Sensitivity of 89–190-GHz Microwave Observations to Ice Particle Scattering

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ABSTRACT

The sensitivity of microwave brightness temperatures (TBs) to hydrometeors at frequencies between 89 and 190 GHz is investigated by comparing *Fengyun-3C* (FY-3C) Microwave Humidity Sounder-2 (MWS-2) measurements with radar reflectivity profiles and retrieved products from the Global Precipitation Measurement mission's Dual-Frequency Precipitation Radar (DPR). Scattering-induced TB depressions (ΔTB s), calculated by subtracting simulated cloud-free TBs from bias-corrected observed TBs for each channel, are compared with DPR-retrieved hydrometeor water path (HWP) and vertically integrated radar reflectivity Z_{INT} . We also account for the number of hydrometeors actually visible in each MWS-2 channel by weighting HWP with the channel's cloud-free gas transmission profile and the observation slant path. We denote these transmission-weighted, slant-path-integrated quantities with a superscript asterisk (e.g., HWP*). The so-derived linear sensitivity of ΔTB with respect to HWP* increases with frequency roughly to the power of 1.78. A retrieved HWP* of 1 kg m^{-2} at 89 GHz on average corresponds to a decrease in observed TB, relative to a cloud-free background, of 11 K. At 183 GHz, the decrease is about 34–53 K. We perform a similar analysis using the vertically integrated, transmission-weighted slant-path radar reflectivity Z_{INT}^* and find that ΔTB also decreases approximately linearly with $(Z_{INT}^*)^{0.58}$. The exponent of 0.58 corresponds to the one we find in the purely DPR-retrieval-based Z_{INT} –HWP relation. The observed sensitivities of ΔTB with respect to Z_{INT}^* and HWP* allow for the validation of hydrometeor scattering models.

1. Introduction

The relationships between passive microwave brightness temperatures (TBs) and the microphysical characteristics of hydrometeors have long been recognized as critical for many aspects of remote sensing, including the derivation of precipitation information (Petty 1994a,b; Smith et al. 1992; Wang et al. 1998; Wilheit 1986). Knowledge of these relationships is essential to develop and assess cloud-resolving models (Tao and Simpson 1993), to improve forward radiative transfer calculations (Smith et al. 1992), to compute vertical diabatic heating profiles (Tao et al. 1993), and to develop precipitation parameter retrieval algorithms (Marzano et al. 1999; Smith et al. 1992). To gain

information about these relationships, many theoretical studies (Smith et al. 1992; Yeh et al. 1990) and experimental studies (Bennartz and Bauer 2003; Heymsfield et al. 1996; Liu and Curry 1998; Simpson et al. 1996) have been conducted.

The simulation of individual scattering properties of nonspherical precipitation-sized ice particles and their effects on radar and passive microwave observations of clouds and precipitation has drawn significant attention in the last years. Tremendous progress has been made in understanding the scattering properties of nonspherical ice particles (Ekelund and Eriksson 2019; Eriksson et al. 2015; Hogan and Westbrook 2014; Hogan et al. 2017; Hong 2007; Hong et al. 2009; Kim 2006; Kuo et al. 2016; Leinonen et al. 2018; Liu 2004; Lu et al. 2016; Mason et al. 2019; Ori et al. 2014; Petty and Huang 2010; Tyynelä et al. 2014). New methods and databases based on these findings are currently being used for various purposes including precipitation retrieval (Ringerud et al. 2019;

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Skofronick-Jackson et al. 2019; Yin and Liu 2019), data assimilation (e.g., Geer and Baordo 2014), and ice water path retrievals (Brath et al. 2018; Buehler et al. 2012; Piyush et al. 2017). The simulated scattering properties of ice particles must ideally and realistically represent the optical properties of actual ice particles in terms of both active and passive microwave remote sensing applications. In this context, closure studies between multifrequency active and passive observations of the scattering signals of ice clouds are of critical importance as they can provide constraints on particles properties in terms of their active scattering properties, radar dual/triple-frequency ratio, for example, Kneifel et al. (2011), and their passive microwave optical properties (Bytheway and Kummerow 2018; Fox et al. 2019; Gong and Wu 2017; Kulie et al. 2010).

Here we use the cross-track scanning Microwave Humidity Sounder-2 (MWS-2) on board *Fengyun-3C* (FY-3C) to globally assess the impact of hydrometeor scattering on observed TBs over the range 89–190 GHz including the 118-GHz oxygen absorption channels. Various observation-based and modeling studies have investigated the impact of hydrometeor scattering on passive microwave brightness temperatures (Bennartz and Petty 2001; Bennartz and Bauer 2003; Bennartz et al. 2002; Burns et al. 1997; Gasiewski 1992; Grody 1991; Skofronick-Jackson and Wang 2000; Spencer 1986; Spencer et al. 1989; Weng and Grody 2000). The combination of 89-, 150-, and 183-GHz observations allows for an improved retrieval of precipitation ice microphysical properties using the differential scattering signature (Skofronick-Jackson and Johnson 2011; Skofronick-Jackson and Wang 2000; Weng and Grody 2000). Bennartz and Bauer (2003) studied the response of brightness temperatures to hydrometeor scattering based solely on simulations at 89, 150, and 183 GHz. Their research shows that 150 GHz is highly sensitive to variations in ice scattering, and water vapor channels ($183.31 \pm X$ GHz) give additional information because of their low sensitivity to surface emission. While these channels have been exploited to understand ice scattering properties, temperature sounding channels ($118.75 \pm X$ GHz) have been relatively unexplored (Bauer and Mugnai 2003). Building conceptually upon these previous findings, we focus on two questions that can be answered by combining active and passive observations. First, at any given frequency, what is the average sensitivity of the brightness temperature to the presence of ice-phase precipitation in the observed atmospheric column? Second, what is the frequency-dependence of this sensitivity in the range 89–190 GHz? To answer these questions, we create a global, ocean-only 1-yr dataset of combined spaceborne

radar observations from the Global Precipitation Measurement mission's (GPM) Dual-Frequency Precipitation Radar (DPR) and FY-3C MWS-2 TBs. Based on this dataset, we study the sensitivity of observed TBs to changes in vertically integrated radar reflectivity and in hydrometeor water path (HWP), which we define as the total amount of all precipitation ice and all liquid in a vertical column. In our analysis we use “scattering-induced brightness temperature depressions” (ΔTB), which is defined as the difference between observed TBs and simulated cloud-free TBs, the latter based on collocated numerical weather prediction (NWP) fields. We pay particular attention to issues related to the absolute calibration of the passive microwave sensor and its ability to reproduce bias-free and cloud-free observed TBs for all channels. To quantify the effects of scattering alone, rather than the effect of combined scattering and gas absorption, we further develop a method to account for the impact of variable gas transmission that most strongly affects the temperature channels near 118 GHz and water vapor absorption channels near 183 GHz. The approach we use here is very similar to Wu et al. (2009) and Gong and Wu (2014) who deal with high-frequency microwave and submillimeter observations of cloud ice water path (IWP).

MWS-2 is particularly well suited for the purpose of the current study as it is the first spaceborne sensor equipped with temperature sounding channels around 118 GHz in addition to the more traditional channels around 89, 150, and 183 GHz (Dong et al. 2009; Zhang et al. 2012). In the near future, the Time-Resolved Observations of Precipitation Structure and Storm Intensity with a Constellation of Smallsats (TROPICS) and EUMETSAT's Microwave Imager (MWI) will also provide observations at 118 GHz (Blackwell et al. 2018; Holmlund et al. 2017; Mattioli et al. 2019). The new channels around 118 GHz are of interest for temperature sounding but also in the context of precipitation retrievals (Bauer and Mugnai 2003; He and Chen 2019).

The rest of this paper begins by describing the MWS-2 dataset, the DPR reflectivity measurements and associated radar-retrieved hydrometeor profiles, and the method to preprocess the data to obtain collocated and coincident matchups between MWS-2 and DPR. We then describe the forward modeling process to obtain cloud-free simulated TBs and to develop a bias correction to the MWS-2 observations. Next, we outline how we account for variable gaseous transmission in the process of comparing HWP and other radar-derived quantities to MWS-2. We then proceed to study the sensitivity of the individual

MWHS-2 channels to the presence of hydrometeors and to evaluate the spectral dependency of this sensitivity.

2. Instruments, data, and methods

a. Description of datasets

The MWHS-2 on board the midmorning polar-orbiting *FY-3C* satellite, launched by the China Meteorological Administration/National Satellite Meteorological Center (CMA/NSMC) on 23 September 2013, is a cross-track, millimeter-wave radiometer. MWHS-2 includes two window channels at 89 and 150 GHz. Five sounding channels are working near 183.31 GHz. MWHS-2 is also the first spaceborne sensor equipped with eight channels near 118.75 GHz and therefore provides new observations that allow for a global assessment of such channels. With these 15 channels (shown in Table 1 along with polarizations), MWHS-2 can retrieve atmospheric temperature profiles, humidity profiles, and water vapor information simultaneously. The spatial resolution is near 16 km at nadir for 183 GHz and 29 km for 89 GHz.

We note that the polarization values listed in Table 1 (as well as in Lawrence et al. 2015) are reversed from the polarization values reported for MWHS-2 in the World Meteorological Organization's Observing Systems Capability Analysis and Review Tool (OSCAR) database (see <https://www.wmo-sat.info/oscar/instruments/view/341> for *FY-3C*'s MWHS-2). The determination of polarization at nadir is a matter of definition, which likely causes this discrepancy. We have numerically verified that the setting reported here align correctly with the values needed in the radiative transfer model we use. MWHS-2 Level-1 files were obtained from the CMA/NSMC (<http://satellite.nsmc.org.cn/portalsite/default.aspx>).

The GPM *Core Observatory* carries both a DPR and a GPM Microwave Imager (GMI) (Hou et al. 2014). The GPM orbit inclination of 65° is such that it enables the orbit to cut across the orbits of sun-synchronous satellites, such as *FY-3C*. It also allows for the gathering of samples at latitudes where most precipitation occurs in terms of absolute amount at various times of the day. The DPR consists of a Ku-band precipitation radar (KuPR at 13.6 GHz) and a Ka-band precipitation radar (KaPR at 35.5 GHz). Data collected from the KuPR and KaPR units provide three-dimensional observations of rain/snow and also provide an accurate estimation of rainfall rate to the scientific community. The radar reflectivity measured by and hydrometeor profiles retrieved from DPR will be compared to the TB of MWHS-2 to study the sensitivity of radiance at microwave high-

TABLE 1. MWHS-2 channel frequencies and polarization at nadir used in RTTOV (H = horizontal; V = vertical).

Channel	Frequency (GHz)	Polarization at nadir used in RTTOV
1	89	H
2	118.75 ± 0.08	V
3	118.75 ± 0.2	V
4	118.75 ± 0.3	V
5	118.75 ± 0.8	V
6	118.75 ± 1.1	V
7	118.75 ± 2.5	V
8	118.75 ± 3.0	V
9	118.75 ± 5.0	V
10	150	H
11	183.31 ± 1.0	V
12	183.31 ± 1.8	V
13	183.31 ± 3.0	V
14	183.31 ± 4.5	V
15	183.31 ± 7.0	V

frequency channels to hydrometeor scattering, which allows us to explore the potential capability of spaceborne observations from these channels for retrieving hydrometeor profiles.

The KuPR has an observation swath of 245 km with 49 beams, each resulting in a circular footprint with a 5.2-km diameter. The KaPR has a swath of 120 km with 49 beams of which 25 beams are matched to the central 25 beams of KuPR. This narrower swath with matched Ka and Ku reflectivities is termed matched scan (MS). The KuPR is more sensitive to heavy and moderate rainfall and the KaPR is more sensitive to light rainfall and snow. The combination of measurements from both radars will allow for a more comprehensive understanding of the ice particle scattering to which the microwave observations are sensitive at different degree levels. Therefore, in this paper we used the 2BCMB product at MS scans, to obtain the reflectivities of both radars and hydrometeor retrievals, including precipitation profiles. The 2BCMB product uses data of KuPR and KaPR, and GMI to determine the precipitation structure that best fits the combined data from these instruments. The resolution of the 2BCMB product is 250 m vertically and 5 km horizontally. To account for the total amount of liquid and ice in a column as fully as possible, we herein consider the HWP to consist of three parts: IWP, precipitation water path (PWP), and cloud liquid water path (LWP), where each of these three is the vertical integral of the corresponding vertically resolved water content. For example, LWP is the vertical integral of liquid water content (LWC). Precipitation water content (PWC) in 2BCMB consists of the mass of all liquid rain and frozen precipitation-sized ice particles and is directly retrieved from GPM observations.

TABLE 2. Symbols and quantities used in this paper.

Symbol	Quantity	Units	Comment
h	Height above surface	m	We use h as a vertical coordinate so as to leave Z reserved for radar reflectivity
f	Frequency	GHz	
θ	Satellite zenith angle	°	
$X = X(h)$	General quantity; X as function of h		X can be any quantity such as IWC, PWC, HWC, or Z
$\tau = \tau(f, h, \theta)$	Transmission (dry gases + water vapor)	—	Transmission between height h and the top of the atmosphere along the satellite's line of sight; in the context of this study it refers to gas transmission, which is defined along the sensor's line of sight and hence depends on the satellite zenith angle
IWC(h)	Ice water content	kg m ⁻³	Cloud ice, being zero in this study; corresponds to variable named "cloudIceWaterCont" in GPM 2BCMB HDF files
LWC(h)	Cloud liquid water content	kg m ⁻³	Cloud liquid; provided by the VER of the Radar Algorithm, which interpolates the JMA analyses and forecasts to the DPR locations/range bins; corresponds to variable named "cloudLiqWaterCont" in GPM 2BCMB HDF files
PWC(h)	Precipitation water content	kg m ⁻³	Precipitation water including liquid-phase and ice-phase precipitation; retrieved from the GPM observations using the GPM combined algorithm; corresponds to variable named "precipTotWaterCont" in GPM 2BCMB HDF files.
HWC(h)	Hydrometeor water content	kg m ⁻³	=IWC + LWC + PWC, combining all hydrometeors.
$Z(h)$	Radar reflectivity	mm ⁶ mm ⁻³	Corrected radar reflectivities; corresponds to variable named "correctedReflectFactor" in GPM 2BCMB HDF files; reflectivities from KuPR and KaPR were investigated individually
IWP	Ice water path	kg m ⁻²	Cloud ice water path, calculated from IWC and Eq. (2), being zero in this study
LWP	Cloud liquid water path	kg m ⁻²	Cloud liquid path, integrated from LWC using Eq. (2)
PWP	Precipitation water path	kg m ⁻²	Liquid precipitation, integrated from PWC using Eq. (2)
HWP	Hydrometeor water path	kg m ⁻³	=IWP + LWP + PWP, combining all hydrometeors; IWP is zero
Z_{INT}	Integrated radar reflectivity	mm ⁶ mm ⁻²	Integrated from $Z(h)$ using Eq. (2)
$\text{HWP}^*(f, \theta)$	Modified hydrometeor water path	kg m ⁻³	Amount of hydrometeors actually visible at a given frequency and zenith angle, calculated by weighting HWP with the channel's cloud-free gas transmission profile and the observation slant path using Eq. (3)
$Z_{\text{INT}}^*(f, \theta)$	Modified integrated radar reflectivity	mm ⁶ mm ⁻²	Amount of Z_{INT} actually observed at a given frequency and zenith angle, calculated by weighting Z_{INT} with the channel's cloud-free gas transmission profile using Eq. (3)

Thus, PWP holds the total mass of precipitating particles per square meter in the column. Note that GPM observations are not sensitive to small ice crystals. Therefore, cloud ice water content (IWC), is assigned to be zero in the 2BCMB product. In consequence, the IWP is also zero and $\text{HWP} = \text{LWP} + \text{PWP}$ in this study. The PWC in the 2BCMB product was retrieved from the GPM observations using the GPM combined algorithm and is composed of liquid-phase and ice-phase precipitation. The cloud LWC in this product differs in that it was provided by the "Vertical Profile Submodule" (VER) of the "Radar Algorithm," which interpolates the Japan Meteorological Agency's (JMA) analyses and forecasts to the DPR locations/range bins (Grecu et al. 2016; Olson et al. 2018). We discuss the impact of LWP on observed brightness temperatures later in the paper. Table 2 provides the symbols, quantities, units, and supplementary information of

these parameters and the others used in this paper. The 2BCMB product was provided by the NASA Precipitation Processing System archived at the NASA GES DISC (<https://doi.org/10.5067/GPM/DPRGMI/CMB/2B/06>).

b. Collocated and coincident measurements

In this study, observations were regarded as collocated and coincident when MWHS-2 brightness temperatures and GPM profiles fell into the same 0.25° latitude $\times$ 0.25° longitude bin and the time difference between the two was less than 15 min. Because the footprint of the MWHS-2 channels 1–9 is about 29 km at nadir, which is approximately equivalent to the grid size (25 km), there were only a few MWHS-2 pixels falling into the same grid (typically fewer than three). Therefore, the measurements from the pixel closest to the center of a grid in distance are extracted to

represent that grid. Contrarily, typically dozens of GPM observations fell into one 0.25° latitude $\times$ 0.25° longitude bin because the DPR footprint (5 km) is much smaller than the grid size. Thus, we averaged all the GPM measurements within one grid to represent the mean over this grid cell. To limit issues caused by the high zenith angles and the low spatial resolution of MWHS-2 at the outer edge of its scan, the five outermost scan positions on each side of the scan were excluded from the collocated dataset. We further limit this study to open oceans only in order to avoid issues related to the simulation of surface emissivity over land. At the end of the collocation process, a total of more than 1.5 million samples over oceans globally for the year of 2017 were generated.

3. Cloud-free radiative transfer simulations and bias correction

a. Atmospheric profiles and surface parameters

ERA-Interim is a third-generation global atmospheric reanalysis provided by the European Centre for Medium-Range Weather Forecasts (ECMWF). It improves on previous versions (e.g., ERA-40) by using an enhanced atmospheric model and assimilation system (Dee et al. 2011). The spatial resolution of the dataset is approximately 80 km (T255 spectral) on 60 vertical levels from the surface up to 0.1 hPa. Among others, it assimilates a significant amount of satellite radiances, (e.g., SSM/I and SSMIS). ERA-Interim data are available since 1979, and it had been continuously updated until 31 August 2019 when it was replaced by ERA5. For this study, 6-hourly surface as well as vertically resolved moisture and temperature field products were used (downloaded from <https://rda.ucar.edu/datasets/ds627.0/>).

Each MWHS-2 observation was matched to the closest ERA-Interim reanalysis profile, leading to maximum time differences between the observations and the reanalysis of 3 h. The ERA-Interim profiles, together with sea surface temperature and surface wind speed, formed the basis of the subsequent cloud-free radiative transfer simulations.

b. Radiative transfer simulations

Radiative transfer simulations were carried out using the TIROS Operational Vertical Sounder Radiative Transfer (RTTOV, version 12.2) Model (Hocking et al. 2019; Saunders et al. 2018, 2007), which is widely used both in the operational and research community. All physical parameterizations including absorption by gases, liquid water, and rough ocean surface scattering were applied unchanged from RTTOV. Radiative transfer

simulations were implemented for all 15 channels listed in Table 1 with RTTOV polarization at nadir as also listed in this table.

c. Bias correction

The bias correction was performed based on histograms of differences between observed and simulated TBs. Figure 1 serves as an example for this approach. For each channel, the histogram of brightness temperature differences was analyzed and the mode of the histogram was used for bias correction. The idea is that the mode of the histogram corresponds to the observations that are least affected by precipitation and therefore produces a more robust estimate of bias than the mean or median.

The process was repeated for each channel and each scan position, leading to the bias-correction values shown in Fig. 2. These values can be directly compared with the values reported in Lawrence et al. (2015, see their Figs. 7 and 12). While they use a different version of RTTOV as well as a different underlying NWP model, their bias-correction values and in particular the dependency of bias correction on scan position are very similar to our values shown in Fig. 2. Significant differences between our results and those reported in Lawrence et al. (2015) exist in particular in channel 14. They report biases between 1 and 3.5 K depending on scan position and we report biases between 0.5 and 6 K with a similar dependency on scan position. It is unclear what these large differences in bias can be attributed to. Lawrence et al. (2015) use a slightly different approach of rejecting precipitation-affected observations. They calculate a scattering index (SI) based on observed and simulated TBs in channels 1 and 2. Whenever the SI exceeded 5 K, the data were rejected and the bias was only estimated on the remaining data, which are only weakly affected by scattering. To test whether this different approach led to different values for the bias correction, we implemented this approach as well. However, the results were not significantly different from the results reported in Fig. 2.

The biases reported in Fig. 2 are subsequently used to correct observed TBs toward the simulations. That is, the bias reported in Fig. 2 is subtracted from the observations before further analysis.

d. Weighting functions under cloud-free conditions

We use the collocated dataset to study the variability of MWHS-2 weighting functions under cloud-free conditions as function of viewing angle, expressed in terms of relative airmass factor, and total column water vapor amount (TCWV), the latter being a proxy for the general type of atmosphere observed. Midlatitude atmospheres typically exhibit TCWV on the order of $20\text{--}30\text{ kg m}^{-2}$, more tropical

FIG. 1. Histograms of observed minus simulated brightness temperatures for all 15 channels of MWHS-2. These histograms contain data for scan position 44 for the entire year of 2017. The different colored vertical lines correspond to the same-colored statistics given in the upper-left part of each panel. The mode of each histogram was used as the basis for bias correction.

FIG. 2. Bias-correction values derived and used in this study as function of scan position for all 15 MWHS-2 channels.

FIG. 3. Peak of cloud-free weighting function for all 15 MWHS-2 channels as a function of viewing angle θ , expressed in terms of relative airmass factor, and TCWV. The values given are averaged over all collocated profiles, binned in intervals of 0.1 airmass factors for airmass factors between 1 (observation zenith angle $\theta = 0^\circ$) and 2 ($\theta = 60^\circ$), and 2.5 kg m^{-2} TCWV intervals in a range between 0 and 60 kg m^{-2} . Isoline labels as well as the color bar are in kilometers. Mean heights μ for each channel and corresponding standard deviations σ are given in each panel's title.

atmospheres around 40 kg m^{-2} or beyond. Figure 3 shows the height of the peak of the weighting function for each channel as function of TCWV and relative airmass factor. Various features can be observed:

- 1) Channel 1 (89 GHz) peaks near the surface. Only for very high TCWV and high sensor zenith angles will the peak of the weighting function move up to about 1 km. Channel 1 will therefore be affected by nonprecipitating warm clouds also in the lower atmosphere.
- 2) The 118.75-GHz channels 2–4 peak in the lower stratosphere or high up in the troposphere at altitudes between 17 and 26 km and will therefore show little sensitivity to the lower atmosphere. Only in exceptional cases would convection reach so high that these channels show a significant signal.
- 3) The 118.75-GHz channels 5 and 6 peak around 10 km altitude and will therefore be affected by most deep convective events.
- 4) The 118.75-GHz channels 7–9 and channel 10 (150 GHz) peak between 4 km and the surface. These channels transition from a stronger sensitivity to observation angle (isolines run more vertically), to a stronger sensitivity to atmospheric moisture (isolines run more horizontally). Thus, for example, the lowest-peaking 118-GHz channel 9 ($118 \pm 5 \text{ GHz}$) and channel 10 show sensitivity to both observation angle and TCWV.
- 5) The 183.31-GHz channels 11–15 peak between 7 and 2 km and show stronger sensitivity to water vapor than to zenith angle.

These findings are generally in agreement with earlier studies (Lawrence et al. 2015) and with the general expectations one would have for a MWS-2-like sounding system under cloud-free conditions. Clouds and precipitation will modify observed brightness temperatures from their cloud-free values only if they reach high enough. For the purpose of precipitation retrievals therefore three findings are relevant. First, the stratospheric channels 2–4 will have little direct information on precipitation because of their high weighting functions. This will be confirmed later. Second, the lower-peaking 118-GHz channels, particularly channels 7 and 8, provide information orthogonal to the 183-GHz water vapor sounding channels, where former are more sensitive to zenith angle changes and the latter more sensitive to TCWV changes. Third, the window channels (1, 9, and 10) have significant sensitivity to the surface and will thus be sensitive also to precipitation and clouds in the lower atmosphere.

We next move on to studying the impact of clouds and precipitation on the observed brightness temperatures. We do this first by studying the difference between observed all-sky brightness temperatures and simulated

cloud-free brightness temperatures. After that, we explicitly calculate the amount of precipitation in the column that is visible for each observation and frequency by explicitly accounting for the atmospheric transmission.

4. Brightness temperature response to clouds and precipitation

a. Observed deviation from cloud-free brightness temperatures

In this study, the scattering-induced brightness temperature depression (ΔTB) is defined as the difference between bias-corrected microwave observations, TB_{obs} , and simulated clear-sky background brightness temperatures, as outlined in section 3, TB_{sim} :

$$\Delta\text{TB} = \text{TB}_{\text{obs}} - \text{TB}_{\text{sim}}. \quad (1)$$

Figure 4 shows an example of the observed brightness temperatures and the resulting scattering-induced brightness temperature depression. The convection cell located around 2°N , 68°E shows significant depressions of brightness temperatures with respect to cloud-free simulated brightness temperatures. Channels 10 and 15 both show ΔTB s exceeding -100 K . In contrast, channel 1 shows a ΔTB of only -20 K . In general, going from lower to higher frequencies, one can expect the scattering signature to increase, as the size parameter of the scatterer increases with decreasing wavelength. Additionally, the oxygen and water vapor sounding channels exhibit a strong dependency on how close each channel is to the center of its corresponding absorption line. This effect is caused by the weighting functions of the sounding channels peaking higher in the atmosphere the closer they are to the center of the absorption line. Ultimately, the channels will peak so high in the atmosphere that even deep convection is not visible at all. This is the case for example for channels 2–4 in Fig. 4.

Addressing the sensitivity of ΔTB to ice particle scattering therefore requires disentangling the effect of broad spectral variations in actual scattering intensity from the effect of gas absorption that attenuates the scattering signal near the absorption lines. Further down, we present an approach to do exactly that. However, before we introduce this approach, we first evaluate the range of variability of ΔTB expected for the entire dataset.

Figure 5 presents two-dimensional probability density distribution functions of TB_{obs} and ΔTB for all 15 MWS-2 channels. Warm TB_{obs} values are mostly from the clear sky, the surface, or from nonscattering clouds, while colder TB_{obs} include scenes with precipitation,

FIG. 4. Example of the scattering signature of deep convective clouds observed by MWHS-2. (a) One full orbit of MWHS-2 data (channel 9; 22 Aug 2016, with start time 0554 UTC). (b) The same data as in (a), enlarged around an area in the Indian Ocean that contains isolated deep convection, visible as the blue and green dots in the center-right of the panel. (c) The spectral variation of ΔTB for the center of one of the convective cells, as highlighted by the arrow to (b); the plot in (c) also shows the location of the 118.75-GHz oxygen absorption line, the 183.31-GHz water vapor absorption line, and the MWHS-2 channel numbers corresponding to Table 1. Note that MWHS-2 sounding channels 2–9 and 11–15 are all double sideband (e.g., channel 9 is 118.75 ± 5 GHz). In (b), their corresponding ΔTB are only shown at the location of the upper subband (e.g., $118.75 + 5$ GHz).

in particular precipitation ice, but can also occur under low-wind, cold and dry, cloud-free conditions. Regardless of how warm the absolute TB_{obs} are, a perfect RTM model with perfect clear-sky input would produce near-zero values in ΔTB for all cloud-free conditions, while large negative ΔTB values are generated in the presence of precipitation that includes the ice phase.

The typical response of high-frequency channels to hydrometeor scattering can be clearly identified for channels 7–15. While most data lie on around the $\Delta TB = 0$ K line, a subset of the data shows a ΔTB decreasing linearly with decreasing TB_{obs} and that the ΔTB eventually decreases to -100 K or more. Channels 5 and 6, which peak higher in the atmosphere, show a similar behavior, but less pronounced.

It is important to note here that TB_{obs} and ΔTB provide independent information. This can be observed for example for channel 9. For TB_{obs} between 200 and 280 K, the probability density plot shows a bifurcation between those data that follow the $\Delta TB = 0$ K line and those data for which ΔTB decreases approximately linearly with decreasing TB_{obs} . If only TB_{obs} was evaluated and a brightness temperature of $TB_{obs} = 220$ K was observed, one would not be able to know whether it was observed under conditions compatible with clear sky (ΔTB around

0 K) or whether it was observed under heavy hydrometeor scattering (ΔTB around -50 K). Also, under precipitation-free conditions, observations in channels 1, 9, and 10 (window channels) might be affected by cloud liquid water from nonprecipitating clouds. Since clouds are not represented in the cloud-free simulations, this causes the ΔTB values to be above the $\Delta TB = 0$ line, for cases where there is no significant precipitation in the column.

The highest-peaking channels 2–4 are not sensitive to hydrometeor scattering and only exhibit slight deviations of ΔTB from zero. Because of their insensitivity to hydrometeor scattering, channels 2–4 will be excluded in subsequent analysis.

In addition to hydrometeor scattering, the emission caused by cloud liquid water and liquid precipitation will also affect brightness temperatures as well as ΔTB for those channels that peak low enough in the atmosphere to be sensitive to it. This emission signal has an effect opposite from the scattering-induced brightness temperature depression. It can be most clearly observed for channel 1 (89 GHz), which is the least-opaque of all channels. For channel 1, one can observe not only ΔTB values below zero, but also a significant portion of ΔTB values above zero, up to $+100$ K. To a lesser extent, channels 7–10 show this behavior, too. Channels

FIG. 5. Two-dimensional probability density distributions of TB_{obs} and ΔTB for all 15 channels of MWHS-2 and for all collocated data. Note the different scales of both the x axis and the y axis.

11–15 (183-GHz water vapor absorption channels) show very little of this effect.

To further understand the effect of ice and liquid in the column, we next evaluate the TB_{OBS} with ΔTB in the context of DPR observations. To avoid any potential issues associated with DPR retrievals, we first look at DPR reflectivity profiles, which are direct measurements and not affected by any assumptions made in DPR retrievals.

b. Spectral sensitivity to precipitation

To compare DPR reflectivity profiles with TB_{OBS} and with ΔTB , we calculate for each radar profile two

integral properties, namely, the total vertically integrated reflectivity as seen by a given MWHS-2 channel and the average height from which this reflectivity originates. This concept will be used further down in a similar manner for HWP and requires some explanation, which is provided next for a general quantity $X(h)$. The vertical integral of any atmospheric quantity $X(h)$ is

$$X_{\text{INT}} = \int_0^{\text{TOA}} X(h) dh. \quad (2)$$

The term $X(h)$ can in principle be any quantity. In this paper, we are mostly concerned with GPM-observed

FIG. 6. Conceptual model relating the different quantities in this section. (left) HWC as an example of a vertically resolved atmospheric parameter. The horizontal line (marked H) shows the average hydrometeor height. (center) The gas transmission for two example channels: one peaking higher in the atmosphere (blue; weighting function peaks at 9 km) and another one peaking lower in the atmosphere (red; weighting function peaks at 4 km). (right) The integrand of Eq. (2). The numbers on the right provide the results of Eqs. (3) and (4) for this case. The values of H^* are also shown as colored horizontal lines in the right panel.

radar reflectivities and hydrometeor products. Consider for example the generic function $X(h)$ to represent cloud LWC in units of kilograms per meter cubed. The corresponding integrated quantity X_{INT} then has units of kilograms per meter squared and is typically called cloud LWP. This principle also applies to PWP and cloud IWC in a vertical column that add up to PWP and IWP, respectively. Note that in the DPR data, cloud ice (IWC) is zero, so that IWP here is also zero, and therefore the HWP in the context of this paper is the sum of LWP and PWP.

As pointed out above, low-frequency passive microwave satellite observations rely on the relative transparency of the atmosphere (window channels) at low frequencies to derive column-integrated quantities, such as LWP. However, sounding instruments, such as MWHS-2, by design, provide several channels at frequencies with strong gas absorption (sounding channels). The weighting functions of these channels peak higher in the atmosphere and lower parts of the atmosphere are obstructed. In addition, the atmosphere is observed at a slant path for any satellite zenith angle larger than zero. The gross impact on weighting functions is discussed in the above section 3a. We now quantify which part of the vertical precipitation column is, in principle, visible to each channel.

Accounting for gas absorption and slant path the definition of X_{INT} can be expanded as follows:

$$X_{\text{INT}}^* = \frac{1}{\cos\theta} \int_0^{\text{TOA}} \tau(h, f, \theta) X(h) dh. \quad (3)$$

Here $\tau(h, f, \theta)$ is the transmission between height h and the top of the atmosphere (TOA). The quantity $X_{\text{INT}}^*(f, \theta)$ now depends on frequency f and zenith angle θ . The symbol X in Eqs. (2)–(4) can in principle be any quantity, as discussed above. Using Eq. (2), we can account for the impact of gaseous absorption and slant path and derive the apparent hydrometeor path visible in a column observed in a sounding channel. This quantity is expressed by HWP* in this paper. For nadir observations in a perfect window channel [i.e., $\theta = 0$ and $\tau(h) = 1$] we have $X_{\text{INT}}^* = X_{\text{INT}}$; that is, $\text{LWP}^* = \text{LWP}$, $\text{PWP}^* = \text{PWP}$, and $\text{HWP}^* = \text{HWP}$.

The transmission-weighted average height H^* can then be calculated as

$$H^*(f, \theta) = \frac{1}{Z_{\text{INT}}^*(f, \theta) \cos\theta} \int_0^{\text{TOA}} h \tau(h, f, \theta) Z(h) dh. \quad (4)$$

The conceptual model shown in Fig. 6 illustrates the above quantities by example of hydrometeor water content [HWC; i.e., the generic quantity $X(h)$ is set to mean HWC(h)]. Thus, in this example X_{INT} represents HWP. While H and HWP (black in Fig. 6) refer to the physical properties of the column of the atmosphere, the properties H^* and HWP* refer to transmission-weighted

average height and the amount of HWP visible for each particular channel. They differ between channels with different vertical weighting function (red and blue in Fig. 6).

Based on the above Eqs. (2)–(4), we now calculate the vertical integral of reflectivity Z_{INT}^* and the radar-retrieved integrated values for HWP* together with their corresponding transmission-weighted average height H^* . Note that this study is performed using Z_{INT}^* from KuPR and KaPR, respectively. Because the findings from both cases do not show significant differences, the results we present in this paper are for KuPR. We first show in Fig. 7 the relation between Z_{INT}^* and HWP* for all channels, as well as the relation between Z_{INT} and HWP, which is independent of any MWS-2 information and is entirely driven by the DPR retrieval. For all channels, the relation between Z_{INT}^* and HWP* follows roughly a power-law relation. The power-law coefficients for the relation between Z_{INT} and HWP are given. Especially at the high end of the observations (approx. $Z_{\text{INT}} > 65$ dB), this relation fits nearly all relations within the range of uncertainty shown around the curve for Z_{INT} versus HWP. We will use this relation later to investigate the agreement between Z_{INT}^* and HWP*, which are related via the DPR retrieval, and the independent observations of ΔTB .

For lower values of Z_{INT} , considerable spread can be observed among the different curves, which again is a manifestation of the DPR retrieval, likely caused by temperature dependency of the retrieval, and likely physically realistic. For higher peaking channels, the observed reflectivities will be at colder temperatures than for lower peaking channels. Thus, size distributions assumed in the DPR retrieval will change and consequently, the retrieved ice water content and HWP* will change. This effect is most clearly identified in channels 12–15 at around $\text{HWP}^* = 1 \text{ kg m}^{-2}$, for which the corresponding Z_{INT}^* decreases monotonically from about 61 dB for the lowest-peaking channel 15 to 56 dB for channel 12.

To understand the relation between MWS-2-derived ΔTB and the radar-derived quantities, we next bin ΔTB by Z_{INT}^* and H^* . The resulting two-dimensional plots of ΔTB as function of Z_{INT}^* and H^* are shown in Fig. 8. In each panel, the x axis (Z_{INT}^*) gives an indication of how much reflectivity in total is seen by each channel. The y axis gives an indication from where in the atmosphere on average the reflectivity stems.

As mentioned above, channels 2–4 peak too high in the atmosphere to see significant scattering. The remaining 12 channels can be split up into two groups with different behavior: The first group includes channel 1 and channels 5–10. The second group includes the five

FIG. 7. Relation between DPR-derived Z_{INT} and HWP (blue dashed line with light-blue uncertainty range). Also shown are the relations between Z_{INT}^* and HWP* for all MWS-2 channels. The linear fit in double-logarithmic space corresponds to $\text{HWP} = 2.82 \times 10^{-4} \times Z_{\text{INT}}^{0.58}$ with HWP in kilograms per meter squared and Z_{INT} in millimeters to the sixth power per meter squared.

water vapor absorption channels 11–15. Channels 1 and 5–10 include the traditional window channels as well as the 118-GHz oxygen absorption channels. Toward large values of Z_{INT}^* (i.e., large amount of precipitation visible to the channel), these channels all behave in a similar manner: The histogram narrows and shows the largest (blue) ΔTB depression near heights H^* of around 3–5 km with only a slight increase in H^* going from the window channels (e.g., channel 10) to the more strongly absorbing oxygen sounding channels (e.g., channel 5). Channels 11–15 behave differently and show more variability in H^* than the oxygen absorption channels. Going from the least-absorbing water vapor channel (channel 15) to the strongest absorbing water vapor channel (channel 11), H^* associated with the largest (blue) ΔTB depression increases from 3 km to about 11 km. In conclusion, the combination of the MWS-2 water vapor and oxygen absorption channels does show sensitivity to hydrometeor scattering throughout the entire troposphere.

Comparing channels 1 (89 GHz) and 10 (150 GHz), one can see that both channels react very similarly to hydrometeor scattering (blue). However, one significant difference between these two channels is that channel 1 shows a significant warming ($\Delta\text{TB} > 0$) for high Z_{INT}^* and low H^* (e.g., around $Z_{\text{INT}}^* = 60$ dB and $H^* = 1.5$ km). This warming is not seen as strongly in any other channel and is likely associated with emission of cloud liquid water and liquid rain in the lower atmosphere. This emission signal is increasingly masked at higher frequencies due to the larger opacity of the atmosphere and the increased hydrometeor scattering. While this emission signal provides physical information on liquid in the atmosphere and is the dominating signal of precipitation

FIG. 8. Binned representation of mean brightness temperature depression ΔTB as a function of transmission-weighted integrated radar reflectivity Z_{INT}^* and transmission-reflectivity-weighted vertical height H^* .

at frequencies below 40 GHz, the overlap of emission and scattering at 89 GHz make this particular channel more difficult to interpret than, for example, the 150-GHz-channel.

Next, we proceed to study the sensitivity of ΔTB of the individual MWS-2 channels to the presence of hydrometeors. In Fig. 9, the top-left panel shows no clear dependency of sensitivity on frequency. This is not surprising because HWP is just the total column hydrometeor water path and, for example, channel 3 peaking at 20.0 km altitude is not expected to be sensitive to precipitation ice and liquid found in the lower troposphere. The variability seen in sensitivity

with respect to HWP is thus driven by two factors. First, the gaseous transmission in each channel, which will depend on the position of the channel relative to the 118 and 183 absorption lines. Second, the actual scattering intensity of hydrometeors, which should be a monotonically increasing function of frequency.

The right-hand panel of Fig. 9 disentangles these two effects by plotting ΔTB against HWP*, and thereby accounting for the portion of hydrometeors actually observed in each channel and allows to assess the response of ΔTB as function of frequency alone. The different curves cluster around the four frequency ranges, 89, 118, 150, and 183 GHz, with significant differences in

FIG. 9. The ΔTB response of the 12 MWHS-2 channels (excluding channels 2–4) to (top left) HWP and (top right) HWP*. Window channels are plotted in black, 118-GHz oxygen absorption channels are plotted in red, and 183-GHz water vapor absorption channels are plotted in blue. (bottom) The corresponding data density for the different plots in the top panels in samples per bin. Note that for the plots on the left-hand side all channels have the same data distributions because HWP is independent of channel number. For plots on the right-hand side, each channel has its unique data distribution because HWP* depends on channel number.

sensitivity between the four. One can identify that over a large range of HWP* the response of ΔTB is approximately linear.

The approximate linearity of ΔTB with HWP* allows for an estimation of the sensitivity of ΔTB to HWP* by calculating the slope of each curve. The results of this analysis are shown in Table 3. The sensitivities of these 12 channels increase in magnitude with frequency from about $11 \text{ K (kg m}^{-2})^{-1}$ at 89 GHz to $35\text{--}53 \text{ K (kg m}^{-2})^{-1}$ at 183 GHz.

This general increase in scattering intensity with frequency is overlaid by a few other effects. Comparing the scattering sensitivity between the different 118-GHz channels (5–9), one can see from Table 3 that the sensitivity decreases with increasing atmospheric opacity. Contrarily, for the water vapor absorption channels (11–15), the sensitivity increases with increasing opacity. We believe that both effects are physical, but further studies are needed in order to fully understand this effect.

Channel 1 (89 GHz) presents a special case. For low HWP*, ΔTB increases with HWP* and the observed ΔTB on average exceeds zero in the range between zero and about 1.5 kg m^{-2} in HWP*. This effect is caused by emission of liquid clouds and precipitation and is only

very weakly observed in the other channels. We therefore calculated the sensitivity of 89 GHz only for values $\text{HWP}^* > 0.74 \text{ kg m}^{-2}$, where visually the relation starts to decrease linearly. The combined effect of emission and scattering at 89 GHz makes this channel particularly difficult to interpret in the context of precipitation retrieval but also provides additional information when combined with higher-frequency channels, where the emission effect is less pronounced or completely absent.

To explain the principal behavior of the curves shown in Fig. 9, we set up a simple conceptual radiative transfer model following the graph shown on the left-hand side of Fig. 10. The upwelling cloud-free brightness temperature $T_{B,CF}$ is modulated by an absorbing/emitting cloud/precipitation layer and a second layer that is purely scattering. The optical depth of the emitting layer is τ_e and the optical depth of the scattering layer is τ_s . To first order, the sum of these two quantities is proportional to HWP, that is, $\text{HWP} \propto \tau_e + \tau_s$, assuming the mass extinction coefficient of the absorbing and scattering materials to be constant. Assuming the scattering layer to be scattering isotropically (asymmetry parameter $g = 0$) and conservatively (single-scattering albedo $\tilde{\omega} = 1$), the diffuse two-stream transmission of that layer is $1/(1 + \tau_s)$, so that Eq. (5) captures the response of

TABLE 3. Sensitivity to ice particle scattering of ΔTB of the different MWHS-2 channels expressed in kelvins per kilogram per meter squared of HWP* and sensitivity to Z_{INT}^* expressed in kelvins per $(\text{mm}^6 \text{m}^{-2})^{0.58}$. The sensitivity was calculated as the slope of the curves shown on the right-hand side of Fig. 9. The minimum and maximum HWP* values listed show the boundaries between which the slope was calculated for each channel. Channels 2–4 provide a too-narrow range in HWP* to calculate the sensitivity. The peak of the weighting function (last column) is reproduced from the values given in Fig. 3. Channels highlighted with italics typically peak at lower than 5 km in the atmosphere and are included in the subsequent discussions about the spectral dependency of sensitivity.

Channel	Min HWP* (kg m^{-2})	Max HWP* (kg m^{-2})	Sensitivity to HWP* [K (kg m^{-2}) $^{-1}$]	Sensitivity to $(Z_{\text{INT}}^*)^{0.58}$ [K $(\text{mm}^6 \text{m}^{-2})^{0.58}$] $^{-1}$]	Peak of weighting function (km)
<i>1</i>	<i>0.74</i>	<i>3.30</i>	<i>-11.33</i>	<i>-2.56×10^{-3}</i>	<i>0.1</i>
2	None	None	None	None	25.9
3	None	None	None	None	20.0
4	None	None	None	None	17.7
5	0.00	0.60	-9.61	-1.42×10^{-3}	11.9
6	0.00	1.00	-11.33	-1.95×10^{-3}	9.6
7	<i>0.00</i>	<i>2.30</i>	<i>-14.93</i>	<i>-3.23×10^{-3}</i>	<i>2.9</i>
8	0.00	2.50	-15.19	-3.40×10^{-3}	2.0
9	0.00	2.50	-17.21	-3.84×10^{-3}	0.9
10	0.00	2.50	-26.46	-6.69×10^{-3}	1.0
11	0.00	0.26	-53.10	-1.79×10^{-2}	7.2
12	0.00	0.35	-52.05	-1.88×10^{-2}	6.2
13	0.00	0.65	-45.26	-1.78×10^{-2}	5.0
14	0.00	1.00	-40.55	-1.42×10^{-2}	4.2
15	0.00	1.20	-34.81	-1.06×10^{-2}	3.2

ΔTB to changes in optical depth as well as to changes in cloud-free brightness temperatures. Other assumptions in this model include neglecting reflected downwelling radiation off the top of the scattering layer. None of the various assumptions made here are fully satisfied in nature, so that the model described here cannot be used to infer any atmospheric properties. However, it does capture in principle the main physical processes leading to the curves observed (ΔTB sensitivity increases with higher frequency) in the top-right panel of Fig. 9.

One can see from Fig. 10 that the magnitude of the positive “bump” in ΔTB depends on the contrast between $T_{B,CF}$ and the temperature T_e of the emission layer. If this contrast is large (black curve, compared to yellow curve, which has a smaller contrast), emission by clouds and precipitation will initially increase ΔTB over the zero line and only as the scattering layer becomes optically thicker, ΔTB will decrease. Large differences between $T_{B,CF}$ and T_e are typically observed in window channels over open ocean situations. As frequency increases, these

FIG. 10. (left) The setup of a simple two-layer model affecting the observed brightness temperatures. This same model is expressed mathematically in Eq. (5). (right) The response of ΔTB to changes in the total optical thickness of the two layers for two different values of cloud-free background temperatures (black and yellow curves) and for another case in which the optical depth of the scattering layer is increased (red curve). The dashed lines show the range for which the response is approximately linear.

FIG. 11. As in Fig. 9, but for $(Z_{\text{INT}}^*)^{0.58}$.

differences become smaller, because surface emissivity increases. Because 89 GHz exhibits the lowest surface emissivity for all MWS-2 channels, the

effect can be observed most strongly in Fig. 9 for 89 GHz and to a much smaller extent for the other channels:

$$\Delta T_B = TB_{\text{OBS}} - TB_{\text{SIM}} = \underbrace{\left[\underbrace{e^{-\tau_e}}_{\text{Transmission of absorbing layer}} \times T_{B,\text{CF}} + \underbrace{(1 - e^{-\tau_e})}_{\text{Emissivity of absorbing layer}} \times T_e \right]}_{TB_{\text{OBS}}} \times \underbrace{\frac{1}{1 + \tau_s}}_{\text{Transmission of scattering layer}} - \underbrace{T_{B,\text{CF}}}_{TB_{\text{SIM}}} \quad (5)$$

Figure 10 also highlights the limited range in ΔTB responds roughly linearly to increases in cloud and precipitation optical depth (dashed lines). One can see that the slope of the yellow and black dashed lines is approximately identical. Only when the scattering optical depth increases (red curve), the slope becomes steeper. This effect is associated with increase in scattering optical depth with frequency and can be observed in the top-right panel of Fig. 9.

We note that the underlying HWP* value itself is derived from DPR observations and therefore also incorporates assumptions about particle habit and size distribution. For observational validation of radiative transfer model simulations of nonspherical

particles, it might be more appropriate to compare the spectral dependency of ΔTB to Z_{INT}^* because, unlike HWP*, Z_{INT}^* is directly observable from radar. This approach was first outlined and applied in Kulie et al. (2010). Figure 11 shows a corresponding analysis as Fig. 7 but with respect to Z_{INT}^* instead of HWP*. Unlike HWP*, Z_{INT}^* does not include any specific assumptions made in the DPR retrieval and is thus directly observable from radar as long as the gaseous transmission of the cloud-free atmosphere is known. In our analysis we found that naively plotting ΔTB against Z_{INT}^* does not provide a linear relation between the two. Rather, $(Z_{\text{INT}}^*)^{0.58}$ provides that same linear relation, on average. Given the relation between HWP and Z_{INT}

FIG. 12. (left) Sensitivity of ΔTB to HWP^* for the four different bands (89, 118, 150, and 183 GHz). The data are shown in a double-logarithmic plot. To avoid issues with channels peaking too high in the atmosphere, only channels with a peak of the weighting function below 5 km were included. Thus, for 118 GHz only channels 7–9 were included, and for 183 GHz only channels 13–15 were included. The error bars at 118 and 183 GHz show the minimum/maximum range of sensitivity over these channels. The red curve shows a power-law fit through the data with parameters given in the inset, with S being the sensitivity [$K (kg m^{-2})^{-1}$] and f being the frequency (GHz). (right) The sensitivity of ΔTB with respect to $(Z_{INT}^*)^{0.58}$.

discussed in Fig. 7, this is an expected result as $(Z_{INT})^{0.58}$ is approximately proportional to HWP^* . HWP^* itself is found from Figs. 9 and 10 to respond linearly to ΔTB .

To better quantify the spectral dependency of ΔTB on HWP^* and Z_{INT}^* , we show in Fig. 12 the sensitivity plotted as function of frequency. For this figure, we have only selected channels peaking below about 5 km (highlighted with italics in Table 3). This selection made to exclude higher peaking channels that show different sensitivities as discussed above. The sensitivity approximately follows a power law with exponent 1.78 for HWP and with an exponent of 2.37 for Z_{INT} . However, the power-law fit for Z_{INT} does not capture the actual behavior as well as the one for HWP .

This observed spectral dependency sheds light on the observed particle scattering behavior over a wide range of frequencies. To first order, the increased response of ΔTB to HWP (and Z_{INT}) is caused by the increased optical thickness of scattering by ice particles. This observed behavior can then be used, statistically, to evaluate the accuracy of theoretical scattering models, which need to reproduce the same spectral behavior. In this context, the above finding of approximate linearity (i) between $(Z_{INT})^{0.58}$ and ΔTB as well as (ii) between HWP and Z_{INT} and (iii) between HWP and ΔTB reinforces the notion that the underlying GPM-DPR

hydrometeor retrieval (which drives the relation between HWP and Z_{INT}) actually is physically consistent. As pointed out above, if it were not, we would not find linearity between $(Z_{INT})^{0.58}$ and ΔTB .

5. Conclusions

In the next few years, several new passive microwave missions will be launched covering the spectral range between 89 and 200 GHz. Some of these missions, as, for example, NASA's TROPICS mission, will also provide channels at 118 GHz but will not provide low-frequency observations below 89 GHz. In this context, it is important to understand the response of passive microwave observations to ice particle scattering.

Here we present an analysis of ice particle scattering behavior in this frequency range as compared to vertically integrated hydrometeor and reflectivity properties derived from collocated GPM-DPR observations. The results presented here build on earlier analysis by Bennartz and Bauer (2003) in which the scattering signal of precipitation was also investigated using simulation studies. In this current study, we overcome a few limitations of the earlier study, which was based solely on radiative transfer simulations and was geared toward understanding the bulk effect of hydrometeor scattering. Here, we

focus on disentangling the effects of gaseous absorption from that of hydrometeor scattering, which allows us to quantify the effects of scattering alone, rather than the effect of combined scattering and gas absorption. We further provide a simple conceptual model that explains the principal dependency of the brightness temperature depression on changes in hydrometeor water path (HWP). With this new approach, we quantify the global response of brightness temperature depressions over a cloud-free background (ΔTB) and its dependency on frequency. Globally averaged, the sensitivity in magnitude of ΔTB to HWP quadruples from about $11 \text{ K (kg m}^{-2}\text{)}^{-1}$ at 89 GHz to more than $53 \text{ K (kg m}^{-2}\text{)}^{-1}$ around 183 GHz. Comparing the two frequencies least affected by gas absorption (89 and 150 GHz, respectively), we find that our observationally derived sensitivities [11 and $26 \text{ K (kg m}^{-2}\text{)}^{-1}$, respectively] are in agreement with the results of Bennartz and Bauer (2003), in which the response at 150 GHz was found to be 2–3 times as strong as at 85 GHz.

Disentangling gaseous absorption from the scattering behavior proves useful for two reasons. First, the profiles of gaseous transmission as well as the cloud-free background brightness temperatures needed for ΔTB can be calculated easily from operational NWP model data. Second, while NWP models do not provide perfectly accurate profiles of temperature and moisture, and collocation errors between NWP models and observed brightness temperatures exist, the simulated cloud-free brightness temperatures are fairly accurate, and their error characteristics are known better than any simulations of precipitation-affected brightness temperatures. Together with the relatively large temperature depressions observed at the frequencies under consideration, the use of ΔTB appears beneficial in the context also of retrievals of hydrometeors. The results and approach reported here will form the basis for such investigations in the future, where we will evaluate the scattering response to different types of precipitation (e.g., deep convective vs frontal) and also address issues related to changes in observed sensitivity within a given band. For example, we found here that different channels within the 118-GHz oxygen band respond differently to scattering than different channels with the 183-GHz water vapor band. In the former band, the sensitivity of ΔTB to HWP decreases with gaseous absorption, whereas in the latter band it increases. Further, from our analysis presented here, it also appears feasible to use the difference between observed brightness temperatures and simulated cloud-free background brightness temperatures to retrieve surface rain rate and hydrometeor water path from passive microwave observations.

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