



Improvement of Airborne Retrievals of Cloud Droplet Number Concentration of Trade Wind Cumulus Using a Synergetic Approach.

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Abstract. The cloud droplet number concentration N is a major variable in numerical weather prediction (NWP) and global climate models (GCMs). In combination with the liquid water content LWC it determines the cloud droplet effective radius r_{eff} , which is used to calculate the radiative effects of clouds. Especially, the solar cloud top reflectivity \mathcal{R} of shallow trade wind cumulus is highly sensitive with respect to N and LWC .

5 In-situ measurements of N are limited by the sampled cloud volume, not covering the significant natural variability. Satellite retrievals of N suffer from large uncertainties and strictly assume adiabatic vertical profiles of cloud properties. Therefore, it is suggested to use the synergy of passive and active airborne remote sensing to reduce the uncertainty of N retrievals and to bridge the gap between in-situ cloud sampling and global averaging.

10 For this purpose, spectral solar radiation measurements above shallow trade-wind cumulus were combined with passive microwave and active radar and lidar observations carried out during the second Next Generation Remote Sensing for Validation Studies (NARVAL-II) campaign with the High Altitude and Long Range Research Aircraft (HALO) in August 2016. The common approach to retrieve N is further developed by including combined measurements and retrievals of cloud optical thickness τ , liquid water path LWP , r_{eff} , as well as cloud base and top altitude. Three approaches are tested and applied to synthetic measurements and two cloud cases of NARVAL-II. Using the new techniques, errors in N due to the adiabatic assumption
15 have been reduced significantly.



1 Introduction

Clouds strongly influence the Earth's radiation budget by reflecting, absorbing, and emitting solar and terrestrial radiation. These effects are typically quantified by the cloud radiative forcing (CRF), which is defined by the difference between the net radiation (irradiance) in cloudy and cloud-free conditions. Depending on cloud type, their optical and microphysical properties, as well as their spatial and temporal occurrence, the CRF can vary significantly (Rosenfeld, 2006). In the tropics, clouds can either cool or warm the atmosphere / surface below the cloud. While for cirrus a warming effect dominates, boundary layer trade wind cumulus typically cool the subjacent atmosphere / surface by efficiently reflecting solar radiation (Warren et al., 1988). Therefore, a realistic representation of these clouds in numerical weather prediction (NWP) and global climate models (GCMs) is essential. Due to their sub-grid scale, internal variability, and boundary layer interactions, trade wind cumulus are not well represented in NWP and GCMs (Kollias and Albrecht, 2010). A major source of uncertainty of these models is caused by insufficient representation of the first aerosol effect (Bony and Dufresne, 2005), which describes the correlation of the cloud droplet number concentration N and the cloud optical thickness τ or cloud top reflectivity \mathcal{R} , commonly known as the Twomey effect (Twomey, 1977). It is most effective for optically thin, low-level clouds such as trade wind cumulus (Platnick and Twomey, 1994; Werner et al., 2014), which are an ubiquitous cloud type in the tropics (Warren et al., 1988; Eastman et al., 2011). Despite their small vertical and horizontal extent they can have fractional cloudiness of more than 25% (Albrecht, 1991) and, therefore, may influence the Earth radiation budget significantly (Chertock et al., 1993). Additional, trade-wind cumulus plays an important role in maintaining the thermodynamic budget in the atmospheric boundary layer. They couple the surface and free atmosphere by transporting latent heat and developing deep convection (Lamer et al., 2015). Another important factor determining the CRF is the number concentration of aerosol particles, in particular the amount of particles which can act as cloud condensation nuclei (CCN) (Werner et al., 2013). Depending on the CCN number concentration, precipitation formation can be promoted or inhibited (Lee and Feingold, 2013). The CCN concentration influences the cloud life cycle and life time (Albrecht, 1989). The magnitude of both effects depend on the individual cloud regime.

Operational NWP models usually do not have the computational capability to consider size-resolved microphysical schemes and, therefore, the usage of simplified parametrizations is inevitable. The most important parameter, which links microphysical and radiative properties of clouds, is the effective radius r_{eff} , which represents the radiatively effective size of a particle population (Pontikis and Hicks, 1992). In NWP and GCMs, r_{eff} is calculated from the cloud droplet number concentration N and the liquid water content LWC . In simple models, assumptions of constant N are applied for different situations, e.g. the separation of polluted and unpolluted air-masses, which is unrealistic. As r_{eff} is calculated from LWC and N , the cloud droplet number concentration is a key parameter for models to calculate reasonable values of r_{eff} and to represent the Twomey effect.

The cloud optical thickness τ of a homogeneous cloud layer, is estimated from LWC and r_{eff} by:

$$\tau = \frac{3 \cdot \int_{h_{\text{base}}}^{h_{\text{top}}} LWC \cdot dz}{2 \cdot \rho_w \cdot r_{\text{eff}}}, \quad (1)$$



following Hansen and Travis (1974) and Stephens (1978), with ρ_w the density of liquid water, h_{base} the cloud base height, and h_{top} the cloud top height. Equation (1) assumes a homogeneous vertical cloud profile, which is not a realistic scenario for most of the observed clouds, as they are usually sub-adiabatic (Breguier et al., 2000; Painemal and Zuidema, 2011; Min et al., 2012).

- 5 To measure N and LWC , airborne in-situ measurements are applied, utilizing different physical methods and instruments (Baumgardner et al., 2011; Wendisch and Breguier, 2013). These are based on optical measurement principles such as forward scattering, phase doppler interferometry and holographic imaging. Beside the uncertainties of the individual measurement techniques, the total sample volume of the instruments is rather limited in comparison to the typical horizontal and vertical extent of clouds. Due to the limited flight time and range, airborne in-situ observations cover the natural variability of N , r_{eff} , and LWC partly. To directly quantify the Twomey effect, co-located measurements of cloud microphysical and radiative properties are required, which was realized only in a few occasions (Ackerman et al., 2000; Siebert et al., 2013; Werner et al., 2014).

- Satellite remote sensing products are developed, for example by Quaas et al. (2009); Minnis et al. (2011); Mace et al. (2016); Bennartz and Rausch (2017). They are a useful tool, as they provide large spatial and temporal data sets. To improve global statistics of estimates of the Twomey effect, several approaches to derive N from satellite observations have been developed (Grosvenor et al., 2018b). Based on passive remote sensing in the solar and terrestrial wavelength range, N is estimated combining the results of bi-spectral retrievals of τ and r_{eff} and cloud top temperature T_{top} by Breguier et al. (2000), Quaas et al. (2006), Zeng et al. (2014), and Bennartz and Rausch (2017). They assume a constant LWC and N throughout the cloud vertical profile, which is not necessarily fulfilled in nature. More complex vertical profile types of LWC and N are applied by Boers et al. (2006), where an inhomogeneous mixing model assumes that entrainment dilutes the air parcel with constant volume radius of the droplets, while r_{eff} follows an adiabatic profile. The retrieved values of N using the homogeneous or the inhomogeneous model deviate within several percent. Further studies show that the inhomogeneous model represents nature more realistically compared to the homogeneous assumption (Boers et al., 1998; Breguier et al., 2000). These methods often use the dependence of τ on N to connect cloud microphysical and radiative properties. However, so far no operational satellite products of N are available and can have uncertainties of up to 80% (Grosvenor et al., 2018b).

- Instead of using τ in the remote sensing of N , LWP from passive microwave sensors can be exploited (Minnis et al., 2011). This has the advantage that LWP is determined at wavelengths, which are not influence by aerosol, sun-glint, or three-dimensional (3D) radiative effects. Further on, active remote sensing techniques have been applied to derive N , e.g., by Austin and Stephens (2001), Mace et al. (2016), who combined r_{eff} vertical profiles derived by cloud radar observations and τ obtained from passive solar remote sensing. Nevertheless, a disadvantage of the radar is that the radar reflectivity Z is mainly determined by large cloud droplets, which biases the results.

- The dependence of τ on N is investigated by Quaas et al. (2009) using satellite measurements. The correlations of τ and N , obtained by satellite are weaker compared to aircraft remote sensing results or in-situ measurements, which primarily is due to the large-scale averaging of the satellite measurement (McComiskey and Feingold, 2008). Analyzing satellite measurements of large-scale averaged N and τ in different thermodynamic conditions and, therefore, varying LWP , updraft velocity and



aerosol particle concentrations, mask the effect of N on τ . As a result, parameterizations derived from satellite observations are not well suited for trade-wind cumulus with their natural variability.

Airborne remote sensing may bridge the gap between in-situ and satellite measurements, as it allows to sample individual clouds under specific conditions and to cover a sufficiently large area to quantify the natural variability of N , r_{eff} , and LWC .

5 Here a method is proposed to combine passive and active airborne remote sensing measurements of cloud vertical profiles of microphysical parameters and cloud radiative properties. Measurements of upward irradiance I_{λ}^{\uparrow} by the Spectral Modular Airborne Radiation measurement system (SMART) are used to determine τ , r_{eff} , and the thermodynamic phase of cloud water close to the cloud top. Observations of the High Altitude and Long range research aircraft Microwave Package (HAMP), which comprises of a multi-channel microwave radiometer and a cloud radar. HAMP provides LWP and cloud geometry,
10 which allows to discriminate between precipitating and non-precipitating clouds. Furthermore, an alternative retrieval to determine r_{eff} from the spectrometer-microwave combination of SMART and HAMP is developed and tested. Lidar measurements of the Water Vapour Lidar Experiment in Space (WALES) are additionally implemented to determine the cloud top height h_{top} , while dropsondes provide estimates of the cloud base height h_{base} .

This paper is structured as follows. In Section 2 the sensitivity of the cloud top reflectivity \mathcal{R} and cloud top albedo α of typical
15 trade-wind cumuli with respect to changes of N is quantified. This is investigated to access the required accuracy of N retrievals and the cloud regime most sensitive to N . The remote sensing instruments utilized in this study are introduced briefly in Section 3. In Section 4 the retrieval of the optical properties and the cloud filtering is described. Subsequently, the three different methods to determine N are presented in Section 5 and applied to synthetic measurements and two exemplary cases of trade wind cumulus. Resulting values of N are correlated with measured cloud top reflectivity \mathcal{R} , separated for different
20 thermodynamic conditions (binned LWP), to show the possibility to obtain parameterizations for the Twomey effect.



2 Sensitivity of the Twomey Effect for Different Cloud Regimes

Cloud optical and cloud microphysical properties depend on the composition and number of aerosol particles as described by the Twomey effect (Twomey, 1977). A sensitivity study was performed to quantify the dependence of cloud top albedo α or cloud top reflectivity \mathcal{R} on N and LWP . To determine which cloud regimes, defined by combinations of N and LWP , is most sensitive to changes of N .

To quantify the Twomey effect for trade wind cumulus with different LWP , radiative transfer simulations (RTS) with the radiative transfer package libRadtran 2.0.2 (Emde et al., 2016) are performed. The solar cloud top albedo was calculated for a homogeneous liquid water cloud located between 1000 m and 1500 m and a solar zenith angle ϑ of 5° . Liquid water path is varied in a range between 10 g m^{-2} and 200 g m^{-2} which was based on in-situ measurements of shallow trade wind cumulus by Siebert et al. (2013).

Figure 1a shows simulated α for all combinations of N and LWP . For constant LWP and increasing N (decreasing r_{eff}), α increases which is described by the Twomey effect. However, this sensitivity is not equal for the different LWP . For constant N and increasing LWP (increasing r_{eff}), α increases with different rates for N . This clearly shows that different cloud regimes react differently to the Twomey effect following Eq. (1), and therefore, LWP , N , and r_{eff} have to be considered to parameterize the radiative properties of trade-wind cumuli.

To quantify the Twomey effect for different cloud regimes, the sensitivity ζ is calculated:

$$\zeta(LWP, r_{\text{eff}}, N) = \frac{d\alpha(LWP, r_{\text{eff}}, N)}{dN}, \quad (2)$$

which represents the change of α with respect to an increase of N and is given in units of cm^3 .

Figure 1b displays ζ as a function of N for different LWP . In general, ζ decreases with increasing N . Clouds with low LWP and low N have a higher ζ compared to clouds with higher LWP but same N . The highest ζ is obtained for clouds with the lowest LWP of 10 g m^{-2} , while thicker clouds with the highest LWP of 200 g m^{-2} have the lowest Twomey effect, lowest ζ .

In Fig. 1c the sensitivity ζ is shown as a function of r_{eff} for clouds of different LWP . Assuming a constant LWP , the effective radius determines N or vice versa following

$$r_{\text{eff}} = \sqrt[3]{\frac{3}{4} \cdot \frac{LWP}{\rho_w \cdot \pi \cdot dz \cdot N}} \cdot k^{-3}. \quad (3)$$

For all LWP cases the sensitivity increases with increasing r_{eff} (decreasing N). This agrees with Fig. 1b where low N have the highest ζ . Clouds with lower LWP show higher ζ and, therefore are more sensitive to changes of r_{eff} compared to clouds with higher LWP .

In Fig. 1d ζ is plotted as a function of τ , which is calculated using Eq. (1) from LWP , N , and r_{eff} used in the simulations. For all clouds with different values of LWP , ζ decreases with increasing τ . This implies that changes in N have larger effects on α for clouds with low τ . As a result, optically thin clouds with low N and large r_{eff} , which is the typical character of shallow trade wind cumulus, are subject to the strongest Twomey effect. Therefore, the Twomey effect on trade wind cumulus is highly

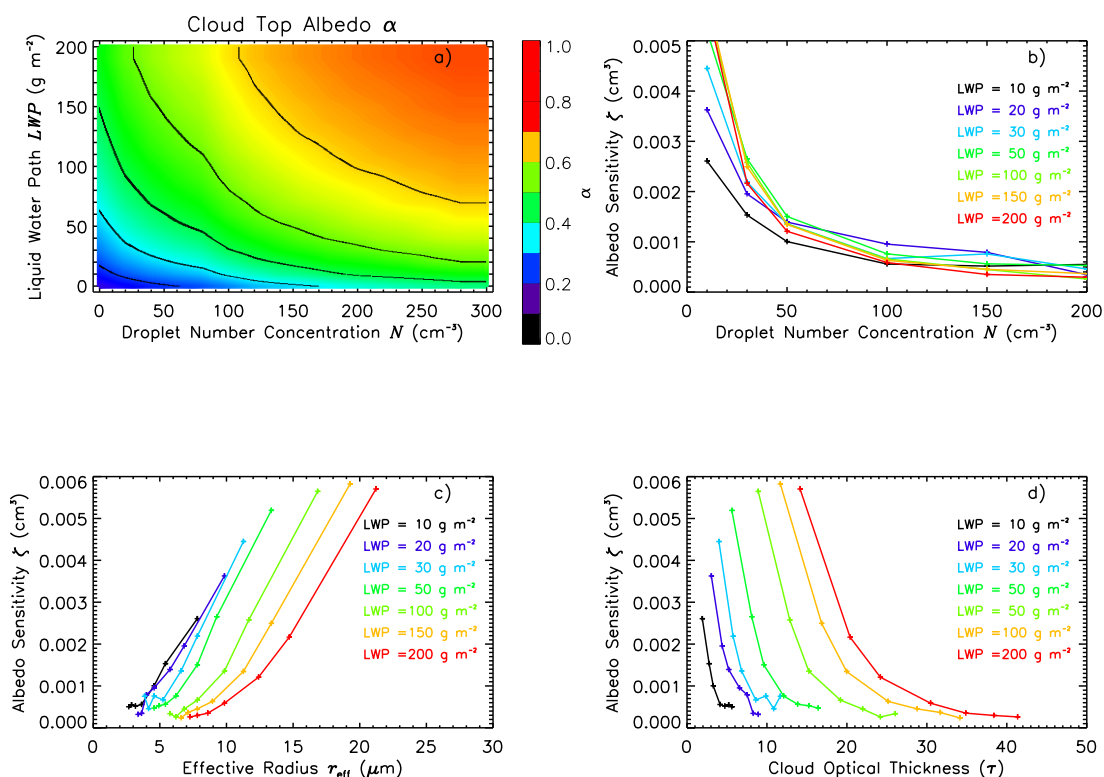


Figure 1. Simulations for a liquid water cloud between 1000 m and 1500 m with liquid water path LWP from 10 g m^{-2} to 200 g m^{-2} and for a solar zenith angle ϑ of 5° . The simulations are integrated over a wavelength range from 250 nm to 2500 nm. Panel a) shows cloud top albedo α for combinations of the cloud droplet number concentration N and LWP . Panel b) shows cloud top albedo sensitivity ζ as a function of N for different LWP . Panel c) and d) display ζ as a function of effective radius r_{eff} and cloud optical thickness τ , respectively.



relevant for NWP and GCMs.

The simulations further illustrate the challenge of estimating α of shallow trade wind cumulus by satellite remote sensing. Typically, satellite retrievals of N can have uncertainties in the range of up to 80% (Grosvenor et al., 2018b). For clouds with low N , e.g., 30 cm^{-3} , and $LWC = 0.1 \text{ g m}^{-3}$ the concentration of N might be biased by up to $\pm 23 \text{ cm}^{-3}$. This would result in a bias of α of ± 0.08 (80 W m^{-2} increased cloud forcing for 1000 W m^{-2} insolation). For clouds with higher N of 200 cm^{-3} the retrieval uncertainties of N naturally increase in absolute values ($\Delta N = \pm 156 \text{ cm}^{-3}$) and still lead to an almost similar uncertainty of $\alpha = \pm 0.07$ even though ζ is reduced for clouds with higher N . This shows, that retrievals of N need to be improved, in order to improve global estimates of N and to reduce the uncertainty of α calculations in NWP and GCM.

3 Observations and Instrumentation

10 Convective shallow trade wind cumulus have been observed by airborne remote sensing during the second Next Generation Remote Sensing for Validation Studies (NARVAL-II) campaign between 8 and 31 August 2016 (Stevens et al., 2018). The High Altitude and Long Range Research Aircraft (HALO) operated from Barbados mostly flying eastward into an area dominated by shallow trade wind cumulus mostly unaffected by anthropogenic influences. HALO was equipped with a set of passive and active remote sensing instruments. Reflected solar radiation was measured by the passive instruments SMART (Wendisch et al., 15 2001, 2016) and Munich Aerosol Cloud Scanner (specMACS) (Ewald et al., 2016), while radiation emitted in the microwave spectral range was measured by the HALO Microwave Package (HAMP). For active remote sensing, HAMP included an cloud radar (Mech et al., 2014). Lidar observations by the WATER vapor Lidar Experiment in Space (WALES) completed the cloud remote sensing instrumentation. WALES measures the backscatter coefficient and depolarization at 532 nm and 1064 nm wavelength, and contains a high spectral resolution lidar channel at 532 nm wavelength (Wirth et al., 2009). Additionally, 20 dropsondes were released from HALO.

All instruments were orientated into nadir direction and synchronized in time. However, the different Field-of-VIEWS (FOV) of the instruments cause a systematic difference in the observed time series. All measured and retrieved quantities from SMART, HAMP, WALES, and the dropsondes are listed in Table 1.

3.1 Spectral Modular Airborne Radiation measurement system

25 During NARVAL-II, SMART measured the spectral upward F_{λ}^{\uparrow} and downward irradiance F_{λ}^{\downarrow} , as well as spectral upward radiance I_{λ}^{\uparrow} . Each quantity was recorded with two separate Zeiss grating spectrometers, one for the visible (VIS) range from 300 nm to 1000 nm and a second one for sampling the near-infrared (NIR) range from 900 nm to 2200 nm. By merging the spectra, about 97% of the solar spectrum is covered (Bierwirth et al., 2009). The spectral resolution defined by the full width at half maximum is 2 - 3 nm for the VIS spectrometer and 8 - 10 nm for the NIR spectrometer.

30 The radiance optical inlet of SMART has an opening angle of 2° . The sampling time t_{int} was set to 0.5 s. For an average aircraft ground-speed of about 220 m s^{-1} and a distance of 10 km between cloud top and the aircraft this results in a FOV of about 100 m x 120 m for an individual pixel.

**Table 1.** Measured and retrieved quantities from SMART, HAMP, WALES, and the dropsondes.

Instrument	Measured / retrieved quantity	Variable	Unit
SMART	Upward radiance	I_{λ}^{\uparrow}	$\text{W m}^{-2} \text{sr}^{-2}$
	Cloud optical thickness	τ	-
	Effective radius	r_{eff}	μm
	Liquid water path	LWP_{A}	g m^{-2}
HAMP	Liquid water path	LWP_{B}	g m^{-2}
	Radar reflectivity	Z	dBz
WALES	Cloud top height	h_{top}	m
Dropsondes	Temperature	T	$^{\circ}\text{C}$
	Dew-point temperature	T_{d}	$^{\circ}\text{C}$
	Lifting condensation level	h_{LCL}	m

The optical inlets for F_{λ}^{\uparrow} and F_{λ}^{\downarrow} mainly consist of integrating spheres, which collect direct and scatter solar radiation from the upper or lower hemisphere. During NARVAL-II the upward-looking inlet was equipped with an active stabilization platform to ensure horizontal alignment of the sensor, which is crucial as F_{λ}^{\downarrow} refers to a horizontal plane (Wendisch et al., 2001).

Prior and after NARVAL-II, SMART was radiometrically calibrated in the laboratory using certified calibration standards traceable to the National Institute of Standards and Technology (NIST). A secondary calibration by a mobile standard was applied during the campaigns to track potential changes of the instrument sensitivity. The total measurement uncertainty of downward irradiance F_{λ}^{\downarrow} and upward radiance I_{λ}^{\uparrow} for typical conditions and observations of shallow cumulus is about 5.4% for the VIS and 8.4% for the NIR range, which is composed of individual errors due to the spectral calibration, the spectrometer noise and dark current, the primary radiometric calibration (Brückner et al., 2014).

10 3.2 HALO Microwave Package

HAMP is a combination of a passive microwave radiometer and an active cloud radar specifically designed for the operation on HALO (Mech et al., 2014). The microwave radiometer includes 26 frequency channels between 22.24 GHz and 183.31 GHz \pm 12.5 GHz. The brightness temperature (BT) measured along the 22.24 GHz and 183.31 GHz rotational water vapor lines provide accurate information on the total column water vapor (Schnitt et al., 2017) and information on its vertical distribution. Liquid water emission increases roughly with the frequency squared. By combining BT in window channels, i.e. 31.4 GHz and 90 GHz, mostly affected by liquid water with those sensitive to water vapor the LWP can be retrieved. This principle is also employed by satellite instruments which provide global climatologies of LWP , which suffer from the coarse footprint of a few 10ths of kilometer (Elsaesser et al., 2017).

The statistical LWP retrieval is based on a large variety of atmospheric profiles with differently structured warm clouds as training data composed from the dropsondes as described by Schnitt et al. (2017). Synthetic BT are simulated from these pro-



files and subsequently used to fit a multi-parameter linear regression model employing higher order terms following Mech et al. (2007). Testing the retrieval algorithm on an independent subsample provides an accuracy of about 30 g m^{-2} for LWP values below 500 g m^{-2} .

The cloud radar MIRA-36 operates at a frequency of 36 GHz and has a similar horizontal resolution as the LWP of about 1000 m and a temporal resolution of 1 s. Vertical profiles are divided into 30 m bins (Mech et al., 2014). The radar provides different parameters linked to the cloud microphysical properties including the radar reflectivity Z , the linear depolarization, and the Doppler velocity and the spectral width of the droplet size distribution. Note, that the latter two are affected by the relative motion of the aircraft to the wind and the antenna width (Mech et al., 2014) and are not used here.

Radar reflectivity is the sixth moment of the drop size distribution and therefore is strongly influenced by the largest particles. In order to calculate the LWC , i.e. proportional to the third moment of the DSD, from Z so-called Z-LWC relations are used which are typically derived from in-situ measurements. According to Khain et al. (2008) there is quite some variability involved and as soon as the transition to drizzle sets in the relation can be off by orders of magnitude. Here the Z-LWC relation

$$LWC_p = LWP \cdot \frac{\sqrt{Z_p}}{\sum_{j=1}^{j=M} \sqrt{Z_j} \cdot \Delta h} \quad (4)$$

following (Frisch et al., 2000) is used to derive vertical profiles of LWC . With the binned LWC_p at height gate p resulting from the vertical resolution of the radar, the LWP of the cloud, is distributed by the weighting of Z_p (Z at height gate p) and $\sum_{j=1}^{j=M} \sqrt{Z_j} \Delta h$ the sum of the Z , over all height gates where a cloud was present.

Brightness temperatures and radar reflectivity profiles are described in more detail by Konow (2018).

3.3 Water Vapour Lidar Experiment in Space (WALES)

The Differential Absorption Lidar (DIAL) called WALES operates at four wavelengths near 935 nm to measure atmospheric water vapor. It provides water-vapor mixing ratio profiles covering the whole atmosphere below the aircraft. The system also contains additional aerosol channels at 532 nm and 1064 nm with depolarization. At 532 nm WALES uses the high-spectral resolution technique, which distinguishes molecular from particle backscatter, to make direct extinction measurements. Within this study only the aerosol channels are used to provide information on the cloud top height. The ranging resolution of the instrument is 15 m. Together with the flight altitude inferred from the HALO on-board positioning system and an appropriate attitude correction the accuracy of cloud top height detection is about 20 m relative to the geoid. Over the sea this can be verified by looking at the surface reflex.

The laser has a beam divergence of 1 mrad which leads to an illuminated spot of 10 m diameter on ground at a flight altitude of 10 km. Laser pulses are emitted with a repetition rate of 100 Hz, but 20 signals are averaged on board to improve the signal to noise ratio, resulting in an along flight track resolution of 44 m at 200 m s^{-1} aircraft speed. Thus the horizontal resolution is considerably smaller than that of SMART and HAMP. Along track this can be easily taken into account by further signal averaging.



4 Measurement Analysis

Trade wind cumulus mostly appear randomly distributed with a tendency to form self-organizing structures (Bony et al., 2015). Typically, the vertical cloud extend is larger than the horizontal extent within an individual cell. This is in strong contrast to stratiform cloud fields to which common retrieval techniques of N are usually applied. Clouds smaller than pixel size covered by the FOV bias the retrieval of the microphysical properties significantly (Oreopoulos and Davies, 1998a, b). With respect to the total cloud volume, trade wind cumuli are stronger effected by turbulent mixing compared to stratiform clouds. The dominance of small-scale cumulus during NARVAL-II, mostly ranging in the horizontal size of a few hundred meters, results in very heterogeneous cloud scenes. This induces several challenges with respect to cloud masking and RTS. Several filters introduced below are applied to the data in order to minimize heterogeneity effects.

10

4.1 Cloud Mask and Precipitation Flag

4.1.1 Cloud Mask

To distinct between cloud and cloud-free measurements over ocean surfaces is linked with the difference in the spectral reflectivity is analyzed. The ratio χ of I_{λ}^{\uparrow} between 858 nm and 648 nm wavelength is calculated in analogy to the MODIS cloud mask (Platnick et al., 2013) by:

15

$$\chi = \frac{I_{858}}{I_{648}}. \quad (5)$$

The cloud mask is based on the relative intensity of I_{λ}^{\uparrow} and χ . Therefore, a single measurement can be identified as cloudy when only a part of the SMART FOV with 100 m x 120 m is covered by a cloud. Masking each measurement point as cloudy or cloud-free, the cloud length l_{cld} is determined. By counting the number n of consecutive cloud masked measurements. Multiplying with the flight speed v_{aircraft} and the constant integration time of SMART of $t_{\text{int}} = 0.5$ s, the cloud length is calculated with:

20

$$l_{\text{cld}} = n \cdot t_{\text{int}} \cdot v_{\text{aircraft}}. \quad (6)$$

For $v_{\text{aircraft}} \approx 220 \text{ m s}^{-1}$ the smallest cloud size which can be resolved is in the range of 120 m along flight track.

The length of trade wind cumulus can be shorter than the FOV of SMART. To identify such cases, an additional homogeneity cloud flag (HCF) is introduced. The cloud is considered as homogeneous (HCF is true) when a single observation and two previous and two following measurements are masked as cloud too. For clouds not surrounded by at least two cloudy pixel, the HCF is set to false. Therefore, the HCF identifies clouds that are large enough to fill the FOV of SMART, HAMP, and WALES at the same time.

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4.1.2 Precipitation Flag

Precipitation is identified using the radar reflectivity Z . Measurements are considered to be affected by precipitation when Z exceeds a threshold of $Z < -20$ dBz within 50 m to 200 m above sea level (Schnitt et al., 2017). This allows to discriminate precipitation events, which affect the LWP measured by the microwave radiometer and retrieved by SMART. Using radar observations to identify rain is prone to deficiencies. The radar reflectivity is very sensitive to the droplet diameter D by $Z \propto D^6$ and only droplets with sufficient size are detected. Hence, measurements with low Z below the detection threshold can still contain slight drizzle.

4.2 Retrieval of Cloud Optical Thickness and Effective Radius

10 Based on the reflected solar radiance I_{λ}^{\uparrow} measured by SMART, a retrieval of τ and r_{eff} is performed, applying the radiance ratio method proposed by Werner et al. (2013). The use of radiance ratios at two different wavelength partly reduces the uncertainties by the radiometric calibration of SMART. For the wavelength ratio used here, an uncertainty of 6% is assumed. Additionally, the use of ratios increases the retrieval sensitivity with respect to r_{eff} by clearly separating the dependency of I_{λ}^{\uparrow} on τ and r_{eff} and, therefore, the retrieval accuracy. Forward simulations of reflected spectral radiance I_{λ}^{\uparrow} were carried out with
15 the libRadtran 2.0.2 package (Emde et al., 2016). The Fortran 77 discrete ordinate radiative transfer solver version 2.0 (FDIS-ORT 2) after Stamnes et al. (2000) is used. The extraterrestrial F_{λ}^{\downarrow} is given by Gueymard (2004) and a marine aerosol profile after Shettle (1989) is selected. Vertical profiles of temperature, pressure, and humidity are obtained from radiosonds released at the Bridgetown International Airport. For the optical properties of liquid water droplets, Mie calculations are performed. The optical thickness and r_{eff} are determined by a modified Look-Up-Table (LUT) method after Nakajima and King (1990).
20 The optical thickness is determined at 870 nm. The effective radius is derived with the radiance ratio method, using a ratio of measurements at 1645 nm and 1050 nm and denoted with subscript "A".
Clouds which do not cover the entire FOV of SMART bias the retrieved optical properties, because they are in direct contradiction to the assumption of plane parallel clouds used in the RTS (Oreopoulos and Davies, 1998a, b). Lower values of I_{λ}^{\uparrow} bias τ towards lower values, whereas r_{eff} is shifting to larger droplet sizes (Cahalan et al., 1995). Further on, the heterogeneous
25 structure of trade wind cumulus may cause 3D radiative effects, like shadowing of parts of the cloud by nearby cloud-towers or enhanced reflectivity due to additional reflection into the FOV. These effects may also bias the retrieval of τ and r_{eff} and the calculation of N . Therefore, the HCF filter is applied to exclude measurements that are influenced by these processes. However, due to the low vertical extend of shallow trade wind cumulus these 3D radiative effects are assumed to be not pronounced for the cases analyzed here.
30 Finally, LWP is obtained directly from the parametrization within libRadtran which calculates "LWP" on the basis of τ and r_{eff} similar to Eq. 1. Liquid water path derived from SMART is denoted with subscript "A". In case of cloud heterogeneity, sun-glint or 3D radiative effects, the retrieval of τ is very likely biased. Following Eq. 1, a bias of τ also influences the retrieval of r_{eff} and therefore, LWP . To omit these effects, measurements of the LWP from HAMP (denoted with subscript "B") are



applied in the forward model of the cloud retrieval. Liquid water path from microwave radiometers are obtained from wavelength not influenced by sun-glint or 3D radiative effects. Using *LWP* from HAMP as a precondition, the LUTs reduce to one absorbing wavelength sensitive to r_{eff} . Therefore, the non-linear dependence between τ and r_{eff} is removed and the retrieval becomes more reliable. Retrieved r_{eff} from combined passive solar radiance and microwave measurements are denoted with subscript "B".

5 Retrieval of Cloud Droplet Number Concentration

The retrieval of N from remote sensing observations is generally based on the relation proposed by Brenguier et al. (2000) and Wood (2006), which links N of a stratiform cloud to τ and r_{eff} by:

$$N_A = \frac{\sqrt{10}}{4 \cdot \pi \cdot \sqrt{\rho_w}} \cdot \sqrt{f_{\text{ad}} \cdot \Gamma_{\text{ad}}} \cdot \frac{\sqrt{\tau}}{\sqrt{r_{\text{eff},A}^5}}. \quad (7)$$

The technique assumes an adiabatic vertical cloud profile, where temperature linearly decreases and *LWC* linearly increases with height. An adiabatic profile implies that the total water mass mixing ratio of the cloud is conserved. This is true when: (i) no water is removed from the cloud (no precipitation or fallout), (ii) no entrainment of dryer air at the cloud edges occurs, and (iii) no evaporation from precipitation happens. As a result, the proposed method should be applied to non-precipitating clouds only, which do not undergo strong vertical convection and mixing. A vertically constant N throughout the cloud layer is assumed. This assumption is verified for stratiform clouds and shallow trade wind cumulus by in-situ measurements, e.g., Reid et al. (1999) and Wendisch and Keil (1999). The main reason for the vertically constant N is the determination by the amount of available CCN at cloud base and their potential to form cloud droplets depending on the degree of supersaturation, which is controlled by temperature, entrainment of dry air and updraft velocity.

The k -parameter,

$$k = \left(\frac{r_{\text{vol}}}{r_{\text{eff}}} \right)^3 \quad (8)$$

relating the effective radius r_{eff} and the volumetric radius r_{vol} , is set to $k = 0.8$ for marine clouds following the suggestion by Martin et al. (1994) and Pontikis (1996). Depending on the cloud type the k -parameter can vary by ± 0.1 (Martin et al., 1994). With help of cloud properties retrieved by airborne remote sensing Eq. (7) can be applied in different complexity to derive N . Three methods are proposed in the following. Method A uses only SMART data, while method B additionally includes HAMP observations of *LWP*, whereas method C also involves measurements by WALES. The obtained parameters and assumptions used by the different methods are summarized in Table 2.

5.1 Method A: Based on Cloud Optical Thickness and Effective Radius

Method A follows the traditional satellite approach to feed Eq. (7) with τ and r_{eff} obtained by a single passive remote sensing instrument. Here the τ_A and r_A retrieved by SMART are applied. The degree of adiabacity is assumed to be 1. This implies,



Table 2. Overview of the cloud droplet number concentration retrievals and applied measurements, retrieval parameters and assumptions.

Method	A	B	C
Instruments and Parameters			
SMART	$\tau, r_{\text{reff},A}$	$r_{\text{reff},B}$	$r_{\text{reff},B}$
HAMP	×	LWP	LWP
WALES	×	×	f_{calc}
Assumptions			
adiabatic cloud-profile	✓	✓	×
adiabatic change of LWC	$f_{\text{ad}} \cdot \Gamma_{\text{ad}} = 2.5 \cdot 10^{-3} \text{ g m}^{-3} \text{ m}^{-1}$		Γ_{calc}
k -parameter	$k = 0.8$	$k = 0.8$	$k = 0.8$
const. N	✓	✓	✓
deep convection	×	×	×
cloud homogeneity	✓	✓	✓
precipitation	×	×	×
min. hori. size	$\approx 150 \text{ m}$	$\approx 150 \text{ m}$	$\approx 150 \text{ m}$

that for trade wind cumuli, which are typically sub-adiabatic, the estimated N is potentially biased. However, similar retrieval are frequently applied to observations from satellite such as MODIS (Grosvenor et al., 2018b).

5.2 Method B: Based on Liquid Water Path and Effective Radius

For adiabatic clouds, Eq. (1) can be solved analytically, which results in a relation that directly links LWP to τ and r_{eff} :

$$5 \quad LWP = \frac{5}{9} \cdot \rho_w \cdot \tau \cdot r_{\text{eff}} \quad (9)$$

following (Brenquier et al., 2000). Equation (9) allows to apply Eq. (7) with an independent measure of LWP instead of τ to calculate N . As given by Wood (2006) combining Eq. (7) and Eq. (9) leads to:

$$N_B = \frac{3 \cdot \sqrt{2}}{4 \cdot \pi \cdot \rho_w} \cdot \sqrt{f_{\text{ad}} \cdot \Gamma_{\text{ad}}} \cdot \frac{\sqrt{LWP_B}}{r_{\text{eff},B}^3}. \quad (10)$$

In method B, LWP measurements by HAMP and derived $r_{\text{eff},B}$ from the combined SMART microwave-radiometer retrieval are applied. The results are denoted with N_B . Exchanging $r_{\text{eff},A}$ by $r_{\text{eff},B}$ takes into account that LWP is determined from HAMP only. This makes the retrieval independent on τ derived by SMART and therefore less sensitive to effects by sun glint.



5.3 Method C: Based on Liquid Water Path, Effective Radius, and Cloud Geometric Thickness

Equations (7) and (10) assume constant values of f_{ad} and Γ_{ad} . Therefore, in method A and B the adiabatic profile of LWC follows the maximum, theoretically possible profile under which liquid water is released due to condensation from upward motion in the atmosphere.

- 5 In-situ measurements of stratocumulus and shallow cumulus clouds, like trade wind cumulus, indicate that a majority of cloud profiles do not follow this adiabatic assumption (Wendisch and Keil, 1999; Merk et al., 2016). In most cases the profiles are sub-adiabatic, meaning a reduced increase of LWC with height, mostly due to entrainment and mixing from dry air at the cloud edges. Entrainment and mixing reduce f_{ad} but not necessarily N . When convection and mixing is moderate, an equilibrium between the droplets and the surrounding air can be assumed. Therefore, the reduced (super-)saturation due to entrainment at
- 10 the cloud edges will cause a shrinking of the droplets but not their complete vanishing. To account for a sub-adiabatic increase of LWC with height in method C, $f_{ad} \cdot \Gamma_{ad}$ is replaced by observations. The increase of LWC with height Γ_{calc} is calculated by:

$$\Gamma_{calc} = \frac{2 \cdot LWP_B}{(dz)^2} \quad (11)$$

- with LWP_B obtained by the microwave radiometer. The cloud geometric thickness $dz = h_{top} - h_{LCL}$ is estimated from a
- 15 combination of the WALES h_{top} observations and h_{LCL} from dropsondes.

- WALES can only derive cloud top height h_{top} when the laser is attenuated by clouds with high τ . As a result, the lidar signal is attenuated soon and the cloud base height is not detectable. Therefore, h_{base} is determined separately from dropsondes, which represent the large-scale thermodynamic structure of the atmosphere. Using the temperature T and dew point temperature T_d at the two lower most points, the lifting condensation level with $h_{LCL} \approx 125 \cdot (T - T_d)$ is approximated (Espy, 1836). Nevertheless,
- 20 uncertainties of estimated h_{LCL} from dropsondes are in the range of ± 35 m not considering additional uncertainties caused by the assumptions in the equation (Romps, 2017). Using the estimated Γ_{calc} , Eq. (10) changes to:

$$N_C = \frac{3 \cdot \sqrt{2}}{4 \cdot \pi \cdot \rho_w} \cdot \frac{LWP_B}{dz \cdot r_{eff,B}^3} \quad (12)$$

- Calculations of N using the estimated dz and resulting Γ_{calc} are denoted with N_C . Alternatively, the cloud top height and cloud base height h_{base} can be determined with the radar for all cloud cases where: (i) the cloud droplets are large enough to
- 25 produce a detectable radar echo, and (ii) no precipitation is present.

5.4 Simulated Synthetic Measurements

- To systematically test the potential of the proposed synergistic retrieval methods, synthetic measurements of spectral upward radiance $I_{\lambda, syn}^{\uparrow}$ are created. In that way, the three different methods are compared omitting the influence by measurement errors.
- 30 Further on, varying environmental conditions, like sea surface albedo, heterogeneous cloud conditions, and 3D cloud radiative effects do not influence the systematic comparison of the retrieval methods. The comparison is based on retrieved cloud droplet



number concentration N with methods A, B, and C and N_{cld} calculated from the model clouds serving as truth value. Six synthetic clouds are simulated. Their respective parameters are listed in Table 3. Cloud droplet number concentrations N_{cld} of 50 cm^{-3} , 100 cm^{-3} , and 200 cm^{-3} represent the typical range of pristine shallow trade wind cumulus (Siebert et al., 2013). For each N_{cld} an adiabatic and a sub-adiabatic cloud profile was set up. Cloud base height is 500 m and cloud top height is 1000 m. For all cloud cases a linear increase of LWC and a constant N_{cld} with height are assumed. In the adiabatic cases (I, III, V) a LWP of 362 g m^{-2} and an adiabatic increase of LWC with height Γ_{ad} of $2.9 \cdot 10^{-6} \text{ kg m}^{-3} \text{ m}^{-1}$, for a surface temperature of $\approx 30^\circ \text{ C}$ are used. For the sub-adiabatic cases (II, IV, VI) Γ is set to $\Gamma_{\text{ad}} \cdot 0.6 = 1.7 \cdot 10^{-6} \text{ kg m}^{-3} \text{ m}^{-1}$ representing a cloud which follows Γ_{ad} by 60% and leads to a LWP of 217 g m^{-2} . To calculate the volumetric radius $r_{\text{vol}}(z)$, the cloud profiles are divided into 20 layers of equal thickness of 25 m. For each layer the parameterization of Martin et al. (1994) is applied:

$$r_{\text{vol}}(z) = \sqrt[3]{\frac{3 \cdot LWC(z)}{4 \cdot \rho_w \cdot \pi \cdot N_{\text{cld}}}}. \quad (13)$$

In the radiative transfer model, the effective radius r_{eff} is used to determine the optical properties of the cloud particles instead of the volumetric radius r_{vol} . To convert $r_{\text{vol}}(z)$ into $r_{\text{eff}}(z)$ a k of 1.0 is applied, what considers the monodisperse droplet size distribution used in the model clouds. The synthetic measurements of $I_{\lambda, \text{syn}}^{\uparrow}$ are calculated with the same simulation set-up as for the cloud retrieval described in Section 4.2.

Simulated synthetic measurements of $I_{\lambda, \text{syn}}^{\uparrow}$ are applied to the retrieval method of τ , r_{eff} , and N of Section 5. All three methods A, B, and C are applied and results are denoted with additional subscript "R". The true values of τ from the RTS (subscript "lib") are calculated directly from the given $r_{\text{eff}, \text{lib}}$, which represents the cloud top r_{eff} of the model cloud. Total cloud optical thickness τ_{lib} and $r_{\text{eff}, \text{lib}}$ from the libradtran radiative transfer simulations are considered to be the reference values which are used to compare the retrieval results and the calculated N . For consistency the labeling of N for the three methods follows Section 5. An overview of all retrieved and calculated parameters is given in Table 3.

The retrieved cloud optical thickness τ_{R} is higher compared to the true value τ_{lib} for all cloud cases. The largest difference of 26% are observed for cloud I. With increasing N_{cld} the absolute and relative differences become smaller. Systematically larger errors are found for the adiabatic clouds. A similar pattern is obtained for $r_{\text{eff}, \text{R}}$ which is always up to 2% smaller than $r_{\text{eff}, \text{lib}}$. The sub-adiabatic clouds show the largest differences. The relative error decreases for higher N_{cld} . The systematic underestimation of $r_{\text{eff}, \text{R}}$, especially for the sub-adiabatic cases, with respect to $r_{\text{eff}, \text{lib}}$ results from the penetration depth of the incident solar radiation into the cloud. For constant LWP , clouds with lower N have a lower τ , which reduces the scattering. Therefore, the incident radiation can penetrate deeper into the cloud compared to clouds with higher N and τ (Platnick, 2000). As a result, I_{λ}^{\uparrow} is more influenced by lower cloud layers and the retrieved $r_{\text{eff}, \text{R}}$ is systematically smaller than $r_{\text{eff}, \text{lib}}$. In this case, $r_{\text{eff}, \text{R}}$ is not representing $r_{\text{eff}, \text{lib}}$ at cloud top. The bias of $r_{\text{eff}, \text{R}}$ from the $r_{\text{eff}, \text{lib}}$ at cloud top feeds back into the retrieval of τ_{R} because of the dependence of τ and r_{eff} and the non-rectangular shape of the Look-Up-Table. The overall underestimation of retrieved $r_{\text{eff}, \text{R}}$, which appears for all passive remote sensing measurements based on reflected solar radiation, generally leads to an overestimation of N , which is intensively discussed e.g. by Brenguier et al. (2000) and Grosvenor et al. (2018b, a) and therefore, not repeated here.



Liquid water path LWP_R is calculated with Eq. (9) from the retrieved τ_R and $r_{\text{eff},R}$ by assuming an adiabatic cloud profile. In all cases, the retrieval overestimates LWP_R by 18% for low N_{cld} up to 27%. The deviation becomes larger for high N_{cld} . The cloud droplet number concentration $N_{A,\text{lib}}$ is calculated with method A by using τ_{lib} , $r_{\text{eff},\text{lib}}$, and assuming an adiabatic vertical profile with Γ_{ad} . This provides a reference for $N_{A,R}$ which applies τ_R and $r_{\text{eff},R}$. By comparing $N_{A,\text{lib}}$ and $N_{A,R}$ the influence of the remote sensing retrieval method (forward simulations and error due to penetration depth) on N for different N_{cld} becomes obvious. In general, $N_{A,\text{lib}}$ and $N_{A,R}$ of all clouds are larger compared to N_{cld} . Differences between $N_{A,\text{lib}}$, $N_{A,R}$ and N_{cld} result from smaller retrieved $r_{\text{eff},R}$ and higher τ_{lib} compared to τ_R . Another reason is the difference between Γ_{ad} used in the model cloud and the assumed LWP parameterization in Eq. (9) which is applied in Eq. (7) to correlate LWP and τ . For all clouds, $N_{A,R}$ is larger than $N_{A,\text{lib}}$ and N_{cld} , because in Eq. (7) N is dominated by $r_{\text{eff}}^{-5/2}$ and less sensitive to $\tau^{1/2}$. Differences between $N_{A,\text{lib}}$ and $N_{A,R}$ vary between 0% and 17%, being largest for cloud I for which the deviation in $r_{\text{eff},\text{lib}}$ and $r_{\text{eff},R}$ is largest. The simulations also show that $N_{A,\text{lib}}$ and $N_{A,R}$ are largest for the sub-adiabatic cloud cases. For method B, $N_{B,\text{lib}}$ and $N_{B,R}$ are larger than N_{cld} with smaller differences for the reference values of $N_{B,\text{lib}}$ and larger differences of $N_{B,R}$ compared to N_{cld} . For method B the deviations of $N_{B,\text{lib}}$ and $N_{B,R}$ compared to N_{cld} are largest for the sub-adiabatic cloud cases. The systematic overestimation of $N_{B,R}$ for all clouds is due to the lower $r_{\text{eff},R}$. The differences reduce for increasing N_{cld} because the differences between $r_{\text{eff},R}$ and $r_{\text{eff},\text{lib}}$ decrease. This clearly shows that a wrong estimation of r_{eff} influences the calculation of N most significantly, while τ contributes to a minor part only, independently which method is used. These results allow to conclude that r_{eff} must be retrieved close top. This is possible if the retrieval applies appropriate wavelength in the infrared, where radiation is effectively absorbed within the upper most part of the cloud. Otherwise systematic overestimation of N occurs. By applying method C the sub-adiabatic nature of the cloud profiles (II, IV, VI) is considered in the estimation of N . The calculated Γ_{calc} is assumed to be correct and identical to the profile of the constructed clouds, with $f_{\text{ad}} = 0.6$ and $\Gamma_{\text{calc}} = \Gamma_{\text{ad}} \cdot 0.6$, respectively. Therefore, it is obvious, that N calculated from method B and C are also identical for adiabatic clouds. In general, $N_{C,R}$ derived from method C is closer to N_{cld} than $N_{B,R}$. However, for the sub-adiabatic clouds (II, IV, VI) results for methods B and C differ. Cloud droplet number concentration $N_{C,\text{lib}}$ is closest to N for all cloud cases and methods. The same pattern is present for $N_{C,R}$ with the best agreement to N_{cld} compared to method A and B. Deviations in $N_{C,\text{lib}}$ and $N_{C,R}$ to N_{cld} are reduced with increasing N . This shows, that a correct assumption of Γ_{calc} , as possible with method C, is crucial for a reliable calculation of N and can compensate biases in N which result from the the sub-adiabatic cloud profile.

5.5 Calculation of Retrieval Uncertainty of Cloud Droplet Number Concentration

Cloud droplet number concentrations calculated with Eq. (7), Eq. (10), and Eq. (12) are mainly effected by uncertainties from τ , LWP and especially r_{eff} , but also depend on the accuracy of k , f_{ad} , and Γ_{ad} . To estimate the uncertainties of retrieved N , it is assumed that the errors are normally distributed and independent from each other. In this case the uncertainty of N_A from



Table 3. Overview of all six synthetic cloud cases. The predefined cloud liquid water path LWP and droplet number concentration N are denoted with subscript "cld". Cloud properties are calculated from the given cloud profile (subscript "lib") and retrieved from synthetic spectral cloud reflectivities (subscript "R"). Calculated N is listed for all three methods A,B, and C once using the predefined cloud properties "lib" and the retrieval results from "R".

	Cloud I	Cloud II	Cloud III	Cloud IV	Cloud V	Cloud VI
	adiabatic	sub-adiabatic	adiabatic	sub-adiabatic	adiabatic	sub-adiabatic
$N_{\text{cld}} [\text{cm}^{-3}]$	50	50	100	100	200	200
$LWP_{\text{cld}} [\text{g m}^{-2}]$	362	217	362	217	362	217
τ_{lib}	35.6	25.5	45.2	32.3	57.3	41.0
$r_{\text{eff,lib}} [\mu\text{m}]$	18.8	18.8	14.9	12.6	11.8	10.0
τ_{R}	37.1	25.7	46.9	32.8	59.4	41.9
$r_{\text{eff,R}} [\mu\text{m}]$	18.3	15.4	14.8	12.3	11.9	9.9
$LWP_{\text{R}} [\text{g m}^{-2}]$	452	264	462	270	471	276
$N_{\text{A,lib}} [\text{cm}^{-3}]$	53	69	106	137	215	274
$N_{\text{A,R}} [\text{cm}^{-3}]$	58	74	111	145	215	288
$N_{\text{B,lib}} [\text{cm}^{-3}]$	52	68	105	134	211	268
$N_{\text{B,R}} [\text{cm}^{-3}]$	57	73	108	143	207	280
$N_{\text{C,lib}} [\text{cm}^{-3}]$	52	53	105	104	211	208
$N_{\text{C,R}} [\text{cm}^{-3}]$	57	57	108	111	207	217

Eq. (7) is calculated by:

$$\Delta N = \sqrt{\left(\frac{\partial N}{\partial k}\right)^2 (\Delta k)^2 + \left(\frac{\partial N}{\partial f_{\text{ad}}}\right)^2 (\Delta f_{\text{ad}})^2 + \left(\frac{\partial N}{\partial \Gamma_{\text{add}}}\right)^2 (\Delta \Gamma_{\text{add}})^2 + \left(\frac{\partial N}{\partial \tau}\right)^2 (\Delta \tau)^2 + \left(\frac{\partial N}{\partial r_{\text{eff}}}\right)^2 (\Delta r_{\text{eff}})^2} \quad (14)$$

and analogous for Eq. (10) and Eq. (12). All uncertainties of N presented in the following sections are based on calculation by this approach. The uncertainties of the single parameters assumed in the calculations are summarized below.

- 5 For method A, B, and C, the uncertainty of k representing the shape of the droplet size distribution is set to $k = 0.8 \pm 0.1$ according to the range of values suggested by Martin et al. (1994) and Pontikis and Hicks (1992).

For methods A and B the degree of adiabaticity f_{ad} is fixed to one. In that case, no uncertainty in a measurement scene is attributed to f_{ad} . For method C, the uncertainty of f_{calc} is determined by the uncertainty of h_{top} , h_{base} and retrieved LWP following (Eq. 11). Cloud top height from WALES is determined with an accuracy of $\Delta h_{\text{top}} = \pm 20$ m. The cloud base height

- 10 is derived from single dropsondes and therefore prone to horizontal variability of T , p and T_{d} . Based on an analysis of different dropsondes in close vicinity, an uncertainty of $h_{\text{LCL}} = 660 \text{ m} \pm 35 \text{ m}$ is assumed. The evaluation of all dropsondes show that the thermodynamic conditions in the selected area stayed constant ($\Delta T < 2 \text{ K}$ and $\Delta p < 4 \text{ hPa}$) during the flight time with $h_{\text{top}} \approx 1800 \text{ m}$, $T_{\text{top}} = 20.2^\circ \text{C}$, and $p_{\text{top}} = 820 \text{ hPa}$. The accuracy of the deployed Vaisala dropsondes RD94 is reported to be



within $\Delta T = \pm 0.2$ K and $\Delta p = \pm 0.4$ hPa. Uncertainties of N_C caused by errors in Γ_{ad} are therefore negligible compared to the influence of τ and r_{eff} .

The adiabatic increase of LWC with height calculated from the Clausius-Clapeyron-Equation depends mostly on cloud top temperature T_{top} and to a lower degree on cloud top pressure p_{top} . Therefore Γ_{ad} depends on T_{top} and p_{top} , too. The cloud droplet number concentration is mostly effected by the assumed T_{top} whereby p_{top} is only of minor contribution. Despite that, the cloud top pressure more strongly affects warm than cold clouds (Grosvenor et al., 2018b). For the uncertainty calculation, a temperature difference of 2 K is considered, which changes Γ_{ad} by $\pm 0.1 \cdot 10^{-3} \text{ g m}^{-3} \text{ m}^{-1}$ for the reference value of $2.5 \cdot 10^{-3} \text{ g m}^{-3} \text{ m}^{-1}$.

The uncertainty of the retrieval of τ and $r_{eff,A}$ result from the measurements uncertainties of SMART which are described in Sec. 3.1. For typical trade wind cumuli uncertainties of ± 0.1 for τ and $\pm 1.1 \mu\text{m}$ for $r_{eff,A}$ are assumed.

Retrieval uncertainties of LWP from HAMP are in the range of $\pm 30 \text{ g m}^{-3}$ and mainly stem from the non-ambiguity of the retrieval problem and instrumental effects described in Section 3.2. Common microwave radiometer retrievals of LWP from satellite observations above 180 g m^{-3} are error-prone because of their large footprint. With the smaller footprint of HAMP the value of 180 g m^{-3} is only for rough orientation and the actual accuracy of LWP is a function of LWP itself and has to be estimated for each instrument. As a result, uncertainties in LWP and from within precipitating cloud cases can be even higher than indicated by the dotted lines (Seethala and Horvath, 2010), which results in a higher uncertainty in retrieved N_B and N_C .

Retrievals of $r_{eff,B}$ from combined measurements of SMART and HAMP are slightly more prone to the uncertainty of the LWP_B measurements and lead to uncertainties of $r_{eff,B}$ of up to $\pm 1.5 \mu\text{m}$. This is a higher uncertainty for r_{eff} than estimated for Method A. However, the uncertainty of N with respect to r_{eff} is lower as the sensitivity of N_B with respect to $r_{eff,B}$ is lower in Eq. (10) compared to Eq. (7). Additionally sun-glint or 3D radiative effects are omitted.

For the exemplary ideal adiabatic case discussed above, the total uncertainties of the three methods are for $\Delta N_A = \pm 7.1 \text{ cm}^{-3}$, $\Delta N_B = \pm 14.1 \text{ cm}^{-3}$ and $\Delta N_C = \pm 15.1 \text{ cm}^{-3}$ in the optimal adiabatic case. In case of sub-adiabatic clouds, the uncertainties of method A and B increase due to the assumption of adiabaticity. The additional error in N results from the increased variability in f_{ad} . Additionally the retrieved r_{eff} is biased to larger values because of the increased penetration depth due to lower optical thickness. Uncertainties for sub-adiabatic clouds increase to $\Delta N_A = \pm 10.9 \text{ cm}^{-3}$, $\Delta N_B = \pm 23.2 \text{ cm}^{-3}$ whereby the uncertainty ΔN_C remains constant.

From the error estimation of the retrieval of N it can be concluded that uncertainties in r_{eff} , LWP , and dz have to be minimized as they influence the retrieval the most. Determination of h_{base} , either from the dropsondes or the radar, and resulting dz have to be accurate within at least $\pm 60\text{m}$.

30 6 Results

The retrieval of N is applied to two measurement cases observed during NARVAL-II. Figure 2 shows the flight track of Research Flight 06 (RF 06) from 19 August 2016 and the flight section (19:24 to 19:39 UTC) of the track for which the remote sensing measurements are analyzed. The satellite image represents the cloud situation at 19:30 UTC. The presence of intense

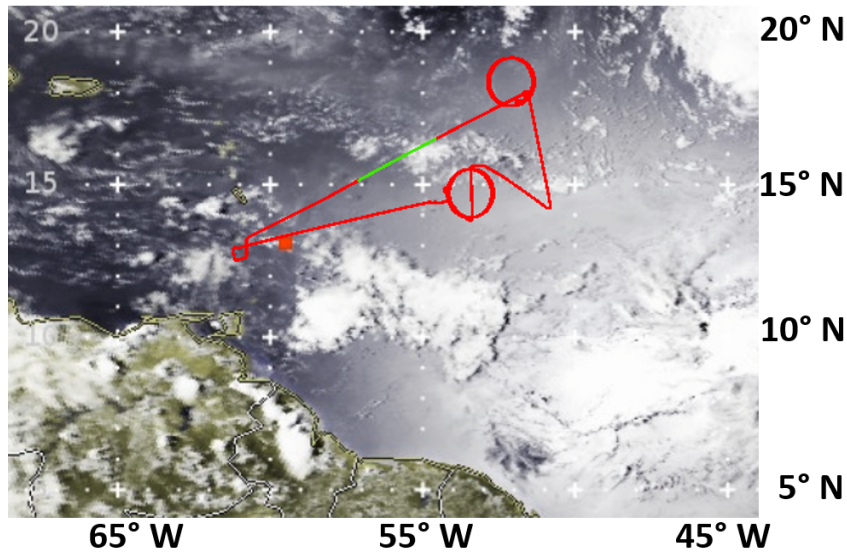


Figure 2. Flighttrack of HALO (red) from RF 06 (19. August 2016) plotted on a METEOSAT satellite composite image from 19:30 UTC. The section for which the remote sensing measurements are analyzed (19:24 UTC to 19:39 UTC) crosses a region with aggregated trade wind cumulus and is plotted in green.

sun-glint is visible, which enhances the reflected radiance I_{λ}^{\uparrow} and influences the cloud detection (low contrast) and the retrieval of τ and $r_{\text{eff},A}$. The analyzed time period is divided into two parts, cloud case #1 and cloud case #2. The north eastern part of the flight track (19:29-19:32 UTC) was dominated by aggregated trade wind cumulus, whereby in the south-western part (19:32-19:36 UTC) shallow cumulus were present. The general weather situation was characterized by moderate convection with low cloud top altitudes. Locally more dense cloud fields formed, at about 10° N and 16° N at 55° W.

Time series of measured and retrieved parameters of both cloud cases are shown in Fig. 3 and Fig. 4. The three methods to calculate N assume, that there is no precipitation. As the radar sensitivity is limited, measured Z is most sensitive to large cloud droplets and it can not be guaranteed that drizzle is completely excluded. In these cases the influence of precipitation is assumed to be negligible. Flight sections which are flagged for precipitation are highlighted by the grey boxes. At the top of Fig. 3 and 4 the cloud mask (blue) and the homogeneity cloud flag HCF (yellow) are indicated. Images of RGB composites by specMACS are given in the lower part of the plots to illustrate the visual cloud characteristics. Data gaps are due to cloud free pixel.



6.1 Cloud Case #1

Case #1 represents a single layer stratiform cloud without any convective areas which is an ideal test case for the retrieval of N . The cloud optical thickness τ shown in Fig. 3a is generally low and ranges between 0 and 2 at the beginning of the section, while τ increases up to 6 with time. The uncertainty of τ is estimated to be ± 0.1 . The effective radius $r_{\text{eff},A}$ (panel b, black line) ranges between $9.6 \mu\text{m}$ and $26.3 \mu\text{m}$ with an uncertainty of $\pm 1.0 \mu\text{m}$, while $r_{\text{eff},B}$ is between $8.3 \mu\text{m}$ and $30 \mu\text{m}$ retrieved with a slightly higher uncertainty of $\pm 1.5 \mu\text{m}$. For the first cloud part, the liquid water path obtained from SMART LWP_A (panel c) is calculated with Eq. (1) using retrieved τ and $r_{\text{eff},A}$. For the first part of the cloud LWP_A is slightly lower than the LWP_B measured by the microwave profiler, while with increasing τ the agreement between both LWP improves. Vertical profiles of Z shown in Fig. 3g are below the detection threshold except for four cloud patches. This indicates, that no precipitation was detected, whereby slight drizzle can not be excluded. Cloud base height is estimated from dropsondes to be around 1500 m, while h_{top} is determined by WALES. The resulting cloud geometric thickness dz (Fig. 3d) varies between 100 m and 420 m. Cloud adiabaticity f_{calc} (Fig. 3e) is mostly below 0.5 indicating a considerable sub-adiabatic cloud. Calculated N_A and N_B are shown in Fig. 3f and range between 5 cm^{-3} and 40 cm^{-3} which results from the low τ , LWP_B , large $r_{\text{eff},A}$, and $r_{\text{eff},B}$. The cloud droplet number concentration N_A shows a peak around 19:34:30 UTC and N_A at 19:35:00 UTC. Cloud droplet number concentration N_C derived by method C is lower than N_A and N_B and does show a reduced variability compared to N_A and N_B . However, the uncertainty of all N is about $\pm 15 \text{ cm}^{-3}$. While in the first part of cloud case #1 the differences in N are large, there is a good agreement between all three methods in the second part where all results are inside the uncertainty range of each method. Mean values of measured and retrieved parameters for cloud case #1 are listed in Table 4.

6.2 Cloud Case #2

The second case represents a more heterogeneous single-layer cloud observed between 19:29 UTC and 19:32 UTC and is shown in Fig. 4. This cloud is in a later state of development and shows moderate convection with slight precipitation. In these areas (highlighted in grey), the criteria for cloud homogeneity is not fulfilled. Despite that and the slight precipitation, calculation of N is performed, knowing that the retrieval of N using method A and B are prone to errors under this circumstances. These results are used to evaluate the improvement of retrieved N by method C which accounts for cloud geometry and sub-adiabaticity. The application of method C provides reasonable N by accounting for cloud-geometry and sub-adiabaticity. By comparing convective and non-convective areas of this cloud case #2, the limitations and advantages of the three methods are investigated. Mean values of the measured and retrieved parameters from the three different methods separated for non-precipitation and precipitation are summarized in Table 4. For the non-precipitating and homogeneous part of cloud case #2, τ does not exceed a value of 30 and $r_{\text{eff},A}$ and $r_{\text{eff},B}$ range between $18 \mu\text{m}$ and $40 \mu\text{m}$ (Fig. 4a, b). The uncertainty of all measured and retrieved parameters, is in a similar range as calculated for cloud case #1. Retrieved LWP from SMART and HAMP (Fig. 4c) agrees within the uncertainty range of HAMP for most parts of the homogeneous cloud sections. Larger differences appear around 19:29:30 UTC where LWP_A is larger than

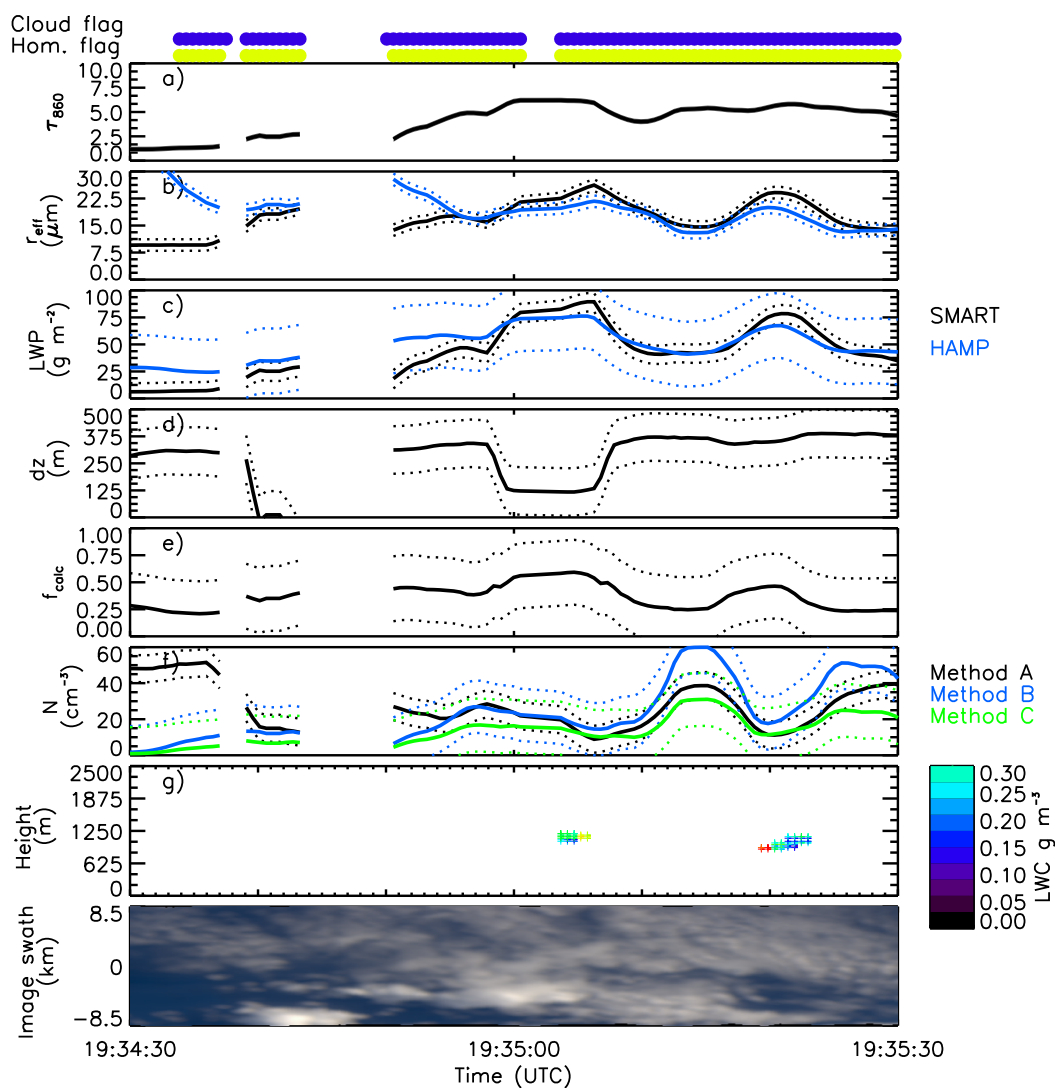


Figure 3. Time series of measured and retrieved cloud properties of cloud case #1 from 19:34:30 to 19:35:30 UTC of RF06. Cloud droplet number concentration N is shown for all three methods A, B, and C. Uncertainty ranges of the individual parameters are indicated by dotted lines. At the top, the cloud mask (blue) and the homogeneity cloud flag (HCF) (yellow) derived by SMART are indicated.



LWP_B . For method C, cloud geometrical thickness is calculated from a combination of HAMP and WALES. Radar reflectivity is above the detection threshold and allows to determine vertical profiles of the LWC and h_{base} with an average value of ≈ 900 m where no precipitation is present. Cloud top height is determined from WALES and ranges between 200 m and 1000 m for the non-precipitating regions. This results in a highly variable f_{calc} , which varies between strongly varies between
5 0.05 and 1.0.

Cloud droplet number concentration from method A and B calculated for cloud case #2 are generally low (see also Table 4) mostly ranging between 20 cm^{-3} and 40 cm^{-3} . Together with large $r_{eff,A}$ and $r_{eff,B}$ these values indicate typical pristine maritime clouds. An exception is observed around 19:29:30 UTC where N peaks up to 120 cm^{-3} for all three methods mostly resulting from a decrease of $r_{eff,A}$ and an increase of τ . The decrease of r_{eff} might result from 3D-radiative effects at the cloud
10 edge overestimating the cloud particle size and can have biased the retrieval of N .

In the areas marked with precipitation, retrieved τ , LWP_A , and LWP_B are higher compared to the precipitation free regions while $r_{eff,A}$ and $r_{eff,B}$ are in the same range as for the non-precipitating areas. In contrast to the homogeneous parts of the cloud, the convective regions show stronger horizontal heterogeneity in all parameters. The optical thickness reaches up to 40 and $r_{reff,A}$ ranges from $20 \mu\text{m}$ to $38 \mu\text{m}$. In these areas the LWP_B from HAMP exceeds 270 g m^{-3} and shows a maximum
15 value up to 500 g m^{-3} . Liquid water path from SMART is in the same range of LWP_B except for the first precipitation section (19:30:30 UTC) where LWP_B is lower than LWP_A . For the precipitating regions the cloud base height h_{base} is assumed to be at the same level as determined for the non-precipitating regions as precipitation makes the cloud base invisible for the radar. The cloud geometric thickness dz is slightly higher for the connective regions and ranges between 800 m and 1300 m. The calculated adiabaticity f_{calc} is lower than 0.5 for the majority of the measurement and shows that most parts of the cloud are
20 sub-adiabatic. For the precipitation regions calculated N are between 10 cm^{-3} and 90 cm^{-3} with the highest concentrations for method B, followed by method A and the lowest N for method C. In the areas with precipitation, N shows a systematic higher variability which is observed by all three methods and likely caused by the variability of r_{eff} retrieved from SMART. One reason for this variability is the relation of r_{vol} to r_{reff} which is assumed to be (i) constant in the retrieval of r_A and r_B and (ii) significantly influenced by formation of precipitation. Therefore, calculated N by all three methods are highly prone
25 to errors for precipitating clouds. The variability of N might also be caused by intense turbulent mixing processes within the cloud. Therefore, it is suggested to filter areas with stronger convection, precipitation, and heterogeneous scenes and analyze the retrieved N with special care.

6.3 Statistical Analysis of Liquid Water Path, Effective Radius and Cloud Droplet Number Concentration

Statistics of retrieved cloud properties are analyzed for measurements between 19:24:00 UTC and 19:39:00 UTC only, where
30 the HCF indicates homogeneous clouds and uncertainties of the retrieved cloud parameters is low. An extension of the analysis to other flights is not possible yet, because the reliable application of the retrieval of N requires careful data selection and good quality data of all individual instruments. However, in total still 700 individual measurements are included which represents a cloud field of 77 km length. The clouds were separated into precipitating (p) and non-precipitating (np) pixel. Mean for each measurement are given in Table 4.

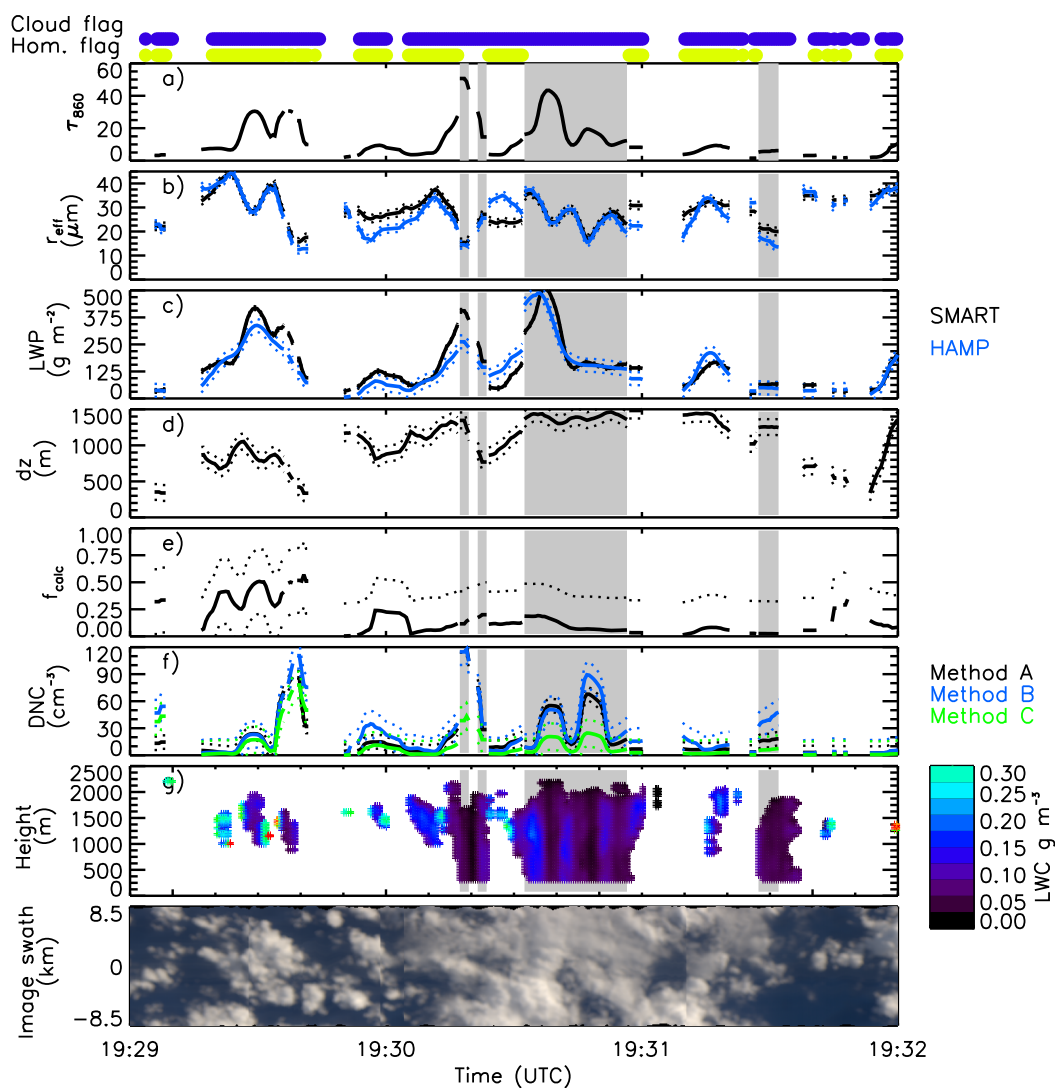


Figure 4. Time series of measured and retrieved cloud properties of cloud case #1 from 19:29 to 19:32 UTC of RF06. Cloud droplet number concentration N is shown for all three methods A, B, and C. Uncertainty ranges of the individual parameters are indicated by dotted lines. At the top, the cloud mask (blue) and the homogeneity cloud flag (HCF) (yellow) derived by SMART are indicated.



Table 4. Mean values of cloud properties of cloud cases #1 and #2.

parameter	Cloud case #1	Cloud case #2 (np)	Cloud case #2 (p)
τ	4.3	3.5	11.3
$r_{\text{eff,A}} [\mu\text{m}]$	17.1	30.4	24.9
$r_{\text{eff,B}} [\mu\text{m}]$	19.2	29.1	23.4
$LWP_A [\text{g m}^{-2}]$	45	135	226
$LWP_B [\text{g m}^{-2}]$	50	120	210
$dz [\text{m}]$	315	959	1315
$N_A [\text{cm}^{-3}]$	27	17	47
$N_B [\text{cm}^{-3}]$	26	25	53
$N_C [\text{cm}^{-3}]$	19	13	40

Figure 5 compares measurements of LWP_A and LWP_B . The data is separated for different $r_{\text{eff,A}}$ split into bins of $5 \mu\text{m}$ size. For the selected time period, LWP_A agrees with LWP_B within the uncertainty range of HAMP of $\pm 30 \text{g m}^{-2}$ indicated by the grey error bars. The differences of LWP_A and LWP_B show a larger variability for clouds with large $r_{\text{eff,A}}$ than for clouds with small $r_{\text{eff,A}}$. For larger cloud droplets, the retrieval uncertainty of τ and $r_{\text{eff,A}}$ increases and, therefore, also LWP_A derived from SMART. Additionally, SMART has a higher sensitivity to droplets at cloud top and the FOV of HAMP is slightly larger compared to SMART what can explain some of the observed variability. Slightly different viewing directions have to be considered too. While for SMART the LWP_A is calculated assuming an adiabatic profile with the retrieved $r_{\text{eff,A}}$ representing cloud top, HAMP obtains an integrated measure of LWP where all cloud layers are more homogeneously weighted and no assumption on the cloud profiles is required. Therefore, a difference between LWP_A and LWP_B indicates that the observed clouds are non-adiabatic. For $LWP_A > LWP_B$ less liquid water is at cloud base than predicted by adiabatic theory and clouds are sub-adiabatic. For $LWP_A < LWP_B$ liquid water at cloud top is reduced, likely by precipitation.

In Fig. 7 the normalized PDF of $r_{\text{eff,A}}$ retrieved from SMART only (method A) and $r_{\text{eff,B}}$ retrieved synergistically from SMART and HAMP from (method B) is presented separately for precipitating and non-precipitating clouds. The mean value for non-precipitating clouds is around $\bar{r}_{\text{eff,A,np}} = 23.2 \mu\text{m}$ and the median is at $r_{\text{eff,A,np,med}} = 21.1 \mu\text{m}$. This droplet size range agrees with in-situ measurements of pristine trade wind cumulus by Siebert et al. (2013) and remote sensing measurements by Werner et al. (2014) in the same geographic region. The distribution shows a bi-modal structure with a first mode around $15 \mu\text{m}$ and a second mode around $32 \mu\text{m}$. The PDF of $r_{\text{eff,A}}$ for precipitation situations shows a similar structure being shifted towards larger $r_{\text{eff,A}}$ with values of $\bar{r}_{\text{eff,A,p}} = 25.1 \mu\text{m}$ and $r_{\text{eff,A,p,med}} = 24.5 \mu\text{m}$. The first mode is at $21 \mu\text{m}$ and the second mode is at $36 \mu\text{m}$. The PDF's of $r_{\text{eff,B}}$ for the np clouds are shifted to larger values by approximately $3 \mu\text{m}$ additionally showing a third mode around $38 \mu\text{m}$. In contrast, the PDF for the p clouds is shifted to lower values by up to $8 \mu\text{m}$ and showing only the bi-modal structure with peaks around $15 \mu\text{m}$ and $33 \mu\text{m}$.

Figures 8a and b show normal normalized PDF of the calculated N for non-precipitating (a) and precipitating regions (b) of the

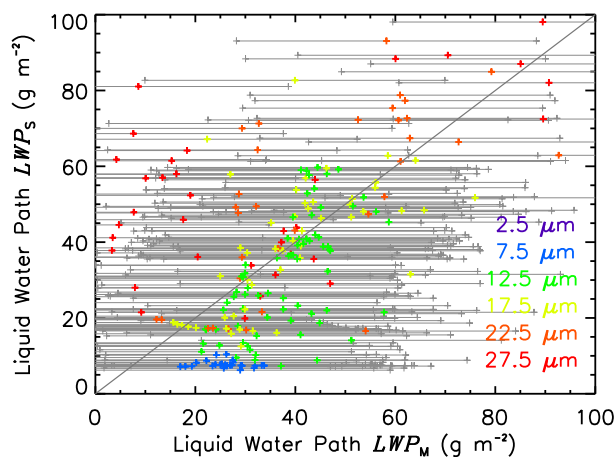


Figure 5. Comparison of liquid water path LWP_B from HAMP microwave radiometer and LWP_A calculated from τ and $r_{\text{eff},A}$ retrieved by SMART. The color code indicates different ranges of $r_{\text{eff},A}$. HAMP uncertainties of LWP ($\pm 30 \text{ g m}^{-2}$) are indicated by grey errors bars.

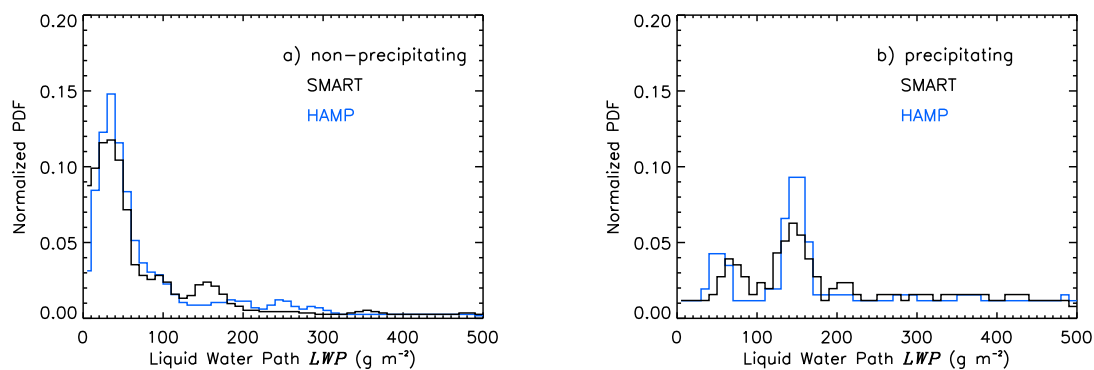


Figure 6. Normalized probability density function (PDF) of measured and calculated liquid water path LWP from HAMP (blue) and SMART (black). Distributions are filtered for non-precipitating a) and precipitating b) clouds.

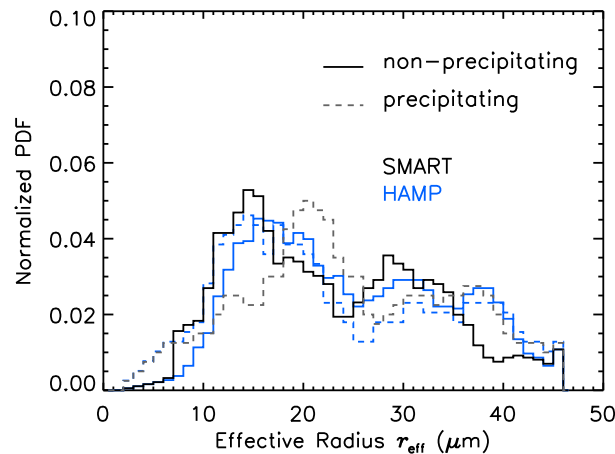


Figure 7. Normalized probability density function (PDF) of the effective radius $r_{\text{eff},A}$ retrieved by using the ratio of 1645 nm to 1050 nm in black and $r_{\text{eff},B}$ from the combined spectrometer-microwave retrieval in blue. Distributions are filtered for non-precipitating (solid line) and precipitating (dashed line) clouds.

selected flight-leg from all three methods A, B and C. For non-precipitating clouds (panel 8a) the distribution of N_A peaks at $\approx 30 \text{ cm}^{-3}$ with a steep decrease towards a concentration of $\approx 100 \text{ cm}^{-3}$. The first local maximum of the N_B distribution is at $\approx 30 \text{ cm}^{-3}$ slowly decreasing for larger N . Only a slight difference between N_A and N_B is present for higher N_A . This can be explained by the slightly higher values of LWP_A retrieved by SMART compared to LWP_B observed by HAMP. The PDFs of N_A and N_B show reasonable results for pristine, maritime clouds with relative large $r_{\text{eff},A}$ and according low N from method A and B. Cloud droplet number concentration from method C are significantly lower as a result of the considered adiabaticity of the individual clouds.

Measurements affected by precipitation compared to Fig. 8a show almost the same distribution with a shift to larger N for all three calculation methods, especially for method C. Filtering for precipitating clouds the statistic might be biased by only considering further developed clouds in which precipitation formation changes and broadens the droplet size distribution. This leads to differences in the means of r_{vol} and r_{eff} , influencing the k -parameter which is assumed to be 0.8 in the N calculation. Retrieving k by passive remote sensing is not possible yet (Wood, 2006).

Figure 9 shows the cloud top reflectivity \mathcal{R}_{532} measured by SMART at 532 nm as a function of N_B retrieved from combined SMART and HAMP measurements. Only measurements of the flight leg where no precipitation was observed are presented. The data is binned for two different LWP . Figure 9a shows clouds with LWP between $0 - 50 \text{ g m}^{-2}$ and Fig. 9b shows clouds in the range between $50 - 100 \text{ g m}^{-2}$. Colors represent $r_{\text{eff},B}$ binned from 5 to $30 \mu\text{m}$ in $5 \mu\text{m}$ steps (label in Fig. 9 refers to the mean bin value). Using \mathcal{R}_{532} as a measure for the reflectivity of the cloud, the sensitivity of \mathcal{R}_{532} changes of N is comparable to the model based sensitivity study in Section 2. Therefore, in Fig. 9 radiative transfer simulations of theoretical

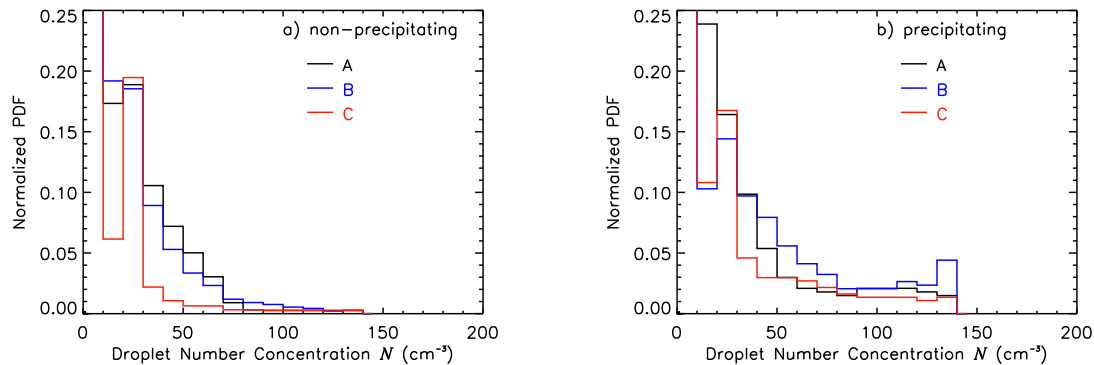


Figure 8. Normalized probability density function of the cloud droplet number concentration N for the selected flight path using method A, B, and C. Distributions are filtered for non-precipitating a) and precipitating b) clouds.

$\mathcal{R}_{532,\text{sim}}$ for clouds of the same LWP are added by the red line. For the thin clouds in Fig. 9a the measured \mathcal{R}_{532} shows a clear increase for higher N_B over the entire measurement range. This correlation is less pronounced for the thicker clouds in Fig. 9b due to a reduced range of \mathcal{R}_{532} and N , that may not cover the entire natural variability. However, for both cloud sub-samples, the measurements follow the theoretical line given by the simulations only that the measured \mathcal{R}_{532} are too low or retrieved N to high. Both might be attributed to measurement biases either the radiometric calibration of SMART or the retrieved LWP_B and $r_{\text{eff},B}$ which feed the calculation of N_B . Additionally, the homogeneous assumption of cloud properties applied in the RTS can lead to an overestimation of $\mathcal{R}_{532,\text{sim}}$ compared to the measurements. The subdivision of data for different $r_{\text{eff},B}$ shows that clouds in an early developing state with low LWP_B (Fig. 9a) are dominated by smaller cloud droplets up to $r_{\text{eff},B} = 17.5 \mu\text{m}$ whereby clouds in a later development state with higher LWP_B (Fig. 9b) are dominated by cloud droplets larger than $r_{\text{eff},B} = 17.5 \mu\text{m}$.

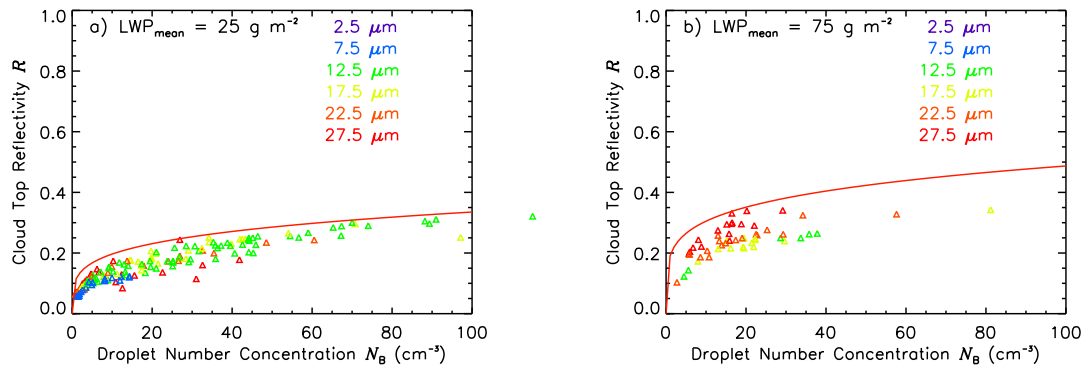


Figure 9. Cloud top reflectivity \mathcal{R}_{532} as a function of cloud droplet number concentration N_B for homogeneous, non-precipitating clouds of different liquid water path LWP_B (panel a: 0 – 50 g m^{-2} , panel b: 50 – 100 g m^{-2}). The droplet effective radius r_{eff} of each measurement is indicated by the color code. The red line represents simulated reflectivity \mathcal{R}_{532} from radiative transfer calculations for clouds with same LWP .

7 Conclusions

- Trade wind cumulus are an ubiquitous cloud type in the tropics, and it influences the Earth radiation budget significantly. Despite that, these clouds are not well represented in numerical weather prediction (NWP) and global climate models (GCMs), causing considerable uncertainties in the radiative calculations. It is well known, e.g., Platnick and Twomey (1994), that the
- 5 albedo α of clouds with low cloud droplet number concentration N and low liquid water path LWP , such as trade wind cumulus, respond very sensitive to changes of N . In order to obtain improved parameterizations and global distributions of N , several methods, including active and passive remote sensing from ground and satellite are developed, but no operational products are available yet. Only a limited number of field campaigns with in-situ measurements of selected cloud cases exist. As a result, the natural variability of trade wind cumulus is poorly covered by appropriate measurements.
- 10 In this paper, it is shown that shallow trade wind cumulus with LWP below 200 g m^{-2} and N below 100 cm^{-3} are very sensitive to changes in N and, therefore, their variable influence on the Earth radiation budget is high. In case of a LWP of 75 g m^{-2} and an increase of N from 50 cm^{-3} to 100 cm^{-3} , the cloud top albedo can increase by 0.1. As a result, the interaction between cloud top albedo α , N , τ , r_{eff} and different thermodynamic conditions (e.g., varying LWP) has to be investigated systematically.
- 15 Applying the common satellite retrieval techniques of N to measurements conducted with a high flying aircraft, such as the High Altitude and Long Range Research Aircraft (HALO), shows the potential of combined airborne passive and active remote sensing instruments. Using aircraft instead of satellite platforms allows to investigate specific cloud types under selected atmospheric conditions, eg.g., T_{top} , $p_{\approx 1}$, LWP .



Table 5. List of symbols, longnames and related units.

Symbol	Longname	Unit
α	Cloud top albedo	-
D	Cloud droplet diameter	m
dz	Cloud geometric thickness	m
f_{ad}	Degree of adiabaticity	-
F_{λ}^{\uparrow}	Spectral upward radiance	$\text{W m}^{-2} \text{nm}^{-1}$
F_{λ}^{\downarrow}	Spectral downward radiance	$\text{W m}^{-2} \text{nm}^{-1}$
Γ_{ad}	Adiabatic rate of liquid water content	$\text{kg m}^{-3} \text{m}^{-1}$
Γ_{calc}	Calculated rate of liquid water content	$\text{kg m}^{-3} \text{m}^{-1}$
h_{base}	Cloud base height	m
h_{LCL}	Lifting condensation level	m
h_{top}	Cloud top height	m
I_{cr}^{\uparrow}	Spectral upward irradiance threshold	$\text{W m}^{-2} \text{nm}^{-1} \text{sr}^{-1}$
I_{λ}^{\uparrow}	Spectral upward irradiance	$\text{W m}^{-2} \text{nm}^{-1} \text{sr}^{-1}$
$I_{\lambda, syn}^{\uparrow}$	Spectral upward irradiance (simulated)	$\text{W m}^{-2} \text{nm}^{-1} \text{sr}^{-1}$
k	k -parameter	-
l_{cld}	Cloud length	m
LWC	Liquid water content	kg m^{-3}
LWP	Liquid water path	kg m^{-2}
LWP_A	Liquid water path from SMART	kg m^{-2}
LWP_B	Liquid water path from HAMP	kg m^{-2}
N	Cloud droplet number concentration	cm^{-3}
N_{cld}	Cloud droplet number concentration of simulated clouds	cm^{-3}
p_{top}	Cloud top pressure	Pa
Q	Extinction coefficient	-
\mathcal{R}	Cloud top reflectivity	-
p_{Top}	Cloud top pressure	Pa
ρ_w	Density of liquid water	kg m^{-3}
r_{eff}	Effective radius	μm
$r_{eff,A}$	Effective radius from SMART	μm
$r_{eff,B}$	Effective radius from SMART & HAMP	μm
r_{vol}	Volumetric radius	μm



Symbol	Longname	Unit
τ	Cloud optical thickness from SMART	-
τ_{lib}	Cloud optical thickness from libradtran	-
T	Temperature	$^{\circ}\text{C}$
T_{d}	Dew-point temperature	$^{\circ}\text{C}$
T_{Top}	Cloud top temperature	$^{\circ}\text{C}$
t_{int}	Integration time of spectrometer	s
v_{aircraft}	Aircraft velocity	ms^{-1}
ϑ	Solar zenith angle	$^{\circ}$
Z	Radar reflectivity	dBz
ζ	Cloud top albedo sensitivity	cm^3

During the second campaign of the Next Generation Remote Sensing for Validation Studies (NARVAL-II), HALO was equipped with a set of passive and active remote sensing instruments. The Spectral Modular Airborne Radiation measurement sysTEM (SMART) measured upward and downward spectral irradiance F_{λ} and upward radiance I_{λ}^{\uparrow} , which allows to calculate α and retrieve τ and $r_{\text{eff,A}}$ at cloud top. The HALO Microwave Package (HAMP) enables to perform retrievals of liquid water path LWP and radar reflectivity Z to discriminate between non-precipitating and precipitating cloud sections. Combining measured values of I_{λ}^{\uparrow} by SMART and LWP by HAMP, alternative values of r_{eff} are retrieved, which are less influenced by sun-glint and 3D cloud radiative effects. Cloud top height is determined by the Water Vapour Lidar Experiment in Space (WALES) while the cloud base height is estimated from dropsondes or radar data.

In case of shallow trade wind cumulus, the heterogeneity of the cloud field has to be considered. This is especially important at the average flight speed of HALO ($\approx 220 \text{ ms}^{-1}$) and different instrument field-of-views (FOV), being in the range of individual clouds. This is indicated by the high occurrence (63%) of clouds with a horizontal size smaller than 300 m. To adapt the calculation of N to this heterogeneous cloud field, a careful cloud masking and filtering for homogeneous cloud sections is crucial. Using cloud flagging and masking to find clouds, which cover the entire FOV, N can be calculated for approximately 55% of all measured clouds.

Three different methods to retrieve N are presented. Possible biases from sun-glint in the solar wavelength range are avoided by using LWP instead of τ . Replacing τ by LWP , possible retrieval biases in method A resulting from sun-glint in the solar wavelength range measured by SMART, are avoided. Retrievals of LWP from HAMP apply wavelengths with several micrometer (20 to 100 GHz), which are not influenced by atmospheric constituents, like aerosol particles. Using the radiance ratio retrieval of SMART to derive r_{eff} from two infrared wavelengths, the remaining sun-glint influence on r_{eff} and possible absolute calibration errors are reduced. Calculating the true cloud geometric thickness dz from combined measurements of HAMP, WALES, and dropsondes the change of LWC with height of the cloud profile is determined and is used as a correction factor, replacing the adiabatic assumption in the common retrieval. Determination of LWP from HAMP further



allows to separate clouds for binned LWP and to untangle the effects of varying LWP on α which is a substantial advantage (McComiskey and Feingold, 2008). Radar reflectivity Z provided by HAMP allows to discriminate between precipitating and non-precipitating clouds.

5 The measurements of LWP must be performed within an uncertainty range of $\pm 30 \text{ g m}^{-2}$ or better to achieve the N accuracy of $\pm 14 \text{ cm}^{-3}$ from SMART. While $\pm 30 \text{ g m}^{-2}$ is chosen as a conservative value of LWP uncertainty from HAMP, a refined uncertainty estimation is ongoing, which will provide the the uncertainty of LWP as a function of LWP . For clouds with higher geometric thickness, measured LWP becomes more representative as information of the entire cloud is used instead of τ and r_{eff} from SMART representing the cloud-top only.

10 To bypass the adiabatic assumption in the calculation of N the possibility to determine the adiabaticity from combined microwave radiometer measurements of LWP , lidar cloud top height h_{top} measurements and cloud base height h_{base} assumptions from dropsondes is investigated. Despite constant thermodynamic conditions in the flight area, concerning temperature and pressure, the distribution of humidity is heterogeneous leading to a variable h_{base} . The cloud base height variability of several tens of meters does not allow to use a fixed h_{base} estimation from calculated lifting condensation level basing on dropsondes to obtain reasonable N .

15 Alternatively, cloud geometric thickness dz can be determined by the radar, despite it is limited to clouds with particles large enough to produce a detectable radar echo. Cloud base height depends on the selected threshold of Z which causes an uncertainty h_{base} in the range of tens of meters. If cloud boundary determination could be achieved in the range of the smallest vertical resolution of the radar with $\pm 30 \text{ m}$ and measurement with $\overline{LWP} = 100 \text{ g m}^{-2} \pm 30 \text{ g m}^{-2}$ are assumed, this results in $\Delta f_{\text{calc}} = \pm 0.1$ which is better than the approach with dz from dropsonde estimations. Nevertheless, a resolution of $\pm 15 \text{ m}$ is
20 only achieved in the optimal case for well-defined cloud edges. Cloud boundaries and thickness determined by radar are more precise but restricted to clouds with well-defined cloud edges, non-precipitating clouds and clouds with droplets detectable by the radar. When the cloud top height and cloud base height can be determined within an uncertainty of $\pm 25 \text{ m}$ by the radar, the uncertainty of the adiabaticity is below ± 0.12 but increasing for clouds with geometric thickness below 500 m . If the cloud base and top heights can not be determined within $\pm 25 \text{ m}$ the adiabatic assumption is more accurate and should be applied.

25 The application of the three presented methods is shown for synthetic measurements of six different clouds with N_{cld} of 50 cm^{-3} , 100 cm^{-3} , and 200 cm^{-3} each following an adiabatic and sub-adiabatic cloud profile. Overall, the sensitivity study leads to the conclusion, that an appropriate retrieval of r_{eff} is the most important factor for the calculation of N . Special care must be taken by considering the penetration depth and the wavelength selection. Depending on τ or LWP and N the penetration depth can reach several tenth meters inside the cloud, not representing r_{eff} at cloud top. Uncertainty of N has
30 to be considered carefully. The synthetic measurements and error analysis clearly indicate that uncertainty in N results from underestimation of r_{eff} due to the penetration depth of the incoming solar radiation into the cloud which depends on τ , N , and LWP . Therefore, clouds with low τ and LWP are most effected by errors of N , reducing for clouds with increasing τ . Replacing τ by the LWP from independent measurements by passive microwave radiometers can improve the accuracy of estimated N . Retrieved N of clouds with low N and τ , like trade wind cumulus, are more effected by errors compared to clouds
35 with higher N and τ . Considering only adiabatic clouds (I,III,V) this can be clearly assigned to errors in retrieved r_{eff} . For



- the sub-adiabatic cloud cases (II,IV,VI) the error in N also includes the shortcoming of assuming an adiabatic cloud profile. Therefore, a further improvement is achieved, when the actual sub-adiabatic profiles of the clouds (II,IV,VI) are considered. By determining the cloud geometry from active radar and lidar measurements and dropsondes, the resulting Γ_{calc} can be determined and used as a correction factor in the calculation of N . The sensitivity study shows that the differences between modeled N_{cld} and retrieved $N_{\text{C,lib}}$ or $N_{\text{C,R}}$ with method C, are significantly reduced comparing to method A or B, for all three cloud cases. This indicates that a correction with Γ_{calc} is vital and necessary for the calculation of N of shallow trade wind cumulus using remote sensing techniques. Otherwise systematic overestimation of retrieved N is present and not feasible. Nevertheless, errors resulting from the cloud boundary determination from the radar and dropsondes will lead to uncertainties in the calculated lapse rate.
- 10 All three methods are applied to a homogeneous and a heterogeneous cloud section. Determination of dz was relatively uncertain and method C was excluded from the statistical analysis. Probability density functions of LWP , r_{eff} , and N of the two cloud cases are presented in Sec. 6. Correlations of \mathcal{R}_{532} to N_{B} for two binned LWP_{B} are shown. These can be used to validate modeled \mathcal{R}_{532} , to describe the sensitivity of \mathcal{R}_{532} with respect to N , and allow to parameterize the Twomey effect better. Further testing of the proposed method to longer flight sections with more homogeneous cloud fields is necessary, to increase the covered natural variability of trade-wind cumulus. Despite remaining uncertainties and assumptions, the application of Γ_{calc} , the separation for different LWP , and the smaller FOV of all instruments, allow to investigate the cloud-radiation interactions better compared to large-scale averaging satellite measurements.
- 15

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