



Introducing hydrometeor orientation into all-sky microwave/sub-millimeter assimilation

Vasileios Barlakas¹, Alan J. Geer², and Patrick Eriksson¹

¹Department of Space, Earth and Environment, Chalmers University of Technology, Gothenburg, Sweden

²European Center for Medium-Range Weather Forecasts, Reading, UK

Correspondence: Vasileios Barlakas (vasileios.baralakas@chalmers.se)

Abstract. Numerical weather prediction systems still employ many simplifications when assimilating microwave radiances in all-sky conditions (clear sky, cloudy, and precipitation). For example, the orientation of ice hydrometeors is ignored, along with the polarization that this causes. We present a simple approach for approximating hydrometeor orientation, requiring minor adaption of software and no additional calculation burden. The approach is introduced in the RTTOV (Radiative Transfer for TOVS) forward operator and tested in the Integrated Forecast System (IFS) of the European Centre for Medium-Range Weather Forecasts (ECMWF). For the first time within a data assimilation (DA) context, this represents the ice-induced brightness temperature differences between vertical (V) and horizontal (H) polarization, the polarization difference (PD). The discrepancies in PD between observations and simulations decrease by an order of magnitude at 166.5 GHz, with maximum reductions of 10–15 K. The error distributions, which were previously highly skewed and therefore problematic for DA, are now roughly symmetrical. The approach is based on rescaling the extinction in V- and H-channels, which is quantified by the polarization ratio ρ . Using dual polarization observations from Global Precipitation Mission microwave imager (GMI), suitable value for ρ was found to be 1.5 and 1.4 at 89.0 and 166.5 GHz, respectively. The scheme was used for all the conical scanners assimilated at ECMWF, with broadly neutral impact on the forecast, but with an increased physical consistency between instruments that employ different polarizations. This opens the way towards representing hydrometeor orientation for cross-track sounders, and at frequencies above 183.0 GHz where the polarization can be even stronger.

1 Introduction

Clouds containing ice hydrometeors are considered among the greatest ambiguities in both climate and numerical weather prediction (NWP) modeling systems (Duncan and Eriksson, 2018). Despite recent significant progress, their interaction with radiation has not been reliably assessed, in part owing to their high heterogeneity in shape and orientation within a cloud.

Over the last decades, several studies explored the use of polarimetric observations at millimeter/sub-millimeter (mm/sub-mm) wavelengths to retrieve microphysical and macrophysical properties of ice hydrometeors (e.g., Czekala, 1998; Prigent



et al., 2005; Xie, 2012; Defer et al., 2014). In particular, and the focus of this study, oriented non-spherical ice hydrometeors are known to cause the observed brightness temperature (T_B) differences between vertical (V) and horizontal (H) polarization. Here, this is denoted as the polarization difference ($PD = T_{BV} - T_{BH}$), owing to different scattering properties between V- and H-polarization (dichroism effect, Davis et al., 2005). Evans and Stephens (1995a, b) simulated the sensitivity of polarized microwave (MW) frequencies (between 85.5 and 340.0 GHz) on non-spherical horizontally oriented ice hydrometeors. They reported a positive polarization signal, which increases with frequency. These findings were further supported by Czekala (1998); by conducting off-nadir simulations including polarization at 200.0 GHz, he reported a PD of about 30 K linked to irregular shaped horizontally oriented hydrometeors. Xie (2012) reported a PD up to 10 K due to oriented snow observed by a ground-based passive radiometry at 150.0 GHz. Studies involving active and passive satellite remote sensing instruments further supported the presence of oriented ice hydrometeors (e.g., Prigent et al., 2001, 2005; Davis et al., 2005; Defer et al., 2014; Gong and Wu, 2017; Gong et al., 2020). Prigent et al. (2001, 2005) focused on passive MW imagers operating at low frequencies (37.0 and 85.0 GHz), i.e., the Tropical Rainfall Measuring Mission's (TRMM) Microwave Imager (TMI) and the Special Sensor Microwave/Imager (SSM/I), and reported a PD due to horizontally oriented liquid at 37.0 GHz or ice hydrometeors at 85.0 GHz. At higher frequencies (up to 166.5 GHz), analogous PDs have been reported by Defer et al. (2014) and Gong and Wu (2017) derived from the Microwave Analysis and Detection of Rain and Atmospheric Structure (MADRAS) and the GPM (Global Precipitation Mission) microwave imager (GMI), respectively.

Defer et al. (2014) and Gong and Wu (2017) tried to interpret the observed PD in a conceptual way. Especially, Gong and Wu (2017) suggested an approximative modeling framework linking the observed PD to the ratio of optical thicknesses in V- and H-polarization (τ_H/τ_V); this ratio is linked to the axial ratio of an oriented hydrometeor. This framework was tested on GMI observations at 89.0 and 166.5 GHz, but also on observations by the airborne radiometer CoSSIR (Compact Scanning Submillimeter-wave Imaging Radiometer) at a frequency of 640.0 GHz. The first rigorous simulations reproducing the observed PD were conducted by Brath et al. (2020). One of their main findings was the strong impact of the assumed shape of the ice hydrometeor on the simulated PD. Presently, GMI is the only operational instrument that measures ice hydrometeors including polarization. The upcoming Ice Cloud Imager (ICI) mission (Eriksson et al., 2020) will cover mm/sub-mm frequencies and will have channels measuring both V- and H- polarization at 243.2 and 664.4 GHz. This will provide further insights regarding the strong polarization signals originated by oriented ice hydrometeors. The observations from ICI will be assimilated in all-sky conditions for weather forecasting, further motivating the need to handle hydrometeor orientation.

All-sky radiance assimilation is the process of assimilating observations under the complete range of atmospheric conditions, i.e., clear sky, cloudy, and precipitation. Microwave humidity satellite observations in particular are gaining weight in forecasts, with the possibility to infer improved dynamical initial conditions (e.g., temperature, winds) from observations of cloud and precipitation (Geer et al., 2018). At the European Centre for Medium-Range Weather Forecasts (ECMWF), all-sky observations comprise around 20 % of the observation impact on forecasts (Geer et al., 2017). There is an ongoing effort to incorporate all possible observations that are sensitive to cloud and precipitation through the development of new capabilities, i.e., including visible, infrared, and sub-mm wavelengths (Geer et al., 2018).



The assimilation of satellite observations necessitates accurate, and at the same time, fast radiative transfer simulations. Currently, the leading radiative transfer models that are employed in global NWP systems for data assimilation (DA) are the Radiative Transfer model for TIROS Operational Vertical Sounder (RTTOV, Saunders et al., 2018) and the Community Radiative Transfer Model (CRTM, Liu and Boukabara, 2014). However, polarization due to the orientation of non-spherical hydrometeors is currently ignored. These forward models apply only “scalar” simulations; one calculation deals with either V- or H-polarization, while in nature scattering by hydrometeors transfers energy between polarizations. In addition, the models use the same ice scattering properties for both computations. It is currently too expensive to move to fully-polarized radiative transfer in an assimilation system. However, this work will show that even with scalar radiative transfer, a significantly improved physical representation can be achieved by correctly varying the scattering properties as a function of polarization.

Following the work of Gong and Wu (2017), we explore the use of a simple scheme to approximate the effect of oriented ice hydrometeors in reproducing the observed PD from conical scanning MW imagers. In this study, a special emphasis is given on quantifying the best fit between model and observations. The performance of such scheme is also tested in cycled DA experiments by utilizing the Integrated Forecast System (IFS) of ECMWF. An ability to model oriented hydrometeors will provide essential information towards assimilating the observations from upcoming satellite missions, e.g., ICI and the Microwave Imager (MWI) that will fly on board the Meteorological Operational Satellite-Second Generation (Metop-SG) operated by the European Organization of Meteorological Satellites (EUMETSAT). Reducing errors in forward modelling could improve the quality of the initial conditions for weather forecasting, as well as further improve weather forecasts.

2 Methods and tools

2.1 Radiative transfer model

Radiative transfer simulations are conducted by means of RTTOV-SCATT, i.e., the part of RTTOV (Saunders et al., 2018) that accounts for multiple scattering (Bauer et al., 2006). RTTOV-SCATT was developed by the EUMETSAT Numerical Weather Prediction Satellite Application Facility (NWP SAF) and is utilized by weather centres and scientists worldwide for fast modelling of all-sky radiances, including as the observations operator in the IFS at ECMWF.

RTTOV-SCATT employs the delta-Eddington approximation (Joseph et al., 1976) to solve the radiative transfer equation in all-sky conditions (clear, cloudy, and precipitating). The simulated T_B is calculated as the weighted mean of the T_B from two independent columns linked to clear-sky and cloudy conditions. The weighting is conducted following a cloud overlap scheme, introduced by Geer et al. (2009). This scheme is considered to be a fast approximation of sub-grid variability and hence, the beam-filling effect (e.g., Barlakas and Eriksson, 2020).

The gaseous absorption component (e.g., oxygen and water vapour) is derived from precalculated regression tables supplied by RTTOV (Saunders et al., 2018). Here, we employ version 12.3 of RTTOV-SCATT, but with two main modifications on the path towards the final version 13 (v13) configuration. First, it considers five hydrometeors and separates convective snow (hereafter graupel) from large scale snow (hereafter snow); a feature not available in version 12. Second, the microphysical configuration is updated as described below. RTTOV-SCATT utilizes pre-calculated bulk optical properties (extinction, single



Table 1. Microphysical setup for all the hydrometeors: PM denotes the particle model, PSD is the particle size distribution, ciw stands for cloud ice water, and clw is the cloud liquid water.

Type	PM	PSD
rain	Mie sphere	Marshall and Palmer (1948)
snow	Sector snowflake (Eriksson et al., 2018)	Field et al. (2007) tropical
graupel	3-Bullet rosette (Liu, 2008)	Field et al. (2007) tropical
clw	Mie sphere	Gamma (see, Geer and Baordo, 2014)
ciw	Large column aggregate (Eriksson et al., 2018)	Heymsfield et al. (2013)

90 scattering albedo, and asymmetry) for each hydrometeor (rain, snow, graupel, cloud liquid water, and cloud ice water), i.e., the hydrometeor tables. For each model level, the bulk properties are derived from the hydrometeor tables given the hydrometeor content and temperature. For details the reader is referred to Geer and Baordo (2014). The microphysical setup employed in this study was found by Geer (2020) using a multi-dimensional parameter search by means of passive monitoring experiments (terminology is given in Sect. 2.5; this is referred to as the intermediate configuration in that paper). One of the main new
 95 aspects is the utilization of ice hydrometeors from the Atmospheric Radiative Transfer Simulator (ARTS) single scattering database (Eriksson et al., 2018). This database is comprised of more realistic hydrometeor shapes (e.g., hail, graupel, and aggregates) covering a greater frequency range (1 to 886.4 GHz) compared to Liu (2008) and hence, ideal for non-spherical hydrometeors; especially, for the upcoming ICI mission. Details on the microphysical setup are found in Table 1.

Over ocean, surface emissivity is calculated by the FAST microwave Emissivity Model-6 (FASTEM-6, Kazumori and En-
 100 glish, 2015). The land surface emissivity is primarily retrieved from satellite observations in surface-sensitive channels following an approach described in Karbou et al. (2010) and extended to all-sky assimilation by Baordo and Geer (2016). In case the retrieval fails, the Tool to Estimate Land Surface Emissivities at Microwave (TELSEM) atlas (Aires et al., 2011) is employed.

2.2 Assimilation system

The ECMWF forecasting system employs a semi-Lagrangian atmospheric model and a 12 h cycling four-dimensional (4D)
 105 variational DA (Rabier et al., 2000) in order to generate global analyses and 10 d forecasts. In this system, the direct assimilation of all-sky radiances from MW imagers went operational in 2009 and was subsequently extended to other sensors including microwave humidity sounders (Geer et al., 2017). Among the other data assimilated are in-situ conventional measurements, radiances from polar orbiting satellites, e.g., Advanced Microwave Sounding Unit-A (AMSU-A), Infrared Atmospheric Sound-
 ing Interferometer (IASI), Advanced Technology Microwave Sounder (ATMS), geostationary radiances, satellite-derived at-
 110 mospheric motion vectors, and surface winds from scatterometers. Note here that microwave imager observations are averaged ("superobbed") in boxes of about 80 km by 80 km to meet the effective resolution in the model (Geer and Bauer, 2010).

The assimilation system is driven by the various observations via the background departure (d) (Geer and Baordo, 2014):

$$d = y^o - y^b. \quad (1)$$



Where y^o and y^b are the actual observations and the equivalent observations simulated from the background forecast:

$$115 \quad y^b = H[M[x^b(t_0)]] + c. \quad (2)$$

The background atmospheric state, $x^b(t_0)$, is a 12 h forecast from an earlier analysis, with t_0 representing the start of the new assimilation window (Geer and Baordo, 2014). The nonlinear forecast model, M , propagates the atmospheric state from $x^b(t_0)$ to the time of the observation and the forward operator H (herein, RTTOV-SCATT) simulates the satellite-observed radiance from the model profile interpolated to the observation location. Finally, c denotes the bias correction term that accounts for
 120 systematic deviations between observations and simulations, which are modelled as a linear function of total column water vapour, surface wind speed, satellite scan angle, and other predictors; this is estimated as part of the DA process (Dee, 2004; Auligné et al., 2007).

The ECMWF assimilation model includes four prognostic hydrometeors, i.e., rain, snow, cloud water and ice, to represent the large-scale cloud processes, along with prognostic cloud and diagnostic precipitation fractions (Tiedtke, 1993; Forbes
 125 et al., 2011). Convective rain and snow are diagnostic variables, derived from a convection scheme. When forming the input to RTTOV-SCATT, convective rain is added to the large-scale counterpart, but the convective snow is treated as a separate graupel hydrometeor type as explained above.

2.3 Observations

This study combines observations from GMI, the Advanced Microwave Scanning Radiometer-2 (AMSR-2), and the Special
 130 Sensor Microwave Imager/Sounder (SSMIS). Table 2 summarizes the channels of these instruments that are actively assimilated in the cycling DA experiments.

2.3.1 Advanced Microwave Scanning Radiometer-2

The AMSR-2 is a conical scanning instrument with 16 channels ranging in frequency between 6.9 and 89.0 GHz, with dual-polarization capabilities. AMSR-2, flies on board the Japan Aerospace Exploration Agency's (JAXA) Global Change Observa-
 135 tion Mission 1st-Water (GCOM-W1) satellite and supplies global observations at a high spatial resolution and an earth incident angle of 55° (Du et al., 2017).

2.3.2 Global Precipitation Mission microwave imager

The GPM core satellite carries a conical scanning MW radiometer, i.e., the GMI. It operates at frequencies between 10.65 and 190.31 GHz (13 channels overall) at a rather high spatial resolution, i.e., between ≈ 26 and ≈ 6 km, depending on the
 140 channel. The earth incident angle is 52.8° for channels 1–9 and 49.1° for channels 10–13. GMI is the only operational passive instrument with dual polarization channels at high frequencies, i.e., 166.5 GHz, offering unique capabilities in sounding ice hydrometeors, among others.



Table 2. Channel characteristics of the Advanced Microwave Scanning Radiometer-2 (AMSR-2), Global Precipitation Mission microwave imager (GMI), and the Special Sensor Microwave Imager/Sounder (SSMIS) considered in this study.

Frequency [GHz]	Polarization	Instruments and No.		
		AMSR-2	GMI	SSMIS
18.7	H	7	3	-
18.7	V	8	4	-
19.35	H	-	-	12
19.35	V	-	-	13
22.235	V	-	-	14
23.8	H	9	-	-
23.8	V	10	5	-
36.5	V	11	6	-
37.0	V	-	-	16
89.0	H	-	-	-
89.0	V	13	8	-
91.655	V	-	-	17
150.0	H	-	-	8
166.5	H	-	11	-
166.5	V	-	10	-
183.31±7.0	V	-	13	-
183.31±6.6	H	-	-	9
183.31±3.0	H	-	-	10
183.31±3.0	V	-	12	-
183.31±1.0	H	-	-	11

2.3.3 Special Sensor Microwave Imager/Sounder

The SSMIS is a conical MW instrument, with an earth incident angle of 53.1° , which is on board the Defense Meteorological
 145 Satellite Program F-17 spacecraft (Sun and Weng, 2012). It consists of 24 channels between ≈ 19.0 and ≈ 190.0 GHz (21
 frequencies in total), three out of which are in both V- and H-polarization, i.e., 19.35, 37.0, and ≈ 91.65 GHz. Compared to
 AMSR-2 and GMI, SSMIS has a lower spatial resolution, with an instantaneous field of view ranging from ≈ 54 km (at the
 lowest frequency channels) to ≈ 14 km (at the highest frequency channels).

2.4 Polarized scattering correction

150 A widely used simplification in radiative transfer modeling is the assumption of totally randomly oriented (TRO) hydrometeors.
 Such hydrometeors are considered as macroscopically isotropic, meaning that there is no favored propagation direction or

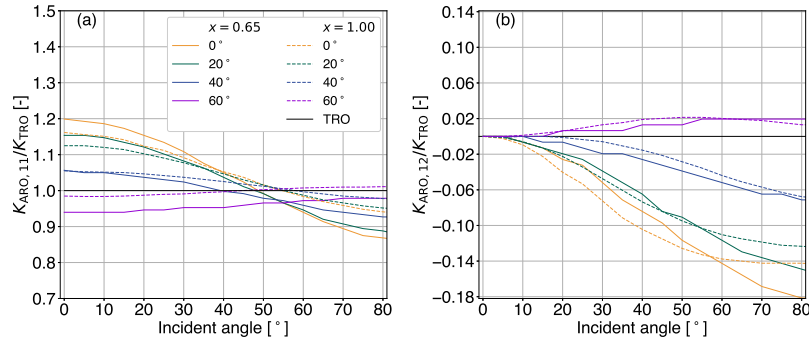


Figure 1. For large plate aggregate, extinction matrix elements of azimuth random orientation (ARO) normalized by the extinction cross section in case of the total random orientation (TRO) as a function of incident angle θ_{inc} at 166.9 GHz. Results are presented in case of (a) $K_{\text{ARO},11}/K_{\text{TRO}}$ and (b) $K_{\text{ARO},12}/K_{\text{TRO}}$ for two size parameters (x) and four tilt angles. In black, the values in case of TRO are depicted.

orientation and any induced polarization signal is only driven by the radiation scattered into the line of sight (e.g., see Barlakas, 2016). However, ice hydrometeors are characterized by non-spherical shapes and thus non-unit aspect ratios and, if they are large enough, they tend to be oriented. This results in anisotropic scattering with viewing-dependent optical properties, meaning different values of extinction at different polarization components of the incident radiation (dichroism effect, Davis et al., 2005). Accordingly, in vector radiative transfer theory, the attenuation between the incident radiation and the hydrometeor is governed by a 4×4 extinction matrix K , depending on the incident direction and the orientation of the hydrometeor (Brath et al., 2020). In case of TRO, the extinction matrix has no angular dependency and is reduced to its first element $K_{\text{TRO}} = K_{11}$, which describes the extinction cross-section. A special case of orientation is the azimuth random orientation (ARO) describing a preferred orientation to the horizon based on the tilt angle, with no favored orientation in the azimuth direction (Brath et al., 2020). In this case, K is reduced to only three independent elements (K_{11} , K_{12} , and K_{34}). K_{12} represents the differences in the extinction between V- and H-polarization (cross section for linear polarization), while K_{34} describes the differences in the extinction between $+45^\circ$ and -45° polarization (cross section for circular polarization), and is not relevant here. For a comprehensive description of ARO, the reader is referred to Brath et al. (2020).

Figure 1 highlights the differences in K between ARO and TRO for the large plate aggregate (LPA) at 166.9 GHz. It displays the first two elements of K normalised by the extinction at TRO as a function of the incident angle (θ_{inc}), for four tilt angles, and two values of the size parameter x :

$$x = \pi \cdot D_{\text{veq}} / \lambda. \quad (3)$$

Where D_{veq} is the the volume-equivalent diameter of the hydrometeor and λ is the wavelength. In case of ARO, the scattering data of Brath et al. (2020) has been utilized. At the tilt angles considered here ($0-60^\circ$), hydrometeor orientation induces differences up to 20% of $K_{\text{ARO},11}$ compared to K_{TRO} . The maximum differences are reported at θ_{inc} of 0° , while the smallest ones are found at around θ_{inc} of 55° and are in the range of $\approx 0-4\%$ (depending on x). In fact for this incidence



angle, the differences between $K_{\text{ARO},11}$ and K_{TRO} are close to zero over all sizes (not shown here). This implies that conical scanning instruments characterized by an earth incident angle around 55° will observe the same mean extinction cross section (K_{11}) for both TRO and ARO hydrometeors.

At small θ_{inc} , the cross section for linear polarization (K_{12}) is close to zero, meaning that the differences in the extinction between V- and H-polarization are negligible. However, these differences are increasing with increasing θ_{inc} up to about $\approx 80^\circ$. The largest differences in the extinction between V- and H-polarization (up to about 18 %) are derived for a tilt angle of 0° . Recall here that results are presented in case of the LPA shape. For other shapes, e.g., plate type, the magnitude of K_{12} is much larger (Brath et al., 2020). In any case, there are large differences in the cross section for linear polarization (K_{12}) between ARO and TRO hydrometeors, even at earth incident angles around 55° . This (rather than any change in K_{11}) is the likely explanation for the polarized scattering observed by microwave imagers.

To that end, a simple scheme has been implemented into RTTOV-SCATT to improve the physical representation of polarized scattering. To model the effect of ARO ice hydrometeors (snow, graupel, and cloud ice) in recreating the observed polarization signal from conical scanning radiometers, the layer optical thickness of TRO hydrometeors (τ) is increased in H-polarized channels and decreased in V-polarized channels by a correction factor α , leading to the polarization ratio (ρ):

$$\rho = \frac{\tau_{\text{H}}}{\tau_{\text{V}}} = \frac{\tau \cdot (1 + \alpha)}{\tau \cdot (1 - \alpha)} = \frac{1 + \alpha}{1 - \alpha}. \quad (4)$$

Where τ_{H} and τ_{V} are the corrected optical thickness at H- and V-channels, respectively. This ratio can be seen as an indirect representation of the axial ratio of an oriented hydrometeor, while α approximates the differences in the extinction between the two channels (K_{12} ; see Fig. 1b). In other words, α forces the extinction at V-polarization (H-polarization) to be weaker (stronger). Note here that the asymmetry parameter and the single scattering albedo have been not modified, as it is not yet clear how to do so.

2.5 Experiments

Data assimilation experiments were conducted utilizing the 46r1 cycle of the IFS of ECMWF. This cycle was operational from June 2019 to June 2020. Compared to the operational system, the microphysical configuration of RTTOV-SCATT has been upgraded as described earlier, and a slightly reduced resolution of about 28 km (labeled Tco399) with 137 vertical levels has been used. Two types of experiments are considered:

1. To scrutinize the performance of the forward operator, a consistent sample of observations should be examined. Consequently, passive monitoring experiments were conducted to simulate GMI radiances. In the monitoring mode, the short-term forecast for all experiments is supplied from a single parent cycling DA experiment. In other words, the background forecasts (sometimes referred to as first guess) comes from the parent experiment and changes to the usage of GMI observations do not feed back; the background atmospheric state is kept constant. Nine experiments have been conducted for a period of one month (6 June to 7 July 2019), with ρ ranging from 1.1 to 1.5, with an increment of 0.05. Additional experiments were run in which the ρ was applied to each ice hydrometeor individually; for details the reader is referred to Sect. 3.3.



2. To assess the potential long term impact of ρ on the forecast, cycled DA experiments were run for a total period of 6 months, i.e., from 15 February to 31 May 2019 and from 13 June to 31 August 2019. Here, the improved forward operator is used for AMSR-2, GMI, and SSMIS as part of the DA process. The resulting analysis provides the initial conditions for the background forecast in the next cycle of DA. Hence, in contrast to the passive experiments, the background forecast and observational usage vary, and the changes can feed back over multiple cycles of DA. These experiments are described further in Sect. 3.4.

For both types of experiments, to highlight whether the forward operator and the forecast are improved or degraded, control experiments were additionally run, with the improved physical representation of polarized scattering being turned off, meaning a ρ value of 1.0 (a correction factor a of 0.0).

215 3 Choosing the best polarization ratio

3.1 Overview

By means of passive monitoring experiments, the ability of ρ to simulate the observed polarization patterns due to oriented ice hydrometeors is explored. Accordingly, a special emphasis is given on the high frequency dual-polarization channels of GMI (89.0 GHz and especially on 166.5 GHz), because scattering due to ice hydrometeors (snow, graupel, and cloud ice) is increasing with frequency. For clarity, we discriminate between the observed (o) and simulated (background, b) polarization difference:

$$\begin{aligned} PD^o &= T_{BV}^o - T_{BH}^o, \\ PD^b &= T_{BV}^b - T_{BH}^b. \end{aligned} \quad (5)$$

One of the challenges encountered was to screen out any surface contamination and particularly, any strong polarization signal that originates from water surfaces (Meissner and Wentz, 2012); strongly polarized surfaces complicate the separation between cloudy and clear sky conditions. Appendix A describes a number of careful screening checks that were used to minimize the surface contribution. As will be seen, the remaining data is almost entirely free from surface contribution.

Figure 2 illustrates the PD as a function of T_{BV} at 166.5 GHz in case of both observations and simulations. Polarized scattering from preferentially oriented ice hydrometeors leads to PD^o up to 10–15 K (in red) centered around 220 K, with increasingly low T_{BV} indicating increasingly cloud-affected scenes. In fact, the arch-like shape of the $PD^o - T_{BV}^o$ (or T_{BH}^o) relation appears to be universal (Gong and Wu, 2017). The existing modelling framework characterized by a polarization ratio of 1.0 (control run; in green) completely fails to reproduce such polarization signal (see Fig. 2a). However, it provides valuable information regarding the surface contribution; if there was any simulated polarization signal due to ocean reflection, it would be visible. The remaining panels in Fig.2, i.e., b–f, depict the ability of ρ to provide realistic simulations of this behaviour, and a first glance indicates that a ρ value between 1.3 and 1.4 could do a reasonable job.

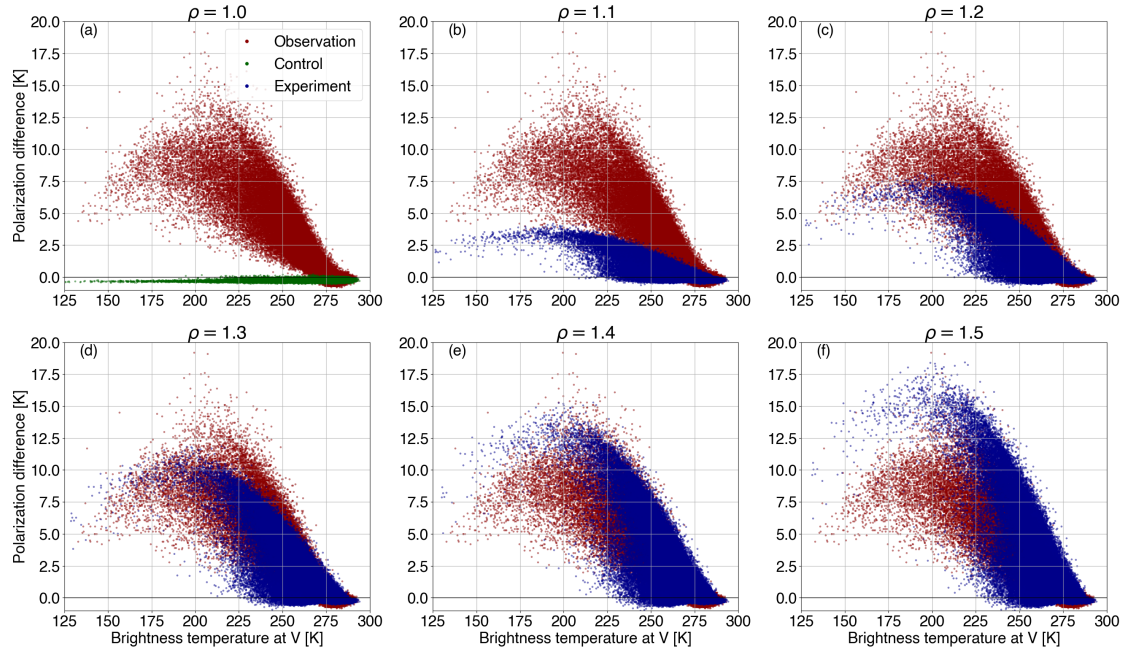


Figure 2. Polarization difference between the vertically and horizontally polarized channels ($PD = T_{BV} - T_{BH}$) as a function of T_{BV} at 166.5 GHz as observed by GMI (PD^o ; in red) and simulated (PD^b ; in blue) by the passive monitoring experiments for a period of one month (6 June to 7 July 2019). Results are presented for polarization ratios (ρ) ranging from 1.0 (control run; in green) to 1.5, with a 0.1 increment.

3.2 Quantifying the fit of model to observations

In order to quantify the best fit between simulations and observations, the commonly used metrics are the mean, the standard deviation (σ), and/or the root mean square error (rmse). Nonetheless, forecast modelling systems are still unable to predict cloud and precipitation at small scales, and thus, are characterized by mislocation errors that could lead to quite large σ and/or rmse (e.g., Geer and Baordo, 2014, and references therein). A more comprehensive assessment of the polarized scattering correction is achieved by measuring the divergence between the distributions of observations and simulations introduced in Geer and Baordo (2014):

$$\Phi = \frac{(\sum_{\text{bins}} \varphi)}{\# \text{bins}} = \left(\sum_{\text{bins}} \left| \log_{10} \frac{\#P_b}{\#P_o} \right| \right) / (\# \text{bins}), \quad (6)$$

where P_b and P_o are the populations of simulations and observations, respectively. The metric of divergence describes how well the experiments approximate the observations by putting a penalty according to the logarithm base ten of the ratio of the populations (φ ; simulations divided by observations) in each bin. To avoid any infinite penalty in case of zero populations, empty bins are replaced by a population of 0.1; varying this limit has a minor effect and increases or decreases the penalty effect. A detailed explanation of the divergence calculation is given later in Fig. 5. To summarise, the statistics used here are

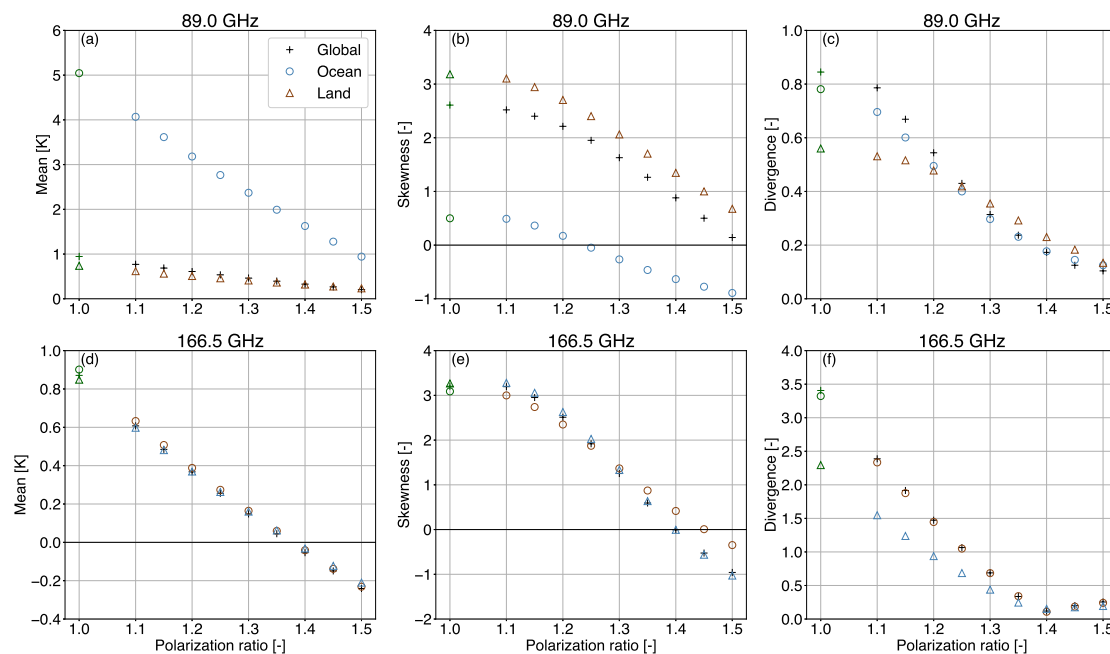


Figure 3. Statistical metrics (mean, skewness, and divergence) describing the differences in polarization differences between observations and simulations, i.e., $PD^o - PD^b$, as a function of the polarization ratio (ρ). Results are presented for 89.0 GHz (top) and 166.5 GHz (bottom) over land (in brown triangles), ocean (in blue circles), and globally (in black crosses) for a period of one month (6 June to 7 July 2019). In terms of the control run ($\rho = 1$), the corresponding differences are highlighted in green.

the mean and skewness of $PD^o - PD^b$ and the histogram divergence, which are non-localised measures of the discrepancy between simulated and observed distributions.

Figure 3 depicts these statistical metrics quantifying the differences in PDs ($PD^o - PD^b$) between observations and simulations with respect to the polarization ratio. Results are reported globally (in black crosses), but the situations over ocean- (in blue triangles) and land-only (in brown dots) are also described. To begin with, the control run, which is highlighted by the corresponding green markers, is unable to provide realistic simulations of the observed PDs at both 89.0 (upper panels) and 166.5 GHz (lower panels); it leads to a mean error of about 1 K (globally) and the histograms (of $PD^o - PD^b$, not shown here) are described by right-skewed distributions that are more asymmetric over the land compared to the ocean. In both frequencies, even a small increase to the polarization ratio above 1.0 reduces discrepancies between observations and simulations and leads to a more symmetrical error; further increasing polarization ratio reduces discrepancies still further up to a point where the best ratio can be found. At 89.0 GHz and over the entire globe, the best polarization ratio is 1.5 (skewness is close to zero). However, over the ocean, a more symmetrical error is achieved by a smaller value, i.e., 1.45. On the contrary, the dependency of 166.5 GHz channel on the polarization ratio is characterized by similar behaviour over both land and ocean. An increase of ρ improves the representation of preferentially oriented ice hydrometeors up to a value of about 1.4; a further increase of

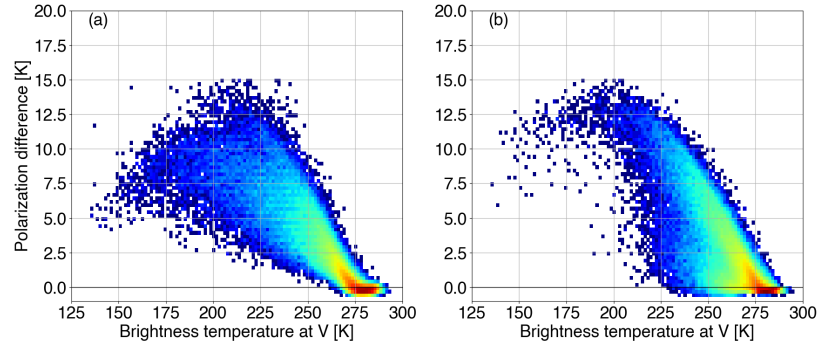


Figure 4. Two-dimensional histograms describing the arch-like relationship between the polarization difference and the brightness temperature at V-polarization at 166.5 GHz as (a) observed ($PD^o - T_{BV}^o$) by GMI and (b) simulated ($PD^b - T_{BV}^b$) for a polarization ratio of 1.4.

the scattering, increases $PD^o - PD^b$ once again, but towards negative values. This implies that the corresponding histogram, from a right-skewed distribution, becomes symmetrical at a polarization ratio of 1.4 and turns into a left-skewed one at higher ρ values. In Fig. 3c the divergence Φ is shown, which is the summation of the absolute value of the logarithm base ten of the ratio of the populations over all bins, divided by the number of bins. The divergence is in line with the conclusions drawn from the other metrics. To conclude, the polarization ratio that models best the oriented non-spherical ice hydrometeors and leads to a more symmetrical error is 1.5 for 89.0 GHz ($\Phi = 0.104$) and 1.4 for 166.5 GHz ($\Phi = 0.119$). This is translated to increasing/decreasing τ_H (τ_V) by 20 % ($a = 0.2$) at 89.0 GHz and by 16.7 % ($a = 0.167$) at 166.5 GHz.

Figure 4 provides an overview of the global performance of the polarization ratio of 1.4 to simulate the observed PDs at 166.5 GHz. To begin with, Fig. 4a illustrates the two-dimensional (2D) histogram describing the observed arch-like relationship of $PD^o - T_{BV}^o$ introduced in Fig. 2 (in red). Most of the points are bundled above ≈ 275 K, which are linked to clear-sky conditions and PDs close to 0 K. The cloud induced PDs are mostly accumulated at T_{BV}^o between 215 and ≈ 275 K. Figure 4b highlights the corresponding simulated $PD^b - T_{BV}^b$ relation for a ratio of 1.4. One can see that such a ρ value approximates quite well the observed PDs at values of T_{BV}^b above ≈ 225 K. For the situation at 89.0 GHz, the reader is referred to Fig. B1 in Appendix B.

A more thorough comparison is found in Fig. 5. Firstly, the left panel shows the histograms of the PDs at 166.5 GHz as observed (PD^o ; in red) by GMI and simulated (PD^b) for a polarization ratio of 1.4 (in blue) and 1.0 (in green; control run). As initially seen in Fig. 2, the control run completely fails to reproduce any positive PDs larger than 0.2 K. On the other hand, a polarization ratio of 1.4 removes most of the discrepancies between observations and simulations. The right panel displays the corresponding divergence (φ), or in other words, the logarithm base ten of the ratio of the populations (see Eq. 6). The experiment with $\rho = 1.4$ slightly overestimates the occurrences of PDs between 0.2 and about 5.0 K, and thus, are penalized with a positive logarithm base ten ratio. From around 5.0 to 10.5 K, a very good agreement is found between simulations and observations, while at negative and at the largest PD values, the largest differences are reported, leading to a Φ value of 0.127.

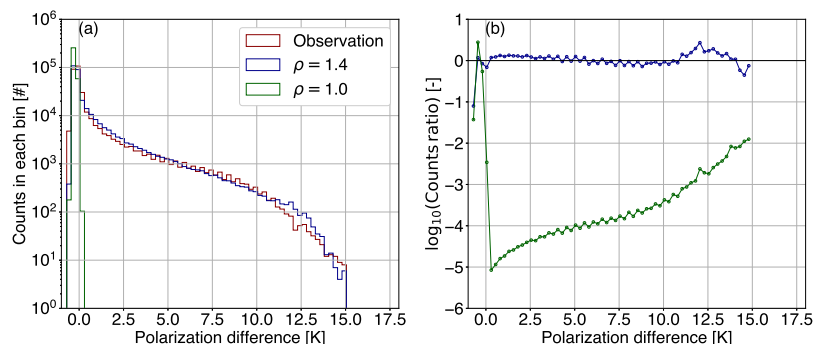


Figure 5. (a) Histograms of the polarization differences at 166.5 GHz as observed (PD^o ; in red) by GMI and simulated (PD^b) for a polarization ratio of 1.4 (in blue) and 1.0 (in green; control run) and (b) corresponding divergence between the distributions of PD^o and PD^b as a function of the polarization difference.

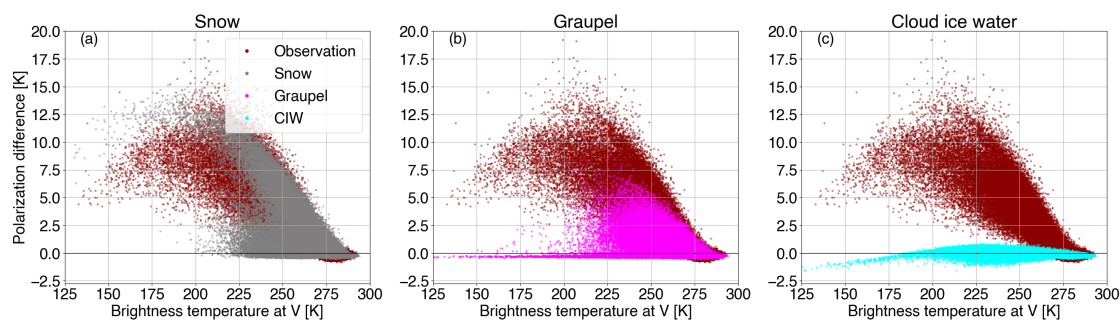


Figure 6. As in Fig. 2, but in case of (a) snow (in grey), (b) graupel (in magenta), and (c) cloud ice water (ciw; in cyan) individually polarized passive monitoring experiments at a polarization ratio of 1.4.

285 In terms of the control run, any φ value corresponding to a PD above 0.2 K results from the penalty effect assigned to avoid infinite values. The control run yields a Φ value of 3.265.

3.3 Polarization differences per hydrometeor type

To discriminate the polarization signal at 166.5 GHz induced by the various ice hydrometeors, additional passive monitoring experiments were conducted, whereby the polarization ratio of 1.4 was applied to snow, graupel, and cloud ice water indi-
 290 vidually. Figure 6 displays the corresponding simulated $PD^b - T_{BV}^b$ arch-like relation compared to the observed one (in red). Preferentially oriented large scale snow produces most of the observed polarization signature, while deep convective snow (graupel) induces only low-to-medium PD values and a much lower peak (≈ 7.5 K). Interestingly, cloud ice produces a minor polarization signal, with negative values at increasingly cold T_{BV} .



To quantify the contribution of each ice hydrometeor to the total polarization difference, the consistency in the distributions
 295 between observations and simulations has been investigated; Figure 7 illustrates the corresponding histograms and their diver-
 gence. Comparing Fig. 5a with Fig. 7a, it follows that snow slightly underestimates PDs between 2 and 11 K, leading to small
 negative penalty values of φ ; the overall divergence Φ is found to be a bit larger (0.165) than the one resulting by applying the
 ρ to all the hydrometeors (0.127). Graupel poorly represents the polarization differences above ≈ 2.5 K, while it completely
 fails to generate any values above ≈ 8.5 K; the underestimation is increasing with increasing PD up to ≈ 8.5 K and, at larger
 300 PD values, any φ value is solely linked to the artificial penalty effect, i.e., empty bins are replaced by a population of 0.1,
 resulting to a Φ value of 1.721. Cloud ice generates a polarization signal between -0.5 and 1.2 K, leading to a much larger Φ
 (3.183); this is close to the one obtained in case of the control run, i.e., 3.265.

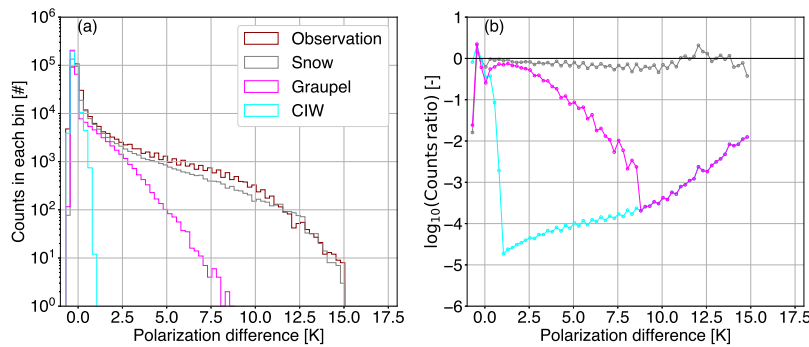


Figure 7. As in Fig. 5, but for snow (in grey), graupel (in magenta), and cloud ice water (ciw; in cyan) individually polarized passive monitoring experiments.

3.4 Forecast impact

As described in Sec. 2.5, cycled DA experiments were run to see whether using polarized scattering in the observation operator
 305 for SSMIS, GMI, and AMSR-2 could improve the quality of the forecast. The polarization ratio $\rho = 1.4$ was used for all ice
 hydrometeors at all frequencies, but it was applied in three different ways: either with equal and opposite perturbations to V-
 and H-channels as in Eq. (4) where $\alpha = 0.167$, or by unilaterally changing the ice hydrometeor optical depth in only H- or only
 V-channels:

$$\rho = \frac{\tau_H}{\tau_V} = \frac{\tau \cdot (1 + \alpha)}{\tau \cdot (1 - \alpha)} = \frac{\tau \cdot (1 + \beta)}{\tau \cdot (1)} = \frac{\tau \cdot (1)}{\tau \cdot (1 - \gamma)} = 1.4. \quad (7)$$

310 These latter experiments are equivalent to increasing τ_H by 40 % ($\beta = 0.4$) or reducing τ_V by 29 % ($\gamma = 0.286$). This is an
 approximate way to explore whether the polarized scattering might provide better results in combination with an adjustment to
 the original ice hydrometeor optical depths, which are not themselves known with high accuracy.

For example, Fig. 8 examines the change in quality of the 12 h forecast as measured by the σ of ATMS background departures
 (using the observations and channels selected for active DA). ATMS is not yet assimilated in all-sky conditions so it provides

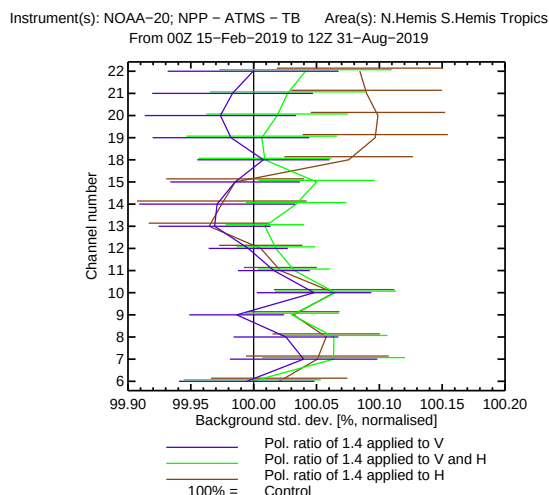


Figure 8. Normalised change in the standard deviation (σ) of background (12 h forecast) fits to independent ATMS observations with a polarization ratio (ρ) of 1.4, applied in different ways to the V- and H-channels. Error bars give the 95 % confidence interval estimated using a paired-difference t-test from per-cycle (every 12 h) statistics, without corrections for multiple comparisons.

an independent reference. The standard deviation, σ , is an appropriate metric of forecast skill here because the distribution of simulated ATMS observations is not expected to change. Starting with the results where $\rho = 1.4$ with equal and opposite perturbations to V- and H-channels, it can be seen that the ATMS tropospheric humidity channels (18–22) and stratospheric channels (11–15) are not significantly affected. However the tropospheric temperature sounding channels (6–10) show a slight but statistically significant increase in σ , suggesting a minor degradation in forecast quality. To counterbalance this, other conventional data (not shown) showed marginal but significant improvements in forecast quality, suggesting overall a neutral impact.

Slightly different results were seen with the unilateral scaling of the layer optical thickness between V- and H-channels. The unilateral increase in H-channels appears to give a significant degradation in forecast skill in tropospheric temperature and humidity channels of ATMS (6–10 and 18–22). In contrast, a unilateral decrease in V-channels appears to leave forecast skill unchanged. As the overall level of hydrometeor optical depth decreases (going from H-only, to V and H, to V-only), it appears that the forecast skill slightly improves. This suggests that it is also important to get the overall level of scattering correct to be able to implement the polarized scattering.

Figure 9 shows the normalised change in σ of background departures in GMI and SSMIS. Here, changes can come from two sources: a change in the quality of the background forecasts (as for ATMS) or directly due to the change in the observation operator (at least, in channels where the scaling was applied). In this second effect, when the observation operator is modified, statistics can be affected by the change in the distributions of simulated brightness temperatures, and hence they are vulnerable to the double penalty effect. For example, adding 40 % to ice hydrometeor optical depths in an H-channel like SSMIS 9 (the brown line) likely explains the 6 % larger σ of background departure. Similarly, reducing scattering by 29 % in V-channels

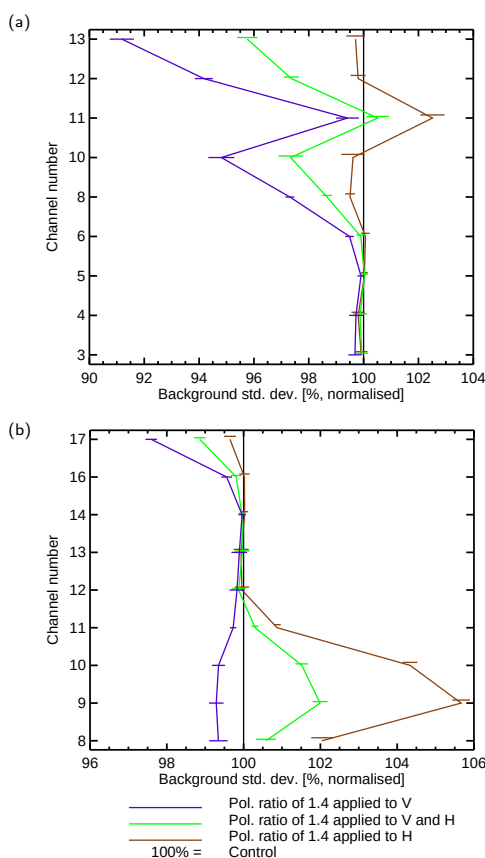


Figure 9. Normalised change in the standard deviation (σ) of background (12 h forecast) fits to (a) GMI and (b) SSMIS observations, with a polarization ratio (ρ) of 1.4, but applied in different ways to the V- and H-channels; other details as in Fig. 8. See Table 2 for channel descriptions.

such as SSMIS 17 (the purple line) likely explains the reduction in σ by 2 %. Where polarization is applied with equal and opposite perturbations to the V- and H- channels (the green line) the change in σ hence appears to be dominated by the double penalty effect. However, the figure still provides an important result on the benefits of polarized scattering. As shown in Table 2, most of the high-frequency channels of GMI are V-polarized (≥ 89.0 GHz, channels 8, 10, 12, and 13) and most of those on SSMIS are H-polarized (channels 8, 9, 10, and 11). Where only the H-channel simulations are modified, the V-channels provide an independent measure of the forecast quality. Likewise when only V-channel simulations are modified, the H-channels are an independent reference. Adding scattering in H-channels (brown line) reduces σ by around 0.5 % in GMI V-channels and in SSMIS channel 17 (91.655 GHz). Reducing scattering in V-channels (purple line) reduces σ in SSMIS H-channels and in GMI channel 11 (166.5 GHz), similarly by around 0.5 %. Results from AMSR-2 (not shown) are equivalent. This suggests that



representing polarized scattering makes the forward modelling more consistent across sensors, as well as between channels. At high frequencies (such as the 183.0 GHz channels) SSMIS will generally see bigger scattering depressions than GMI due to the different polarizations; if this effect is not represented in the observation operator, it degrades the consistency of the DA.

4 Discussion

At mm/sub-mm wavelengths, the interaction between ice hydrometeors and radiation is chiefly driven by scattering (high single scattering albedo, see, e.g., Eriksson et al., 2011). Consequently, such interaction induces a considerable polarization signature that strongly depends on the shape and orientation of the hydrometeors. In other words, non-spherical ice hydrometeors are characterized by non-unit aspect ratios, i.e., the ratio of the longest to the shortest axis, and, if they are large enough, they tend to be oriented in the atmosphere. This results in viewing-dependent scattering properties, leading to brightness temperature differences between V- and H-polarized channels, i.e., the polarization difference.

4.1 Polarization differences due to oriented ice hydrometeors

In the GMI observations examined here, the high frequency dual-polarization channels (89.0 GHz and, particularly, 166.5 GHz) show a global arch-like relationship between the polarization difference and the brightness temperature at V-polarization, confirming the findings of Gong and Wu (2017) and Brath et al. (2020) over the same frequencies and Defer et al. (2014) at the slightly lower frequency of MADRAS, i.e., 157.0 GHz. The arch-like relationship is generally attributed to a two-fold scattering effect. Increasing the number of hydrometeors leads to an increasing scattering, and hence, a stronger polarization signal is produced. However, the final polarization state results from only the first few orders of scattering (similar effects are seen at visible frequencies, see, e.g., Barlakas (2016) and references therein). Hence, an increasing multiple-scattering process due to the presence of enough hydrometeors will lead, at first, to a saturation (plateau) of the polarization and, at second, to a further decrease of the PD. Accordingly, the lowest PD values are found at warm T_{BV} (≈ 275 K), corresponding to very thin clouds, while the largest values are linked to intermediate cold T_{BV} and medium thick clouds, particularly in the anvil regions of convection; within deep convective cores (at very low T_{BV}), the multiple-scattering process is enhanced, leading to medium values of PDs ($\approx 4 - 10$ K). At 166.5 GHz, for example, scattering from preferentially oriented ice hydrometeors yields a positive polarization difference of about 10–15 K that peaks at T_{BV} 200–225 K.

In contrast to 89.0 GHz (see Fig. A1a), the arch-like shape of PD– T_{BV} at 166.5 GHz is characterized by a greater dynamical range (≈ 140 – 275 K compared to ≈ 170 – 275 K), in line with the results reported by Gong and Wu (2017) and Defer et al. (2014) and peaks at larger PD values (≈ 14 K compared to ≈ 10 K); however, Gong and Wu (2017) reported a lower peak at 166.5 GHz (≈ 10 K), but still larger compared to the one found at 89.0 GHz. Any differences at 89.0 GHz could be attributed to the different methods applied to screen out any surface induced polarization, but also due to the "superrobbing"; recall here that observations have been averaged in order to meet the effective resolution of the model. The smaller PD range found at 89.0 GHz could be explained by the fact that this channel is more sensitive to larger ice hydrometeors, while the lower PD



values are due to its multi-sensitivity aspect; a frequency of 89 GHz is quite sensitive to water vapor, water cloud, and rain
 375 droplets that diminish the polarization signal.

4.2 Approximate treatment of polarized scattering from oriented ice hydrometeors

In the current framework of RTTOV-SCATT, only TRO hydrometeors are considered, which fails to reproduce the observed
 PDs, leading to errors in polarized scattering up to about 10–15 K. To model the effect of preferentially oriented (ARO) ice hy-
 hydrometeors, a simple correction scheme has been implemented in RTTOV-SCATT that reduces TRO extinction in V-polarized
 380 channels and increases it in H-polarized ones. This is quantified by the ratio of extinction in H- over that in V-channels, i.e.,
 the polarization ratio ρ . This ratio holds an indirect microphysical representation of the aspect ratio of non-spherical oriented
 hydrometeors. Hence, TRO hydrometeors (control run) are characterized by a ρ of 1.0 and cannot induce polarization in our
 modelling framework. For the dual polarization channels, the value of ρ that approximates best the orientation of non-spherical
 ice hydrometeors and yields a more symmetrical error is 1.5 and 1.4 for 89.0 and 166.5 GHz, respectively. For example, the
 385 error in the differences in PD between simulations and observations ($PD^o - PD^b$) is diminished by an order of magnitude
 at 166.5 GHz. These values of ρ are equivalent to adjusting the extinction of the control run by $\pm 20\%$ at 89.0 GHz and by
 $\pm 16.7\%$ at 166.5 GHz in order to match the extinction at the two polarization components. In addition, they approximate
 quite well the magnitude of the K_{12} element of the extinction matrix at a tilt angle of 0° (see Fig 1b; in case of large plate
 aggregates), which describes the actual differences in the extinction between the H- and V-polarization. These findings are in
 390 line with those reported by Brath et al. (2020). The values of ρ are valid at a global scale (over both land and ocean), with a
 higher confidence at 166.5 GHz. Figure 4 shows that the simple polarized correction scheme in RTTOV-SCATT is able to do a
 reasonable job of reproducing the observed scattering arches, but compared to observations it tends to over-favour larger PDs
 at the peak of the arch (e.g., at 200 K), and PDs do not drop low enough at lower T_{BV} (e.g., 150 K). RTTOV-SCATT does
 not simulate the full arch-like relationship because it cannot transfer energy from one polarization to the other (the multiple
 395 scattering effect). However, it does reproduce some of the drop in polarization in strongly scattering scenes which is likely due
 to saturation of the scattering; the differences between τ_H and τ_V become irrelevant.

Going back to the optimum choice of the ρ , at 89.0 GHz, a more symmetrical error is achieved at a ratio of 1.45 over the
 ocean, but 1.5 overall. The IFS simulates this channel quite differently over land and ocean, partly due to its aforementioned
 multi-sensitivity aspect that complicates radiative transfer. Also, some polarization signal measured at this frequency could
 400 originate from inland water, e.g., large lakes or flooding, that have not been perfectly screened. Nevertheless, it is known that
 lower frequencies are more sensitive to larger hydrometeors (e.g., Buehler et al., 2007). In addition, the axial ratio and thus
 the orientation increases with size. Considering that the PSD by Field et al. (2007) manifests a greater preference for larger
 hydrometeors (with increasing hydrometeor content, Ekelund et al., 2020), a frequency of 89.0 GHz samples the PSD at larger
 sizes compared to 166.5 GHz, leading to a bit larger ρ . Similar but lower values of the ratio (1.3) are reported by Gong and
 405 Wu (2017) at the same dual polarization channels of GMI. In the same direction are the findings of Davis et al. (2005), but
 at the lower frequency channel of 122 GHz of the Earth Observing System (EOS) Microwave Limb Sounder (MLS). On the
 other hand, Defer et al. (2014) suggested an aspect ratio of 1.6 to be the most realistic one in reproducing the PDs observed



by MADRAS in both 89.0 and 157.0 GHz. Note here that in both Davis et al. (2005) and Defer et al. (2014), the aspect ratio has the actual microphysical sense. The various ratios among the studies are subject to the microphysical representation of the hydrometeors employed, so a hydrometeor type with a different scattering efficiency would result in a different ratio. The differing resolutions of the observations are also likely to affect the results; note that we apply "superobbing" and hence the results represent scales of order 80 km.

4.3 Linking the polarization difference to a hydrometeor type

The polarization signal prompted by preferentially oriented hydrometeors strongly depends on their microphysical representation. To further interpret this signal, ρ was applied to all ice hydrometeors individually. Horizontally oriented large scale snow and deep convective snow (graupel) are responsible for causing most of the observed PDs at 166.5 GHz, in consistency with Brath et al. (2020), with snow being associated with the largest PD values as also addressed by Defer et al. (2014) and Gong and Wu (2017). Comparing the snow- and graupel-only scaled simulations, the former one produces PD values at the entire range, while the latter one is responsible for generating mostly low to medium PDs at warm to intermediate cold T_{BV} (270 to 210 K). This is thought to come from the representation of convection in the forecast model and RTTOV-SCATT as occupying only 5% of the model grid-box. Even if a convective column generates strongly polarized scattering in RTTOV-SCATT, this limits its effect on the whole-scene brightness temperatures. In contrast, precipitation generated by the large-scale scheme typically occupies a large fraction of the grid-box and is able to generate stronger PDs in the complete scene.

In reality, similar behaviors are observed, but driven by different processes (Gong and Wu, 2017; Gong et al., 2020). In case of snow, the two-fold scattering effect described in Sect. 4.1 can explain the observed polarization patterns. A plausible interpretation of the different polarization signal induced by graupel is its higher weight; for the same amount of hydrometeors it will be less oriented, resulting in lower PD values (especially at rather deep convective situations).

Although one could assume that cloud ice hydrometeors are too small to induce any polarization at 166.5 GHz, the cloud ice-only scaled simulations conducted here suggest a minor PD (less than 1.2 K) with a preference to negative values at low brightness temperatures. Representing the cloud ice by the large column aggregate habit and the PSD introduced by Heymsfield et al. (2013) can potentially lead to large enough hydrometeors, and hence, to enough scattering to induce a visible polarization signal. This supports the suggestion by Gong et al. (2020) that some smaller PDs at 166.5 GHz may come from cloud ice. At warmer temperatures, negative PDs at 166.5 GHz were reported by Gong and Wu (2017); they linked this to clear sky conditions and instrumental noise. At colder temperatures, they also saw negative PDs, without providing any clear explanation. Here, negative PD values are simulated in deep convective areas (TB below 180 K). The most likely explanation is that RTTOV-SCATT is representing cloud ice above deep convection with relatively low single-scattering albedos, and hence, there is an emission signal that is stronger in the H-channels due to the stronger extinction. In reality, negative PDs are also measured (Davis et al., 2005; Prigent et al., 2005), with the most dominant interpretation being the vertical orientation of hydrometeors. A likely explanation of the vertical orientation is lightning activities at cloud top (hydrometeor electrification) of deep convective systems (Prigent et al., 2005). However, RTTOV-SCATT does not represent vertically oriented hydrometeors,



so if it is able to simulate a negative PD through absorption, this suggests the electrification hypothesis may only be one part of the story.

5 Conclusions

Herein, an effort has been carried out to improve the physical representation of polarized scattering in RTTOV-SCATT (Radiative Transfer model for TOVS that accounts multiple scattering) and to explore whether such an improvement would have an impact on the forecast of the European Centre for Medium-Range Weather Forecasts (ECMWF). To approximate the effect of oriented ice hydrometeors in reproducing the observed brightness temperature (T_B) differences between vertical (V) and horizontal (H) polarization, i.e., the polarization difference ($PD = T_{BV} - T_{BH}$), from conical scanning sounders, the layer optical thickness of these hydrometeors is increased in H- and decreased in V-channels. This results in the polarization ratio (ρ).

By optimising measures of fit between dual-polarization observations from the Global Precipitation Mission (GPM) microwave imager (GMI) and simulations from the Integrated Forecast System (IFS) of ECMWF, it follows that the value of ρ that models best the orientation of ice hydrometeors is 1.5 at 89.0 GHz and 1.4 at 166.5 GHz. With these settings, RTTOV-SCATT is capable of simulating the effect of oriented hydrometeors and it generates a reasonable representation of the observed arch-like relationship between PD and T_{BV} (or T_{BH}). This reduces maximum modelling errors in PD by about 10–15 K. Although the simulated PD is not perfect, the discrepancies in PD between observations and simulations are ameliorated, by an order of magnitude at 166.5 GHz, and the remaining errors are now approximately symmetrical. In the context of data assimilation (DA), assigned observation errors are quite large (e.g., up to 25 K for the GMI 166.5 GHz channels in deep convective areas), but a 15 K error reduction would still be a significant improvement.

Applying ρ to each hydrometeor type individually, we demonstrated that the polarization signal strongly depends on their microphysical representation. Assuming that IFS gives a fair representation of the real atmosphere, it is suggested that snow and graupel are responsible for causing most of the observed polarization signal, with snow producing PD values at the entire range and graupel generating mostly low to medium PDs at warm to intermediate cold brightness temperatures. Where only cloud ice was polarized, simulations give a negative polarization signal over deep convection or heavy precipitation, that could be potentially linked to enhanced emission effects in the H-channels.

Cycling DA experiments were used to examine the impact of representing polarized scattering in the observation operators for the conical scanners: GMI, Special Sensor Microwave Imager/Sounder (SSMIS), and Advanced Microwave Scanning Radiometer-2 (AMSR-2). Validation against independent references (i.e., instruments employed in DA, but where the observation operator was unchanged) showed mixed but broadly neutral changes to forecast quality – in other words, the improved modelling of polarization difference does not appear to affect the broader forecast quality. However, when the desired ρ was achieved with a unilateral scaling, e.g., in V-channels, it was possible to see positive impacts in the forecast skill measured by the conical sounder channels in H-channels, and vice versa. This suggests that representing polarization makes the forward operator more consistent between V- and H-channels of the same instrument, but also between instruments with different polarizations. For example at high frequencies (e.g., ≈ 183.0 GHz), SSMIS will generally see bigger scattering depressions than



GMI due to the different polarizations. So, if SSMIS and GMI both observe the same feature (even at different times during
 475 the assimilation window, or in different windows), then the representation of polarized scattering would allow the DA scheme
 to make more consistent corrections to the forecast. This likely improves the consistency of the forecast in areas of cloud and
 precipitation (e.g., frontal areas).

A second implication from the cycling DA experiments came from noting that a unilateral decrease of scattering in V-
 channels gave marginally better results (versus independent humidity-sensitive observations), whereas the unilateral increase
 480 in H-channels gave marginally worse results. This suggests the need for tuning the overall level of scattering in RTTOV-SCATT.
 This is particularly necessary since the selection of hydrometeor assumptions used for this work was tuned against instruments
 with predominantly H-polarization channels at high frequencies (e.g., Geer, 2020, used SSMIS). Hence, Geer (2020) carried
 out a further retuning of the microphysical assumptions, taking account of the polarization scheme introduced here, with a
 default ρ value of 1.4 for all ice species at all channels. That final retuning, along with the new polarization scheme, provides
 485 the final configuration for RTTOV-SCATT version 13. This configuration is also aimed at implementation in a future cycle of
 the IFS.

The performance of this polarization scheme has been tested only with conical scanning sounders and it is likely valid only
 at earth incident angles around 55° , where hydrometeor orientation does not change the overall level of extinction (the K_{11}
 term in the Brath et al. (2020) database, as presented in Sec. 2.4). To model the effect of hydrometeor orientation at other angles
 490 will require significant further work, since both the overall extinction and the degree of polarization must vary as a function
 of the earth incident angle. For cross-track microwave sounders this will be particularly difficult, since they are not currently
 equipped with high-frequency dual-polarization channels with which to validate the results, and the polarization rotates across
 the swath. A subsequent step would be to explore the performance of ρ at higher sub-millimeter wavelengths; the upcoming Ice
 Cloud Imager (ICI) mission, with dual-polarization channels at frequencies that are sensitive to the column scattering due to ice
 495 hydrometeors (243.2 and 664.4 GHz), will provide further insights regarding the polarization signal due to small oriented ice
 hydrometeors. So far, the issue under scrutiny in most studies is either totally randomly or horizontally oriented hydrometeors.
 The recently developed scattering database of azimuthally oriented ice hydrometeors (Brath et al., 2020) gives a path towards
 further developments in RTTOV-SCATT and the IFS.

Code and data availability. The web address for RTTOV-SCATT is nwp-saf.eumetsat.int/site/software/rttov, where the model can be ob-
 500 tained through registration. Due to the large volume of data generated by the assimilation experiments of the IFS, its permanent archiv-
 ing/curation is not possible. But, reanalysis data, e.g., ERA-5 (Fifth generation of ECMWF atmospheric reanalyses of the global climate), is
 available.



Appendix A: Screening method

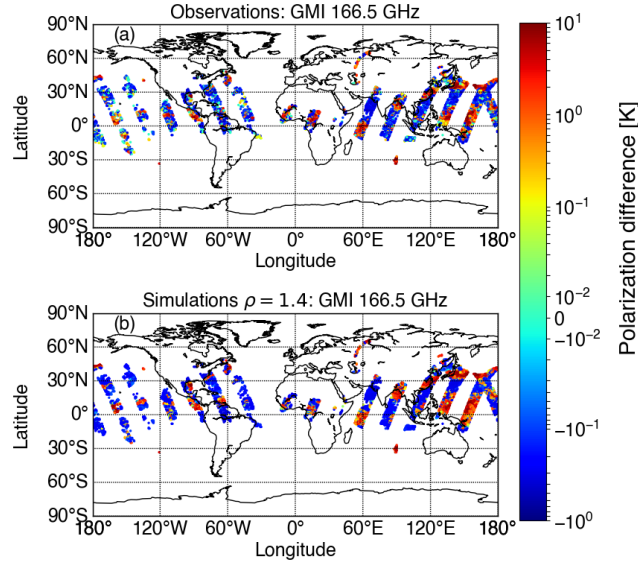


Figure A1. Performance of the screening method at an example scene measured by GMI on 6 July 2019 at a frequency of 166.5 GHz. Polarization difference, i.e., brightness temperature differences between the vertical and horizontal polarization ($PD = T_{BV} - T_{BH}$), due to oriented ice hydrometeors as (a) observed (PD^o) by GMI and (b) simulated (PD^b) for a polarization ratio (ρ) of 1.4.

To minimize the surface contribution, the clear-sky surface-to-space transmittance (t) simulated by RTTOV has been employed.
 Recall here that the superscript b corresponds to the simulated (background) quantities. Accordingly, the surface contribution to the polarization can be approximated (ignoring the downward radiance term) by:

$$PD_{sfc}^b = T_{BV,sfc}^b - T_{BH,sfc}^b \simeq (e_V - e_H) \cdot T_{sfc} \cdot t_{166.5}. \quad (A1)$$

Where e_V and e_H are the surface emissivities at V- and H-polarization, respectively, T_{sfc} is the surface temperature, and $t_{166.5}$ denotes the transmittance at 166.5 GHz. An upper threshold $t_{166.5}$ value of 0.05 has been adopted to mask out the surface contribution at both dual-polarization channels. This can be translated (via Eq. A1) to a maximum surface induced polarization difference of 2.9 K. Lower values of the clear-sky $t_{166.5}$ have been also tested, i.e., 0.01–0.04 (0.58 K–2.32 K in PD_{sfc}^b), to ensure that the polarization signature originates from oriented ice hydrometeors and not from clear-sky conditions and thus, from the surface. To isolate cloudy GMI pixels, we additionally defined the hydrometeor impact as:

$$\begin{aligned} \Delta T_B^o &= T_{B,cloudy}^o - T_{B,clear}^b, \\ \Delta T_B^b &= T_{B,cloudy}^b - T_{B,clear}^b. \end{aligned} \quad (A2)$$

ΔT_B^o (ΔT_B^b) represents the hydrometeor impact defined by the observed (simulated) brightness temperature in all-sky conditions minus the simulated equivalent atmosphere without clouds. Any situation with a hydrometeor impact above 0 K is



rejected; this is done on a channel-wise basis and the rejection is done if either observations or simulations cross this threshold. The bias correction is included at 166.5 GHz, but excluded at 89.0 GHz because it led to a mismatch between observations and simulations. This two-fold screening method does not completely eliminate surface polarization signatures at 89.0 GHz. Hence, an additional basic filtering has been applied to this channel only. First, pixels where the control run yields a PD^b greater than 1 K are masked out, since the control run does not produce any polarization due to preferentially oriented hydrometeors. Second, pixels where the observed PD (PD^o) is greater than 5 K and $T_{BV}^o > 265$ K are excluded (this is the visible rectangular bite out of the distribution in Fig. B1a).

Figure A1a displays the performance of the screening method at an example scene measured by GMI on 6 July 2019 at 166.5 GHz. This method minimises any surface contamination, albeit it is rather strict and screens out thin cirrus clouds, especially over the mid-latitudes. Figure A1b demonstrates the very good performance of ρ value of 1.4 to simulate PD.

Appendix B: Two-dimensional histograms of the arch-like relationship at 89.0 GHz

Figure B1 displays the two-dimensional histogram of the arch-like relationship between the polarization difference and the brightness temperature at V-polarization at 89.0 GHz as observed by GMI (Fig. B1a) and simulated for a polarization ratio of 1.5 (Fig. B1b). Simulations fail to reproduce the full arch-like relationship. Due to the screening method, most moderately cloudy scenes over the ocean have been excluded, leading to a rather small sample. However, both the main PD^o branch ($230 < T_{BV} < 275$ K) and the PD^o peak (≈ 10 K) have been modeled quite well by a ρ value of 1.5.

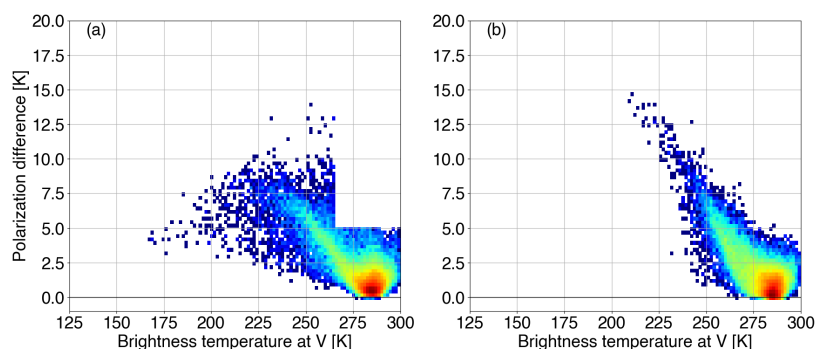


Figure B1. Two-dimensional histograms describing the arch-like relationship between the PD and the brightness temperature at V-polarization at 89.0 GHz as (a) observed ($PD^o - T_{BV}^o$) by GMI and (b) simulated ($PD^b - T_{BV}^b$) for a polarization ratio of 1.5.

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