2 Cloud top turbulence parameterization**

300 SLW layers at cloud top are a few hundreds of meters deep - i.e. of comparable width
 301 or even thinner than common atmospheric model layers - and they are characterized by
 302 a vigorous turbulence that is critical to generate and maintain the SLW. This turbulence
 303 should be represented in models. However, cloud tops are regions of sharp vertical gra-
 304 dients of atmospheric properties which are difficult to simulate with the current verti-
 305 cal resolutions of models. The turbulent mixing at cloud top - or entrainment - has been
 306 and is still an active subject of research especially for warm stratocumulus found over
 307 the tropical oceans (e.g., Stevens, 2002; Mellado, 2017). In particular, representing the
 308 buoyancy flux and the subsequent top-down convection associated with cloud top radia-
 309 tive cooling and to a lesser extent, with the evaporation or sublimation of condensates
 310 (see for instance large eddy simulation studies in Brient, Couvreur, Villefranche, Rio,
 311 & Honnert, 2019), requires specific parameterizations (Lenderink & Holtslag, 2000).

Some studies using 1-order turbulent mixing schemes proposed to adapt the ver-
 tical profiles of the eddy-diffusivity coefficient between the ground and the cloudy boundary-
 layer top depending on the radiative and evaporative cooling (Lock et al., 2000; Wilson,
 2015; Ghonima et al., 2017). However, such schemes do not properly apply for mid-tropospheric
 clouds. In this study, we follow the approach of Guo et al. (2019) based on the pioneer-
 ing ideas of Deardoff (1972), Lock (1998) and Grenier and Bretherton (2001). This study
 includes a specific parameterization for the TKE production term associated with the
 buoyancy flux at the top of a liquid cloud. Briefly, this parameterization accounts for
 the buoyancy flux associated with the fraction of the radiative flux divergence that is not
 explicitly resolved by the model due to its too coarse vertical resolution. This additional
 TKE production term P_R can read:

$$P_R = \mathcal{F}(q_c, p) \frac{g}{\theta_v} \frac{\Delta_z F_{LW} \Delta z}{c_p \rho \Pi} \quad (3)$$

where g is the acceleration of gravity, θ_v is the virtual potential temperature, ρ is the air density, c_p is the air heat capacity, Π is the Exner function, $\Delta z F_{LW}$ is the longwave radiative flux vertical divergence at cloud top and Δz is the cloud top model layer depth. $\mathcal{F}(q_c, p)$ is a function of the cloud liquid water content q_c and pressure p and is bounded between 0 and 1. Because estimating \mathcal{F} for a mixed-phase cloud would be much more complex, we decide to follow a simplified approach:

$$P_R = \phi \frac{g}{\theta_v} \frac{\Delta z F_{LW} \Delta z}{c_p \rho \Pi} \quad (4)$$

312 with ϕ is tuning coefficient ranging between 0 and 1. By default, we set $\phi = 0.05$
 313 (value that gives reasonable cloud top liquid content and turbulence, see next section)
 314 but the sensitivity to this value will be assessed.

315 **3 Results**

316 **3.1 Brief description of the evolution of clouds and precipitation from** 317 **observations**

318 The synoptic conditions at 00 UTC, 15 February 2018 in the ctrl WRF simulation
 319 are plotted in Figure 1. A synoptic weather system manifesting as a minimum of 500-
 320 hPa geopotential height sets at the north-west of Mawson, advecting warm and moist
 321 oceanic air towards the ice sheet along its eastern flank. In particular, a zonally elon-
 322 gated tongue of integrated condensed water content (shading) is moving towards the sta-
 323 tion and the ship (blue dot). This tongue preceding a warm sector (temperature in cyan
 324 contours) corresponds to the warm front of the system. During the 15 and 16 February,
 325 the warm front moves to the south-east of the station and dissipates. The ship thus en-
 326 ters the warm sector while the cold front remains far from the coast over the Southern
 327 Ocean. Meanwhile, the extra-tropical cyclone progressively weakens at the west of Maw-
 328 son and disappears during the second half of the 16 February.

329 Figure 2 shows the time-height plot of the MWACR reflectivity (panel a), Doppler
 330 velocity (panel b) and Doppler spectral width (panel c) above the ship during the event.
 331 Note that the radar ceased functioning between 13 and 17 UTC, 15 February. In panel
 332 a, black contours indicate regions identified as SLW cloud layers using the MPL data.
 333 Panel a indicates a pre-precipitation virga period (16 UTC, 14 February to 02 UTC, 15
 334 February) during the arrival of the warm front above the ship and is characterized by

335 significant reflectivity values in altitude but not at the surface. This period is followed
 336 by actual surface precipitation within the warm sector - with high reflectivity values at
 337 the first radar gate - which is followed by a post-precipitation phase (06 to 17 UTC, 16
 338 February) when the extra-tropical cyclone dissipates. Such temporal structure (pre-precipitation
 339 virga, surface precipitation, post-precipitation virga) associated with the passage of a
 340 warm front above the station was shown to be representative of the precipitation events
 341 affecting the coast of Adélie Land (Jullien et al., 2020), East Antarctica. From the li-
 342 dar data, clear SLW layers are particularly identified:

- 343 1. at the top of boundary-layer stratocumulus upstream of the warm front in the cool
 344 sector, within the first 1500 m a.s.l. and between 15 and 22 UTC, 14 February;
- 345 2. at the top of the first high frontal clouds (altocumulus), just above pre-precipitation
 346 iced-virga between 17 and 21 UTC, 14 February;
- 347 3. at the end of the event, sitting on top of post-precipitation ice virga between 11
 348 and 13 UTC, 16 February;

349 The Doppler velocity field shows that where SLW is present, weakly-negative or even pos-
 350 sive values of the mean vertical velocity are measured (see Figure 2b and the Doppler
 351 velocity distribution conditioned to SLW patches in Figure S1b). Below SLW layers, one
 352 can point out rapid alternations of strongly and weakly negative Doppler velocities. Sim-
 353 ilarly, the Doppler spectral width - that strongly depends on turbulence - exhibits large
 354 values within and in the few hundred meters below SLW layers (Figure 2c). The creation
 355 and resilience of SLW at the top of the frontal mixed-phase clouds thus appears related
 356 to the dynamics of cloud-top convective cells (A. V. Korolev & Mazin, 2003; A. Korolev
 357 et al., 2017) as within mid-latitude altocumulus (Heymsfield et al., 1991; Smith et al.,
 358 2009; P. A. Barrett et al., 2020). It is important to note that the lidar signal may be to-
 359 tally attenuated by precipitation, so there could have been SLW between 15 February
 360 02 UTC and 16 February 06 UTC. As a matter of fact, the highest LWP values estimated
 361 from the microwave radiometer were measured between 09 and 19 UTC 15 February (see
 362 next section). This suggests the presence of SLW layers or patches within or at the top
 363 of the deep nimbostratus during this period, especially within or at the summit of lay-
 364 ers with both high values of Doppler velocity and Doppler spectral width. The visual
 365 inspection of Doppler spectra indeed confirms the occurrence of elevated SLW layers dur-
 366 ing the precipitation period (Figure S4).

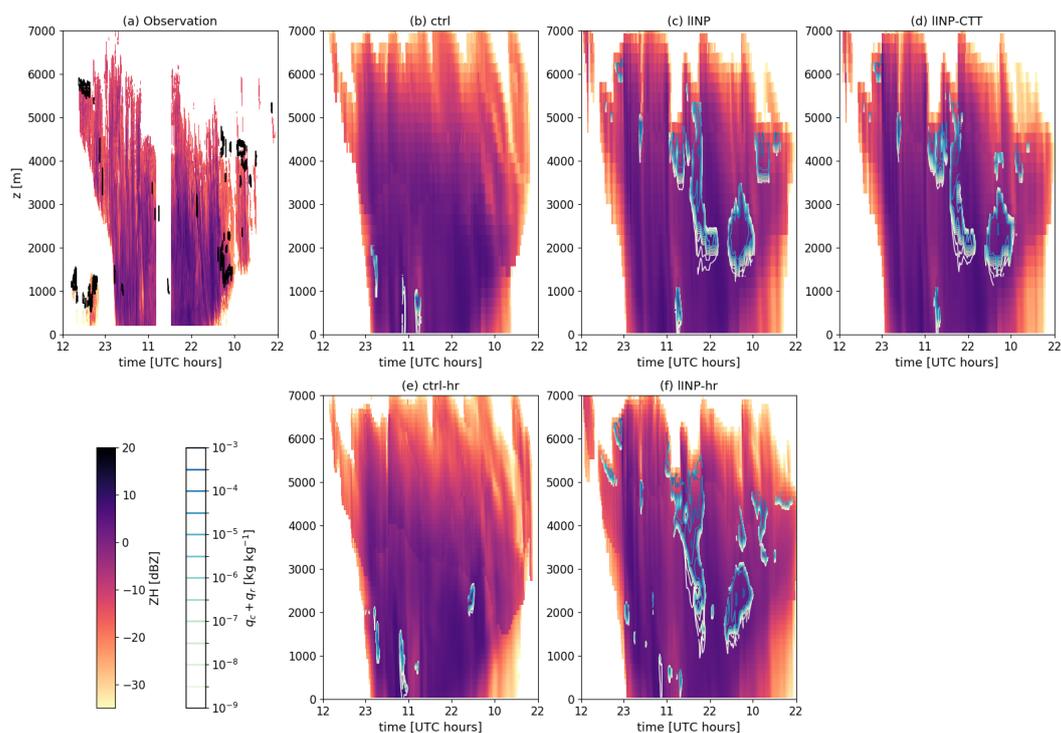


Figure 4. Time-height plot of the W-band reflectivity in MWACR observation (panel a) and as calculated from WRF simulations with the CR-SIM radar simulator (panels b-f). In panel a, black outlines locate regions where the MPL detects SLW. In panels b-f, yellow-to-blue contours show the mass mixing ratio of cloud liquid water (sum of cloud and rain droplets).

367

3.2 Simulating the vertical structure of liquid-topped frontal mixed-phase clouds

368

369

We now assess the ability of WRF to reproduce the observed cloud vertical structure. Unlike the control (ctrl) simulation with the standard Morrison microphysical scheme, simulations using the empirical high-latitude Southern Ocean ice nucleation parameterization with a lower - but more realistic - INP concentration, are named ‘IINP’. Simulations accounting for the cloud top turbulence parameterization are named with the ‘-CTT’ suffix. Likewise, simulations run with the refined vertical grid in the troposphere are designated with the ‘-hr’ suffix.

375

376

Figure 4 shows that the ctrl simulation reproduces the timing and the overall structure of the system reasonably well. However the local low-level clouds preceding the passage of the warm front are absent in the simulation above the ship position but similar

377

378

379 local clouds form a few kilometers away (not shown). In addition, the model generally
380 overestimates the cloud top height particularly owing to the excessive ice nucleation at
381 cold temperatures. More importantly, Figures 4b and 5m show that the ctrl configura-
382 tion produces almost fully glaciated clouds and refining the vertical grid in the mid-troposphere
383 (ctrl-hr simulation) barely improves the production of liquid droplets. Note that chang-
384 ing the microphysical scheme to the one from Thompson et al. (2008) - that together with
385 the Morrison scheme yields the best cloud liquid water content and surface radiative fluxes
386 in previous Antarctic studies with WRF (Listowski & Lachlan-Cope, 2017; Hines et al.,
387 2019) - leads to the same conclusion (not shown). Note also that replacing the INP for-
388 mulation with the one from DeMott et al. (2010) in the deposition/condensation freez-
389 ing nucleation parameterization leads to slightly more SLW in the lowest part of the clouds
390 (where the temperature is greater than -15°C) but its overall amount remains strongly
391 underestimated. It is also worth mentioning that unlike WRF in its standard configu-
392 ration, the recent ERA5 reanalysis produces some cloud liquid content during this event,
393 but not the correct amount nor at the correct location (at too low altitude and too warm
394 temperature, see Sect. 3 of the supporting information).

395 The empirical INP formulation leads to a significant increase in cloud liquid wa-
396 ter content throughout the event (Figures 4c and 5m). In addition, the IINP simulation
397 exhibits sharp vertical gradients of condensate mixing ratio in the uppermost part of the
398 clouds. Subsequently, the cloud top radiative cooling is stronger and the resulting buoy-
399 ancy flux triggers and enhances significant turbulence that is absent in the ctrl simula-
400 tion (Figure 6).

401 Panels a-f and g-l of Figure 5 show vertical profiles of atmospheric variables for two
402 particular times with clear liquid-topped altocumulus identified in observations (see ver-
403 tical green lines in Figure 1a-c). During the arrival of the warm front at 1730 UTC, 14
404 February 2018, the new INP parameterization makes WRF able to reach the saturation
405 with respect to liquid in a layer around 5600 m whatever the vertical resolution employed.
406 A thin SLW layer is therefore simulated at cloud top but its height is slightly underes-
407 timated compared to lidar observations. Below this layer, ice crystals grow by vapor de-
408 position and sediment (Figure 7). The presence of liquid droplets at cloud top also en-
409 hances the radiative cooling, leading to an almost neutral vertical profile of potential tem-
410 perature in agreement with radiosonde observation (Figure 5a). However, with the coarse
411 vertical resolution employed in IINP, the liquid layer does not persist in time. When re-

412 fining the vertical resolution (IINP-hr simulation), the resilience of the SLW layer dur-
413 ing the warm front arrival (Figure 4f) is better reproduced - in agreement with the 1D-
414 simulations of A. I. Barrett et al. (2017b). The altitude of the liquid layer gradually de-
415 creases owing to the drying effect associated with cloud-top turbulent entrainment.

416 The IINP simulation does not reproduce the cloud top turbulence during this spe-
417 cific period (Figure 5f) but IINP-hr exhibits both a resilient SLW layer and vigorous mix-
418 ing. Activating the additional parameterization for cloud top turbulence enhancement
419 in the low-resolution configuration (IINP-CTT simulation) helps generate turbulence in
420 the upper part of the altocumulus. A similar conclusion can be drawn for the middle phase
421 of the event (around 1200 UTC, 15 February, see Figure 6d). However, this parameter-
422 ization does not improve the persistence of the SLW layer through time. Turbulence tends
423 to thin the SLW layer out by mixing it with underlying and overlying drier air. As ex-
424 pected, increasing the ϕ parameter increases the TKE and ϵ but for $\phi \geq 0.1$, the mix-
425 ing becomes too intense - with respect to the vertical resolution used - for SLW to sur-
426 vive over more than a few time steps (see Figure S6).

427 Similarly to the IINP-CTT simulation, the turbulence in the IINP-hr simulation
428 tends to thin the SLW layer by mixing with drier air (Figure 5d). Interestingly, SLW does
429 not form in the atmospheric layer where turbulent mixing cools the air (Figure 7). This
430 is somewhat contradictory with the conceptual model of supercooled droplet condensa-
431 tion within adiabatically cooled turbulent updrafts (P. A. Barrett et al., 2020). This as-
432 pect will be discussed in Sect. 4.3.

433 Analysis of profiles at 1230 UTC, 16 February (Figure 5g-l) generally concurs with
434 our main inferences regarding the performances of WRF at the beginning of the event.
435 We can still notice the absence of turbulence between 3000 and 4600 m in the IINP-hr
436 simulation which is explained by the cloud being too deep (see Figure 4f and Figure 5h)
437 - so an overestimated cloud top height and underestimated radiative cooling between 4000
438 and 4500 m (Figure 3k) - at this specific time. We do not have a clear explanation for
439 this bias but it seems that the deep nimbostratus stays too long over the ship location
440 and, interestingly, a thin SLW layer at around 4600 m is simulated during the end of the
441 16 February (Figure 4f). In absence of turbulence, the SLW layer in the IINP and IINP-
442 hr simulations - at this specific time - is too thick. It becomes more realistic later in the
443 day (see Figure 4).

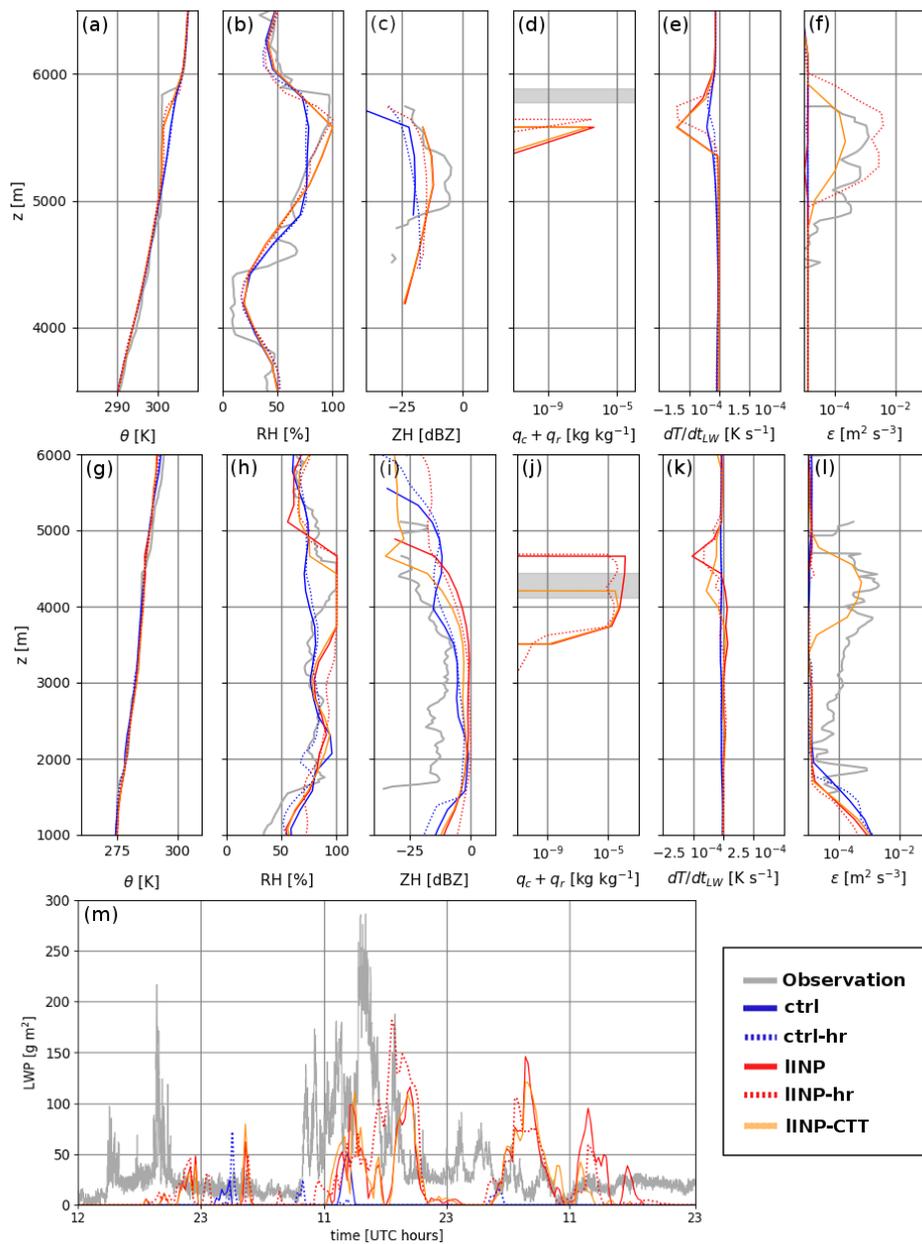


Figure 5. Panels a-l: Vertical profiles of the potential temperature (a and g), relative humidity with respect to liquid (b and h), W-band radar reflectivity (c and i), liquid water content (sum of cloud droplets and rain drops, d and j), temperature tendency due to longwave radiative warming (e and k) and rate of turbulent kinetic energy dissipation (f and l) in observations (grey lines) and WRF simulations. Panels a-f refers to the 14 February 2018 at 1730 UTC while panels g-l refers to the 16 February 2018 at 1230 UTC. In panels a, b, g and h, observational data are from the closest-in-time radiosounding. In panels c, f, i and l, observations are from MWACR data. In panels d and j, the grey shading indicates the altitude range where the MPL detects SLW. Panel m: 14-16 February 2018 LWP time series above the Aurora Australis position from radiometer estimations (grey line) and from WRF simulations.

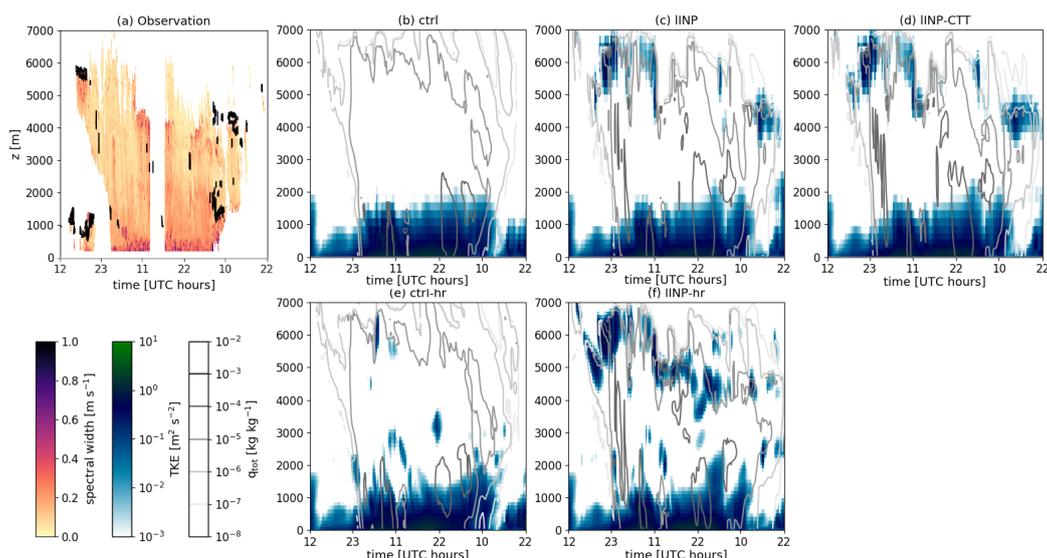


Figure 6. Panel a: Time-height plot of the spectral width in MWACR observation. Black outlines locate regions where the MPL detects SLW. Panels b-f: Time height plots of the TKE (color shading) and of the mass mixing ratio of cloud condensates (contours, sum of cloud droplet, cloud ice, snow, graupel and rain species, q_{tot}) above the ship position for different WRF simulations.

444 As previously mentioned, Sotiropoulou et al. (2020) suggest that secondary ice pro-
 445 duction through ice particle collisional break-up might be an important process in coastal
 446 Antarctic clouds. We have assessed the model sensitivity to this process on our study
 447 case (details in Sect. 4 of the supporting information). Collisional break-up significantly
 448 modifies the ice particle number concentration at temperature greater than -25°C but
 449 the available observational dataset does not enable us to state whether this is truly ben-
 450 efiticial to our simulations or not. In any case, this parameterization is not detrimental
 451 to the simulation of SLW layers which is our main scope here.

3.3 Cloud radiative effect

452
 453 Achieving the simulation of SLW layers substantially impacts the cloud radiative
 454 effect (CRE) during the event. Figure 8 shows the difference in CRE averaged over the
 455 whole study case between the IINP-hr (the configuration with the most realistic SLW
 456 layers) and ctrl simulations. At the top of the atmosphere (TOA), the IINP-hr simula-
 457 tion exhibits more reflected shortwave radiation than the ctrl simulation (panel a), es-

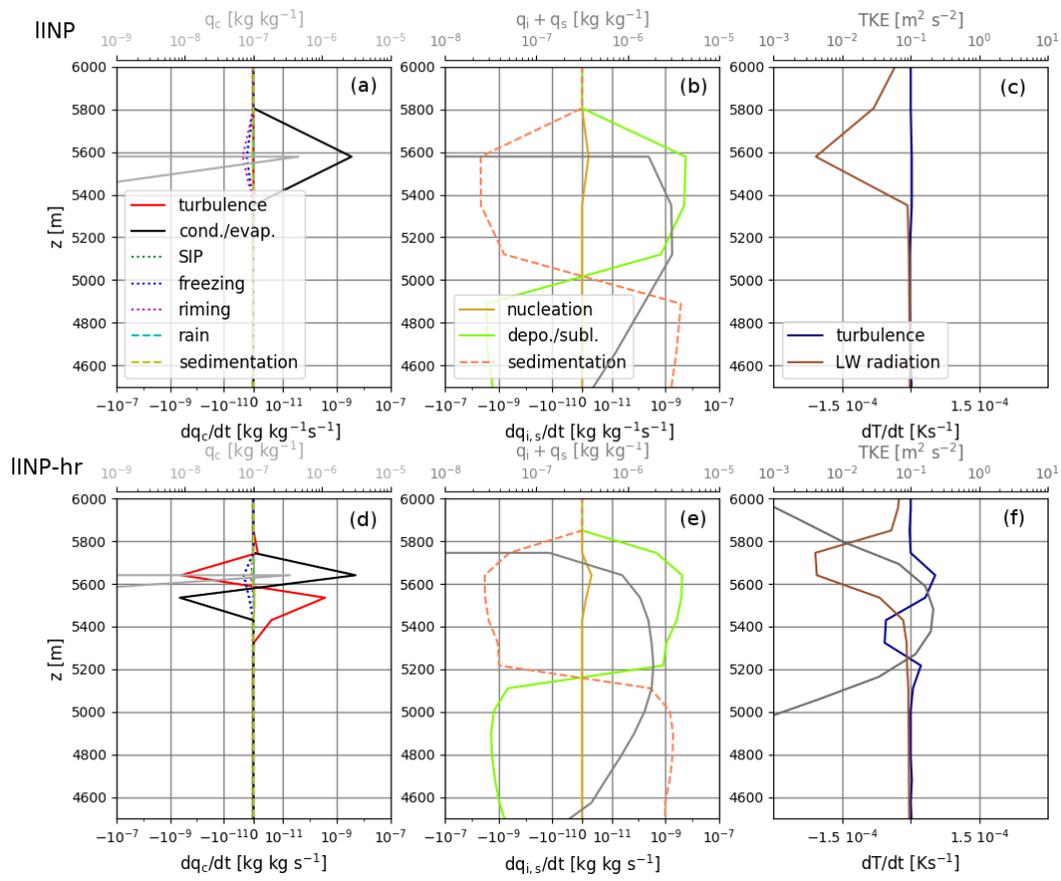


Figure 7. Vertical profiles at 1730 UTC, 14 February 2018, of different model variables for the IINP (panels a-c) and IINP-hr (panels d-f) WRF simulations. Panels a and d show the cloud liquid water content (q_c , light grey line, top x-axis) and the different source/loss terms of cloud liquid water (colored lines, note the logarithmic scale on x-axis). 'turbulence' refers to turbulent mixing; 'cond./evap.' refers to droplet condensation or evaporation in a saturated/unsaturated atmosphere; 'SIP' refers to secondary ice production by splintering of droplets accreted on iced hydrometeors; 'riming' refers to the riming of iced precipitation; 'freezing' refers to the ice-nucleation through freezing (loss term for droplets), 'rain' refers to the autoconversion to rain and 'sedimentation' refers to the sedimentation of droplets. Panels b and e show the mass mixing ratio of the ice and snow species ($q_i + q_s$, grey line, top x-axis) and the relative tendencies due to ice nucleation (solid gold line), vapor deposition or sublimation (solid green line) and sedimentation (dashed orange line). Panels c and f: TKE (dark grey line, top x-axis), longwave radiative (brown) and turbulent (blue) heating rates. Note that the model does not simulate any TKE in panel c.

458 pecially over the Southern Ocean because of an increase in cloud albedo, while the albedo
 459 discrepancy over snow and ice covered areas over the continent is less significant. This
 460 increase in albedo is also responsible for a decrease in the amount of shortwave radia-
 461 tion that reaches the ground surface (Figure 8d). On the other hand, the outgoing long-
 462 wave radiative flux towards space diminishes due to colder cloud tops. Importantly, as
 463 liquid-bearing clouds are optically thicker, the IINP-hr simulation shows a much higher
 464 downward radiative flux, leading to a significant surface warming over the ice sheet sur-
 465 face with respect to the ctrl simulation (panel f). The same conclusions can be drawn
 466 for the IINP and IINP-CTT simulations. It is also worth noting that although our new
 467 parameterizations targeted mid-level clouds, inspection of vertical profiles of cloud prop-
 468 erties over the whole simulation domain shows that boundary-layer clouds are also - but
 469 to a lesser extent because of the warmer temperatures at lower altitude - modified with
 470 higher SLW content. Figure 8 thus integrates combined effects from changes on both mid-
 471 level and low-level clouds. Comparison of the surface downward longwave radiative flux
 472 above the ship also reveals a better agreement when the new INP parameterization is
 473 activated. The mean downwelling longwave flux value between 1200 UTC, 14 February
 474 and 2200 UTC, 16 February equals 292.2 W m^{-2} in the observations, and 227.5 W m^{-2} ,
 475 237.6 W m^{-2} , 238.0 W m^{-2} , 241.1 W m^{-2} in the ctrl, IINP, IINP-CTT and IINP-hr sim-
 476 ulations respectively. The value is however significantly underestimated in all the sim-
 477 ulations. Inspection of flux time series (not shown) reveals that this is mostly due to the
 478 absence of local low-level clouds just above the ship position and preceding the warm
 479 front. Such clouds indeed have a particularly strong warming effect. Comparison with
 480 model grid points in the vicinity of the ship that contains low-level clouds shows a sub-
 481 stantially higher (up to 40 W m^{-2}) downward longwave radiative flux at the arrival of
 482 the warm front.

483 **4 Discussion**

484 **4.1 Remaining shortcomings in our simulations**

485 Despite improvements regarding the simulation of SLW layers, shortcomings remain
 486 in our simulations whatever the physical configuration used. Amongst the most strik-
 487 ing biases, Figures 4 and 5h evidence an insufficient low-level sublimation during the last
 488 day of the event that is associated with an overestimation of the relative humidity when
 489 comparing with radiosoundings. This aspect can be improved when accounting for sec-

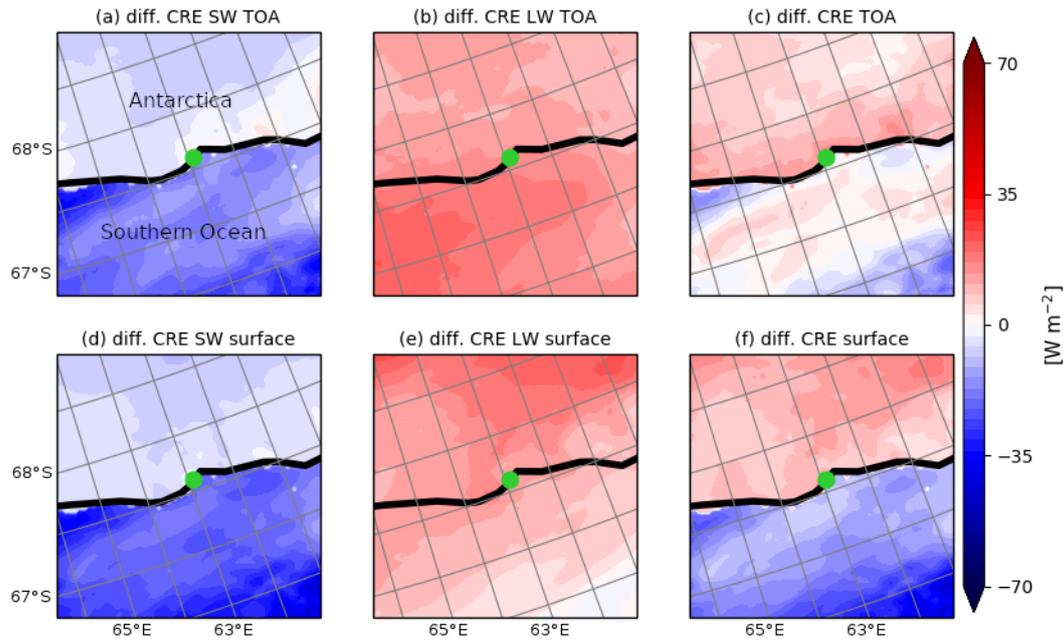


Figure 8. Difference in cloud radiative effect (CRE) averaged over the whole duration of the study case at the top of the atmosphere (TOA, top row) and at the surface (bottom row) between the IINP-CTT and ctrl WRF simulations (3-km resolution innermost domain). Panels a and d show the shortwave (SW) component, panels b and e the longwave (LW) component, and panels c and f the total difference. The black line is the Antarctic landfall and the green dot locates Mawson station.

490 ondry ice production through ice particle break-up (see Sect. 4 of the supporting in-
491 formation).

492 Figure 5m also reveals an overall underestimation of the LWP in all the simulations
493 as well as issues concerning the timing of the LWP peaks. Changing the intensity of the
494 27-km domain nudging or adding a nudging term on the temperature and/or the humid-
495 ity fields did not alleviate those biases (not shown). A bias propagation from the ERA5
496 forcings into our inner simulation domains can thus not be excluded. Moreover, the ab-
497 sence of lidar measurements during the strong precipitation phase that coincides with
498 the highest LWP values prevents us from precisely evaluating the SLW representation
499 during this period. Warm frontal systems often exhibit SLW layers or patches within deep
500 nimbostratus associated with embedded convective cells (Keppas et al., 2018). The MWACR
501 data shows high values (in magnitude) of the Doppler velocity and of the Doppler spec-
502 tral width at the top of the nimbostratus between 9 and 13 UTC, 15 February, suggest-
503 ing that intense cloud-top turbulent updrafts may explain a significant part of the SLW
504 production during this period. In addition to turbulence, Gehring et al. (2020) show that
505 within a nimbostratus over Korea, the large scale ascent corresponding to the warm con-
506 veyor belt of an extra-tropical cyclone can be sufficient to create and sustain SLW. Com-
507 paring Figure 4 and Figure 6 shows that the IINP and IINP-hr simulations also exhibit
508 SLW patches in the middle of the nimbostratus (between about 2000 and 3500 m) i.e.
509 in an altitude range with low values of TKE in the model and low values of spectral width
510 in the MWACR data. Further inspection of the resolved vertical velocity field in the model
511 reveals that those SLW patches coincide with significant ascents (around $+0.1 - 0.2 \text{ m s}^{-1}$,
512 not shown) but their realism cannot be assessed by comparison with our observational
513 dataset. Hence, it remains difficult to disentangle whether the remaining biases in SLW
514 quantity and timing during the precipitation period are due to a poor representation of
515 the turbulence at the top of the nimbostratus and/or to the modeling of the large-scale
516 ascent associated with synoptic dynamics.

517 **4.2 INP, turbulence, vertical resolution: what matters the most for achiev-** 518 **ing the simulation of SLW layers?**

519 Our results highlight that without a realistic ice nucleation parameterization that
520 accounts for the particularly low INP concentration over the high-latitude Southern Ocean,
521 the representation of thin turbulent SLW layers and realistic SLW contents cannot be

522 achieved. This conclusion holds whatever the vertical resolution tested, with or without
523 additional subgrid turbulent mixing at cloud top. From the present analysis, the nature
524 of the heterogeneous ice nucleation parameterization in atmospheric models, especially
525 the representation of the limited INP numbers concentrations over this region, is an es-
526 sential prerequisite to simulate the liquid phase in frontal mid-level mixed-phase clouds
527 at high southern latitudes. Furthermore, it makes the model produce significant TKE
528 near cloud top - which is missing in the ctrl simulation - due to enhanced radiative di-
529 vergence. In our IINP and IINP-hr simulations, the persistence of the saturation with
530 respect to liquid - and of the resulting SLW layer - mostly depends on a subtle compe-
531 tition between air cooling (primarily due to radiative divergence, see Figure 7) and mois-
532 ture removal associated with the growth of ice crystals. Increasing the vertical resolu-
533 tion usually helps maintain the saturation because newly formed crystals get more eas-
534 ily separated from the liquid layer while falling. One can refer to A. I. Barrett et al. (2017b)
535 for further discussion on the link between SLW resilience and model vertical resolution.
536 In IINP, although SLW continues to form at 1730 UTC, the ice particle growth (Figure
537 7) makes the air under-saturated with respect to liquid after a few minutes. When the
538 liquid layer disappears, the precipitating ice crystals falling towards the lower layer are
539 not replaced by newly formed crystals and the total cloud water content decreases. The
540 reappearance of SLW becomes impossible if other moistening processes (through advec-
541 tion for instance) do not come into play or until the temperature reaches the dew point
542 through radiative cooling. In contrast in IINP-hr, the atmospheric layer between 5600
543 and 5750 m shows lower ice crystal concentration, a weaker vapor deposition on ice and
544 significant radiative cooling (Figure 7), enabling the persistence of the SLW layer for sev-
545 eral hours.

546 Regarding the representation of turbulence, the underestimation of the occurrence
547 and intensity of cloud top mixing at coarse vertical resolution could be anticipated and
548 motivated the implementation of an additional source term in the TKE equation. The
549 latter parameterization leads to better agreement with ϵ estimations from Doppler radar
550 measurements during the front arrival and during the course of the event. However and
551 unlike the increase in vertical resolution, this parameterization does not help sustain the
552 SLW layer and conversely it can amplify its depletion if the ϕ coefficient is set to a too
553 high value. This apparent second role of turbulence for SLW resilience may be co-incidental
554 since the state-of-the-art MYNN local turbulent mixing scheme is likely inadequate for

555 reproducing the top-down convection at mid-level cloud top. This may even question the
 556 physical representation of cloud droplet formation and growth in the model (see next sec-
 557 tion).

558 **4.3 The pressing need of revisiting the parameterization of cloud top**
 559 **turbulence**

560 One aspect that particularly deserves further discussion is the representation of cloud
 561 top turbulence in the model. We have shown that in some cases, an additional source
 562 term in the TKE equation, compensating for the incomplete reproduction of the radia-
 563 tive cooling, helps obtain some TKE at cloud top. However the *local* TKE generation
 564 by buoyancy fluxes in the IINP-CTT and IINP-hr (and to a lesser extent in the IINP)
 565 simulations lead to a patch of TKE (or ϵ) that is vertically centered around cloud top
 566 liquid and that unrealistically diminishes the temperature inversion (Figure 5a,f and 7f).
 567 Even though our estimation of ϵ only applies where the radar detects signal in the cloud,
 568 the sharp temperature inversion in the observations suggests that turbulent motions mostly
 569 occur within and below the cloud. Using turbulence data from aircraft measurements,
 570 P. A. Barrett et al. (2020) show that the TKE maximum occurs several hundred meters
 571 below typical mixed-phase altocumulus top. Indeed, the turbulence structure within al-
 572 tocumulus consists of shallow small-scale eddies at cloud top below which an organized
 573 Rayleigh Bénard-type convection takes place with negatively buoyant air parcels that
 574 descend through the cloud layer in coherent downdrafts and force upward motion through
 575 mass continuity (Schmidt et al., 2014; P. A. Barrett et al., 2020). Subrotor circulations
 576 associated with ice virga shafts may also participate in the mixing below the cloud. Over-
 577 all the organized convection triggered at cloud-top cannot be represented by the typi-
 578 cal local turbulent mixing schemes used in atmospheric models like MYNN or all the cur-
 579 rent 1.5-order planetary boundary layer schemes in WRF. Moreover, the adiabatic cool-
 580 ing, the saturation with respect to liquid and the growth and vertical transport of droplets
 581 only occurs within updrafts. Considering each model layer as homogeneous in terms of
 582 temperature and humidity necessarily prevents the proper representation of the dynam-
 583 ics of turbulent mixed-phase clouds. Albeit satisfactory compared to simulations with
 584 the standard version of WRF, the representation of SLW layers in the IINP-hr config-
 585 uration may result from a partially non-physical interplay between turbulence and mi-
 586 crophysics. Adapting a non-local turbulent mixing parameterization based on a mass-

587 flux scheme that treats separately a 'lifting' fraction and a 'subsiding' fraction of each
 588 mesh (see Hourdin et al., 2019 for instance) might be an interesting approach to tackle
 589 this issue in the future. Such types of scheme are already active in many atmospheric
 590 models to parameterize the mixing in convective ground-based boundary layers but they
 591 are not active aloft.

592 **5 Conclusions**

593 By using remotely-sensed measurements obtained during the MARCUS campaign,
 594 we have evaluated the ability of the WRF regional atmospheric model to reproduce the
 595 thin and turbulent layers of SLW at the top of frontal mixed-phase clouds over the high-
 596 latitude Southern Ocean.

597 While the control simulation did not exhibit any cloud liquid water above the bound-
 598 ary layer, we found that modifying the ice nucleation parameterization through the im-
 599 plementation of a truly representative INP concentrations measured around the time of
 600 the event considerably improved our simulation results. We can thus infer that adapt-
 601 ing the ice nucleation parameterization to the particularly pristine conditions prevail-
 602 ing over the Southern Ocean is essential for atmospheric models running over this re-
 603 gion, in agreement with the conclusions of Vergara-Temprado et al. (2018). Refining the
 604 vertical resolution in the troposphere led to slightly higher liquid water content, but, first
 605 and foremost, it allowed us to simulate more stable-in-time SLW layers and to simulate
 606 vigorous and frequent turbulence within clouds. At coarse vertical resolution, the enhanced
 607 cloud-top radiative cooling associated with the cloud droplet production still made it pos-
 608 sible to simulate some turbulence in mid-level clouds. An additional parameterization
 609 for cloud-top turbulence generation further led to more realistic comparison with radar
 610 estimations of the TKE dissipation rate during specific periods like during the arrival
 611 of the warm front, but it does not help sustain the SLW layer at altocumulus top.

612 Our changes in the model physics considerably modified the simulated CRE dur-
 613 ing the event. Amongst the most prominent signals, we could point out a pronounced
 614 decrease in CRE at the ocean surface due to more shortwave radiation reflected toward
 615 space by the more realistic SLW layers and an increase in CRE at the ice sheet surface
 616 owing to an enhanced downward longwave radiative flux. Despite improvements regard-
 617 ing the simulation of SLW, the timing and the correct quantity of the LWP were still not

618 satisfactorily reproduced, questioning the representation of cloud-top liquid layers and/or
619 embedded liquid patches within clouds during the precipitation period.

620 Albeit very promising, our new ice nucleation parameterization based on an INP
621 concentration formulation that only depends on temperature cannot be fully satisfac-
622 tory since it does not account for the true link between aerosol populations and ice nu-
623 cleation. This calls for a future more accurate aerosol-aware formulation for INPs in the
624 high-latitude Southern Ocean.

625 Importantly, the way turbulent mixing at cloud top is represented - and hence the
626 physical representation of liquid droplet condensation and growth in mixed-phase clouds
627 - remains questionable since the local 1.5 order turbulent mixing parameterization does
628 not properly account for non-local convective transport and since it does not treat sep-
629 arately the respective evolution of rising and subsiding air parcels. This invites further
630 parameterization development targeting the top-down convection at cloud-top, taking,
631 for instance, inspiration from mass-flux schemes used to treat the mixing by thermal plumes
632 in convective boundary-layers.

633 Last but not least, our work does not enable us to draw any robust conclusions about
634 the ability of WRF to reproduce the low-level mixed-phase clouds which have strong ra-
635 diative effects at the surface in our study case and which explain the major part of the
636 radiative bias over the Southern Ocean in CMIP models. Future studies are thus needed
637 to broach this aspect, tackling in particular the coupling - or decoupling - between clouds
638 and the ocean surface, the effect of surface evaporation and the interactions with the boundary-
639 layer dynamics.

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 656 reanalyses can be freely downloaded from the Copernicus climate data store ([https://](https://cds.climate.copernicus.eu)
 657 cds.climate.copernicus.eu). The original WRF code is open source. The new WRF
 658 topography file from the Reference Elevation Model of Antarctica dataset will be made
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