DISSERTATION

Airborne Passive Remote Sensing of Optical Thickness and Particle Effective Radius of Cirrus and Deep Convective Clouds

Submitted by

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Abstract:
Within this PhD thesis, the optical thickness and particle effective radius of cirrus and deep convective clouds (DCCs) are retrieved using passive remote sensing techniques. For this purpose, airborne and satellite spectral solar radiation measurements combined with extensive radiative transfer simulations have been conducted. Data analyzed in this thesis were collected during the ML-CIRRUS and the ACRIDICON-CHUVA campaigns, which aimed to study mid-latitude cirrus and DCCs using the German High Altitude and Long Range Research Aircraft (HALO), respectively. During the two campaigns, HALO was equipped with a comprehensive set of remote sensing and in situ instruments. In particular flights, closely collocated measurements with the overpasses of the Moderate Resolution Imaging Spectroradiometer (MODIS) aboard of the Aqua satellite were carried out. A cirrus located above liquid water clouds and a DCC topped by an anvil cirrus are investigated.

In general, the frameworks of this thesis can be divided into four parts. In the first part, the spectral upward radiance measured by the Spectral Modular Airborne Radiation Measurement System (SMART)-Albedometer are compared with that measured by MODIS. In the second part, a radiance ratio technique assuming a vertically homogeneous cloud is applied to obtain the cloud optical thickness and particle effective radius based on the measurements of SMART-Albedometer and MODIS. Multiple near-infrared wavelengths with different absorption characteristics are applied in the retrieval to study the vertical structure of particle sizes in the cloud. In the third part, the retrieved cloud properties are compared with those derived from the MODIS cloud products. For the cirrus case, the retrieved particle effective radius is further compared with in situ data measured by the Cloud Combination Probe (CCP) using a vertical weighting method. Although the comparison results in a good agreement, retrievals using this conventional technique only provides information on cloud particle sizes at the upper cloud layers, even if spectral measurements have been applied. The retrieved particle effective radius represents a vertically weighted value, where the upper cloud layers are weighted at most.
Within the fourth part of this thesis, an extended method based on a Bayesian optimal estimation has been developed to obtain the full vertical extent of particle effective radius. For this purpose, a parameterization assuming the shape of the vertical profile of particle effective radius with respect to a vertical coordinate within the cloud is applied. The analysis of the Shannon information content of SMART-Albedometer measurements is carried out to identify wavelengths that bring the most information pertaining to each retrieval parameters. The new retrieval technique is applied to the cirrus case to infer the profile of particle effective radius as a function of cloud optical thickness. The comparison between the retrieved and the in situ profile shows a good agreement with a deviation of about 5% at the cloud top and increases to values of up to 15% at the cloud base. The new retrieval technique has shown an excellent skill in improving the study of the vertical profile of cloud microphysical properties, which can be applied in the future generation of airborne and satellite retrievals based on the measurements of passive remote sensing.
## Contents

1 Introduction 1
   1.1 Development, occurrence, and properties of cirrus and deep convective clouds 1
   1.2 Cloud radiative impact .................................................. 7
   1.3 Remote sensing of clouds ................................................ 9
      1.3.1 Importance of airborne and satellite comparisons .......... 10
      1.3.2 Problems in remote sensing of cloud properties .......... 12
   1.4 Objectives and outline ............................................... 14

2 Definitions 17
   2.1 Radiative quantities ....................................................... 17
      2.1.1 Single scattering properties of individual particles .......... 21
      2.1.2 Volumetric optical properties .................................. 23
      2.1.3 Cloud microphysical properties ................................. 24
   2.2 Radiative transfer equation .......................................... 26

3 Measurements 29
   3.1 Airborne campaigns ...................................................... 29
   3.2 Remote sensing measurements ......................................... 31
      3.2.1 Spectral Modular Airborne Radiation Measurement System .... 31
      3.2.2 Moderate Resolution Imaging Spectroradiometer .......... 35
   3.3 In situ observations .................................................... 36
      3.3.1 Cloud Combination Probe ....................................... 36
      3.3.2 Water Vapor Analyzer ........................................... 37

4 Comparison of upward radiance 39
   4.1 Spectral and spatial resolution adjustments ......................... 39
   4.2 Data filter .................................................................... 42
   4.3 Results of upward radiance comparison .............................. 44

5 Retrieval of cloud optical thickness and particle effective radius 49
   5.1 Radiative transfer simulation and radiance ratio retrieval ........ 49
      5.1.1 Radiative transfer simulation .................................. 49
      5.1.2 Discriminating the properties of multilayer clouds .......... 51
      5.1.3 Radiance ratio retrieval ....................................... 53
## Contents

5.2 Sensitivity studies ........................................... 57  
5.2.1 Impact of underlying liquid layer clouds on the cirrus retrieval ................ 57  
5.2.2 Impact of ice crystal habit .................................. 58  
5.2.3 Retrieval uncertainties ....................................... 61  
5.2.4 Impact of underlying liquid water cloud on the cloud phase index ............... 62  
5.3 Vertical photon transport ...................................... 65  
5.3.1 Modelling vertically inhomogeneous clouds .................................. 65  
5.3.2 Vertical weighting function .................................... 66  
5.3.3 Impact of surface albedo on the vertical weighting function ..................... 69  
5.3.4 Impact of underlying liquid water cloud on the vertical weighting function .......... 71  
5.3.5 Vertical penetration depth ..................................... 73  

6 Comparison of cloud optical thickness and particle effective radius ................. 79  
6.1 Comparison of SMART-Albedometer and MODIS retrievals .......................... 79  
6.2 Comparison of particle effective radius between retrievals and in situ measurements .................................................. 82  

7 Retrieval of the vertical profile of particle effective radius ............................... 87  
7.1 Basic ideas ..................................................... 87  
7.2 Retrieval methodology and information content ....................................... 89  
7.2.1 Bayesian optimal estimation retrieval .................................. 89  
7.2.2 Shannon information content and wavelength selection ......................... 92  
7.2.3 Estimation of the forward model uncertainties .................................. 95  
7.3 Result and validation ........................................................................ 98  

8 Summary and conclusion ........................................................................ 103  
8.1 Airborne campaigns ..................................................................... 103  
8.2 Comparison of upward radiance .................................................. 103  
8.3 Retrieval of cloud optical thickness and particle effective radius .................. 104  
8.4 Comparison of cloud optical thickness and particle effective radius ............... 107  
8.5 Retrieving the vertical profile of cirrus properties .................................. 108  

Appendix A Matlab code to generate the cloud profiles ................................. 109  
Appendix B Impact of the vertical profile of particle effective radius on the spectral downward irradiance ................................................................. 113  

Bibliography ........................................................................ 115  
List of Symbols .................................................................... 133  
List of Abbreviations ............................................................. 137
1 Introduction

One of the most challenging issues of current climate research is understanding the impact of clouds and radiation interactions, which govern the global energy budget (e.g., IPCC, 2013). In this context, cloud retrievals have been developed to infer cloud optical and microphysical properties using passive remote sensing techniques. One assumption, which is applied by the common retrieval techniques is about the vertical profile of the cloud particle sizes. It is known from in situ measurements that cloud particle sizes considerably vary with altitude. In non-precipitating ice clouds, ice crystal sizes generally decrease as a function of altitude. Conversely for non-precipitating liquid water clouds, droplet sizes typically increase with increasing altitude. However, these vertical variabilities are commonly ignored despite their importance for the cloud radiative energy budget. This thesis is focused on investigating the cloud properties and their vertical distribution using measurements of passive remote sensing. Parts of this thesis have been published by Krisna et al. (2018). This introduction will discuss some of the fundamentals of cirrus and deep convective clouds (formation, occurrence, and properties), cloud radiative impacts, remote sensing of clouds, importance of airborne and satellite comparison, and problems in remote sensing of cloud properties. This introduction is enclosed with the objectives and the outline of the thesis.

1.1 Development, occurrence, and properties of cirrus and deep convective clouds

Cirrus is defined as a cloud forming in the Earth’s upper troposphere at temperatures somewhat below -30°C, composed of ice crystals and forming long and wispy trails (Lynch et al., 2002). This characteristic shape, in the form of a curl of hair, results from evaporation in vertical shear of horizontal winds. The International Satellite Cloud Climatology Project (ISCCP) defines cirrus as high-level clouds with cloud top pressures below 440mb [i.e., above approximately 6 km altitude, Rossow et al. (1996)]. However, the maximum altitude is modulated by the height of the tropopause, thus, leading to latitudinal cloud top height variations between 4–20 km (Dowling and Radke, 1990). There are a number of types of cirrus clouds, with the most frequent ones occurring in layers or sheets with horizontal dimensions of hundreds or even thousands of kilometers. Compared to liquid water clouds, which contain relatively high concentrations of small liquid water droplets, cirrus clouds...
typically have lower concentrations of larger ice crystals with various nonspherical shapes (Dowling and Radke, 1990). Sassen et al. (2002) and Sassen and Wang (2008) summarized that the generating mechanisms responsible for cirrus formation involve several factors, such as synoptics (jet stream, fronts, etc.), injection cirrus (thunderstorm anvil-derived), mountain-wave updraft, cold trap (tropopause-topped thin layer), and contrail cirrus (rapid cooling of aircraft exhausts). Ice particle formation in the upper troposphere is realized by two different processes: homogeneous or heterogeneous ice nucleation. The former process describes the spontaneous freezing of pure or highly diluted supercooled liquid water droplets at temperatures below approximately -36° to -40° C. At higher temperatures, ice particle formation can only be triggered by the presence of aerosol particles acting as ice nuclei (heterogeneous ice nucleation). Besides these two primary mechanisms, there are so-called secondary or ice multiplication processes (Heymsfield, 2005).

Satellite observations have shown that cirrus are largely widespread over the entire globe (Sassen and Wang, 2008). Fig. 1.1a shows the annually averaged global distributions of identified cirrus clouds (both single and multiple layers) based on 5° longitude by 5° latitude (up to the 85° latitude satellite viewing limit) grid boxes. The figure indicates that the cirrus coverage is minimum over the polar high latitudes and the upper mid-latitudes bounding the tropical belt. Over land areas, the deserts and desert-like regions (e.g., northern Africa, western China, central Australia, and the southwestern United States) cirrus occurs less

**Figure 1.1:** Latitude versus longitude display mean cirrus (a) and deep convective clouds (b) frequency of occurrence derived from CALIPSO and CloudSat data (Sassen and Wang, 2008; Sassen et al., 2009). For the cirrus, the data are 1-year averaged, while for the deep convective cloud is 2-year averaged and of single and multilayer clouds.
1.1. Development, occurrence, and properties of cirrus and deep convective clouds

Frequent. With values of up to 60%, the maximum coverage of cirrus occurs in the tropical belt (\(\sim \pm 15^\circ\) latitude) at relatively high altitudes as a result of anvil produced by deep convective systems associated with the Intertropical Convergence Zone (ITCZ) and seasonal monsoonal circulations. Another source of tropical thin-to-subvisual tropopause cirrus is called as "cold trap", which has unique characteristics: they occur under very cold (\(-70^\circ\) to \(-90^\circ\) C) and high (15-20 km) conditions, which are rarely encountered outside the tropics. According to Heymsfield (1986) and McFarquhar et al. (1999), these tenuous layers are composed of relatively small ice crystals and are maintained by moisture supply from deep convection.

Less but still significant amounts of cirrus are observed in the northern and southern mid-latitude, which are mostly formed in connection with frontal and low-pressure systems as well as with the jet streams (Lynch et al., 2002). Thus, they are mostly present in the storm track regions at somewhat lower altitudes (Sassen and Wang, 2008). In addition, the formation of cirrus can be induced by the dense air traffic (IPCC, 1999; Schumann, 2005).

A cirrus contrail (a term introduced for "condensation trail" in 1942 by British pilots) is a visible cloud forming behind plane, mainly due to water vapor emissions from the aircraft engines in the upper troposphere. Contrails form thermodynamically according to the Schmidt-Appleman criterion (Schmidt, 1941; Appleman, 1953) when the relative humidity (RH) in the plume of exhaust gases mixing with ambient air temporarily reaches or exceeds the saturation water vapor pressure over a liquid surface, so that liquid droplets form on cloud condensation nuclei (CCN) and soon freeze to ice particles. From Fig. 1.1a, the cirrus coverage over Europe can be as high as about 30%, indicating further increases due to dense air traffics in this region. The properties of cirrus clouds are highly variable, which is caused by the interaction of a multitude of generating mechanisms, by their temperatures, potentially distinct updraft strengths, and availability of cloud-forming nuclei.

On the other hand, deep convective clouds (DCCs) are significant meteorological phenomena because they are associated with severe thunderstorms and heavy precipitation causing serious human hazards. DCCs are a major risk for aviation because airplanes are endangered by strong vertical motion, turbulence, and icing (Mecikalski et al., 2007). DCCs are optically thick and can extend to the tropopause and be topped by an anvil developing in the outflow. Their life cycle is determined by complex microphysical processes including different cloud particle growth and shrinking mechanisms, changes of the thermodynamic phase with respective latent heat release and consumption, and the development of precipitation. Thus, improving knowledge of DCCs is a fundamental issue in atmospheric science. The study of DCCs has been pioneered by Byers (1949) demonstrating the life cycle of DCCs using aircraft and radar measurements. Among others, Byers (1949) summarized that the life cycle of DCCs comprises of a developing stage, a mature stage, and a dissipating stage. The dominant cloud types at the three stages are cumulus, cumulonimbus, and anvil cirrus, respectively. A convective system is initiated with the Sun that heats the ground, which in turn warms the air directly above it. The warmer air expands,
becoming less dense than the surrounding air mass. Since warm air has a lower density than cool air, the warm air rises within cooler air. Clouds start to form as relatively warmer air carrying moisture rises. When the moist air rises, it cools causing some of the water vapor in the rising packet of air to condense. When the moisture condenses, it latent heat that allows the rising packet of air to cool less than its surrounding air, accelerating the ascent. If enough instability is present in the atmosphere, this process will continue long enough for cumulonimbus clouds to form, which support lightning and thunder.

The development of DCCs is controlled by a variety of atmospheric processes. It has been generally accepted that the initiation of cloud particles (droplet nucleation), the transition from liquid water particles to ice crystals (ice nucleation), and the thermodynamic conditions are key processes determining the evolution of DCCs (e.g., Andreae and Rosenfeld, 2008; Rosenfeld et al., 2008b). The thermodynamic range of conditions is mostly captured by the cloud base temperature. Warmer cloud base means larger vertical distance for the development of warm rain below the freezing level. Droplet and ice nucleation are influenced by properties and concentration of cloud condensation nuclei (CCN) and ice nucleating particle (INP). Along with the thermodynamic condition, the availability of CCN and INP in the atmosphere will impact the vertical development, microphysical properties, cloud-top height, and electrification of DCCs (Rosenfeld et al., 2008a; Tao et al., 2012; Li et al., 2011; Albrecht et al., 2011; Morrison and Grabowski, 2013; Fan et al., 2013). Considering the variety of aerosol conditions ranging between pristine and highly polluted, containing small and large concentrations of CCN/INP, investigations of aerosol-cloud interactions are currently in the focal point of the atmospheric science community (Heintzenberg and Charlson, 2009). In this regard, the ice nucleating capability (ratio between CCN and INP) affects the mixed and ice phase processes. The impact of ice nucleation on the cloud development comprises of two major issues: (a) the latent heat release in conjunction with the phase transition from liquid to ice water, and (b) the formation of precipitation.

The precipitation type of DCCs differs depending on the life stage (Inoue et al., 2009; Byers, 1949). Convective rain is dominant during the developing and mature stages, whereas stratiform rain is dominant at the decaying stage of the DCCs life time. Warm rain processes (excluding ice phase) dominate the precipitation formation in tropical DCCs (Phillips et al., 2002; Reisin et al., 1996). However, in extratropical convection which develops in colder environments, the ice phase is increasingly important for precipitation formation. The latent heat release by freezing particles accelerates the vertical motion in the cloud, which results in a high altitude and large horizontal extent of the anvil. The stronger the heat release, the larger the updraft speed and the possibility that the cloud top reaches the tropopause region. Overshooting convection (convection penetrates through the tropopause) is a major source for stratospheric water vapor and aerosol particles (de Reus et al., 2009; Chaboureau et al., 2007). In general, the phase transition from liquid water to ice is sensitive to the dominant liquid droplet and ice nucleation processes and the amount of available CCN and INP (Phillips et al., 2005; Leroy et al., 2009; Carrio et al., 2007). It was found that homogeneous
droplet freezing is the source of almost all of the ice crystals in the anvil (Phillips et al., 2005). The primary liquid droplets were formed by aerosol particle activation far above cloud base in the interior of the cloud (so-called secondary droplet nucleation). As shown by Phillips et al. (2002), ice crystals tend to be smaller and more numerous in the anvil with increasing CCN concentrations (increasing droplet nucleation). The corresponding sensitivity with respect to INP concentrations is much lower and shows a considerable impact on the cloud properties only for unrealistic high concentrations typical for plumes of desert dust (Phillips et al., 2007).

Fig. 1.1b shows the frequency of occurrence of DCCs in tropics ($\sim \pm 30^\circ$ latitude). Sassen et al. (2009) reported that the mean DCCs frequency of occurrence in the tropics is about 5% with the maximum of up to 15% found near equator (e.g., eastern Indonesia and northwestern Brazil). The geographical distribution pattern of DCCs is quite similar to that of the cirrus coverage. Thus, the general occurrence of cirrus is well correlated with the locations of DCCs in the tropical region. Cirrus frequencies are typically lowest (less than 10%) where deep convection is essentially absent. This implies that the formation mechanisms of the two cloud types are related, either directly through anvil spreading, or indirectly. The indirect connections may involve cirrus occurrence changes due to the radiative effects from lower, colder cloud tops versus the warm surface, tropopause transitional layer (TTL) humidification via penetrating turrets, regional updrafts via a pileus cloud analog, and gravity waves spawned by convection (Wang and Feingold, 2009). Some regional exceptions to the cirrus/DCC linkage occur, such as off the coasts of Ecuador and equatorial western Africa, where deep convection ends along the coastlines, and what appears to be anvil blow-off continues to the northwest over ocean.

According to Sassen and Wang (2008), the properties of cirrus and DCCs anvil depend on temperature, updraft strengths, and available supply of ice particle forming nuclei. Due to the complex formation, there is an extreme natural variation of cirrus and anvil properties with regard to spatial extent and cloud structure, ice particle number concentrations, sizes and shapes/habits (Dowling and Radke, 1990). Among other, Baran (2009) summarized that ice crystal sizes in cirrus range from $\leq 10 \mu$m to several thousand micrometers. Ice particle sizes in synoptically generated cirrus in mid-latitudes range between approximately 10–4000 $\mu$m but due to relatively low updraft velocities largest crystals are typically less than 1500 $\mu$m (Baum et al., 2005). Due to growth processes and gravitational settling smallest crystals are observed at cloud top, largest towards cloud base (e.g., Francis et al., 1998; Gayet et al., 2004). In tropical cirrus associated with deep convection and strong updrafts, larger crystals can exist at cloud top close to the convective core (Heymsfield, 2003), particles up to centimeters in size are reported (Baum et al., 2005). During cirrus anvil outflow these large crystals settle gravitationally so that at cloud top small crystals are observed (Garrett et al., 2003) and crystal sizes increase towards cloud base (Heymsfield, 2003). In optically thin tropical cirrus small crystals of only a few micrometer in size are common (McFarquhar and Heymsfield, 1996). However, it is important to note that in situ
measurements of ice particle sizes and number concentrations are technically challenging and a number of instrumental uncertainties exist, especially in the detection of small ice crystals (Gayet et al., 2002; Brenguier et al., 2013). In situ measurements are affected by shattering of the ice crystals on themicrophysical probes, which leads to an overestimation of the number concentration of small crystals (Heymsfield, 2007; Brenguier et al., 2013). In general, the complexity of ice crystal shapes increase toward the cloud top due to e.g., sedimentation, aggregation, and riming processes.

Fig. 1.2 shows images of ice crystals as a function of altitude (temperature), as reported in Heymsfield (2003). The data were sampled in mid-latitude regions on 25 November 1991, near Coffeyville, Kansas, USA during the National Aeronautics and Space Administration
1.2 Cloud radiative impact

(NASA) First ISCCP Regional Experiment II (FIRE-II). Despite the large variability of observed crystal shapes, which are influenced by dynamic and thermodynamic processes, some general features can be summarized. At low temperatures and dynamically undisturbed cloud top regions, small and pristine crystals in the form of hexagonal plates, bullets and bullet-rosettes are most common (e.g., Heymsfield and Platt, 1984; Heymsfield, 2003). Under these conditions deposition nucleation is the dominant ice particle formation process. Hexagonal shapes are favored because such structures maximize attraction forces and minimize repellent forces within the ice crystal lattice. Ice crystals with a maximum dimension smaller than about 30 µm are often of quasi-spherical shape (Korolev and Isaac, 2003). They are called droxtals (artificial word based on droplet and crystal), which are believed to form, either when supercooled liquid droplets freeze so rapidly that hexagonal planes cannot build up, or when hexagonally shaped ice crystals start to melt from the edge.

In mid-latitudinal cirrus, the most typical ice crystal shapes are bulletrosettes and non-symmetric aggregates, while hexagonal plates and columns are observed less frequently (Korolev et al., 2000; Gayet et al., 2004; Lawson et al., 2006). For tropical anvil cirrus, planar crystals, rosettes, and irregulars that occasionally build chain-like aggregates, are reported (Baran, 2009). However, turbulences may lead to uncoordinated crystal growth. With their long persistence of several hours, after convection has ceased, high tropical cirrus clouds are of particular interest for further studies from a chemical, water-cycle, and radiative energy budget point of view (Edwards et al., 2007). Sedimentation of ice crystals leads to a dehydration of the atmosphere and might explain the observed low relative humidities in the tropical stratosphere (e.g., Hartmann et al., 2001; Holton and Gettelman, 2001). A recent study has been performed by Järvinen et al. (2016) investigating ice crystal shapes and sizes in the anvil of tropical DCCs at altitudes around 12 km. Overall, they classified 23% of the imaged ice particles as frozen droplets and 19% as small (≤50 µm) irregular ice particles. Smaller observed fractions are comprised of plates (9%), bullet rosettes (14%), columns (3%), and aggregated ice particles (15%).

1.2 Cloud radiative impact

Covering about 75% of the Earth, clouds are a crucial component in modulating the Earth’s energy budget (Wylie et al., 2005; Kim and Ramanathan, 2008; Stubenrauch et al., 2013). They influence the radiative energy budget by scattering and absorption of solar radiation (in the wavelength range $\lambda = 0.2-4 \, \mu m$), as well as additional emission of terrestrial radiation ($\lambda = 4-100 \, \mu m$) in the atmosphere. Wielicki et al. (1995) reported as globally averaged that clouds tend to cool the climate due to the strong albedo effect (reflection of incoming solar radiation), which leads to the solar cooling (about $-50 \, W \, m^{-2}$) dominating over the terrestrial warming due to the greenhouse effect (about $+30 \, W \, m^{-2}$) with an uncertainty of less than 10% (Loeb et al., 2009). While thin cirrus are expected to warm the atmosphere and surface below the cloud, thick cirrus may rather cool (e.g., Liou, 1986; Wendisch et al.,
Thick cirrus clouds reflect a large amount of incoming solar radiation back to the space, reducing the transmitted radiation through the cloud. In this context, a cooling is expected. In turn, if the clouds are thin, the fraction of transmitted radiation is enhanced. Given that cirrus clouds are situated at high altitudes (below freezing level), the terrestrial radiation emitted by the atmosphere and the Earth’s surface are absorbed and further re-emitted (greenhouse effect) enhancing the terrestrial warming. Thus, when the Sun is low (e.g., Arctic winter or during night time), thereby the solar forcing is highly reduced, cirrus clouds may result in a warming.

On the other hand, the ways DCCs regulate the radiative energy budget in the atmosphere are more complex. Not only the reflection of solar radiation and absorption or emission of terrestrial radiation, the changes of liquid and ice water and hydrometeor profiles also have significant roles (Jensen and Del Genio, 2003; Sherwood et al., 2004; Sohn et al., 2015). DCCs typically have high optical and geometrical thicknesses (low-warm cloud base and high-cold cloud top). Due to the high optical thickness, they strongly reflect the incoming solar radiation back to the space, which lead to a significant solar cooling at the atmosphere and the surface below the cloud. A higher thermal emission is expected due to a warmer temperature at the cloud base. Given that the cloud top is high, the greenhouse effect of DCCs is considerably large to enhance the terrestrial warming. Through the development of DCCs, ice crystals are often formed in the top layers (anvil cirrus). By this condition, the solar cooling is strengthened because ice crystal particles typically have a higher albedo.
Overall, tropical DCCs are expected to cool the atmosphere and the surface below the cloud because the solar cooling largely dominates over the terrestrial warming.

Whether clouds result in a warming or cooling, it depends on their optical and microphysical properties, thermodynamic phase, altitude, underlying surface type, geometry, as well as on the ice crystal shape in case of ice clouds (e.g., Shupe and Intrieri, 2004). Two important cloud properties, which govern the radiative properties, and thereby the cloud radiative impact, are the cloud optical thickness $\tau$ and the effective radius $r_{\text{eff}}$ of the cloud particle population (liquid water droplets or ice crystals). Fig. 1.3 shows the total radiative forcing $\Delta F_{\text{tot}}$ (solar + terrestrial forcing) of ice clouds (a) and liquid water clouds (b) calculated at the Earth’s surface ($z = 0$) as a function of $\tau$ and $r_{\text{eff}}$. The $\Delta F_{\text{tot}}$ is calculated based on simulations for $\tau$ between 0.1 and 25 and $r_{\text{eff}}$ between 5 and 25 µm. The simulations are performed in a condition with a solar zenith angle $\theta_0$ of 36° and by assuming a typical surface albedo of ocean. The results show that $\Delta F_{\text{tot}}$ varies with $\tau$ and $r_{\text{eff}}$. Under this specific condition, clouds tend to cool the Earth’s surface due to the domination of the solar cooling over the terrestrial warming. Enhancing $\tau$ increases the cooling as more incoming solar radiations are reflected back to the space. On the other hand, increasing $r_{\text{eff}}$ results in a more warming due to larger absorption and re-emission by the cloud particles. For the ranges of $\tau$ and $r_{\text{eff}}$ analyzed here, the values of $\Delta F_{\text{tot}}$ are in the range of -8 to -579 W m$^{-2}$ for ice clouds and -2 to -494 W m$^{-2}$ for liquid water clouds. While $\tau$ and $r_{\text{eff}}$ values for both clouds are the same, it is obvious that ice clouds lead to a stronger cooling compared to liquid water clouds. This is caused by the different cloud thermodynamic phases, where a ice crystals have a higher albedo compared to liquid water droplets.

1.3 Remote sensing of clouds

Investigating cloud properties using in situ measurements remains challenges (e.g., Freud et al., 2008; Heymsfield, 2005; Rosenfeld et al., 2006). Due to the limitations in time and space, in situ measurements can only give a snapshot of the complexity of cloud properties. Apparently, this became one important reason why airborne and satellite remote sensing techniques have been developing. While satellite remote sensing offers statistical data and continuous measurements to determine long-term evolution of cloud properties, airborne remote sensing enables to perform closer measurements to the cloud. Thus, the uncertainties due to interferences by atmospheric gases and molecules between the cloud and the airborne sensor are smaller. Passive remote sensing techniques of cloud properties are well-established technique (e.g., Stephens and Kummerow, 2007). These techniques commonly rely on reflected solar or emitted terrestrial spectral radiance or reflectivity measured either by satellite or aircraft. Cloud properties are obtained by an inverse technique, realized either by interpolating a set of measurements within pre-calculated lookup tables (Nakajima and King, 1990) or by an optimal estimation technique (Rodgers, 2000).
Lately, satellite active sensors have been established. The Cloud-Aerosol Lidar with Orthogonal Polarization (CALIOP) on the Cloud-Aerosol Lidar and Infrared Pathfinder Satellite Observations (CALIPSO) satellite uses lidar to derive vertical profiles of cloud properties. Nevertheless, it is limited to optical thin clouds due to the strong attenuation of the laser beam inside clouds (Winker et al., 2009). Additionally, a radar system is applied on board of CloudSat to derive profiles of cloud liquid or ice water content and precipitation (Stephens et al., 2002, 2018). While using active sensors allows to study the profiles of cloud properties, both lidar and radar are limited in time (polar orbiting satellite) and space (narrow field of view). Thus, active remote sensing techniques are less convenient for investigating cirrus and DCCs. To extend the spatial coverage, passive imaging sensors on polar orbiting satellites have to be used, e.g., the Moderate Resolution Imaging Spectroradiometer (MODIS) on Aqua and Terra satellites. To add the temporal component for continuous observations, passive imaging sensors on geostationary orbiting satellites, e.g., the Advanced Very High Resolution Radiometer (AVHRR) on the National Oceanic and Atmospheric Administration (NOAA) weather satellites or the Spinning Enhanced Visible and Infrared Imager (SEVIRI) on Meteosat second generation, can be an option (Rosenfeld and Lensky, 1998; Kox et al., 2014).

1.3.1 Importance of airborne and satellite comparisons

The performance of post-launch validation activities is crucial to verify the quality of satellite measurement systems. It is essential to assess all components, i.e., sensors, algorithms, primary measured radiative quantities, and derived cloud products, as well as to continue validation activities throughout the satellite lifetime (Larar et al., 2010). In general, the performance of satellite instruments (e.g., MODIS) rely on extensive pre-launch calibration of the sensors and further monitoring of the sensor stability (Xiong et al., 2009; Heidinger and Stephens, 2002). For this purpose, on-board calibrations by a standard lamp or a solar diffuser are applied. However, the aging of calibration instruments may result in an imperfect calibration (Xiong and Barnes, 2006).

Studies have shown that the performance of satellite sensor will continuously decrease following the satellite lifetime (Rossow and Schiffer, 1991; Rao, 1999). Heidinger (2003) estimated a drift in the radiances measured by AVHRR of about 3 % per year. The response of the Geostationary Operational Environmental Satellite (GOES)-8 imager decreases by 4–6 % annually, depending on the chosen time period (Knapp and Vonder Haar, 2000; Le Borgne et al., 2004). For the Meteosat series, Rigollier et al. (2002) showed that the calibration coefficients alter up to 40 % between the years 1983 and 1997. By comparison of different corrections applied to the Meteosat images, biases in the corrected radiances of up to 6 % were found. For MODIS sensors, Barnes et al. (2003) analyzed the degradation of the two sensors for nearly three years of global data in 36 spectral bands. MODIS has experienced failures in a power supply and a formatter and is muting on the redundant units. Trending
of 2.5 years data has revealed decreasing sensitivity on the solar diffuser between 1 and 8\%.

To perform continuous validation, different techniques have been developed. Traditional methods use radiometrically stable surfaces like deserts (Rao et al., 2001; Heidinger, 2003), oceans (Moulin and Schneider, 1999; Grau et al., 2002), stars (Bremer et al., 1998; Weinreb et al., 1999), or the Moon (Xiong et al., 2003; Godden and McKay, 1998) to monitor the relative degradation of sensors during the satellite lifetime. Other methods derive absolute calibration coefficients of the sensors for measurements over ocean (Knapp and Vonder Haar, 2000) and/or deserts (Arriaga and Schmetz, 1999). These studies compare the raw data of the sensors to results of radiative transfer models. This implies an accurate knowledge on the present conditions (incoming solar radiation, viewing angle, aerosol particle properties) which affect the reflected solar radiation. However, this approach remains challenges, such as bidirectional effects, spectral and seasonal behavior of the target surface and the difficulty to obtain a dense temporal sampling.

An alternative method for deriving an absolute calibration is to apply a cross calibration between visible channels of several meteorological satellites against well-calibrated instruments of research satellites Minnis et al. (2008, 2002); Liu et al. (2004). Similar to this approach, well-calibrated airborne measurements are applied by Heidinger (2003) to assess the calibration of AVHRR. For cross-calibrations, bright cloud tops are remarked as a suitable target (Doelling et al., 2004; Fougnie et al., 2007; Aumann et al., 2007). Subsequently, (Mu et al., 2017) have also concluded that high-bright clouds are convenient for evaluating the long term radiometric stability of MODIS measurements. Doelling et al. (2004) developed a method to calibrate AVHRR radiances by analyzing scenes with DCCs. This technique is independent of solar zenith angle and consistent with inter-calibration of co-incident AVHRR and MODIS measurements. Those confirm the sufficiency of such natural diffuser as an appropriate calibration reference.

For MODIS instruments, Xiong et al. (2003) reported an estimated uncertainty of about 1-5\% for measurements in the reflective solar bands (RSBs). This uncertainty determine mainly the retrieval uncertainties. In addition to the measurement uncertainty, the retrieval uncertainties are influenced by many other factors e.g., the assumptions of surface albedo and ice crystal shape (in case of ice or mixed-phase clouds), as well as by three-dimensional (3-D) radiative effects. All these factors propagate in the retrieval and may amplify the retrieval uncertainties. An inaccurate assumption of surface albedo can lead to uncertainties of up to 83\% for $\tau$ and 62\% for $r_{eff}$ (Rolland and Liou, 2001; Fricke et al., 2014; Ehrlich et al., 2017). Uncertainties of up to 70\% for $\tau$ and 20\% for $r_{eff}$ were obtained by Eichler et al. (2009) when an inappropriate shape is assumed in cirrus retrievals. Additionally, King et al. (2013) noted that three-dimensional (3-D) radiative effects should be considered in the retrieval uncertainties. In order to assess and reduce these uncertainties, collocated measurements i.e., airborne and satellite remote sensing accompanied with in situ observations are necessary. The similar geometry of airborne and satellite radiation
sensors allows for a direct comparison of primary measured quantities. Furthermore, the in situ data can be applied for stringent validations of retrieval algorithms and results.

1.3.2 Problems in remote sensing of cloud properties

Retrievals of cloud properties using measurements of passive remote sensing remain problems. The standard retrieval technique, such as that applied by MODIS, commonly assumes a priori that there is only a single homogeneous cloud layer with a specific thermodynamic phase in the atmosphere, either liquid water or ice (Platnick et al., 2017). However, studies of Hahn et al. (1984) and Warren et al. (1985) using ground-based observations reported that the existence of multilayer clouds (e.g., cirrus above liquid water clouds) is about 50% from all cases. Chang and Li (2005) and Sourdeval et al. (2015) have demonstrated that omitting the low liquid water cloud in the retrieval algorithm introduces significant uncertainties in the retrieved cirrus properties. Another issue arises from the cloud vertical structures, which are not well represented by the existing retrieval technique. Clouds are commonly assumed as a vertically homogeneous column. In reality, cirrus and DCCs have considerably vertical variabilities, as shown by in situ measurements. As a consequence, this may result is significant discrepancies on the resulting cloud radiative impact.

Platnick (2000) and van Diedenhoven et al. (2016) discussed that $r_{\text{eff}}$ retrieved from reflected radiation measurements depends on the vertical penetration of reflected radiation into the cloud. At a wavelength with higher absorption by cloud particles, the probability of photons being scattered back out of the cloud without being absorbed decreases. Thus, using different wavelengths with different absorption by cloud particles in the retrieval algorithm yields different values of $r_{\text{eff}}$ related to different cloud altitudes. Chang and Li (2002) and Chang and Li (2003) have shown that the conventional airborne and satellite retrievals assuming in-cloud vertically homogeneous result in a vertically weighted value $r_{\text{eff}}$ from the entire cloud layer (bulk property). The contribution of each layer to the absorption, thereby to the retrieved $r_{\text{eff}}$, is a function of the cloud profile itself and the applied wavelength. Thus, it should be kept in mind that $r_{\text{eff}}$ retrieved by this technique does not represent a particle size at a single layer only, as well as the real vertical profile of $r_{\text{eff}}$. By in situ measurements, the particle effective radius is sampled at a specific altitude $r_{\text{eff}}(z)$.

The different points of view make difficulties when comparing retrieved and in situ $r_{\text{eff}}$ that may lead to a systematic error. A direct comparison at a certain altitude is problematic because it is not clear to what level the retrieved $r_{\text{eff}}$ corresponds to the value measured by the in situ instrument. King et al. (2013) demonstrated that the comparison between retrieved and in situ values can only be made if the full vertical profile of $r_{\text{eff}}$ is measured by in situ cloud probes. The study by Painemal and Zuidema (2011), which compared $r_{\text{eff}}$ values retrieved from MODIS observations with averaged values of $r_{\text{eff}}$ measured by in situ near the top of liquid water clouds, showed absolute deviations of up to 20%. King et al. (2013) concluded that there is no apparent link between the variation of $r_{\text{eff}}$ retrieved using
different near-infrared wavelengths of MODIS and the vertical structure of $r_{\text{eff}}$ measured by in situ methods. The deviation depends on the wavelength chosen, but it tends to be smaller for high absorbing wavelengths. Painemal and Zuidema (2011) identified four potential reasons for reveal this deviation: the variability of droplet size distributions, the formation of precipitation, water vapor absorption, and viewing-geometry-dependent biases. Zhang et al. (2010) and Nagao et al. (2013) noted that the discrepancy between remote sensing and in situ measurements is also influenced by the simplification of the cloud vertical structures.

![Figure 1.4](#)

**Figure 1.4:** (a) is the solar (solid lines) and the terrestrial (dashed lines) forcing computed by assuming a realistic profile (black) and a vertically homogeneous profile (red) of $r_{\text{eff}}$. (b) is the total radiative forcing $\Delta F_{\text{tot}}$ (solar + terrestrial forcing). The results represent a condition with $\theta_0 = 36^\circ$, standard atmosphere profile of mid-latitude summer, and surface albedo of ocean.

In order to reconstruct the full vertical profile of $r_{\text{eff}}$, the peak of the weighting function should be distributed evenly throughout the profile (Wang et al., 2009; King and Vaughan, 2012). Both studies found that the conventional technique only provides information on particle sizes at the upper layers. Deeper in the cloud, there still exist large portions of the profile where there is seemingly very little information, even if spectral measurements have been applied in the retrieval. To infer the full profile using measurements of passive remote sensing, hence, it is required to assume the shape of the profile. This allows portions of the profile for which the measurement contains sufficient information to be utilized to derive the parameters determining the profile throughout (e.g., Chang and Li, 2002; Wang et al., 2009).
Improving current passive remote sensing techniques to study the vertical profile of $r_{\text{eff}}$ is necessary. Assuming different profiles of $r_{\text{eff}}$ produce discrepancies in the resulting radiative properties, particularly at the wavelengths that are sensitive to absorption by cloud particles. Apparently, such discrepancies will bias the cloud radiative impact. To signify the impact, the radiative forcing is calculated by assuming two different vertical profiles of $r_{\text{eff}}$ of ice clouds: (1) a realistic profile that represent cirrus in reality and (2) a vertically homogenous profile that is commonly applied by conventional techniques. For this analysis, the values of $\tau$ are fixed to 0.1-5. Fig. 1.4a shows the solar (solid lines) and the terrestrial (dashed lines) forcing at the Earth’s surface ($z = 0$) computed using the realistic profile (black) and the vertically homogeneous profile (red), while Fig. 1.4b is the corresponding total radiative forcing $\Delta F_{\text{tot}}$ (solar + terrestrial forcing). Significant discrepancies are obtained in the solar forcing, which range between 0.3 and 4.2 W m$^{-2}$ increasing with $\tau$. For the terrestrial forcing, the discrepancies are lower (0.07-0.6 W m$^{-2}$). These discrepancies clearly show the impact of particle size distribution throughout the cloud vertical extent. $\Delta F_{\text{tot}}$ values resulted by the two profiles differ between 0.2 and 3.6 W m$^2$, which are equivalent to a relative deviation between 1.3 and 2.5 %. Such magnitudes are considerably high and can not be underestimated Shupe and Intrieri (2004).

1.4 Objectives and outline

The objectives of this thesis are adapted from one of the scientific goals of the Mid-Latitude Cirrus (ML-CIRRUS) and the Aerosol, Cloud, Precipitation, and Radiation Interaction and Dynamic of Convective Clouds System - Cloud Processes of the Main Precipitation Systems in Brazil: A Contribution to Cloud Resolving Modelling and to the Global Precipitation Measurement (ACRIDICON-CHUVA) campaigns, which is validating remote sensing measurements (Wendisch et al., 2016; Voigt et al., 2017). The thesis objectives capture four major aspects:

1. Comparison of upward radiance measurements.
2. Retrieval of cloud optical thickness and particle effective radius.
3. Comparison of retrieval results.
4. Retrieval of the vertical profile of particle effective radius.

To pursue these objectives, spectral upward radiances were measured by the Spectral Modular Airborne Radiation Measurement System (SMART)-Albedometer installed aboard of

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*Detailed specifications of the vertical profiles of $r_{\text{eff}}$ for calculating the radiative forcing in Fig. 1.4 are given in Table 5.2. The homogeneous profile is retrieved based on the simulated upward radiance from the realistic profile using $\lambda = 645$ nm and 1640 nm. Thus, both profiles produce same upward radiances at both wavelengths.*
the German High Altitude and Long Range Research Aircraft (HALO) during the ML-
CIRRUS and the ACRIDICON-CHUVA campaigns. HALO with its long endurance of up
to 8 hours and high ceiling of up to 15 km altitude is optimally suited to fly above cirrus
and DCCs. In high altitude, measurements of upward radiance (cloud-reflected) are only
marginally affected by scattering and absorption of atmospheric gases and molecules. Des-
ignated flights above clouds have been carried out during the two campaigns, which were
closely collocated with overpasses of MODIS-Aqua satellite. Additionally, comprehensive
in situ instruments installed on HALO were deployed to sample cloud micro- and macro-
physical properties along the cloud vertical extent. This unique setup allows to evaluate
primary solar radiation measurements, as well as retrieval methodologies and products.

The outline of this thesis is as follows. The fundamentals of radiative and cloud properties
are given in Chapter 2. Chapter 3 describes the airborne campaigns along with the instru-
mentation used in this study. The comparison of upward radiance measured by SMART-
Albedometer and MODIS is given in Chapter 4. Chapter 5 describes the retrieval of $\tau$ and
$\text{reff}$. Here, extensive sensitivity studies based on radiative transfer simulations are also
presented. The comparison of retrieval results is given in Chapter 6. Not only comparing
the retrieval results from SMART-Albedometer and MODIS, in situ data are also utilized
to validate the retrievals. Chapter 7 describes the advanced retrieval technique to infer the
full vertical profile of $\text{reff}$. Further, the retrieved profile is validated with the corresponding
in situ data. This thesis is enclosed with summary and conclusion in Chapter 8.
2 Definitions

This chapter introduces the fundamental radiometric properties, as well as micro- and macrophysical quantities used in this work. The definitions follow the explanations in several textbooks related to the topic, such as Bohren and Clothiaux (2006), Petty (2006), and Wendisch and Yang (2012).

2.1 Radiative quantities

The radiant energy flux $\Phi$ (in units of $\text{W}$) is the energy of electromagnetic (EM) radiation $E_{\text{rad}}$ (in units of $\text{W s}$) emitted or received per unit time $dt$:

$$\Phi = \frac{dE_{\text{rad}}}{dt}. \quad (2.1)$$

Based on $\Phi$, radiance and irradiance, the two major radiative quantities to measure quantitatively the strength of the EM radiation field, are derived.

The radiant energy flux density or irradiance $F$ (in units of $\text{W m}^{-2}$) is a measure of radiant energy flux received on a plane surface with unit area $dA$ and orientation $\vec{n}$:

$$F = \frac{d\Phi}{dA}. \quad (2.2)$$

The orientation of the reference plane can be random but for describing the radiative transfer in the atmosphere, it is considered to be horizontal (see Fig. 2.1). $F$ is weighted with the cosine of the incidence angle $\theta$ (zenith angle) on the horizontal surface with $\theta = 0^\circ$ referring to perpendicular incidence.

While $F$ describes $\Phi$ propagating through a plane unit surface from all possible directions of a hemisphere, the radiance $I$ (in units of $\text{W m}^{-2} \text{sr}^{-1}$) refers to a certain direction $\vec{\Omega}$. $I(\vec{\Omega})$ is defined as radiant energy flux $\Phi$ transported through a plane unit area $dA$ within the solid angle element $d\Omega$ centered around the direction of propagation of the EM radiation $\vec{\Omega}$:

$$I(\vec{\Omega}) = \frac{d^2 \Phi}{\cos \theta \, dA \, d\Omega}. \quad (2.3)$$
The solid angle element $d\Omega$ is expressed in terms of the two directional angles $\theta$ (zenith angle) and $\varphi$ (azimuth angle) and defined as:

$$d\Omega = \sin \theta \, d\theta \, d\varphi,$$  

(2.4)

where $d\Omega$ is in units of steradian $\text{sr}$. On the basis of Eq. 2.2 and 2.3, the integration of $I(\hat{\Omega})$ over $d\Omega$ yields $F$:

$$F = \int\int I(\hat{\Omega}) \cdot \cos \theta \, d\Omega = \int_0^{2\pi} \int_0^\pi I_\lambda(\theta, \varphi) \cdot \cos \theta \cdot \sin \theta \, d\theta \, d\varphi.$$  

(2.5)

![Figure 2.1: Illustration of the geometry to define radiance and irradiance.](image)

Separate consideration of the lower $[\theta = (\pi/2 \cdots \pi) \, \text{rad}, \varphi = (0 \cdots 2\pi) \, \text{rad}]$ and upper hemisphere $[\theta = (0 \cdots \pi/2) \, \text{rad}, \varphi = (0 \cdots 2\pi) \, \text{rad}]$ which are divided by the horizontally oriented surface, leads to upward ($\uparrow$) and downward ($\downarrow$) irradiance that are defined as:

$$F^\downarrow_\lambda = \int_0^{2\pi} \int_0^{\pi/2} I_\lambda(\theta, \varphi) \cdot \cos \theta \cdot \sin \theta \, d\theta \, d\varphi,$$  

(2.6)

$$F^\uparrow_\lambda = -\int_0^{2\pi} \int_0^{\pi/2} I_\lambda(\theta, \varphi) \cdot \cos \theta \cdot \sin \theta \, d\theta \, d\varphi.$$  

(2.7)
Eq. 2.6 shows how radiance determines irradiance. In the case of an isotropic field of radiation radiance is independent of direction \( I(\vec{\Omega}) = I_{\text{iso}} \) and it follows:

\[
F_{\text{iso}}^\downarrow = F_{\text{iso}}^\uparrow = I_{\text{iso}} \cdot \pi \text{ sr}. \tag{2.8}
\]

The downward irradiance \( F_{\text{iso}}^\downarrow \) propagating through the atmosphere consists of a direct \( F_{\text{dir}}^\downarrow \) and an indirect component \( F_{\text{diff}}^\downarrow \) so that:

\[
F_{\downarrow} = F_{\text{dir}}^\downarrow + F_{\text{diff}}^\downarrow. \tag{2.9}
\]

\( F_{\text{dir}}^\downarrow \) refers to the solar radiation which has not encountered scattering processes in the atmosphere yet. \( F_{\text{diff}}^\downarrow \) describes the part of the radiation which has been scattered, absorbed, and emitted by atmospheric molecules and particles or is reflected from the surface back into the atmosphere. The upward irradiance is diffuse only (\( F_{\uparrow} = F_{\text{diff}}^\uparrow \)).

**Figure 2.2:** Downward solar irradiance \( F_{\downarrow} \) and absorption bands in the visible and near-infrared spectral range. The calculations are performed at the Earth’s surface \((z = 0)\) and the top of atmosphere \(\text{(TOA)}\).

The visible (VIS) and near-infrared (NIR) spectral ranges are relevant spectral ranges that are used for numerous applications in this study. Fig. 2.2 shows the spectral downward irradiance at \( \lambda = 200\text{-}2200 \) nm computed at the Earth’s surface \((z = 0)\) and the top of atmosphere \(\text{(TOA)}\). The absorption features are given by the labels. In the visible range, there are a few absorption bands of water vapor \((\text{H}_2\text{O})\) and oxygen \((\text{O}_2)\), but they become more pronounced in the near-infrared range. Between 1350 and 1400 nm, as well as between 1800 and 1900 nm, absorption by water vapor are obvious. The fluctuations in the ultraviolet and visible spectral ranges are caused by absorption from the gases in the Sun’s photosphere (Fraunhofer lines).
Irradiance is the relevant quantity for calculating the cloud radiative forcing $\Delta F$. Shupe and Intrieri (2004) defined $\Delta F$ at altitude $z$ as the radiative impact that clouds have relative to cloud-free cases (atmosphere without clouds). The difference between downward and upward irradiances ($F^\downarrow - F^\uparrow$) is called as the net irradiance. By this definition, $\Delta F$ can be computed by subtracting the net irradiance in the presence of cloud (index "c") with the the net irradiance without cloud (index "0") as follows:

$$\Delta F(z) = \left[ F^\downarrow(z) - F^\uparrow(z) \right]_c - \left[ F^\downarrow(z) - F^\uparrow(z) \right]_0.$$ (2.10)

$\Delta F$ can be positive or negative. A negative forcing indicates that clouds reduce the amount of radiation (cooling), while a positive forcing indicates warming. $\Delta F$ is often considered separately, the solar forcing $\Delta F_{\text{sol}}(\lambda = 300-4000 \text{ nm})$ and terrestrial forcing $\Delta F_{\text{ter}}(\lambda = 4000-90000 \text{ nm})$. The sum of both is called as the total radiative forcing $\Delta F_{\text{tot}}$ as that shown in Fig. 1.3.

Relating radiative quantities to an infinitesimal wavelength interval $[\lambda, \lambda + d\lambda]$, results in spectral quantities such as spectral radiances $I_{\lambda}(\Omega)$ (in units of $W \text{ m}^{-2} \text{ sr}^{-1} \text{ nm}^{-1}$) and spectral irradiances $F_{\lambda}$ (in units of $W \text{ m}^{-2} \text{ nm}^{-1}$). By integrating the spectral quantities over a wavelength interval, broadband quantities are obtained. The following definitions are valid for both broadband or spectral quantities.

Another important radiative quantities are the reflectivity $R$ and the albedo $\rho$. At a certain altitude, $R(z)$ and $\rho(z)$ are defined as:

$$R_{\lambda}(z) = \frac{I^\downarrow_{\lambda}(z)}{F^\downarrow_{\lambda}(z)} \cdot \pi \text{ sr},$$ (2.11)

$$\rho_{\lambda}(z) = \frac{F^\downarrow_{\lambda}(z)}{F^\downarrow_{\lambda}(z)},$$ (2.12)

where $I^\downarrow(z)$ denotes the upward radiance in the nadir direction ($\theta = \pi \text{ rad}$). In an isotropic condition (Eq. 2.8), $R(z)$ and $\rho(z)$ are identical. While values of $R$ can be larger than 1, values of $\rho$ range between 0 and 1 to ensure the conservation of radiant energy. Furthermore, cloud top reflectivity $R_c$ will be of importance. They are defined for $z \geq z_{\text{top}}$, with $z_{\text{top}}$ being the cloud top. The surface albedo $\rho_{\text{surf}} = \rho(z = 0)$ describes how much of the incident irradiance is reflected by the surface with the extremes of 0 and 1 referring to no reflection and total reflection, respectively. At $z = 0$, the amount that is not reflected $(1 - \rho_{\text{surf}})$ is absorbed by the surface. The surface albedo strongly depends on the surface type (e.g. water, forest, snow, etc.) and the wavelength.
2.1.1 Single scattering properties of individual particles

The interaction between atmospheric radiation and individual particles within the atmosphere (gas molecules, cloud, and aerosol particles) is described by the extinction cross section $C_{\text{ext}}$, the single scattering albedo $\tilde{\omega}$, and the scattering phase function $\mathcal{P}$, which are defined by the mass/cross sectional area, the spectral complex index of refraction $\tilde{n}$, the particle shape, size, and orientation.

![Figure 2.3:](image)

**Figure 2.3:** (a) is the single scattering albedo $\tilde{\omega}$ of different ice crystal shapes by Baum et al. (2014) and Yang et al. (2013) and the liquid water droplet Wiscombe (1980). The calculations are made for $r_{\text{eff}} = 25 \mu m$. Imaginary part of refractive index $n_i$ of ice and liquid water droplet is shown in (b).

For spherical liquid water droplets, Mie theory yields an analytical solution for these quantities (Mie, 1908; Bohren and Huffman, 1998). The single scattering properties of non-spherical particles, such as ice crystals and aerosol particles, cannot be described by an analytical solution and numerical methods have to be applied. Spectral single scattering properties of different ice crystal shapes and sizes have been published by e.g., Yang et al. (2013), Baum et al. (2014), and others.

The extinction cross section $C_{\text{ext}}$ (in units of $m^2$) defines how effective an individual particle attenuates atmospheric radiation. It is defined by the radiant energy flux subject to extinction ($\Phi_{\text{ext}}$) relative to the incident $F_{\text{inc}}$:

$$C_{\text{ext}} = \frac{\Phi_{\text{ext}}}{F_{\text{inc}}}.$$  \hspace{1cm} (2.13)
It can also be derived by adding up the scattering cross section $C_{\text{sca}}$ and the absorption cross section $C_{\text{abs}}$:

$$C_{\text{ext}} = C_{\text{sca}} + C_{\text{abs}}.$$ (2.14)

The dimensionless particle single scattering albedo $\tilde{\omega}$ characterizes the probability of atmospheric radiation being absorbed or scattered by a particle. It is derived from the optical cross sections by:

$$\tilde{\omega} = \frac{C_{\text{sca}}}{C_{\text{sca}} + C_{\text{abs}}}.$$ (2.15)

$\tilde{\omega}$ ranges between 0 and 1, where values close to 1 indicate weak absorption by the particle. Both $C_{\text{abs}}$ and $\tilde{\omega}$ define the probability of absorption and are closely related to the imaginary part of the refractive index $\tilde{n}_i$. Fig. 2.3 shows $\tilde{\omega}$ (a) and $\tilde{n}_i$ (b) as a function of wavelength $\lambda$ for different ice crystal habits: general habit mixtures (GHM) by Baum et al. (2014), aggregated columns and plates by Yang et al. (2013), as well as the liquid water droplet by Wiscombe (1980) for $r_{\text{eff}} = 25 \ \mu$m. All the ice crystal habits represent particles with high surface roughness corresponding to those observed and applied for the applications in this study. With decreasing $\tilde{\omega}$, the absorption by cloud particles increases. The spectral absorption between ice crystals and liquid water droplet yields differences. For the ice crystals, the absorption peaks at wavelengths around 1500 nm and 2000 nm, while for the liquid water droplet, the maximum is found at wavelengths around 1450 nm and 1900 nm. It is obvious that both quantities ($\tilde{\omega}$ and $\tilde{n}_i$) are closely related because the probability of absorption by cloud particles increases (thereby decreasing $\tilde{\omega}$) with increasing $\tilde{n}_i$.

For the scattering processes, the dimensionless scattering phase function $\mathcal{P}$ defines the angular probability distribution of scattered atmospheric radiation from the incident direction $(\mu_{\text{inc}}, \varphi_{\text{inc}})$ to all directions $(\mu, \varphi)$, with $\mu = \cos \theta$. $\mathcal{P}$ is normalized to $4\pi$:

$$\int_0^{2\pi} \int_{-1}^1 \mathcal{P}([\mu_{\text{inc}}, \varphi_{\text{inc}}] \rightarrow [\mu, \varphi]) \, d\mu \, d\varphi = 4\pi \text{ sr}.$$ (2.16)

For particles that are symmetric in the azimuth direction (such as liquid cloud droplets) or for azimuthal averaging (in case of complex ice crystal forms) the relation between the incident and scattered direction in the scattering plane is given by the scattering angle $\vartheta$:

$$\cos \vartheta = \mu_{\text{sca}} \cdot \mu_{\text{inc}} + \sqrt{1 - \mu_{\text{sca}}^2} \cdot \sqrt{1 - \mu_{\text{inc}}^2}.$$ (2.17)

Fig. 2.4 illustrates the scattering phase function $\mathcal{P}$ of different ice crystal habits and the liquid water droplet corresponding to the data shown in Fig. 2.3. Differences arise mainly from different shapes. In general, $\mathcal{P}$ of the ice crystals are smoother than $\mathcal{P}$ of the liquid
water droplet, which comes from the azimuthal averaging in case of the ice crystals. The phase function of the ice crystals exhibits a strong forward peak, while the forward peak is truncated for the liquid water droplet.

The asymmetry factor \(g\) of the phase function of an individual particle is a coarse measure of the angular distribution of the radiation scattered by the particle. It is defined using the phase function by:

\[
g = \int_{-1}^{1} \int_{0}^{2\pi} p(\vartheta) \cdot \cos \vartheta \, d^2 \Omega = \frac{1}{2} \int_{-1}^{1} \cos \vartheta \cdot P(\cos \vartheta) \, d \cos \vartheta,
\]

where the phase function \(p\) represents the relative angular distribution of the scattered radiation and the assumption that \(P(\cos \vartheta)\) is independent of the scattering azimuthal angle. A value of \(g = 1\) means pure forward scattering \((\vartheta = 0, \cos \vartheta = 1)\), while \(g = -1\) means specular reflection in the backward direction \((\vartheta = 180, \cos \vartheta = -1)\). A value of \(g = 0\) denotes that the scattered radiation in the forward direction is the same as that in the backward direction, such in case of Rayleigh scattering.

\subsection*{2.1.2 Volumetric optical properties}

The volumetric (bulk) cloud optical properties are derived from the single scattering properties and the number size distribution \(\frac{dN}{dD}(D)\) of particles within the respective size bin.
\( D + dD. \) According to Wendisch et al. (2005), the volumetric cloud properties are calculated by integrating the single scattering properties, weighted by \( \frac{dN}{dD}(D) \). The spectral volumetric extinction coefficient \( b_{\text{ext}} \) is defined as:

\[
b_{\text{ext}} = \int C_{\text{ext}} \cdot \frac{dN}{dD}(D) \, dD,
\]

in units of \( \text{m}^{-1} \). The cloud optical thickness \( \tau \) is derived by integration of \( b_{\text{ext}} \) from the cloud base altitude \( z_{\text{base}} \) to the cloud top altitude \( z_{\text{top}} \):

\[
\tau = \int_{z_{\text{base}}}^{z_{\text{top}}} b_{\text{ext}}(z) \, dz.
\]

The volumetric single scattering albedo \( \langle \tilde{\omega} \rangle \) is calculated by:

\[
\langle \tilde{\omega} \rangle = \frac{1}{b_{\text{sca}}} \int \tilde{\omega}(D) \cdot C_{\text{ext}} \cdot \frac{dN}{dD}(D) \, dD.
\]

In the same way, the volumetric phase function \( \langle P \rangle \) is derived from:

\[
\langle P(\vartheta) \rangle = \frac{1}{b_{\text{sca}}} \int P(\vartheta, D) \cdot C_{\text{ext}} \cdot \frac{dN}{dD}(D) \, dD.
\]

As for the single scattering events, the numerical solution of the radiative transfer equation requires the Legendre expansion of \( \langle P \rangle \). Here the expanded volumetric phase function is calculated via:

\[
\langle P(\cos \theta) \rangle = \sum_{i=0}^{\infty} \langle b_i \rangle \cdot P_i(\cos \theta),
\]

where the volumetric Legendre moments \( \langle b_i \rangle \) are given by:

\[
\langle b_i \rangle = \frac{1}{b_{\text{sca}}} \int b_i(D) \cdot C_{\text{ext}} \cdot \frac{dN}{dD}(D) \, dD.
\]

### 2.1.3 Cloud microphysical properties

According to Hansen and Travis (1974), the effective droplet radius \( r_{\text{ef}} \) in units of \( \mu \text{m} \) is defined as the ratio of the third to the second moment of the droplet number size distribution \( \frac{dN}{dD}(D) \) (with droplet diameter \( D \)) and characterizes the mean radius weighted by \( \frac{dN}{dD}(D) \) within an ensemble of cloud droplets that can be expressed as:
2.1. Radiative quantities

\[ r_{\text{eff}} = \frac{1}{2} \int D^3 \frac{dN(D)}{dD} dD \]

Equation (2.25)

It represents the ratio of the volume of a cloud particle to its surface area. This definition becomes important when discussing \( r_{\text{eff}} \) or ice crystals since different authors may have their own definition (McFarquhar and Heymsfield, 1998). Following Yang et al. (2000), \( r_{\text{eff}} \) is defined by the maximum dimension of an ice crystal \( D_{\text{max}} \), its volume \( V_D \), and its projected area \( A_D \), as shown in Eq. 2.26. \( V_D \) and \( A_D \) are derived by calculating the diameter a sphere with the same volume and surface area as:

\[ r_{\text{eff}} = \frac{3}{2} \int \frac{V_D(D)}{D^2} \frac{dN(D)}{dD} dD \]

Equation (2.26)

The liquid water content \( LWC \) in units of \( \text{g m}^{-3} \) is defined as the mass concentration of liquid water droplets in the cloud volume:

\[ LWC = \varrho_w \cdot \frac{4}{3} \cdot \pi \cdot \int \left( \frac{D}{2} \right)^3 \frac{dN}{dD}(D) dD \cdot \frac{1}{1 \text{ m}^3} \]

Equation (2.27)

where \( \varrho_w \) is the density of liquid water (\( \approx 1 \text{ g cm}^{-3} \)). For a vertically homogeneous cloud, \( LWC \) can be parameterized using \( \tau \) and \( r_{\text{eff}} \) from cloud base \( z_{\text{base}} \) to cloud top \( z_{\text{top}} \) as:

\[ LWC = \frac{2}{3} \cdot \varrho_w \cdot \int_{z_{\text{base}}}^{z_{\text{top}}} r_{\text{eff}} \cdot \tau \frac{dz}{z} \]

Equation (2.28)

Replacing \( \frac{2}{3} \) with \( \frac{5}{9} \) yields the relation for adiabatic clouds (Wood and Hartmann, 2006). The liquid water path \( LWP \) in units of \( \text{g m}^{-2} \) is the vertical integration of \( LWC \) from \( z_{\text{base}} \) to \( z_{\text{top}} \) expressed by:

\[ LWP = \int_{z_{\text{base}}}^{z_{\text{top}}} LWC(z) dz \]

Equation (2.29)

The ice water content \( IWC \) and ice water path \( IWP \) of ice clouds are quantified by:

\[ IWC = \varrho_{\text{ice}} \cdot \int V_D(D) \cdot \frac{dN(D)}{dD} dD, \]  

Equation (2.30)

\[ IWP = \int_{z_{\text{base}}}^{z_{\text{top}}} IWC(z) dz, \]

Equation (2.31)

where \( \varrho_{\text{ice}} \) is the density of ice (\( \approx 0.9168 \text{ g cm}^{-3} \)). Integration limits are not indicated, where theoretically they extend from 0 to \( \infty \). Practically, they are set to a finite upper
and nonzero lower limit chosen in a way that guarantees integration over the entire size spectrum.

2.2 Radiative transfer equation

The radiative transfer equation (RTE) is determined by the quantities $b_{\text{ext}}$, $\tilde{\omega}$ and $\mathcal{P}$. The attenuation of direct solar radiance $I_{\text{dir}}$ along a path through the atmosphere, with $\tau$ as a vertical coordinate and $\mu_0 = \cos \theta_0$, is characterized by the law of Beer, Lambert, and Bouguer as following:

$$ I_{\text{dir}}(\tau, \mu_0, \varphi_0) = \frac{S_0}{4\pi \text{sr}} \cdot \exp \left( -\frac{\tau}{\mu_0} \right). \quad (2.32) $$

$S_0$ is the solar constant (the extraterrestrial irradiance arrived at the TOA), while $\mu_0 = \cos \theta_0$ and $\varphi_0$ define the position of the sun. Eq. 2.32 shows that $I_{\text{dir}}$ decreases exponentially with $\tau$. Thus, $I_{\text{dir}}$ is strongly attenuated in the presence of clouds, allowing to describe the solar radiative transfer in clouds by the diffuse radiance $I_{\text{diff}}$ only. According to Chandrasekhar (1950), the one-dimensional (1D) radiative transfer equation (RTE) assuming a plane-parallel, horizontally homogeneous atmosphere is expressed by:

$$ \mu \frac{dI_{\text{diff}}(\tau, \mu, \varphi)}{d\tau} = I_{\text{diff}} - (J_{\text{dir}} + J_{\text{diff}}). \quad (2.33) $$

Here, $\mu$ and $\varphi$ define the direction of propagation of $I_{\text{diff}}$. $J_{\text{dir}} + J_{\text{diff}}$ are two source terms that characterize the radiation scattered into the viewing direction. $J_{\text{dir}}$ is the single scattering term, specifying the amount of direct solar radiation which is scattered into the viewing direction:

$$ J_{\text{dir}} = \frac{\tilde{\omega}(\tau)}{4\pi \text{sr}} \cdot S_0 \cdot \exp \left( -\frac{\tau}{\mu_0} \right) \cdot \mathcal{P}(\tau, [-\mu_0, \varphi_0] \rightarrow [\mu, \varphi]). \quad (2.34) $$

Incoming solar radiation is attenuated according to the the law of Beer, Lambert, and Bouguer. The attenuated radiation is scattered into the viewing direction, depending on the amount of absorption ($\tilde{\omega}$) and the scattering phase function $\mathcal{P}$.

$J_{\text{diff}}$ is the multiple scattering term, describing the amount if diffuse radiation which is scattered into the viewing direction:

$$ J_{\text{diff}} = \frac{\tilde{\omega}(\tau)}{4\pi \text{sr}} \int_0^{2\pi} \int_{-1}^{1} I_{\text{diff}}(\tau, \mu_{\text{inc}}, \varphi_{\text{inc}}) \cdot \mathcal{P}(\tau, [\mu_{\text{inc}}, \varphi_{\text{inc}}] \rightarrow [\mu, \varphi]) \, d\mu_{\text{inc}} \, d\varphi_{\text{inc}}, \quad (2.35) $$
where $\tilde{\omega}$, $\mathcal{P}$ and $\tau$ are the defining quantities to describe $J_{\text{diff}}$. 
3 Measurements

This Chapter gives an overview of the two HALO campaigns, ML-CIRRUS and ACRIDICON-CHUVA, as well as remote sensing and in situ instruments used for this work. Sec. 3.1 describes the two HALO campaigns. In Sec. 3.2, the characteristics of remote sensing instruments will be elaborated in detail. Subsequently, the in situ instruments are briefly introduced in Sec. 3.3.

3.1 Airborne campaigns

HALO and a suite of ground-based instruments were deployed during the ML-CIRRUS and the ACRIDICON–CHUVA campaigns. HALO (see http://www.halo.dlr.de/) is an ultra-long-range business jet G550 (manufactured by Gulfstream) that is similar to the U.S. High-Performance Instrumented Airborne Platform for Environmental Research (HIAPER) (Laursen et al., 2006). With its high ceiling altitude (up to 15 km) and long endurance (up to 8 hours), HALO is capable of collecting airborne in situ and remote sensing measurements of cloud microphysical and radiative properties, aerosol characteristics, and chemical tracer compounds in and around mid-latitude cirrus and tropical DCCs, which are needed to study the scientific objectives during the two campaigns. Protracted problems of high cirrus and deep convective clouds measurements from previous campaigns have been remarked, such as limited ceiling to reach the top of cirrus and deep convective clouds, insufficient endurance to study the cloud life cycle, and icing of aircraft when penetrating the clouds. HALO provides opportunities to cope these difficulties although aircraft icing still remains a problem in extreme DCCs.

Between 21 March 2014 and 15 April 2014, the ML-CIRRUS campaign performed 16 research flights over Europe and the Atlantic ocean to study nucleation, life-cycle, and climate impact of natural cirrus and aircraft induced contrail cirrus (Voigt et al., 2017). The campaign was conducted over Europe out of Oberpfaffenhofen (48°5’N, 11°17’E) in southern Germany in March and April 2014. The location is well suited to reach cirrus over all parts of Europe and to access regions with high air traffic abundance, with more than 30,000 commercial flights over Europe per day. A mission in midlatitudes provides the unique opportunity to measure cirrus clouds linked to a large variety of dynamical weather regimes, such as frontal systems, air pressure ridges, high pressure systems, jet streams, mountain
waves, and convection at HALO cruise altitudes. In particular, a case study of cirrus formation in the outflow of a warm conveyor belt (WCB) by Spichtinger et al. (2005) motivated flight planning for ML-CIRRUS. The spring season was chosen to combine a high abundance of both WCBs (Madonna et al., 2014) and contrail cirrus.

![Image of flight paths](image-url)

**Figure 3.1:** HALO flight paths during the ML-CIRRUS (a) and the ACRIDICON-CHUVA (b) campaigns. Only scientific flights are shown.

Between 1 September 2014 and 4 October 2014, the ACRIDICON-CHUVA campaign performed 14 research flights combined with satellite and ground-based observations over the Brazilian Amazon rainforest to quantify aerosol-cloud-precipitation interactions and
their thermodynamic, dynamic, and radiative effects of tropical DCCs (Wendisch et al., 2016). The campaign was based in Manaus, a city of two million people. Manaus is an isolated urban area in the central Amazon basin situated at the confluence of the two major tributaries of the Amazon River. Outside this industrial city, there is mostly natural forest for over 1000-2000 km in every direction. This makes it possible to study the impact of local pollution on cloud evolution by taking measurements upwind and downwind of the city. ACRIDICON–CHUVA was intentionally planned to take place at a time of year (September–October) when the nonlinear interactions between modified cloud microphysical properties (by higher concentrations of CCN) and thermodynamic conditions (by land cover contrasts) were amplified. It is during the transition between dry to wet season (September–October) that the gradual large-scale advection of humidity in the troposphere increases the conditional thermodynamical instability, while biomass burning peaks just before first rainfalls. This allows the separation of the individual effects on deep convection.

3.2 Remote sensing measurements

3.2.1 Spectral Modular Airborne Radiation Measurement System

The SMART-Albedometer instrument was first developed by (Wendisch et al., 2001). The instrument was designed for airborne measurements of spectral radiometric quantities in the solar spectral range. This is realized by connecting specific optical inlets, which are pointing in upward or downward direction, to a set of up to six grating spectrometers that are connected to a computer control system. Depending on the scientific objective, the optical inlets, which determine the measured radiometric quantities, can be chosen. Originally, SMART-Albedometer was used to measure upward and downward irradiances in order to determine the surface albedo, hence it is called as an Albedometer (Wendisch et al., 2004). For that application, an active stabilization system keeping the optical inlets in horizontal position during aircraft movements of up to \( \pm 6 \) from the horizontal plane was developed to obtain credible irradiance measurements. Furthermore, Ehrlich et al. (2008) developed the optical inlets for radiance measurements. The original spectral range covered by the SMART-Albedometer was 290–1000 nm. It was extended by Bierwirth (2008) to 2200 nm, so it covers approximately 97% of the entire solar spectrum. However, due to the decreasing sensitivity of the spectrometers at small and large wavelengths obtained during the two campaigns, the reasonable wavelength range is restricted to 400-1800 nm for applications in this study. Different types of spectrometers (manufactured by Carl Zeiss Jena GmbH, Jena, Germany) are used in the SMART-Albedometer. The Multi Channel Spectrometer (MSC 55 UV/NIR) with a 1024 pixel photodiode array (PDA) and a spectral resolution (Full Width at Half Maximum, FWHM) of 2–3 nm covers the 200–1000 nm wavelength range including the entire visible (VIS) wavelength range (380–700 nm) and therefore, it is named VIS. The
PGS 2.2 (Plane Grating Spectrometer 2.2) operating in the near-infrared wavelength range (900–2200 nm, from now on called NIR) has a 256 pixel PDA and a FWHM of 8-10 nm. The full characteristics of the two types of spectrometers are listed in Table 3.1.

During the ML-CIRRUS and the ACRIDICON-CHUVA campaigns, the SMART-Albedometer has been deployed to measure the spectral upward radiance $I_{s,\lambda}^{↑}$, as well as the spectral upward and downward irradiances, $F_{s,\lambda}^{↑}$ and $F_{s,\lambda}^{↓}$, respectively. Here, the index "s" refers to measurements by SMART-Albedometer, while $\lambda$ indicates spectral quantities in units of nm$^{-1}$. In most application of this study, the radiance measurements are used for the comparison with MODIS observations and cloud retrievals, while the irradiance measurements are applied to determine the spectral surface albedo. The nadir radiance measured by SMART-Albedometer is comparable to measurements of MODIS reflective solar bands (RSBs) in the band numbers 1-19, and 26 ranging between 410-2130 µm (Xiong and Barnes, 2006). A sketch of the setup of the SMART-Albedometer applied during the ML-CIRRUS and the ACRIDICON-CHUVA airborne observations is shown in Fig. 3.2. The irradiance inlets were placed at the top (upward looking) and the bottom (downward looking) of HALO to measure the downward (3) and the upward radiances (2), respectively. The radiance inlet (downward looking) was placed at the bottom-back of the aircraft to measure the upward radiance (1).

Table 3.1: Characteristics of the two different spectrometers of the SMART-Albedometer used during the two HALO campaigns.

<table>
<thead>
<tr>
<th>Name</th>
<th>Type</th>
<th>Spectral Range (nm)</th>
<th>Spectral Resolution (nm)</th>
<th>Number of pixels</th>
</tr>
</thead>
<tbody>
<tr>
<td>VIS</td>
<td>MCS 55 UV/NIR</td>
<td>200-1000</td>
<td>2-3</td>
<td>1024</td>
</tr>
<tr>
<td>NIR</td>
<td>PGS 2.2</td>
<td>900-2200</td>
<td>8-10</td>
<td>256</td>
</tr>
</tbody>
</table>

According to Ehrlich (2009), the radiance optical inlet has a field of view (FoV) $\Theta$ of 2.1°, which was hold for the radiance measurements during the two campaigns. This opening angle depends on the collimator lens and the optical fiber connected to the radiance inlet. $\Theta$ significantly controls the instantaneous footprint diameter $d$ of the radiance inlet (see Fig. 3.3a). With increasing distance between the surface and the radiance inlet installed aboard the aircraft, $d$ increases following Eq. 3.3. For sensor altitudes of about 100 m/1000 m/10,000 m/14,000 m above surface, this will result in $d$ of 3.7 m/36.7 m/367 m/513 m, respectively. During the ML-CIRRUS and the ACRIDICON-CHUVA campaigns, typically HALO flew several hundred meters to 1 km above clouds, so that $d$ was on the order of 20-100 m (see Fig. 3.3b). The spatial resolution of the measurements also depends on aircraft true air speed (TAS) and the integration time $T_{\text{int}}$. During the two campaign, the temporal resolution was set to 0.2-0.4 s depending on the prevailing conditions. Assuming a TAS of 200 m s$^{-1}$, an averaging of 20-80 m along the flight track is obtained.
3.2. Remote sensing measurements

The instrument is calibrated radiometrically before, during, and after each campaign using certified calibration standards traceable to the National Institute of Standards and Technology (NIST) and by the secondary calibration using a travelling standard. The calibration procedures have been discussed in detail by, e.g., Eichler et al. (2009). Several parameters add to the total uncertainty of the upward radiance measurements. The

\[ d = 2z \cdot \tan \left( \frac{\Theta}{2} \right) \]  

(3.1)
individual components are assumed to be randomly distributed, and therefore independent. The composite of relative uncertainty (in units of %) is determined by Gaussian error propagation. According to Werner et al. (2014), the relevant measurement uncertainties consist of the uncertainty of absolute calibration, spectrometer signal, and transfer calibration.

Figure 3.3: Illustration of (a) the instantaneous footprint diameter \( d \) of the radiance optical inlet and (b) of \( d \) as function of altitude \( z \).

Table 3.2: Uncertainties of upward radiance measurements of the SMART-Albedometer in different spectral ranges. Note that only the spectral range between 400 and 1800 nm is used in this study.

<table>
<thead>
<tr>
<th>Wavelength (nm)</th>
<th>Absolute calibration (%)</th>
<th>Spectrometer signal (%)</th>
<th>Transfer calibration (%)</th>
<th>Total (%)</th>
</tr>
</thead>
<tbody>
<tr>
<td>350-400</td>
<td>5.1</td>
<td>5.7</td>
<td>12.9</td>
<td>14.99</td>
</tr>
<tr>
<td>400-800</td>
<td>4</td>
<td>0.5</td>
<td>0.5</td>
<td>4.1</td>
</tr>
<tr>
<td>810-1550</td>
<td>8</td>
<td>1.8</td>
<td>0.7</td>
<td>8.3</td>
</tr>
<tr>
<td>1550-1800</td>
<td>9</td>
<td>2.2</td>
<td>1.0</td>
<td>9.4</td>
</tr>
<tr>
<td>1800-2200</td>
<td>10</td>
<td>18.7</td>
<td>1.1</td>
<td>21.3</td>
</tr>
</tbody>
</table>

The individual uncertainties are summarized in Table 3.2. The absolute calibration accuracy for \( \lambda \leq 400 \text{ nm} \) consists of the uncertainty of the 1000 W calibration lamp and the reflectivity panel used for the calibration. Both are certified by the manufacturer. They are on the order of 5 % and less than 1 %, respectively. For wavelengths \( \lambda > 400 \text{ nm} \), the uncertainty of the integrating sphere was used, which is in the range of 5-10 %. The uncertainty of the spectrometer signal includes the signal-to-noise ratio (SNR) of the spectrometers and the uncertainty of the wavelength calibration. The SNR was determined
3.2. Remote sensing measurements

with the standard deviation of the dark signal in the visible to near-infrared (VNIR) and the shortwave-infrared (SWIR) wavelength ranges. It is in the range of 2 digital counts in the VNIR with no spectral dependence and 10-20 digital counts in the SWIR. The spectrometers have a range of 0-32, 700 digital counts. A usual signal in the middle of the range has a maximum of 15,000 digital counts in the VNIR and 5000 digital counts in the SWIR. The uncertainty in the spectrometer signal, therefore, amounts to 1-2.2%. At $\lambda \leq 400 \text{ nm}$ and $\lambda \geq 1800 \text{ nm}$, the sensitivity of the spectrometers reduces, and therefore, the uncertainty considerably increases. The uncertainty in the wavelength calibration is determined by the FWHM and it is $< 1\%$. The uncertainty in the transfer calibration is obtained from the standard deviation of the transfer calibrations. Except for $\lambda \leq 400 \text{ nm}$, where the uncertainty of the integrating sphere is higher, they are in the range of about $1\%$. Applying Gaussian error propagation gives the total measurement uncertainty which ranges from $4\%$ in the VNIR (400-800 nm) to $9.4\%$ in the SWIR (800-1800 nm).

3.2.2 Moderate Resolution Imaging Spectroradiometer

The satellite data used in this study stem from the Level 1B Moderate Resolution Imaging Spectroradiometer (MODIS) - Aqua collection 6, namely MYD021KM. Detailed instrument specifications and features of MODIS have been described by, e.g., Platnick et al. (2003), Xiong and Barnes (2006). The data contain calibrated and geolocated radiances and reflectances/reflectivities for 36 spectral bands distributed between 0.41 and 14.2 $\mu$m, including 20 reflective solar bands (RSBs) and 16 thermal emissive bands (TEBs) with a nadir horizontal resolution of about 1 km. The radiances are generated from MODIS Level 1A scans of raw radiance. The solar reflectance values are based on a solar diffuse panel for reflectance calibration up through the RSBs. A diffuser stability monitor is available for assessing the stability of the diffuser of up to 1 $\mu$m (Platnick et al., 2003). The spectral response is determined by an interference filter overlying a detector array imaging a 10 km along track scene for each scan (40, 20, and 10 elements arrays for the 250 m, 500 m, and 1 km bands, respectively). Onboard instruments used for in-orbit radiometric calibration were discussed by Xiong et al. (2003) and Sun et al. (2007).

To process the radiance data, the geolocation files are needed. The MODIS geolocation product, namely MYD03, contains geodetic coordinates, ground elevation, and solar and satellite zenith, and azimuth angles for each MODIS 1 km sample. These data are provided as a companion data set to the Level 1B calibrated radiances along with the Level 2 data sets to enable data processing, such as the spatial aggregation and the interpolation of the radiance values. These geolocation fields are determined using the spacecraft attitude and orbit, instrument telemetry, and a digital elevation model. All the MODIS data used in this study are downloaded from the official webpage https://ladsweb.modaps.eosdis.nasa.gov under the data sharing policies. In addition to the radiances and the geolocation files, the calibration data including the relative spectral re-
3. Measurements

Table 3.3: MODIS bands used in this study along with the central wavelength and the bandwidth. Spatial resolutions are 250 m for bands 1-2, 500 m for bands 3-7, and 1000 m for bands 8-19 and 26.

<table>
<thead>
<tr>
<th>Band</th>
<th>Central (nm)</th>
<th>Bandwidth (nm)</th>
<th>Band</th>
<th>Central (nm)</th>
<th>Bandwidth (nm)</th>
</tr>
</thead>
<tbody>
<tr>
<td>1</td>
<td>645</td>
<td>620 - 670</td>
<td>11</td>
<td>531</td>
<td>526 - 536</td>
</tr>
<tr>
<td>2</td>
<td>858</td>
<td>841 - 876</td>
<td>12</td>
<td>551</td>
<td>546 - 556</td>
</tr>
<tr>
<td>3</td>
<td>469</td>
<td>459 - 479</td>
<td>13</td>
<td>667</td>
<td>662 - 672</td>
</tr>
<tr>
<td>4</td>
<td>555</td>
<td>545 - 565</td>
<td>14</td>
<td>678</td>
<td>673 - 682</td>
</tr>
<tr>
<td>5</td>
<td>1240</td>
<td>1230 - 1250</td>
<td>15</td>
<td>748</td>
<td>743 - 753</td>
</tr>
<tr>
<td>6</td>
<td>1640</td>
<td>1628 - 1652</td>
<td>16</td>
<td>869</td>
<td>862 - 877</td>
</tr>
<tr>
<td>7</td>
<td>2130</td>
<td>2105 - 2155</td>
<td>17</td>
<td>905</td>
<td>890 - 920</td>
</tr>
<tr>
<td>8</td>
<td>421</td>
<td>405 - 420</td>
<td>18</td>
<td>936</td>
<td>931 - 941</td>
</tr>
<tr>
<td>9</td>
<td>443</td>
<td>438 - 448</td>
<td>19</td>
<td>940</td>
<td>915 - 965</td>
</tr>
<tr>
<td>10</td>
<td>488</td>
<td>483 - 493</td>
<td>26</td>
<td>1375</td>
<td>1360 - 1390</td>
</tr>
</tbody>
</table>

Response are available on https://mcst.gsfc.nasa.gov/calibration/parameters.

3.3 In situ observations

3.3.1 Cloud Combination Probe

The Cloud Combination Probe (CCP) incorporates two separate instruments, the Cloud Droplet Probe (CDP) and the greyscale Cloud Imaging Probe (CIPgs) (Weigel et al., 2016; Brenguier et al., 2013). The instruments were installed below the right wing of the aircraft (forward facing) as shown in Fig. 3.2. The CCP overall covers a diameter range from 2 µm to 960 µm, including large aerosol particles, liquid cloud droplets, and small frozen hydrometeors (Klingebiel et al., 2015). The CDP detects the forward-scattered laser light, when cloud particles cross the CDP laser beam (Lance et al., 2010). Thus, the CDP provides an improved replacement for the Forward Scattering Spectrometer Probe (FSSP) (Dye and Baumgardner, 1984; Baumgardner et al., 1985). Molleker et al. (2014) showed, that the CDP exhibits a nominal limit for cloud particle diameters from 3 µm up to 50 µm. The CIPgs records two-dimensional shadow images of cloud particles in a size range from 15 µm up to 960 µm with an optical resolution of 15 µm (Klingebiel et al., 2015; Weigel et al., 2016). Special algorithms are used to process and analyze the captured images in order to estimate particle number concentrations, particle size distributions, and to identify particle shapes (Korolev, 2007).
The CCP measurements are employed to derive $r_{eff}$ for the comparison with the retrieval products from SMART-Albedometer and MODIS. The $r_{eff}$ from the CCP is derived from the geometrical properties and number of detected particles. The accuracy of the cloud particle sizing is estimated to be about 10% for spherical particles (Molleker et al., 2014). The sizing uncertainty increases as a function of particles shape complexity (i.e., when dendrites or particles with elevated aspect ratio were predominating). The size bin limits of the CCP cloud particle data are adapted to reduce ambiguities due to the Mie curve, particularly for cloud particles with sizes less than 5 µm. The instrument sample volume is calculated as a product of the aircraft TAS and the effective detection area. All concentration data are corrected concerning the air compression upstream of the underwing cloud probe at the high flight speeds (Weigel et al., 2016). The robust performance of the specific CCP instrument used in this study was demonstrated by Frey et al. (2011) for tropical convective outflow, Molleker et al. (2014) for polar stratospheric clouds, Klingebiel et al. (2015) for low level mixed-phase clouds in the Arctic, as well as Braga et al. (2017) and Cecchini et al. (2017) for tropical convective clouds.

### 3.3.2 Water Vapor Analyzer

Water vapor concentrations were measured by the Water Vapor Analyzer (WARAN), which is a tunable diode laser hygrometer based on the absorption of a laser beam by gaseous water molecules at $\lambda = 1370$ nm (Voigt et al., 2014; Kaufmann et al., 2014). The WARAN is installed on the forward-facing HALO trace gas inlet (TGI) as shown in Fig. 3.2. The instrument measures total water, i.e., gas phase plus enhanced ice water content $IWC$, in the range between 50 - 40,000 ppm with an accuracy of about ± 50 ppm or 5% of reading. Detailed descriptions about the measurement strategy and uncertainties in the data processing are discussed by Afchine et al. (2017). $IWC$ is derived from the difference between the amount of total enhanced water ($H_2O_{tot}$) and the amount of gas phase water ($H_2O_{gas}$) (Kaufmann et al., 2016). Due to the enhancement factor (Voigt et al., 2006) at the HALO-TGI, which is about 20-35, the minimum detectable $IWC$ is in the range between 1 - 2000 ppm ($1 - 2000 \times 10^{-2}$ mg m$^{-3}$). For the applications in this study, the vertical profile of $IWC$ is then utilized to infer the profile of cloud optical thickness $\tau(z)$. 
4 Comparison of upward radiance

This Chapter is related to the first part of the thesis. Sec. 4.1 describes spectral and resolution adjustments applied to SMART-Albedometer and MODIS measurements. Adjustments are needed to allow a fair comparison between SMART-Albedometer and MODIS due to different spectral and spatial resolutions. In Sec. 4.2, data filters are applied to choose appropriate cloud cases. The comparisons of upward radiance are given in Sec. 4.3. The majority of the materials shown in Chapter have been published in Krisna et al. (2018).

4.1 Spectral and spatial resolution adjustments

It should be emphasized that SMART-Albedometer and MODIS have different spectral resolutions. MODIS measures in broad spectral bands (see Table 3.3), while SMART-Albedometer measures in much narrower spectral with FWHM between 2-10 nm (see Table 3.1). Fig. 4.1 shows the relative spectral response \( RSR(\lambda) \) of MODIS bands 1, 5, and 6 centered at \( \lambda = 645 \) nm, 1240 nm, and 1640 nm, respectively. The spectral response of an instrument describes its relative sensitivity to energy of different wavelengths. To allow fair comparisons, the spectral upward radiance measured by SMART-Albedometer \( I_{s,\lambda}^{\uparrow} \) must be convoluted with the MODIS relative spectral response. The convoluted radiance of SMART-Albedometer \( I_{s,\lambda}^{\uparrow} \) is calculated by:

\[
I_{s,\lambda}^{\uparrow} = \frac{\int_{\lambda_1}^{\lambda_2} I_{s,\lambda}^{\uparrow} \cdot RSR(\lambda) \, d\lambda}{\int_{\lambda_1}^{\lambda_2} RSR(\lambda) \, d\lambda}.
\]  

(4.1)

In this study, upward radiances centered at the MODIS band 6 (\( \lambda = 1640 \) nm) will be primarily used to retrieve \( \tau \) and \( r_{eff} \), in addition to the MODIS band 5 (\( \lambda = 1240 \) nm). However, it is known that 15 of the 20 detectors of the MODIS-Aqua band 6 are either nonfunctional or noisy. Using the remaining pixels to obtain the radiance values for the applications in this study is not possible, because only limited number of pixels are left. According to Wang et al. (2006), the MODIS radiance band 6 \( I_{M,B6} \) can be retrieved using band 7 \( I_{M,B7} \) (\( \lambda = 2130 \) nm). This technique was originally developed and tested on the basis of snow surfaces deeming that the spectral characteristics of the snow reflectivity between MODIS bands 6 and 7 do not change significantly for different snow types. Given the fact that snow and ice clouds have similar optical properties, the same approach can therefore be applied for ice
clouds. For this purpose, forward simulations are performed to calculate spectral upward radiances for cirrus clouds with different values of $\tau$ and $r_{\text{eff}}$. Furthermore, a polynomial fit is applied to quantify the relation between $I_{M,B6}$ and $I_{M,B7}$, resulting in the following parameterization:

$$I_{M,B6} = -81.033 \cdot I_{M,B7}^2 + 3.257 \cdot I_{M,B7} + 0.002. \quad (4.2)$$

Fig. 4.2a shows the scatter plot of the original-simulated upward radiance (indexed with "ori") between $I_{M,B6}$ and $I_{M,B7}$. Here, the points represent the resulting upward radiance from the forward simulations assuming different $\tau$ and $r_{\text{eff}}$. Before developing the parameterization, the correlation between simulated $I_{M,B6}$ and $I_{M,B7}$ are analyzed. As shown in Fig. 4.2a, the original values of upward radiance at the two bands are highly correlated with a correlation coefficient $R^2$ of 1. This justifies the feasibility of using $I_{M,B7}$ to retrieved $I_{M,B6}$. Based on the simulated $I_{M,B7}$, the retrieval of $I_{M,B6}$ is performed using the parameterization given by Eq. 4.2. Fig. 4.2b shows the scatter plot between the original-simulated $I_{M,B6}$ and the result of $I_{M,B6}$ retrieval (indexed with "ret"). Again, the analysis shows that the original and the retrieved values have a robust correlation with $R^2 = 1$. The validity of $I_{M,B6}$ retrieval is further tested using the remaining pixels of $I_{M,B6}$ from actual measurements of MODIS above cirrus clouds, as shown in Fig. 4.2c. For this purpose, the MODIS cloud flag is utilized to indicate the measurements of cirrus clouds. Here, the linear regression analysis between the original and the retrieved values of $I_{M,B6}$ results in $R^2 = 0.95$ and only differs below 5 % (slope of 0.95 and zero bias). To some degree, this confirms the applicability of $I_{M,B6}$ retrieval for ice clouds. For liquid water clouds, the result might be different due to different optical properties of ice and liquid water particles.
The geolocation files of MODIS data are calculated for each 1 km instantaneous field of views (IFOV), where the geographic (longitude, latitude, and altitude) and ancillary information correspond to the intersection of the centers of each IFOV from the detectors in an ideal 1 km band on the Earth surface. However, the spatial resolution of SMART-Albedometer highly varies depending on the flight altitude and the temporal resolution. At a flight altitude of 10 km, SMART-Albedometer has a swath of approximately 349 m at the Earth surface (see Fig. 3.3). Therefore, this difference has to be considered in the data analysis.

![Figure 4.2](image)

**Figure 4.2:** (a) is the scatter plot of the simulated radiance bands 6 and 7. (b) is the scatter plot of the simulated band 6 (original and retrieved). (c) is the scatter plot of MODIS band 6 (original/remaining pixels and retrieved). The dashed lines represent the one-to-one line.

![Figure 4.3](image)

**Figure 4.3:** Illustration of the binning technique applied for SMART-Albedometer measurements. A 1 km MODIS pixel is bordered by points A - D, where E is the center of the pixel. The red arrow illustrates a flight path of HALO, where SMART-Albedometer measures three times (red crosses) within the MODIS pixel.

In order to decrease biases resulting from comparing individual measurements, the comparison of SMART-Albedometer and MODIS measurements is conducted using a binning technique illustrated in Fig. 4.3. The black rectangle describes a MODIS pixel with a 1
km² horizontal resolution centered at the point "E" while a HALO flight leg crossing the MODIS pixel is illustrated by the red line. During this overpass, SMART-Albedometer performs three measurements indicated by the red crosses. For doing the binning, initially the points A-D must be determined using a spatial aggregation technique. Subsequently, the four points are applied as the bound for indicating whether the measurements of SMART-Albedometer are within the prescribed MODIS pixel. When the measurements are classified in the predetermined pixel, the values are averaged and then the resulting mean value can be compared with the MODIS data.

4.2 Data filter

Only clouds with a top altitude higher than 8 km are selected for this study. The higher proximity to TOA reduces the influence of scattering and absorption by atmospheric molecules and aerosol particles above cloud. Consequently, no correction for the influence of the atmospheric layer above HALO is needed. To assure a similar viewing zenith angle of SMART-Albedometer and MODIS, only nadir observations in the center of MODIS swath are selected for the comparison. Werner et al. (2013) discussed that off-nadir measurements of less than 5° may lead to a bias in the retrieved \(\tau\) and \(r_{\text{eff}}\) of up to 1% and 5%, respectively. To minimize this bias, SMART-Albedometer measurements with roll and pitch angles larger than 3° are discarded. Additionally, only straight flight legs with altitude changes of less than 50 m are analyzed.

<table>
<thead>
<tr>
<th>Flight</th>
<th>Date</th>
<th>Cloud Type</th>
<th>Appearance</th>
<th>(z_t) (km)</th>
<th>Time - UTC (HH:MM:SS)</th>
<th>(v) (m s(^{-1}))</th>
<th>(\theta_0) (°)</th>
</tr>
</thead>
<tbody>
<tr>
<td>AC-18</td>
<td>09/28/2014</td>
<td>DCC topped by anvil</td>
<td>Inhomogeneous</td>
<td>8</td>
<td>17:56:00 - 17:57:30</td>
<td>9</td>
<td>26</td>
</tr>
</tbody>
</table>

The nadir point of MODIS moves much faster than the aircraft. Thus, it is impossible that SMART-Albedometer and MODIS measure exactly above each other along the joint flight. To analyze the effects caused by the time shift between SMART-Albedometer and MODIS measurements, data from the ML-CIRRUS and ACRIDICON-CHUVA are divided into groups within and without a threshold \(|\Delta t|\) of 500 s for the cirrus and 300 s for the DCCs. Scatter plots of SMART-Albedometer and MODIS radiance centered at \(\lambda = 645\) nm are shown in Fig. 4.4a for the cirrus and Fig. 4.4b for the DCC. For the cirrus (Fig. 4.4a), \(I_{S,645}^{\uparrow}\) and \(I_{M,645}^{\uparrow}\) are in a better agreement for \(|\Delta t| < 500\) s with a correlation coefficient \(R^2 = 0.96\), while for \(|\Delta t| > 500\) s deviations are larger with \(R^2 = 0.58\). The large scatter for
$|\Delta t| > 500$ s is mainly caused by the fast horizontal wind speed during the measurements (see Table 4.1). Additionally, the wind direction also becomes the key factor causing a significant cloud drift for the larger time shift. For the DCC (Fig. 4.4b), the scatter is profoundly larger compared to the cirrus for the given threshold of $|\Delta t| < 300$ s and even worse for the threshold of $|\Delta t| > 300$ s with $R^2 = 0.79$ and -0.09, respectively. In this case, the horizontal wind speed is smaller but the fast cloud evolution is the major issue. Luo et al. (2014) and Schumacher et al. (2015) reported that tropical DCCs located at altitudes between 6 - 8 km typically have an updraft velocity of about 2 - 4 m s$^{-1}$. According to these findings, the comparison are restricted to $|\Delta t| < 500$ s for the cirrus while for the DCC, the threshold is tightened to $|\Delta t| < 300$ s.

![Figure 4.4: Scatter plots of upward radiance at $\lambda = 645$ nm measured by SMART-Albedometer ($I_{S,645}^\uparrow$) and MODIS ($I_{M,645}^\uparrow$) within a threshold of 500 s for the cirrus (a) and 300 s the DCC (b). Blue circles and red triangles represent data within and without the predetermined threshold. The dashed line represents the one-to-one line.](image)

After the filtering, two suitable cases are selected, which fulfill most requirements of the analysis. The first case is a cirrus located above low liquid water clouds as part of ML-15 (ML-CIRRUS, 04/13/2014) between 13:56:20 - 13:57:35 UTC (Fig. 4.5a). The cloud top altitude $z_t$ of the cirrus was about 12 km, while HALO flew at about 12.3 km altitude. The second case, a DCC topped by an anvil cirrus is selected from AC-18 (ACRIDICON-CHUVA, 09/28/2014) between 17:56:00 - 17:57:30 UTC as presented in Fig. 4.5b. The $z_t$ of the selected DCC was about 8 km while HALO flew at about 8.3 km altitude. Flight descriptions and atmospheric conditions during the cloud measurements are summarized in Table 4.1. The selected time periods extend to 75 s for the cirrus case and 90 s for the DCC case. By assuming a TAS of HALO of about 200 m s$^{-1}$, flying at a constant altitude, those correspond to horizontal distances of about 15 km and 18 km, respectively.

In addition to the issues discussed above, the cloud edges should also be considered. The cloud edges are associated with sharp changes of $I_\lambda^\uparrow$ and significant 3-D radiative effects.
In order to avoid biases ascribed to the cloud edges, the measurements in vicinity of the cloud edges are discarded. For the measurements of SMART-Albedometer, the cloud mask algorithm by Ackerman et al. (1998) is applied to identify the clear and cloudy pixels, as well as where the cloud edges are. For the MODIS data, the first and the last pixel of MODIS cloudy pixels are discarded in the data analysis.

### 4.3 Results of upward radiance comparison

Upward radiances measured by SMART-Albedometer and MODIS are compared for the two selected cloud cases. Fig. 4.6 shows time series of upward radiance measured by SMART-Albedometer \( I_{S,L} \) and MODIS \( I_{S,L} \) centered at \( \lambda = 645 \text{ nm} \) (a), 1240 nm (b), and 1640 nm (c) for the cirrus case, while Fig. 4.7 shows the same for the DCC case. Those three wavelengths will be primarily utilized to retrieve the cloud properties in this study. Time series of upward radiances in Fig. 4.6 and Fig. 4.7 illustrate, that the cirrus is more homogeneous along the flight legs compared to the DCC. For the DCC case, the cloud anvil is observed between 17:56:00 - 17:56:20 UTC. Later, \( I_{645} \) increases sharply corresponding to the DCC core and decreases again towards the cloud edge. The results show, that the differences are larger for the DCC case, which are caused by the remaining effects of the cloud evolution. For the cirrus case, the differences are smaller because high cirrus typically change less rapidly.

Fig. 4.8 shows the comparison of mean spectral upward radiance measured by SMART-Albedometer and MODIS for the cirrus (a) and DCC case (b). The solid line represents
4.3. Results of upward radiance comparison

Figure 4.6: Time series of $I^\uparrow_{\lambda}$ centered at $\lambda = 645$ nm (a), 1240 nm (b), and 1640 nm (c) measured by SMART-Albedometer (black) and MODIS (red) for the cirrus case. Shaded areas are measurement uncertainties. Gaps on the time series indicate when the shutter of SMART-Albedometer closed for dark current measurements.

The spectral radiance measured by SMART-Albedometer $I^\uparrow_{s,\lambda}$, while $I^\uparrow_{s,\lambda}$ is the convoluted radiance of SMART-Albedometer using Eq. 4.1, and $I^\uparrow_{M,\lambda}$ is the radiance measured by MODIS.

The values of mean standard deviation $\eta$ at each spectral wavelength are summarized in Table 4.2. Note, that all standard deviation values in this thesis refer to the $\pm$ values. To quantify the agreement, the normalized mean absolute deviation $\zeta$ is calculated by:

$$\zeta = \frac{1}{n} \sum_{i=1}^{n} \left| \frac{x_i - \overline{x}}{\overline{x}} \right|,$$

where $n$ is the number of observed values, $x_i$ are the individual values, and $\overline{x}$ is the mean value of the radiances measured by SMART-Albedometer and MODIS along the selected time series. For the cirrus case, $\zeta_{645}$ is found to be 0.04 %, while $\zeta_{1240}$ and $\zeta_{1640}$ are 7.68 % and 1.36 %, respectively. For the DCC case, $\zeta_{645}$ yields a value of 4.25 %, while $\zeta_{1240}$ and $\zeta_{1640}$ are 6.72 % and 5.61 %, respectively. The good agreement between SMART-Albedometer $I^\uparrow_{s,1640}$ and MODIS $I^\uparrow_{M,1640}$ justifies the application of the retrieval of MODIS band 6 using the parameterization given by Eq. 4.2. Overall, all the values of $\zeta$ in Table 4.2 lie within the measurement uncertainties. The comparison yields a better agreement for the cirrus than for the DCC case. The larger deviations in case of DCC are not only...
Figure 4.7: Same as Fig. 4.6 but for the DCC case.

Figure 4.8: Comparison of mean $I_\lambda^\uparrow$ measured by SMART-Albedometer and MODIS for the cirrus case (a) and the DCC case (b) at $\lambda$ between 400 - 1800 nm. Error bars represent measurement uncertainties. Wavelengths centered at $\lambda = 645$ nm, 1240 nm, and 1640 nm are indicated by dashed lines while grey bands correspond to the interval of MODIS relative spectral response $RSR(\lambda)$ for the respective wavelengths.


Table 4.2: Comparison of SMART-Albedometer $I_{S,\lambda}^{\uparrow}$ and MODIS $I_{M,\lambda}^{\uparrow}$ for the cirrus (ci) and DCC case. $\eta$ is the mean standard deviation with a subscript of "S" for SMART-Albedometer and "M" for MODIS. $\zeta$ is the normalized mean absolute deviation between SMART-Albedometer and MODIS measurements.

<table>
<thead>
<tr>
<th>$\lambda$ (nm)</th>
<th>$\eta_{S,ci}$</th>
<th>$\eta_{M,ci}$</th>
<th>$\zeta_{ci}$ (%)</th>
<th>$\eta_{S,DCC}$</th>
<th>$\eta_{M,DCC}$</th>
<th>$\zeta_{DCC}$ (%)</th>
</tr>
</thead>
<tbody>
<tr>
<td>421</td>
<td>0.231 ± 0.014</td>
<td>0.234 ± 0.011</td>
<td>0.81</td>
<td>0.295 ± 0.122</td>
<td>0.251 ± 0.013</td>
<td>8.06</td>
</tr>
<tr>
<td>469</td>
<td>0.266 ± 0.018</td>
<td>0.265 ± 0.014</td>
<td>0.20</td>
<td>0.335 ± 0.149</td>
<td>0.351 ± 0.050</td>
<td>2.34</td>
</tr>
<tr>
<td>555</td>
<td>0.229 ± 0.018</td>
<td>0.224 ± 0.013</td>
<td>1.19</td>
<td>0.290 ± 0.135</td>
<td>0.303 ± 0.047</td>
<td>2.12</td>
</tr>
<tr>
<td>645</td>
<td>0.193 ± 0.016</td>
<td>0.193 ± 0.012</td>
<td>0.04</td>
<td>0.241 ± 0.117</td>
<td>0.263 ± 0.042</td>
<td>4.25</td>
</tr>
<tr>
<td>858</td>
<td>0.125 ± 0.011</td>
<td>0.128 ± 0.008</td>
<td>1.29</td>
<td>0.162 ± 0.069</td>
<td>0.167 ± 0.018</td>
<td>1.47</td>
</tr>
<tr>
<td>905</td>
<td>0.096 ± 0.008</td>
<td>0.104 ± 0.007</td>
<td>4.36</td>
<td>0.124 ± 0.059</td>
<td>0.129 ± 0.016</td>
<td>1.96</td>
</tr>
<tr>
<td>936</td>
<td>0.048 ± 0.005</td>
<td>0.056 ± 0.005</td>
<td>7.49</td>
<td>0.069 ± 0.043</td>
<td>0.080 ± 0.018</td>
<td>7.95</td>
</tr>
<tr>
<td>940</td>
<td>0.062 ± 0.006</td>
<td>0.071 ± 0.005</td>
<td>7.18</td>
<td>0.084 ± 0.047</td>
<td>0.099 ± 0.018</td>
<td>8.26</td>
</tr>
<tr>
<td>1240</td>
<td>0.052 ± 0.004</td>
<td>0.061 ± 0.004</td>
<td>7.68</td>
<td>0.057 ± 0.029</td>
<td>0.065 ± 0.009</td>
<td>6.72</td>
</tr>
<tr>
<td>1375</td>
<td>0.005 ± 0.001</td>
<td>0.005 ± 0.001</td>
<td>3.24</td>
<td>0.004 ± 0.004</td>
<td>0.004 ± 0.003</td>
<td>6.17</td>
</tr>
<tr>
<td>1640</td>
<td>0.024 ± 0.002</td>
<td>0.025 ± 0.001</td>
<td>1.36</td>
<td>0.016 ± 0.010</td>
<td>0.018 ± 0.001</td>
<td>5.61</td>
</tr>
</tbody>
</table>

Influenced by the cloud evolution, but also by 3-D radiative effects. Liang et al. (2009), Zhang and Platnick (2011), and King et al. (2013) estimated the influence of 3-D radiative effects using the cloud heterogeneity index $\sigma_{\text{sub}}$. The $\sigma_{\text{sub}}$ is defined as a ratio between the standard deviation and the mean value of MODIS radiance band 2 ($\lambda = 858$ nm and 250 m spatial resolution). For the cirrus case, $\sigma_{\text{sub}}$ is about 0.1. Higher heterogeneities are found for the DCC case with $\sigma_{\text{sub}}$ of about 0.4. This shows, that 3-D radiative effects are potentially larger for the DCC case, and therefore have to be considered when interpreting the retrieval results from different instruments.
5 Retrieval of cloud optical thickness and particle effective radius

This Chapter is related to the second part of the thesis. Sec. 5.1 discusses the radiative transfer simulation, the technique applied to handle the multilayer cloud condition, and the radiance ratio retrieval. Extensive sensitivity studies are performed in Sec. 5.2. Among others, they are conducted to study the impacts of underlying liquid water clouds and the choice of crystal shape on cirrus retrievals, the retrieval uncertainties due to the applied wavelengths, the impacts of underlying liquid water clouds on the phase index. The vertical photon transport is of a big concern in this study. Studies on the vertical weighting function and its applications, such as to investigate impacts of surface albedo assumption and underlying liquid water clouds, as well as to calculate the vertical penetration depth, are carried out. Parts of this Chapter have been published in Krisna et al. (2018).

5.1 Radiative transfer simulation and radiance ratio retrieval

5.1.1 Radiative transfer simulation

One-dimensional (1-D) radiative transfer simulations are performed to calculate the spectral upward radiance by the discrete ordinate radiative transfer solver (DISORT) version 2 (Stamnes et al., 2000) in the radiative transfer package LibRadtran 2.0.2 (Mayer, 2005; Emde et al., 2016). The extraterrestrial spectral irradiance is taken from Gueymard (2004). The standard atmospheric profiles of gases and constituents are adapted from Anderson et al. (1986). The "mid-latitude summer" is applied for the ML-CIRRUS campaign, while the "tropical summer" profile if for the ACRIDICON-CHUVA campaign. The standard profiles are adjusted to the concurrent radio sounding data (temperature and humidity), measured by closest stations from the measurement area. The standard aerosol particle profile for "spring/summer condition" of "maritime aerosol type" (Shettle, 1989) is applied according to the condition in the measurement area. For the molecular and gas absorption, the parameterization of the low atmospheric transmission (LOWTRAN) by Pierluissi and Peng (1985) as adopted from the Santa Barbara DISORT atmospheric radiative transfer (SBDART) (Ricchiazzi and Gautier, 1998) is used.
The spectral surface albedo $\rho$ is an important component in the forward simulation, particularly for simulating optically thin clouds located above highly reflecting surfaces. During the two campaigns, $\rho$ was obtained from upward- and downward irradiances measured by the SMART-Albedometer. The detailed technique for measuring $\rho$ is described in Wendisch et al. (2004). For the cirrus case, $\rho$ of ocean was derived from low-level flights performed in the vicinity of the cirrus case. Given that the ocean surface was homogeneous, the measured $\rho$ can be applied in the simulations along the selected flight leg. Nevertheless, the same approach can not be implemented for the DCC case as the variability of the surface albedo in the Amazon rainforest is very high. In this area, forested and deforested areas are located side by side. This implies that a representative assumption of a homogeneous surface along the flight leg is inappropriate. Since there were no corresponding SMART-Albedometer measurements at low altitude covering exactly the same flight leg, $\rho$ is derived from the MODIS Bidirectional Reflectance Distribution Function (BRDF)/Albedo product (Strahler et al., 1999) to include the horizontal variability of the surface albedo. Other than that, the output altitude of the upward radiance must be specified in the simulation. The upward radiance is simulated at the flight altitude of HALO for representing the measurements of SMART-Albedometer and at the top of atmosphere (TOA) for the MODIS observations. Since the measurements were performed at a high altitude, no significant differences are obtained.

For simulating the atmosphere with clouds, an input file specifying the profile of cloud water content ($LWC$ or $IWC$) (in units of g m$^{-3}$) and $r_{\text{eff}}$ (in units of µm) must be defined in the simulation. For simplification, a vertically homogeneous cloud is commonly assumed. Forward simulations can be performed either in the liquid water or ice mode, depending on the thermodynamic phase of the cloud. To decide which mode is used, a cloud phase index $I_p$ is determined using the spectral slope method according to Jäkel et al. (2013). $I_p$ is defined by the slope of SMART-Albedometer radiance measurements at $\lambda = 1550$ nm and 1700 nm (see Eq. 5.1), which typically results in a positive value for ice clouds and a negative value for liquid water clouds.

$$I_p = \frac{I_{\uparrow}^{1700} - I_{\uparrow}^{1550}}{I_{\uparrow}^{1700}}$$  \hspace{1cm} (5.1)

In this study, a threshold of 0.15 is applied to classify ice and liquid water clouds. For the cirrus case, time series of $I_p$ yield values larger than 0.25 indicating pure ice clouds. It means that for this multilayer case, $I_p$ is mostly sensitive to the thermodynamic phase of the upper cloud layer (cirrus), while the lower cloud layer (liquid water clouds) only have a limited influence within the wavelength range used to determine $I_p$. This situation holds true as long as the cirrus $\tau$ is sufficiently thick, such that investigated in this study. For the DCC case, the values of $I_p$ highly vary between 0.2 and 0.4 with a mean value of about 0.23. According to the high $I_p$ values, the forward simulations, and thereby the retrievals for both analyzed cloud cases, are performed in the ice mode.
For the simulations of liquid water clouds, the optical properties of liquid water droplets are derived from Mie calculation (Wiscombe, 1980; Mie, 1908) while for ice clouds, the choice of ice crystal habit relies on the information given by the in situ measurements. During the ML-CIRRUS campaign, the in situ measurements indicated the presences of mixture particles with a high surface roughness (Voigt et al., 2017), thus a representative habit of general habit mixtures (GHM) based on severely roughened aggregates by Baum et al. (2014) is applied for the cirrus case. During the ACRIDICON-CHUVA campaign, ice crystals composed of aggregated plates with a high surface roughness were most found at the anvil of the DCCs (Järvinen et al., 2016). Thus, an ice crystal habit of aggregated plates with a high surface roughness by Yang et al. (2013) is applied for simulating the DCC case.

5.1.2 Discriminating the properties of multilayer clouds

Due to the multilayer cloud situation, a liquid water cloud layer has to be included in the forward simulation of the cirrus case. One difficulty caused by such conditions is that in the simulation, it is required to specify the properties of the low liquid water cloud, jointly with those for the cirrus. In this study, the properties of the liquid water cloud are estimated by comparing the average value of spectral upward radiance measured by SMART-Albedometer with simulations assuming different combinations of the properties of the two clouds. In the simulation, the cirrus layer is set between 10 and 12 km while the liquid water cloud is between 1.5 and 2 km, according to observations. An optimal combination of cloud properties will produce a good fit in the entire range of spectral upward radiance, particularly at the wavelengths that contain significant information on the properties of the two layer clouds. Fig. 5.1 illustrates the technique implemented for determining the properties of the liquid water cloud. In this example, only relevant combinations are shown.

In the first step (see Fig. 5.1a), the optical thickness of the liquid water cloud \( \tau_{\text{liq}} \) is varied between 6 and 10 with the purpose to estimate the true value of \( \tau_{\text{liq}} \). For this reason, the particle effective radius of the liquid water cloud \( r_{\text{eff}, \text{liq}} \) is fixed to 10 µm. The result shows that changes of \( \tau_{\text{eff}} \) will influence the entire spectral, where the largest impact is obtained at the wavelengths dominated by scattering (co-albedo \( \approx 0 \)). Here, a simulation with \( \tau_{\text{liq}} = 8 \) results in the best fit with the measurement.

In the second step (see Fig. 5.1b), \( r_{\text{eff}, \text{liq}} \) is varied between 6 and 14 µm in order to estimate the true value of \( r_{\text{eff}, \text{liq}} \). According to the result obtained in the first step, \( \tau_{\text{liq}} \) is fixed to 8. Here, a simulation with \( r_{\text{eff}, \text{liq}} = 10 \) µm yields the best fit with the measurement. Unlike \( \tau_{\text{liq}} \), the impact of changing \( r_{\text{eff}, \text{liq}} \) to the entire spectral is smaller. But, it can still be observed that there is a particular region, which is sensitive to the changes of \( r_{\text{eff}, \text{liq}} \). The highest sensitivity is found at wavelengths with high absorption by cloud particles, such as at \( \lambda = 1450-1650 \) nm. Note that all the simulations presented here are made by assuming \( \tau_{\text{ci}} = 3 \) and \( r_{\text{eff}, \text{ci}} = 18 \) µm, which are found as the most-likely value. From the first two steps, a simulation with \( \tau_{\text{liq}} = 8 \) and \( r_{\text{eff}, \text{liq}} = 10 \) µm shows the best agreement with the
measurement, particularly in the absorption bands by water vapor (H\textsubscript{2}O), e.g., at $\lambda = 940$ nm and 1135 nm, and oxygen-A (O\textsubscript{2}-A) at $\lambda = 762$ nm, as shown in Fig. 5.1c. Rozanov and Kokhanovsky (2004) and Wind et al. (2010) have described that these bands are sensitive for discriminating the properties of multilayer clouds, particularly for $\tau$. Assuming over- or underestimated values of $\tau_{\text{liq}}$ and $\tau_{\text{ci}}$ will lead to apparent discrepancies and therefore, they must be distributed correctly.

![Figure 5.1](image_url)

**Figure 5.1**: The technique for specifying the properties of the liquid water cloud. (a) is by varying $\tau_{\text{liq}}$ with fixed $r_{\text{eff,liq}} = 10$ $\mu$m, while (b) is by varying $r_{\text{eff,liq}}$ with fixed $\tau_{\text{liq}} = 8$. (c) is the best fit with $\tau_{\text{liq}} = 8$ and $r_{\text{eff,liq}} = 10$ $\mu$m. (d) is when the total $\tau (\tau_{\text{ci}}+\tau_{\text{liq}})$ is fixed to 11, but $\tau_{\text{ci}}$ and $\tau_{\text{liq}}$ are distributed proportionally.

The third step of this technique is aimed to analyze the configuration in the absorption bands by water vapor and O\textsubscript{2} – A. While the majority of the spectra might have a good fit, it can be noticed as a spurious solution if the measurement and the simulation do not
have a good fit in those bands. For this purpose, forward simulations are performed with a fixed total optical thickness $\tau_c$ of 11 ($\tau_{ci} + \tau_{liq}$), but $\tau_{ci}$ and $\tau_{liq}$ are varied proportionally. The value of $\tau_c$ is defined according to the result from the first step. Fig. 5.1d shows if prescribed $\tau_{liq}$ is overestimated and $\tau_{ci}$ is thereby underestimated, the resulting upward radiance in the absorption bands is lower. In this case, more radiation are transmitted by the thin cirrus and further being absorbed by water vapor and $O_2$-A below the cirrus. Conversely, the thicker the $\tau_{ci}$, the higher the upward radiance. In this case, the incoming solar radiation is largely reflected by the cirrus while the amount of transmitted radiation absorbed by water vapor and $O_2$-A below the cirrus gets to be smaller.

Indeed, the absorption by water vapor also affects the upward radiance at wavelengths between 1360 and 1420 nm. However, the absorption here is not influenced by water vapor solely. In addition to water vapor, carbon dioxide ($CO_2$) also contributes largely. Given that the profile of $CO_2$ was not measured during the campaign, it is difficult to reproduce the measurement using the simulation. Additionally, since the absorption bands are also sensitive to the cloud geometrical altitudes, they must be defined correctly in the forward simulation. In this study, those information are provided by the lidar measurement of the water vapor lidar experiment in space (WALES) aboard of HALO (Wirth et al., 2009). The approach for discriminating the properties of the two layer clouds applied here is similar to the result of simultaneous retrieval of ice and liquid water cloud properties, as presented by, e.g., Sourdeval et al. (2015). However, developing this type of retrieval technique is beyond the scope of this thesis.

### 5.1.3 Radiance ratio retrieval

A radiance ratio technique (Werner et al., 2013) is applied in order to retrieve $\tau$ and $r_{eff}$ assuming vertically homogeneous clouds. In the radiometric calibration processes, one source of uncertainties originates from the applied radiation source. This type of uncertainty will equally influence the entire spectral radiance. When the ratio is applied, the measurement uncertainties are reduced because the uncertainty from the radiation source cancel. Consequently, the radiance ratio technique is also capable to reduce the retrieval uncertainties. Retrievals of $\tau$ and $r_{eff}$ require wavelengths, which sensitive to each individual retrieval parameters. To analyze the wavelengths, which have enough sensitivities in deriving $\tau$ and $r_{eff}$, forward simulations are run for various combinations of $\tau$ and $r_{eff}$ that assume ice clouds. Fig. 5.2a shows the ratio of standard deviation and mean of upward radiance while Fig. 5.2b is the the co-albedo ($1-\tilde{\omega}$) of GHM by Baum et al. (2014) for $r_{eff}$ between 10 and 40 $\mu$m as a function of wavelength. For the sensitivity test of $\tau$ (red line Fig. 5.2a), $\tau$ is varied between 0.2 and 7 but $r_{eff}$ is fixed to 30 $\mu$m. Whereas for the sensitivity of $r_{eff}$ (black line in Fig. 5.2a), $r_{eff}$ is varied from 10 to 60 $\mu$m and $\tau = 3$ is fixed. The results in Fig. 5.2a confirms that at wavelengths where the scattering dominates, thereby the co-albedo $\approx 0$, the ratio increases which represent an enhancing sensitivity of upward
radiance with respect to \( \tau \). On the other hand, at wavelengths where the absorption is more pronounced, thereby the co-albedo is higher, the value increases correspondingly. This yields an increasing sensitivity of upward radiance in term of \( r_{\text{eff}} \). It should be remarked that the wavelengths selected for the retrieval are beyond gaseous and molecular absorption.

\[ \text{Figure 5.2: (a) shows the standard deviation } \sigma \text{ of radiance as a function of wavelength. For the sensitivity test of } \tau \text{ (red line), } \tau \text{ is varied between 1 and 8 while } r_{\text{eff}} \text{ is fixed to 30 µm and } r_{\text{eff}} \text{ (black). Whereas for the sensitivity test of } r_{\text{eff}} \text{ (black line), } r_{\text{eff}} \text{ is varied from 5-60 µm and } \tau = 3 \text{ is fixed. (b) shows the co-albedo } (1-\tilde{\omega}) \text{ of GHM by Baum et al. (2014) with } r_{\text{eff}} = 10-40 \mu m.} \]

In the retrieval algorithm, the upward radiance of MODIS bands 1 (\( \lambda = 645 \text{ nm} \)), 5 (\( \lambda = 1240 \text{ nm} \)), and 6 (\( \lambda = 1640 \text{ nm} \)) are applied to calculate the following radiance ratios: \( R_{1240} = \frac{I_{1240}^\uparrow}{I_{645}^\uparrow} \) and \( R_{1640} = \frac{I_{1640}^\uparrow}{I_{645}^\uparrow} \). \( I_{645}^\uparrow \) is selected due to its high sensitivity to \( \tau \) while \( I_{1240}^\uparrow \) and \( I_{1640}^\uparrow \) are sensitive to \( r_{\text{eff}} \) and more importantly, they are covered by the spectral range of SMART-Albedometer and MODIS. The lookup tables along with the measurements of SMART-Albedometer and MODIS for the cirrus case are shown in Fig. 5.3a and 5.3b whereas Fig. 5.3c and 5.3d are for the DCC case. In the retrieval algorithm, the upward radiance at a scattering wavelength \( I_{645}^\uparrow \) (\( \lambda = 645 \text{ nm} \)) is combined with \( R_{1240} \) (combination 1 / so-called C1) and with \( R_{1640} \) (combination 2 / so-called C2). For the cirrus case, the lookup tables are computed for \( \tau \) between 1 and 5 with steps of 1 and \( r_{\text{eff}} \) between 5 and 60 µm with steps of 3 µm. Note that for this case, a liquid water cloud layer is included in the forward simulations due to the multilayer cloud condition. The properties of
the liquid water cloud have inferred by the approach discussed in Sec. 5.1.2 that results in \( \tau_{\text{liq}} = 8 \) and \( r_{\text{eff,liq}} = 10 \) µm and they are assumed to be constant along the flight leg.

5.1. Radiative transfer simulation and radiance ratio retrieval

For the DCC case, the lookup tables cover \( \tau \) between 6 and 40 with steps of 1 for \( \tau \) between 6 and 22 and steps of 2 for \( \tau \) between 24 and 40 while \( r_{\text{eff}} \) spans between 5 and 90 µm with steps of 3 µm for \( r_{\text{eff}} \) between 5 and 56 µm and steps of 4 µm for \( r_{\text{eff}} \) between 60 and 90 µm. It is obvious that \( I_{645} \) increases with \( \tau \) due to enhanced reflected radiation by the cloud. Given that the water vapor absorption is weak at this wavelength, only a relatively small dependence of the upward radiance to \( r_{\text{eff}} \) is observed. Contrarily, there is a strong sensitivity of \( R_{1240} \) and \( R_{1640} \) with respect to \( r_{\text{eff}} \). The decrease in \( R_{1240} \) and \( R_{1640} \) with increasing \( r_{\text{eff}} \) is due to increased absorption for larger ice crystals. While the isolines are more distinct for larger values of \( \tau \) and \( r_{\text{eff}} \), they become more ambiguous for small values of \( \tau \), such as when \( \tau < 1 \) (not shown here). Platnick (2000) has discovered that this is caused by both decreasing asymmetry parameter and absorption for smaller \( r_{\text{eff}} \).

Figure 5.3: Radiance lookup tables for the cirrus case (a,b) and DCC case (c,d). (a) and (c) are using C1 (\( I_{645} \) and \( R_{1240} \)) while (b) and (d) are using C2 (\( I_{645} \) and \( R_{1640} \)). For the cirrus case, the simulations are performed with \( \theta_0 = 37^\circ \) and assuming GHM (Baum et al., 2014) while for the DCC case, \( \theta_0 = 26^\circ \) and the ice habit of aggregated plates (Yang et al., 2013) are applied. The measurements of SMART-Albedometer and MODIS are illustrated by symbols.
In this way, forward simulations assuming smaller particles can result in similar upward radiances with those assuming larger particles. As a consequence, the retrieval can yield multiple solutions of $r_{\text{eff}}$. To avoid this issue, the retrievals are restricted only for $\tau$ larger than 1.

For the C1, which is based on $I_{1240}^\uparrow$, the MODIS data do not match the lookup table solution space. The comparison of upward radiance in Section 4.3 indicates that $I_{M,1240}^\uparrow$ are higher than $I_{S,1240}^\uparrow$ by about 15%. For the cirrus case in Fig. 5.3a, all the retrievals of $r_{\text{eff}}$ using the original $I_{M,1240}^\uparrow$ (blue circles) fail since the measurements lie far outside the solution space. For the DCC case, it is found that the retrieval failure is smaller (see Fig. 5.3c). Enhancing retrieval failure in the cirrus case is caused by the larger $\theta_0$, therefore the upward radiance becomes more insensitive to the changes of $r_{\text{eff}}$ resulting in tighter lookup tables.

In order to gain meaningful cloud properties, a correction of $I_{M,1240}^\uparrow$ is necessary. Following Lyapustin et al. (2014), the correction factor $c$ is defined by the slope of the linear regression between $I_{M,1240}^\uparrow$ and $I_{S,1240}^\uparrow$ which results in $c = 0.88$ for the cirrus case and $c = 0.90$ for the DCC case. The corrected $I_{M,1240}^\uparrow$ (red circles) are added in Fig. 5.3 and now match the solution space. Thus, in the following, all the MODIS retrievals for the two investigated cloud cases use the corrected $I_{M,1240}^\uparrow$.

\[
\chi = \sum_{i=1}^{n} \frac{[y_i - F_i(x)]^2}{\sigma_i^2}
\]  

(5.2)

In the retrieval, the solution is inferred by minimizing the cost function $\chi$ given by Eq. 5.2 through a successive interpolation technique (inverse problem). Here, $i$ represents the number of measurement and $\sigma$ is the standard deviation. The retrieved state $x$ is obtained if the cost function is small enough or when the difference between the measurement $y$ and the forward model knowing the state $F(x)$ is considerably close so that it can be explained by the uncertainties. In the radiance ratio method, the measurement uncertainties are estimated to be about 4% for $I_{645}^\uparrow$ and 6% for $R_{1240}$ and $R_{1640}$. The retrieval uncertainties are estimated by considering the measurement uncertainties expressed by its double standard deviation $2\sigma$. The retrieval is run by varying each measurements separately by adding and subtracting $2\sigma$ which results in four solutions. The median of the four solutions is assumed as the retrieved value while the standard deviation is used to represent the retrieval uncertainties, $\delta\tau$ for $\tau$ and $\delta r_{\text{eff}}$ for $r_{\text{eff}}$. Note that by using C1, retrievals of $r_{\text{eff}}$ will introduce larger uncertainties compared to by using C2 because the absorption at $\lambda = 1240$ nm is smaller compared to that at $\lambda = 1640$ nm (see, Fig. 5.2). As the result, the upward radiance becomes more invariant to the changes of $r_{\text{eff}}$. A more detailed investigation about the retrieval uncertainties is given in Sec. 5.2.3.
5.2 Sensitivity studies

5.2.1 Impact of underlying liquid layer clouds on the cirrus retrieval

For the cirrus case, the properties of the low liquid water cloud are assumed to be constant along the flight legs. This assumption might not hold in reality and affect the retrieved cirrus properties. Therefore, the sensitivity of the cirrus retrieval with respect to the assumed properties of the low liquid water cloud is quantified using radiative transfer simulations. Spectral upward radiances are simulated for different combinations of liquid water cloud and cirrus properties. The liquid water cloud is varied for \( \tau_{\text{liq}} = 6 - 10 \) (steps of 1) and \( r_{\text{eff,liq}} = 6 - 14 \) µm (steps of 2 µm), while the cirrus is changed for \( \tau_{\text{ci}} = 2 - 8 \) (steps of 1 µm) and \( r_{\text{eff,ci}} = 10 - 40 \) µm (steps of 5 µm). These simulated upward radiances are used as synthetic measurements and analyzed with the retrieval algorithm using C2 (\( I_{\lambda_{145}} \) and \( R_{\lambda_{1640}} \)) which assumes a liquid water cloud with \( \tau_{\text{liq}} = 8 \) and \( r_{\text{eff,liq}} = 10 \) µm. The comparison of synthetically retrieved and original \( \tau_{\text{ci}} \) and \( r_{\text{eff,ci}} \) is shown in Fig. 5.4. Above one-to-one line is when the retrieval is run with an underestimation, while below one-to-one line is with an overestimation of the properties of the low liquid water cloud. The retrieved \( \tau_{\text{ci}} \) are analyzed in Fig. 5.4a for different \( \tau_{\text{liq}} \), while \( r_{\text{eff,ci}} \) and \( r_{\text{eff,liq}} \) are fixed to 20 µm and 10 µm, respectively. Similarly, the retrieved \( r_{\text{eff,ci}} \) are analyzed in Fig. 5.4b for different \( r_{\text{eff,liq}} \) but for a fixed combination of \( \tau_{\text{ci}} = 3 \) and \( \tau_{\text{liq}} = 8 \).

Figure 5.4: Comparison of synthetically retrieved \( \tau_{\text{ci}} \) (a) and \( r_{\text{eff,ci}} \) (b). Calculations in (a) are performed by changing \( \tau_{\text{liq}} \), while the original value is 8 and \( r_{\text{eff,ci}} = 20 \) µm and \( r_{\text{eff,liq}} = 10 \) µm are fixed. In (b), \( r_{\text{eff,liq}} \) is changed, while the original value is 10 µm and \( \tau_{\text{ci}} = 3 \) and \( \tau_{\text{liq}} = 8 \) are fixed.

In general, the results show that an overestimation of \( \tau_{\text{liq}} \) leads to an underestimation of \( \tau_{\text{ci}} \) because in this case, the liquid water cloud contributes more strongly to the reflected radiation than in reality. Therefore, a smaller \( \tau_{\text{ci}} \) is required to match the measurement, and
vice versa. For the range of $\tau_{ci}$ analyzed here, the retrieved $\tau_{ci}$ is found to be overestimated or underestimated by 1.3 when in reality $\tau_{liq}$ is 6 or 10 while the retrieval assumes $\tau_{liq} = 8$. These biases show that $\tau_{liq}$ needs to be estimated accurately because an inappropriate assumption of $\tau_{liq}$ almost directly propagates to the uncertainties of $\tau_{ci}$. A similar result is found in the retrieval of $r_{eff,ci}$ where an overestimation of $r_{eff,liq}$ leads to an underestimation of $r_{eff,ci}$, and vice versa. Assuming larger liquid droplets than in reality implies that these droplets contribute more strongly to the measured absorption at $\lambda = 1640$ nm, and therefore the ice crystals contribute less, represented by a smaller $r_{eff,ci}$. Fig. 5.4b illustrates that the impact of $r_{eff,liq}$ is strongest when small liquid droplets ($r_{eff,liq} \leq 8$ µm) are present. For larger liquid droplets ($r_{eff,liq} > 10$ µm), the impact is considerably reduced. The maximum uncertainties of $r_{eff,ci}$ within the range of $r_{eff,ci}$ and $r_{eff,liq}$ analyzed here are about 8 µm for an underestimation of $r_{eff,liq}$ showing a tendency of higher uncertainties for higher $r_{eff,ci}$.

Figure 5.5: Same as Fig. 5.5b but it is calculated for $\tau_{ci} = 5$ (a) and 8 (b).

Fig. 5.5 shows the same as Fig. 5.4b but it is calculated for $\tau_{ci} = 5$ (a) and 8 (b). For $\tau_{ci} = 5$ (Fig. 5.5a), the bias on the retrieved $r_{eff,ci}$ is found in the range between 1 and 4 µm showing a tendency of increasing uncertainty for larger $r_{eff,ci}$. Meanwhile for $\tau_{ci} = 5$ as shown in Fig. 5.5b), the uncertainty is significantly reduced in the range between 0.5 and 1.9 µm with the same tendency. If $\tau_{ci}$ exceeds the value of 10, the impact of $r_{eff,liq}$ to the retrieval of $r_{eff,liq}$ becomes negligible. Overall, it can be concluded that the retrieval of $r_{eff,ci}$ is less affected by $r_{eff,liq}$ when the cirrus layer is sufficiently thick since the cirrus layer will dominate the reflected radiation in the absorption bands applied for the retrieval. Thus, for the DCC case analyzed in this study with $\tau_{ci} \geq 13$, the influence of $r_{eff,liq}$ is not an issue.

5.2.2 Impact of ice crystal habit

In this study, a representative ice crystal habit of general habit mixture Baum et al. (2014) is assumed in the forward simulation of the cirrus case while for the DCC case, a habit of
aggregated plates Yang et al. (2013) is applied. However, it should be taken into consideration that assuming a single habit can introduce uncertainties since in reality, the ice crystal habit evolves following the development of the cloud. According to van Diedenhoven et al. (2014) and Sassen and Wang (2008), the variability of ice crystal habit is strongly governed by dynamic and thermodynamic state of the cloud, as well as the available supply of ice particle forming nuclei. The information given by in situ measurements capture at a certain state of the cloud only. Therefore, a sensitivity study is carried out to analyze uncertainties resulting from the assumption of ice crystal habit. According to Platnick et al. (2017), the discrepancy on the retrievals of $\tau$ between assuming two different habits (indexed with "A" and "B") can be approximated by:

$$\frac{\tau_A}{\tau_B} \approx \frac{1 - \tilde{\omega}_{0,B} \cdot g_B}{1 - \tilde{\omega}_{0,A} \cdot g_A},$$  \hspace{1cm} (5.3)$$

where $\tilde{\omega}_0$ is the single scattering albedo and $g$ is the single scattering asymmetry parameter. Since retrievals of $\tau$ commonly uses a scattering wavelength (e.g., $\lambda = 645$ nm) characterized with a single scattering albedo $\tilde{\omega}_0 \approx 0$, the expressions on the right hand side of Eq. 5.3 can be simplified to $1 - g$. Fig. 5.6 shows $g$ and the co-albedo $(1 - \tilde{\omega}_0)$ of GHM (red) by Baum et al. (2014), aggregated plates (blue), and aggregated columns (black) by Yang et al. (2013) as a function of $r_{eff}$ calculated at three wavelengths: 645 nm (left), 1240 nm (middle), and 1640 nm (right). All the habits analyzed for this study are modeled with a high surface roughness (so-called severely roughed surface), in accordance with the information given by the in situ measurements during the campaigns (Järvinen et al., 2016; Voigt et al., 2017). It is found that the aggregated columns have the least magnitude of $g$ ($\sim 0.75$) compared to the other habits. Due to smaller $g$, the incoming solar radiation is scattered more to the backward direction, relative to the forward direction, enhancing the upward radiance measured by the sensor above the cloud. As a result, retrievals of $\tau$ assuming a habit with smaller $g$ will result in a smaller $\tau$, and vice versa. Considering the three habits analyzed here, this leads to $\tau_{agg.\text{columns}} < \tau_{GHM} < \tau_{agg.\text{plates}}$.

In Fig. 5.6d-5.6f, it is obvious that each habit has distinct values of co-albedo that vary with $r_{eff}$. To approximate the discrepancies for assuming different habits, Platnick et al. (2017) and Holz et al. (2016) used the co-albedo value of habit "A" with a predetermined $r_{eff}$ and then interpolates this value onto habit "B" to find a $r_{eff}$ that gives the same co-albedo value. By this approach, assuming a habit with a smaller co-albedo will result in a larger retrieved $r_{eff}$, and vice versa. Nevertheless, this approach seems to be inappropriate because if this is the case, the cloud is then considered to consist of only one particle. In reality, it is known that a cloud is composed of many particles. Additionally, it should be kept in mind that retrievals of $r_{eff}$ are not only influenced by the co-albedo solely, but also by $g$. Assuming a habit with a smaller $g$ will result in a larger retrieved $r_{eff}$. Due to a smaller $g$, the resulting simulated upward radiance is higher, thus a larger $r_{eff}$ which gives more absorption is required to match the measurement, and vice versa. Based on these
facts, a more comprehensive approach in analyzing the discrepancies on the retrievals of \( r_{\text{eff}} \) should into account both the co-albedo and \( g \).

**Figure 5.6:** Single scattering asymmetry parameter \( g \) as a function of \( r_{\text{eff}} \) at \( \lambda = 645 \text{ nm} \) (a), 1240 nm (b), and 1640 nm (c) for different ice crystal habits: GHM (red) by Baum et al. (2014), aggregated plates (blue), and aggregated columns (black) by Yang et al. (2013) (red). (d), (e), (f) are the co-albedo (1 - \( \tilde{\omega}_0 \)) at corresponding wavelengths.

**Figure 5.7:** Comparison of synthetically retrieved \( \tau \) (a) and \( r_{\text{eff}} \) (b). Synthetic measurements are generated assuming three habits, GHM (red cross), aggregated plates (blue diamond), and aggregated column (black star). The retrievals are performed by assuming GHM.
5.2. Sensitivity studies

Fig. 5.7 shows the comparison of synthetically retrieved $\tau$ (a) and $r_{\text{eff}}$ (b). For this purpose, synthetic measurements are generated via forward simulations by assuming three different habits, GHM (red cross), aggregated plates (blue diamond), and aggregated columns (black star) which cover $\tau$ between 1 and 8, and $r_{\text{eff}}$ between 10 and 45 µm. For the retrieval of $\tau$ in Fig. 5.7a, the $r_{\text{eff}}$ is fixed to 25 µm while for the retrieval of $r_{\text{eff}}$ in Fig. 5.7b, the $\tau$ is fixed to 3. All the retrievals are made by assuming GHM and using the combination 1 (645 nm and $\Re_{1640}$). The result in Fig. 5.7 clearly shows that retrievals of $\tau$ assuming GHM result in larger retrieved $\tau$ when in reality, the ice crystals are composed of aggregated columns. On the other hand, smaller $\tau$ are obtained when in reality the ice crystals consist of aggregated plates. This condition is profoundly influenced by $g_{\text{GHM}} > g_{\text{agg.columns}}$ and $g_{\text{GHM}} > g_{\text{agg.plates}}$. The resulting relative difference between the original and retrieved values for the aggregated columns yields values between 16 % and 19 % that increases with $\tau$ while for the aggregated plates, it ranges between 20 % and 30 % which decreases with $\tau$. The result in Fig. 5.7b yields that retrievals of $r_{\text{eff}}$ assuming GHM result in larger $r_{\text{eff}}$ when in reality, the ice crystals are composed of aggregated plates. Conversely, smaller $r_{\text{eff}}$ are obtained if in reality, the ice crystals are comprised of aggregated columns. The resulting relative difference between the original and retrieved values for the aggregated columns is between 13 % and 16 % that increases with $r_{\text{eff}}$ while for the aggregated plates, it ranges between 30 % and 49 % which decreases with $r_{\text{eff}}$. The results reveal clearly that the discrepancies on the retrievals of $\tau$ and $r_{\text{eff}}$ are mainly influenced by $g$. When the co-albedo gets more distinct, e.g., between GHM and aggregated plates for $r_{\text{eff}} > 20 \mu$ (see Fig. 5.7f), the influence of co-albedo on the retrievals of $r_{\text{eff}}$ is more pronounced but it is considerably small compared to the impact of $g$.

5.2.3 Retrieval uncertainties

The resulting retrieval uncertainties of by using C1 ($I_{645}^\uparrow$ and $\Re_{1240}$) and C2 ($I_{645}^\uparrow$ and $\Re_{1640}$), will be analyzed with different solar zenith angle $\theta_0$. Spectral upward radiance are simulated for cirrus clouds with different $\tau$ and $r_{\text{eff}}$. The simulations are carried out by assuming $\theta_0 = 10$ and 36° and using the ice crystal habit of GHM by Baum et al. (2014). To analyze the uncertainties of $\tau$ ($\delta\tau$), synthetic measurements with $\tau = 2$-4 and $r_{\text{eff}} = 25 \mu$m are chosen, while for analyzing the uncertainty of $r_{\text{eff}}$ ($\delta r_{\text{eff}}$), synthetic measurements with $r_{\text{eff}} = 30$-40 µm and $\tau = 3$ are applied. All the retrievals are run by pondering measurements uncertainties of 4 % for $I_{645}^\uparrow$ and 6 % for the ratios $\Re_{1240}$ ($I_{1240}^\uparrow/ I_{645}^\uparrow$) and $\Re_{1640}$ ($I_{1640}^\uparrow/ I_{645}^\uparrow$).

Table 5.1 summarizes the retrieval uncertainties ($\delta\tau$ and $\delta r_{\text{eff}}$) caused by using combinations C1 and C2, in different conditions with $\theta_0 = 10$ and 36°. Firstly, the resulting uncertainties from a condition with $\theta_0 = 10°$ (high Sun) are analyzed. Note that in the following discussion, the uncertainties is presented in the form of fractional uncertainty (in units of %). By using C1, the retrievals of $\tau$ result in $\delta\tau$ values of 10-14 % while for $r_{\text{eff}}$, the values of $\delta r_{\text{eff}}$ differ between 50 % and 60 %. For C2, the values of $\delta\tau$ are between 9 % and 10 % while
the values of $\delta r_{\text{eff}}$ are in the range of 5-8%. In a condition with $\theta_0 = 36^\circ$ (lower Sun), the values of $\delta \tau$ of C1 are found to be 9-10% while the values of $\delta r_{\text{eff}}$ differ between 49% and 58%. There are no significant differences found in the resulting uncertainties from the two conditions of $\theta_0$. Only for $\theta_0 = 36^\circ$, the resulting uncertainties from C1 are slightly smaller compared to $\theta_0 = 10^\circ$. Therefore, for the cases analyzed here, $\theta_0$ does not play a significant role in changing the retrieval uncertainties. In cases when the solar zenith angle influences the retrieval uncertainties, it is beyond the scope of this study.

Table 5.1: The retrieval uncertainty of $\tau$ ($\delta \tau$) and $r_{\text{eff}}$ ($\delta r_{\text{eff}}$) calculated in two conditions with $\theta_0 = 10^\circ$ and $36^\circ$. The retrievals are run using combinations 1 (C1) and 2 (C2). For calculating $\delta \tau$, $r_{\text{eff}}$ is fixed to 25 µm while for $\delta r_{\text{eff}}$, $\tau$ is set to 3.

<table>
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<tr>
<th>$\theta_0$</th>
<th>$\delta \tau_{\text{C1}}$</th>
<th>$\delta \tau_{\text{C2}}$</th>
<th>$\delta r_{\text{eff},\text{C1}}$</th>
<th>$\delta r_{\text{eff},\text{C2}}$</th>
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<td>2</td>
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<tr>
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<td>0.10</td>
<td>20.13</td>
<td>1.89</td>
</tr>
</tbody>
</table>

*Note that the retrieval uncertainties are calculated for a given 4% measurement uncertainty for $I_{645}$ and 6% for both $R_{1240}$ and $R_{1640}$, according to the uncertainties of the radiance ratio method. It is obvious that using C1 in the retrieval introduces larger $\delta r_{\text{eff}}$ compared to by using C2 because of lowering sensitivity with respect to $r_{\text{eff}}$ at $\lambda = 1240$ nm. As the result, the C1 lookup tables of $r_{\text{eff}}$ are tighter. These large uncertainties (up to 60%) need to be considered when utilizing low absorption wavelengths in the retrieval. Due to the non-orthogonality in the lookup tables, the values of $\delta \tau$ is also influenced by the retrieved $r_{\text{eff}}$. The lookup tables of C1 and C2 tend to tilt to the right (see Fig. 5.3). As a consequence, retrievals will result in larger $\tau$ for larger $r_{\text{eff}}$ and vice versa. This clearly reveals why $\delta \tau$ of C1 is larger than C2.

5.2.4 Impact of underlying liquid water cloud on the cloud phase index

Identifying the cloud thermodynamic phase is a crucial step for the retrieval algorithm. During the ML-cirrus, multilayer cloud situations were frequently observed with liquid water clouds located below cirrus. The condition is more complex because the measurements were performed above the a high absorbing surface (ocean), which adds more characteristics of water absorption in the measured upward radiance. These issues can introduce biases on the resulting $I_p$. To investigate such biases, a sensitivity study is performed based
on radiative transfer simulations assuming pure ice clouds and multilayer clouds with a $\theta_0 = 36^\circ$, a surface albedo of ocean, and a standard atmospheric profile of mid-latitude summer (Anderson et al., 1986). For simulating pure ice clouds, a cirrus layer is placed between 10 and 12 km while for multilayer conditions, a liquid water cloud layer is added below the cirrus at 1.5-2 km altitude. The sensor is located at $z = 13$ km looking downward (nadir-viewing). The setup applied here is in line with the measurements of the cirrus case. In this analysis, $\tau_{ci}$ is varied between 0.01 and 20 while $\tau_{liq}$ is varied between 0.5 and 8. The values of $r_{eff,ci}$ and $r_{eff,liq}$ are fixed to 30 $\mu$m and 10 $\mu$m, respectively. For simplification, the assumption of in-cloud vertically homogeneous is applied.

**Figure 5.8:** Cloud phase index $I_p$ of pure cirrus (black solid line) and when liquid water clouds present below cirrus (dashed lines). Red represents $I_p$ values indicated as a liquid phase (negative) while blue are those indicated as an ice phase (positive).

Fig. 5.8 shows $I_p$ as a function of $\tau$. The black solid line represent the $I_p$ values for pure cirrus (without the presence of low liquid water cloud). The dashed lines show results when liquid water clouds with $\tau_{liq} = 0.5-8$ are present below cirrus with $\tau_{ci} = 0.5, 1, 1.9, 3, 5, \text{ and } 10$ (see the labels). For the multilayer case, $\tau$ is the sum of $\tau_{ci}$ and $\tau_{liq}$. Blue denotes $I_p$ values indicated as an ice phase (positive) while those indicated as a liquid phase (negative) are shown in red. For thin cirrus with $\tau_{ci}$ less than 1.9, $I_p$ always negative, no matter whether a low liquid water cloud is present or not. Here, the information is misleading since the cirrus is identified as a liquid water cloud. This abnormal condition highlights that for optically thin clouds, $I_p$ is largely influenced by the surface. The spectral albedo of ocean gives a similar signature with the absorption given by liquid water droplets imprinted on the upward radiances at $\lambda = 1550$ and 1700 nm. For $\tau_{ci}$ larger than 1.9, the absorption by ice crystals are more dominant minimizing the influence of surface albedo. Compared to pure cirrus, the present of low liquid water cloud will reduce $I_p$. Given that the co-albedo
of liquid water droplet at $\lambda = 1550$ nm is smaller, and thereby less absorbing, the resulting upward radiance at this wavelength is enhanced when there is a liquid water cloud presents below cirrus. Consequently, it will reduce the slope. When the cirrus is sufficiently thick ($\tau_{ci} \geq 10$), the influence of underlying liquid water cloud on the $I_p$ is completely vanished. For simulations with larger $r_{eff,ci}$, the resulting $I_p$ is larger due to stronger absorption by larger ice crystals at $\lambda = 1550$ nm increasing the slope. Thus, the influences of underlying liquid water clouds and surface albedo are reduced when $r_{eff,ci}$ is larger.

![Figure 5.9: Cloud phase index $I_p$ of pure cirrus (blue line) and liquid water clouds (red line). The black solid line indicates where the influence of surface albedo disappears for the $I_p$ of cirrus (at $\tau_{ci} \approx 1.7$).](image)

Fig. 5.9 describes $I_p$ as a function of $\tau$ computed for pure cirrus (blue) and liquid water (red) clouds. For this calculation, $r_{eff}$ of both clouds is fixed to 30 µm. The simulations are performed with $\theta_0 = 36^\circ$ and a spectral surface albedo of ocean. While $I_p$ values of liquid water cloud are always negative, the values for thin cirrus are ambiguous because they lie in the region of liquid phase (negative). For cirrus with $r_{eff,ci} = 30$ µm, the $I_p$ results in negative values for $\tau_{ci}$ less than 1.7, which is reduced to 1.3 for $r_{eff,ci} = 40$ µm (not shown here). Thus, identifying the cloud thermodynamic phase using $I_p$ solely is not enough under this condition, which therefore requires a more advance technique (e.g., Wind et al., 2010). Above these thresholds, the values are constantly positive due to sufficient absorption characteristics given by ice crystals. For considerably thick cirrus with $\tau_{ci}$ larger than 15, the values are nearly steady showing same sensitivities at $\lambda = 1550$ and 1700 nm with increasing $\tau_{ci}$. In addition to ocean, the simulation is also performed by assuming a spectral surface albedo of densely forest, which produces similar results (not shown here). Thus, the results presented here are also valid for measurements above densely forest.
5.3 Vertical photon transport

5.3.1 Modelling vertically inhomogeneous clouds

It is known from measurements, that cloud particle sizes can significantly vary with altitude. For nonprecipitating cirrus clouds, the particle sizes increase rapidly in a very small vertical scale near the cloud base, and then decrease towards the cloud top due to sedimentation processes (e.g., Heymsfield et al., 2017; Wang et al., 2009). For simplifying the retrieval algorithm, however, a vertically homogeneous profile of \( r_{\text{eff}} \) is commonly assumed. Nevertheless, this assumption is somewhat questionable considering the variation throughout the vertical extent, which might influence the retrieved \( r_{\text{eff}} \). To quantify the impacts of such simplification, a simulated measurement is calculated from a realistic profile of \( r_{\text{eff}} \). Then, the resulting upward radiance is applied as the input for a retrieval assuming a homogeneous profile of \( r_{\text{eff}} \). The deviation between the retrieved \( r_{\text{eff}} \) and the original profile \( r_{\text{eff}} \) can be used to signify the impact of such simplification.

An analytical profile of cirrus \( r_{\text{eff}} \) as a function of geometrical height is modeled as following. Firstly, an exponential decrease of \( r_{\text{eff}} \) with altitude is calculated by:

\[
r_{\text{eff}}(z, h) = a_0 - \left( a_1 - a_2 \cdot \frac{z}{h} \right)^{1/k}.
\]

(5.4)

The altitude \( z \) is defined as 0 at the cloud base and increases to \( h \) at the cloud top. The parameters \( a_0 = r_{\text{eff}, t} + r_{\text{eff}, b} \), \( a_1 = r_{\text{eff}, t}^k \), and \( a_2 = r_{\text{eff}, t}^k - r_{\text{eff}, b}^k \) are determined from prescribed boundary condition of the particle effective radius at the cloud top \( r_{\text{eff}, t} \) and at the cloud base \( r_{\text{eff}, b} \). Here, \( k \) is defined as the shape parameter, which typically ranges between 1 and 5. For \( k = 1 \), such as that applied by Wang et al. (2009), \( r_{\text{eff}} \) will decrease linearly towards the cloud top. Increasing \( k \) will result in a profile with a more steep curve. In this study, the value is fixed to \( k = 3 \). For generating the vertical profile \( r_{\text{eff}} \) that more represents cirrus in reality, a geometric transformation is applied to mirror the profile with an exponential decrease over the profile with a linear decrease. A more detailed descriptions and the MATLAB code to generate the profile are given in Appendix A.

The vertical profile of \( r_{\text{eff}} \) is coupled with the vertical profile of ice water content \( IWC \) that is assumed to decrease linearly with altitude. The \( IWC \) profile is generated using Eq. 5.4, assuming \( k = 1 \). Fig. 5.10a and 5.10b show the profiles of \( r_{\text{eff}} \) that represent a cirrus (cloud A) and a DCC composed of ice particles only (cloud B). For the implementation in the forward simulation, the profiles are divided into 20 layers where each layer is assigned to a homogeneous thin \( \tau \) that increases linearly from the cloud top down to the cloud base. All the parameters used to set up clouds A and B are summarized in Table 5.2. For the given total optical thickness \( \tau_c \) of 3 for cloud A and 15 for cloud B, it will result in \( d\tau = 0.15 \) for cloud A and 0.75 for cloud B. The spectral upward radiance above the cloud is
calculated using an adding-superposition technique from the cloud top towards the cloud base, similar to that by Platnick (2000).

Table 5.2: Total optical thickness $\tau_c$, effective radius at the cloud top $r_{\text{eff},t}$ and cloud base $r_{\text{eff},b}$, $IWC$ from the cloud base to the cloud top. $z_b$ and $z_t$ are the altitude of the cloud base and cloud top, respectively. Retrieved $r_{\text{eff},\text{ret}}$ is compared to the weighting-estimate $r_{\text{eff},\text{w}}^*$ for two near-infrared wavelengths at $\lambda = 1240$ nm and 1640 nm.

<table>
<thead>
<tr>
<th>Cloud</th>
<th>$\tau_c$</th>
<th>$r_{\text{eff},b}$</th>
<th>$r_{\text{eff},t}$</th>
<th>$k$</th>
<th>$IWC$</th>
<th>$z_b$</th>
<th>$z_t$</th>
<th>$r_{\text{eff},\text{w}}^*$ (µm)</th>
<th>$r_{\text{eff},\text{ret}}$ (µm)</th>
</tr>
</thead>
<tbody>
<tr>
<td>A</td>
<td>3</td>
<td>40</td>
<td>10</td>
<td>3</td>
<td>0.1 - 0.04</td>
<td>10</td>
<td>12</td>
<td>18.3</td>
<td>17.3</td>
</tr>
<tr>
<td>B</td>
<td>15</td>
<td>50</td>
<td>20</td>
<td>3</td>
<td>0.2 - 0.1</td>
<td>6</td>
<td>8</td>
<td>26.6</td>
<td>24.1</td>
</tr>
</tbody>
</table>

5.3.2 Vertical weighting function

According to Platnick (2000), the vertical weighting function $w_m$ describes the total fraction of reflected photon penetrating into each cloud layer considering multiple scattering. Therefore, it can be used to characterize the cloud level where the retrieved $r_{\text{eff}}$ is most representative. For nadir observation, $w_m$ as a function of $\tau$ is expressed by:

$$w_m(\lambda, \tau, \tau_c, \mu_0, r_{\text{eff}}) = \left| \frac{dI(\lambda, \tau, \mu_0, r_{\text{eff}})}{d\tau} \right| \cdot \frac{1}{\int_0^{\tau_c} \frac{dI(\lambda, \tau, \mu_0, r_{\text{eff}})}{d\tau} d\tau} \cdot \frac{1}{\int_0^{\tau_c} \frac{dI(\lambda, \tau, \mu_0, r_{\text{eff}})}{d\tau} d\tau}.$$ (5.5)

$I$ is the radiance above the cloud and $\tau_c$ is the total cloud optical thickness. Additionally, Platnick (2000) showed that $w_m$ can be used to estimate the retrieved value of effective radius $r_{\text{eff},\text{w}}$ (so-called weighting-estimate) from a given profile of $r_{\text{eff}}(\tau)$ by:

$$r_{\text{eff},\text{w}}^*(\lambda, \tau_c, \mu_0, r_{\text{eff}}) = \int_0^{\tau_c} w_m(\lambda, \tau, \tau_c, \mu_0, r_{\text{eff}}) r_{\text{eff}}(\tau) \, d\tau.$$ (5.6)

The $w_m$ calculated for cloud A and B are shown in Fig. 5.10c and 5.10d, respectively. For cloud A with $\tau_c = 3$, it is found that $w_m$ at $\lambda = 1240$ nm and 1640 nm are almost homogeneously distributed throughout the profile. This exposes that each layer has nearly an equal contribution to the absorption and therefore, to the retrieved $r_{\text{eff}}$. Whereas for cloud B with $\tau_c = 15$, the upper layers contribute most to the absorption, and thereby to the retrieved $r_{\text{eff}}$. The results show that for $\lambda = 1640$ nm, the maximum is found closer to the cloud top, while for $\lambda = 1240$ nm it is located in a deeper layer. This describes that retrievals of $r_{\text{eff}}$ using $\lambda = 1640$ nm will result in $r_{\text{eff}}$ that represent particle sizes located at a higher altitude.
compared to by using $\lambda = 1240$ nm. For the two idealized cloud cases (cloud A and B), this would in general lead to $r_{\text{eff, 1640}} < r_{\text{eff, 1240}}$.

Fig. 5.11a shows the $w_m$ calculated for cloud A at $\lambda = 1000 - 2000$ nm, while Fig. 5.11b is the co-albedo of GHM with $r_{\text{eff}}$ of 10 µm and 15 µm. The co-albedo describes the degree of absorption by cloud particles as a function of wavelength. Increasing co-albedo corresponds to increasing absorption, and vice versa. The result in Fig. 5.11a shows clearly that the $w_m$ at each cloud layer introduces a spectral dependence. At a wavelength with a higher co-albedo, the maximum of $w_m$ is located closer to the cloud top. In contrast, for a wavelength with the co-albedo $\approx 0$, the $w_m$ in the lower layers increases and consequently, the maximum is reduced. The result in Fig. 5.11a also shows that spectral measurements in the near-infrared wavelengths offers more information on the particle sizes located in different cloud altitudes.

It is found, that $w_m$ is a function of the cloud profile itself. Assuming a vertically homogeneous profile in the forward simulation results in different $w_m$ compared to assuming
a realistic profile. This may consequently lead to discrepancies on the $r_{\text{eff}}$ retrieved using both assumptions. With the help of $w_m$, possible impacts are investigated by comparing the weighting-estimate $r_{\text{eff},w}^*$ and the retrieved $r_{\text{eff},\text{ret}}$ at $\lambda = 1240$ nm and 1640 nm. For doing this, the upward radiances above clouds A and B that are calculated for the entire cloud layer $I_{\lambda,\tau,c}^i$ serve as synthetic measurements in the radiance ratio retrieval using combinations C1 (1240 nm) and C2 (1640 nm). The resulting $r_{\text{eff},w}^*$ and $r_{\text{eff},\text{ret}}$ are summarized in Table 5.2. The absolute deviation between $r_{\text{eff},\text{ret},1240}$ and $r_{\text{eff},w,1240}^*$ is 0.4 µm for cloud A and 0.5 µm for cloud B. Between $r_{\text{eff},\text{ret},1640}$ and $r_{\text{eff},w,1640}^*$, the absolute deviation is 0.4 µm for cloud A and 0.1 µm for cloud B. The $r_{\text{eff}}$ retrieved by using measurements at $\lambda = 1640$ nm is consistently smaller than $\lambda = 1240$ nm which agree with a condition where the particle size decreases towards the cloud top.

![Figure 5.11](image)

**Figure 5.11:** (a) The $w_m$ calculated for cloud A at $\lambda = 1000 - 2000$ nm. The color represents the weighting. (b) Single scattering albedo $\tilde{\omega}_0$ of GHM (Baum et al., 2014) with $r_{\text{eff}} = 10$ µm (dashed line) and 15 µm (solid line).

The comparisons between $r_{\text{eff},w}^*$ and $r_{\text{eff},\text{ret}}$ for clouds A and B yield a systematic deviation. It is found, that retrievals using a vertically homogeneous assumption result in a slight underestimation of $r_{\text{eff},\text{ret}}$ compared to $r_{\text{eff},w}^*$ which assumes a realistic profile with decreasing particle size toward the cloud top. For the clouds A and B, larger particles with higher absorption are located in the lower layers. Consequently, $w_m$ at the lower layers is slightly higher while at the upper layers, it is slightly smaller compared to a vertically homogeneous cloud profile (not shown here). Apart from that, the results show that different assumptions in the profile of $r_{\text{eff}}$ yield similar values on the retrieved $r_{\text{eff}}$ since what is required in the retrieval is the total absorption given by the whole layer which is not influenced largely by the shape of the profile. The impact of such assumptions is decreased
when the retrieval is run using wavelengths having higher absorption by cloud particles, such as $\lambda = 1640$ nm.

### 5.3.3 Impact of surface albedo on the vertical weighting function

Fig. 5.12a shows a photo of Amazon surface taken from HALO during a flight. Several surface types are classified in a small scale such as (1) forest, (2) dry-land, (3) water body, and (4) wet-land associated with different surface albedos. It is conceivable that for measurements above this area, the surface albedo will change suddenly in a very small scale along the flight path. Platnick (2000) discussed that changes of the surface albedo $\rho$ will alter the $w_m$ accordingly. When the underlying surface has $\rho = 0$ (black surface), the transmitted radiation will be completely absorbed by the black surface. Conversely, if it is not a black surface ($\rho > 0$), some portions of the transmitted radiation will be reflected back by the surface to the cloud. In this way, the reflected radiation by the surface will change the $w_m$. Therefore, it is important to implement a correct surface albedo in the forward simulation. Otherwise, a bias on the retrieved $r_{\text{eff}}$ might be misinterpreted.

Figure 5.12: (a) A picture of Amazonian surface taken during the AC-18 flight. Four surface types are classified such as (1) forest, (2) dry-land, (3) water body, and (4) wet-land. (b) $w_m$ calculated at $\lambda = 1240$ nm (black) and 1640 nm (red). The dashed lines represent $w_m$ calculated by assuming the spectral surface albedo of forest measured by SMART-Albedometer while the solid lines are by using the MODIS BRDF/Albedo product.

In Sec. 5.1.1, it has been introduced that for simulating the DCC case, the MODIS BRDF/Albedo (MCD43A3) product is applied. According to Strahler et al. (1999), both MODIS Terra and Aqua are used to generate the surface albedo in 500 meter resolution which combines registered, multi-date, multi-band, atmospherically corrected surface reflectance data from the MODIS and the multi-angle imaging spectroradiometer (MISR) instruments to fit a BRDF in seven spectral bands consisting of three visible bands centered at $\lambda = 460$
5. Retrieval of cloud optical thickness and particle effective radius

nm, 555 nm, and 645 nm, and four near-infrared bands centered at \( \lambda = 865 \) nm, 1240 nm, 1640 nm, and 2130 nm. Fig. 5.13 shows the spectral surface albedo derived from the MODIS BRDF/Albedo product \( \rho_{M,\lambda} \) centered at \( \lambda = 645 \) nm (a), 858 nm (b), 555 nm (c), 1240 nm (d), 1640 nm (e), and 2130 nm (f). The black arrows illustrate the flight leg of HALO when measuring the DCC. From Fig. 5.13, it obvious that the DCC was situated above a heterogeneous surface. This justifies that assuming a homogeneous surface along the whole leg is therefore inappropriate.

Changes of the \( w_m \) due to different surface albedo assumptions are investigated. For this purpose, cloud B (see Table 5.2) is chosen to represent a DCC topped by an anvil cirrus. The \( w_m \) is calculated twice by considering two approaches in determining the spectral surface albedo \( \rho_{\lambda} \). For the first approach, the spectral surface albedo of forest measured by SMART-Albedometer \( \rho_{S,\lambda} \) during the ACRIDICON-CHUVA campaign which yields a value of \( \rho_{S,645} = 0.04, \rho_{S,1240} = 0.30, \) and \( \rho_{S,1640} = 0.13 \). As the second approach, \( \rho_{M,\lambda} \) is employed. For the comparison with the first approach, the values of \( \rho_{M,\lambda} \) are averaged along the selected flight leg which results in a value of \( \rho_{M,645} = 0.04, \rho_{M,1240} = 0.14, \) and \( \rho_{M,1640} = 0.08 \). Fig. 5.12b shows \( w_m \) computed at \( \lambda = 1240 \) nm (black) and 1640 nm (red). The dashed lines represent \( w_m \) calculated using \( \rho_{S,\lambda} \) while the solid lines are using \( \rho_{M,\lambda} \). When implementing a higher value of \( \rho_{\lambda} \), it is found in general that the maximum at the upper layers is reduced and shifted to the lower altitude while at the lower layers, it is
more weighted due to the enhanced absorption of the reflected radiation from the surface. Thus, by assuming a higher \( \rho \lambda \), the retrieved \( r_{\text{eff}} \) is more influenced by the absorption at the lower layers. For cloud B with decreasing particle size towards the cloud top, assuming a higher surface albedo will result in a larger retrieved \( r_{\text{eff}} \) than by assuming a smaller surface albedo. The opposite result is expected for clouds, where the particle size decreases toward the cloud top, such as adiabatic liquid water clouds.

By means of the \( w_m \), the differences in the weighting estimate \( r_{\text{eff}}^* \) are calculated using Eq. 5.6. This then allows to quantify how large the discrepancies in the retrieved \( r_{\text{eff}} \) for implementing the two approaches of surface albedo. \( r_{\text{eff}} = 27.5 \, \mu m \) for \( \lambda = 1240 \, nm \) and \( r_{\text{eff}} = 24.2 \, \mu m \) for \( \lambda = 1640 \, nm \) are obtained when implementing the measured surface albedo of forest (SMART-Albedometer). On the other hand, when using the surface albedo from the MODIS BRDF/Albedo product, \( r_{\text{eff}} = 27 \, \mu m \) for \( \lambda = 1240 \, nm \) and \( r_{\text{eff}} = 24.1 \, \mu m \) \( \lambda = 1640 \, nm \) are acquired. According to these findings, the two approaches in determining the surface albedo seemingly do not give a significant impact on the retrieved \( r_{\text{eff}} \) of cloud B. The resulting differences are only 0.5 \( \mu m \) for \( \lambda = 1240 \, nm \) and 0.1 \( \mu m \) for \( \lambda = 1640 \, nm \). Given that the cloud B is optically thick (\( \tau = 15 \)), the radiation is largely attenuated by the cloud itself. Consequently, only small amount of radiation are transmitted to the surface which minimizes the impact of reflected radiation from the surface. The impact is more relevant for thin clouds with \( \tau < 5 \) (not shown here) but it can be minimized by using wavelengths with a higher absorption, such as \( \lambda = 1640 \, nm \).

### 5.3.4 Impact of underlying liquid water cloud on the vertical weighting function

The changes of the \( w_m \) due to the presence of liquid water clouds below clouds A and B are investigated. Therefore, the calculations of \( w_m \) for clouds A and B presented in Section 5.3.2 are repeated by adding a liquid water cloud underneath the cirrus. For cloud A, the liquid water cloud is located between 1.5 - 2 km with \( \tau = 8 \) and \( r_{\text{eff}} = 10 \, \mu m \), which represent a cirrus above a low liquid water cloud. For cloud B, the liquid water cloud is located between 5 and 6 km with \( \tau = 15 \) and \( r_{\text{eff}} = 15 \, \mu m \) representing a DCC topped by an anvil cirrus where the lower core of DCC is composed of a liquid water cloud. For simplification, the profiles of liquid water cloud are assumed to be vertically homogeneous. For comparison, \( w_m \) are calculated and normalized for the ice cloud only. Fig. 5.14a and 5.14b show \( w_m \) at \( \lambda = 1240 \, nm \) (black) and 1640 (red) nm calculated for clouds A and B in a condition with (solid line) and without (dashed line) the presence of the liquid water cloud. Additionally, the co-albedo of GHM (blue lines) and liquid droplets (red lines) with \( r_{\text{eff}} \) of 10 \( \mu m \) (dashed lines) and 15 \( \mu m \) (solid lines) is displayed in Fig. 5.14c.

According to Platnick (2000), it is expected that the low liquid water cloud changes \( w_m \) similar to the impact of a bright surface where the maximum will be reduced and shifted to a lower altitude due to the enhanced reflection of transmitted radiation back to the
cloud base eventually reaching the sensor above cloud top. Consequently, this will result in a larger retrieved $r_{\text{eff}}$ for cirrus clouds with decreasing particle size towards the cloud top. The results in Fig. 5.14a and 5.14b show, that this indeed holds for the $w_m$ at $\lambda = 1240$ nm where scattering by cloud particles dominates. For cloud A and B, the maximum of $w_m$ is shifted to a lower altitude due to multiple reflections of radiation between the liquid water cloud and the ice cloud. However, the $w_m$ at $\lambda = 1640$ nm changes differently when adding a liquid water cloud below the ice cloud. For cloud A, the changes of $w_m$ are significantly larger that are obviously influenced by the the small $\tau_c$. Therefore, large amount of radiation are transmitted by the cirrus. For optically thick cloud B with $\tau_c = 15$, the cirrus does not transmit sufficient radiation to have a strong interaction with the low level cloud which leads to a similar $w_m$.

![Graphs showing $w_m$ for different conditions](image)

**Figure 5.14**: (a) The $w_m$ calculated for cloud A at $\lambda = 1240$ nm and 1640 nm, while (b) is for cloud B. Solid line and dashed line describe $w_m$ calculated with and without the presence of underlying liquid water cloud, respectively. (c) Single scattering albedo $\tilde{\omega}_0$ of GHM and liquid water droplets with $r_{\text{eff}}$ of 10 µm and 15 µm.

For optically thin cloud A with $\tau_c = 3$, $w_m$ at the upper layers is modified due to the reflected radiation from the underlying liquid water cloud. Here the different particle phase and size of the liquid water cloud layer lead to a reduction of the upward radiance when an ice cloud layer is added to the simulations. Given that small liquid droplets have a smaller co-albedo at $\lambda = 1640$ nm, the liquid water cloud alone reflects stronger than together with the ice cloud which adds large ice crystals characterized by a larger co-albedo reducing the total
vertical radiance. Decreasing upward radiance strongly contributes to the $w_m$ close to the cloud top, while at about $\tau = 1$ the minimum of $w_m$ is observed where the upward radiance changes only slightly. Below $\tau = 1$ (lower altitudes), the impact of the liquid water cloud vanishes and scattering by the ice particles increases the upward radiance again corresponding to higher $w_m$ towards cloud base. In general, a similar behavior is imprinted in the $w_m$ of cloud B but not relevant for the entire $w_m$ due to the higher $\tau_c$ of the ice cloud and the higher $\theta_0$. These findings justifies that for optically thick clouds such as the DCC case investigated in this study, a retrieval assuming only ice cloud can be applied to retrieve $r_{\text{eff}}$ of the upper most cloud layer, even if liquid water clouds are present below the ice cloud layer.

5.3.5 Vertical penetration depth

Investigating the vertical penetration depth is important to know from which layer the retrieved $r_{\text{eff}}$ come from and how deep the photons penetrate into the cloud layers. These information are crucial for i.e., a reliable retrieval of particle number concentration, where it requires precise knowledge of $r_{\text{eff}}$ at the cloud top (Painemal and Zuidema, 2011; Wood, 2006). The vertical penetration depth can be evaluated in terms of optical thickness $\tau_w$ and geometrical thickness $h_w$. For $\tau_w$, the computation is more straightforward since $\tau$ is defined as zero at the cloud. In this context, the information on the cloud geometrical altitudes, and thereby the cloud geometrical thickness $h$, can be omitted. To calculate $\tau_w$, the weighting optical thickness $\tau^*_w$ needs to be calculated first by:

$$
\tau^*_w(\lambda, \tau_c, \mu_0, r_{\text{eff}}) = \int_0^{\tau_c} w_m(\lambda, \tau, \tau_c, \mu_0, r_{\text{eff}}) \tau \, d\tau.
$$

The resulting $\tau^*_w$ gives the information on $\tau$, where the retrieved $r_{\text{eff}}$ pertains. It should be located around the peak of $w_m$. Due to $\tau_{\text{top}} = 0$, $\tau^*_w$ can also be identified $\tau_w$. To obtain $h_w$, the weighting altitude $z^*_w$ has to be calculated first by:

$$
z^*_w(\lambda, \tau_c, \mu_0, r_{\text{eff}}) = \int_0^{\tau_c} w_m(\lambda, \tau, \tau_c, \mu_0, r_{\text{eff}}) z(\tau) \, d\tau.
$$

Subsequently, $h_w$ can be inferred by subtracting the cloud top altitude $z_{\text{top}}$ with $z^*_w$ as following:

$$
h_w(\lambda, \tau_c, \mu_0, r_{\text{eff}}) = z_{\text{top}} - z^*_w(\lambda, \tau_c, \mu_0, r_{\text{eff}}).
$$

Due to different absorption characteristics at each wavelength, the penetration depth will spread accordingly. From the result in Fig. 5.11a, it can be seen that for less absorbing wavelengths, the retrieved $r_{\text{eff}}$ is located at a lower altitude because the lower layers
are more weighted than at more absorbing wavelengths. This implies that less absorbing wavelengths have a higher penetration depth, and vice versa. Note that the penetration depth is measured from the cloud top pondering that the sensor and the radiation source are located above the cloud.

Interpreting \( h_w \) should be done carefully because it significantly varies with changes of \( h \). To investigate this issue, two vertical profiles of cirrus are applied for calculating \( h_m \). The two profiles have have same values of \( \tau \) and \( r_{\text{eff}} \) but with different \( h \), as illustrated in Fig. 5.15. Cloud I is located between 10 and 12 km altitudes (\( h = 2 \) km) while cloud II is between 10 and 11 km altitudes (\( h = 1 \) km). Both have \( r_{\text{eff}} = 40 \mu m \) at the cloud base and decreases to 10 \( \mu m \) at the cloud top, similar to cloud A in Fig. 5.10a. The simulations are performed by assuming \( \theta_0 = 36^{\circ} \), the atmospheric profile of "mid-latitude" summer by Anderson et al. (1986), \( \rho = 0 \), and the ice crystal habit of GHM by Baum et al. (2014).

\[ \text{Figure 5.15: The setup of the clouds for quantifying the penetration depths. Cloud I is located between 10 and 12 km altitude while cloud II is between 10 and 11 km altitude. Both represent cirrus with identical optical and microphysical properties. Cloud III represents a liquid water cloud located between 2 and 4 km altitude. Detailed descriptions are given in the text.} \]

Fig. 5.16 shows the resulting \( \tau_w \) (a) and \( h_w \) (b) at \( \lambda = 1000, 1240, 1500, 1640, 2130, \) and 3700 nm for \( \tau_c \) between 2 and 40. All the wavelengths are covered by the SMART-Albedometer, except for \( \lambda = 3700 \) nm that represents the MODIS band 20. The solid lines represents the results calculated using cloud I while the dashed lines are those for cloud II. In Fig. 5.16a, it can be seen that \( \tau_w \) values are nearly identical. Only small differences are observed at \( \lambda = 3700 \) nm for large \( \tau_c \) (> 15). These results shows that the penetration depth is not affected by \( h \). Instead, it is \( \tau_c \) that largely determines the penetration depth. Wavelengths with stronger absorption (e.g., \( \lambda = 1500, 2130, \) and 3700 nm) have a lower penetration depth and vice versa. From the result in Fig. 5.16b, significant discrepancies are observed when the penetration depth is interpreted using \( h_w \). Cloud II apparently has lower \( h_w \) by the factor
of two compared to those for cloud I. Given that the resulting $w_{\text{in}}$ is nearly identical, thus, in this case the discrepancies are caused by $h$ of the two profiles.

$$r_{\text{eff}}(z, h) = \left( a_1 - a_2 \cdot \frac{z}{h} \right)^{1/k}$$  \hspace{1cm} (5.10)

The computations are repeated by assuming a liquid water cloud, as illustrated by cloud III in Fig. 5.15. The liquid water cloud is placed between 2 and 4 km altitudes ($h = 2$ km) with $r_{\text{eff}} = 6 \mu$m at the cloud base that increases exponentially to $18 \mu$m at the cloud top following an adiabatic assumption (Platnick, 2000). The parameterization to build the vertical profile of $r_{\text{eff}}$ is given by Eq. 5.10 \(^a\). The comparison of the penetration depths is shown in Fig. 5.17 where (a) is for $\tau_w$ and (b) is for $h_w$. The solid lines represent the results for cloud I (cirrus) while the dashed lines are for cloud III (liquid water clouds). It is obvious that the penetration depths from both cloud types are distinct. The discrepancies are caused by the different single scattering properties of ice crystals and liquid water droplets. In general, the penetrations depths for liquid water clouds are higher, except for $\lambda = 3700$ nm. The order between $\lambda = 1500$ and 2130 nm is inverted. In cirrus, the penetration depths at $\lambda = 1500$ nm is lower than those at $\lambda = 2130$ nm.

![Figure 5.16](image-url)

**Figure 5.16:** The vertical penetration depth in terms of optical thickness $\tau_w$ (a) and geometrical thickness $h_w$ (b). Solid lines represent those calculated using cloud I ($h = 2$ km) while dashed lines are using cloud II ($h = 1$ km). The simulations are run by assuming $\theta_0 = 36^\circ$, GHM by Baum et al. (2014), and $\rho = 0$ (black surface). Note that the penetration depth is defined from the cloud top.

The interpretation of penetration depths has complexities because they highly vary with $\tau_c$ and $h$. Saying $\tau_w = 1$ can be noted as a high penetration depth if the value of $\tau_c$ is only 2. This means that the radiation interact strongly with 50\% of the $\tau_c$. In turn, for a cloud with $\tau_c = 40$, $\tau_w = 1$ is only a subject of 2.5\% from $\tau_c$. Similarly, saying $h_w = 0.4$ km is

\(^a\)The definitions of all variables applied in Eq. 5.10 are identical with those in Eq. 5.4. Both equations are similar but not the same (there is no $a_0$ in Eq. 5.10)
considerably high for $h = 1$ km as it is equivalent to a 40% of the $h$. For this reason, the penetration depth needs to be normalized with respect to the total cloud thickness, either $\tau_c$ or $h$. The advantage of using this approach is that the results (both $\tau_w$ and $h_w$) will fall within the same order of magnitude because both are calculated based on the same $w_m$.

![Figure 5.17](image)

**Figure 5.17:** Same with Fig. 5.16 but it is calculated using cloud I (solid lines) to represent cirrus and cloud III (dashed lines) for liquid water clouds.

Table 5.3 summarizes the normalized vertical penetration depth in units of % (so-called effective penetration depth $h_{\text{eff}}$) calculated for cirrus and liquid water clouds. In general, $h_{\text{eff}}$ decreases with increasing $\tau_c$. This is due to the fact that the probability of radiation transmitted to the lower layers is smaller for larger $\tau_c$. For six wavelengths analyzed here, $\lambda = 1000$ and 1240 nm seemingly have similar values of $h_{\text{eff}}$, which are about 47% for $\tau_c = 2$ and decrease to between 21% and 29% for $\tau_c = 40$. For cirrus clouds, the values at $\lambda = 1640$ and 2130 nm are similar, which are about 42% for $\tau_c = 2$ and decreases to be about 10% for $\tau_c = 40$. For liquid water cloud, similarities are obtained between $\lambda = 1500$ and 2130 nm with $h_{\text{eff}}$ of about 46% for $\tau_c = 2$ and reduces to 12% for $\tau_c = 40$. Retrievals using wavelengths with similar $h_{\text{eff}}$ do not give significant information on the vertical structure of $r_{\text{eff}}$. The feature of $h_{\text{eff}}$ at $\lambda = 3700$ nm between cirrus and liquid water clouds is more complex. For $\tau_c$ less than 12, $h_{\text{eff}}$ of liquid water clouds is higher but for $\tau_c$ larger than 12, it is lower compared to cirrus clouds. For the two cloud types, $\lambda = 3700$ nm introduces the lowest penetration depth due to very strong absorption by cloud particles. $h_{\text{eff}}$ values for cirrus clouds are within 7% and 30%. Whereas for liquid water clouds, they vary in the range of 4-45% decreasing with $\tau_c$. Thus, for sufficiently thick clouds, retrievals of $r_{\text{eff}}$ using $\lambda = 3700$ nm correspond to particle sizes at the very cloud top.

Given that $w_m$ is also determined by the geometry, changes of $\theta_0$ will consequently alter the vertical penetration depth. To investigate this issue, the calculation of $w_m$ for cloud I (see Fig. 5.15) is repeated by assuming $\theta_0 = 60^\circ$. Fig. 5.18 shows $w_m$ at $\lambda = 1500$, 1640, and 2130 nm calculated using $\theta_0 = 36^\circ$ (solid lines) and $\theta_0 = 60^\circ$ (dashed lines). It is obvious
that for larger $\theta_0$ (lower sun), the weighting at the upper layer increases while at the lower layers, it decreases. It is conceivable that if the sun is low, the slant of incident radiation is larger. In this way, the cloud is more illuminated from the side. Thus, the radiation interact stronger with the upper parts of the cloud reducing the amount of radiation penetrating to the lower layers. This leads to increasing absorption from the upper layers and conversely decreasing absorption from the lower layer. In a condition with larger $\theta_0$, the retrieved $r_{\text{eff}}$ will represent particles sizes at a higher cloud altitude compared to when $\theta_0$ is smaller. Therefore, the vertical penetration depth is lower for larger $\theta_0$ and higher for smaller $\theta_0$. For two values of $\theta_0$ analyzed here, the differences on $h_{\text{eff}}$ at $\lambda = 1500$, 1640, and 2130 nm are obtained of about 4%. 

Figure 5.18: Vertical weighting function $w_m$ at $\lambda = 1500$, 1640, and 2130 nm. Solid lines represent $w_m$ calculated by assuming $\theta_0 = 36^\circ$ while dashed lines use $\theta_0 = 60^\circ$. The applied profile for calculating $w_m$ here stems from cloud I to represent cirrus (see Fig. 5.15).
Table 5.3: Effective penetration depth $h_{eff}$ (in units of %) calculated for cirrus and liquid water clouds in a condition with $\theta_0 = 36^\circ$ and $\rho = 0$ (black surface). For cirrus clouds, GHM by Baum et al. (2014) is assumed. Note that the penetration depth is defined from the cloud top.

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6 Comparison of cloud optical thickness and particle effective radius

This Chapter is related to the third part of the thesis. In Sec. 6.1, the retrieval results based on SMART-Albedometer and MODIS measurements will be compared for the two cloud cases. The comparison also includes the cloud properties derived from the MODIS cloud product that helps to verify the MODIS cloud product algorithm. In order to validate the retrieval, the retrieved particle effective radius will be compared with the in situ data for the cirrus case, which is discussed in Sec. 6.2. Most parts in this Chapter have been published in Krisna et al. (2018).

6.1 Comparison of SMART-Albedometer and MODIS retrievals

Time series of $\tau$ and $r_{\text{eff}}$ retrieved from SMART-Albedometer and MODIS radiance measurements, along with the MODIS cloud product, are compared for the two cloud cases. The MODIS cloud product, namely MYD06_L2, provides three different $r_{\text{eff}}$ (so-called $r_{\text{eff},L,1640}$, $r_{\text{eff},L,2130}$, and $r_{\text{eff},L,3700}$) which are retrieved using respective near-infrared wavelengths centered at $\lambda = 1640$ nm, 2130 nm, and 3700 nm (Platnick et al., 2017). However, the information of $r_{\text{eff},L,1640}$ is very limited due to problems of the detectors and therefore, it cannot be used in this comparison. Due to the similar ice crystal absorption at $\lambda = 1640$ nm and 2130 nm, both wavelengths have an almost identical $w_m$ (Wang et al., 2009; Zhang et al., 2010). For typical cloud profiles analyzed in Sec. 5.3.2, the differences of $r_{\text{eff}}$ retrieved using $\lambda = 1640$ nm and 2130 nm are less than 1 $\mu$m. Therefore, $r_{\text{eff},L,2130}$ can be compared with SMART-Albedometer and MODIS $r_{\text{eff}}$ retrieved using C2 (1640 nm). For observations over land, the MODIS algorithm combines the reflectivity at $\lambda = 645$ nm and 2130 nm (combination 3 - C3). While over ocean, it combines the reflectivity at $\lambda = 858$ nm and 2130 nm (combination 4 - C4).

Time series of cirrus optical thickness and effective radius retrieved using C1, $\tau_{\text{ci},C1}$ and $r_{\text{eff,ci},C1}$, are presented in Fig. 6.1a and 6.1b, respectively. The $\eta$ describes the mean standard deviation of the corresponding cloud properties along the selected time series with the
subscript of "S" for SMART-Albedometer and "M" for MODIS. To quantify the agreement of the retrieved cirrus properties based on SMART-Albedometer and MODIS, the normalized mean absolute deviation $\zeta$ is calculated. A $\zeta_{\tau_{ci,C1}}$ of 1.2% and a $\zeta_{r_{eff,ci,C1}}$ of 0.7% are obtained. Fig. 6.1c and 6.1d show time series of cirrus optical thickness and effective radius retrieved using C2, $\tau_{ci,C2}$ and $r_{eff,ci,C2}$, respectively. A $\zeta_{\tau_{ci,C2}}$ of 0.5% and a $\zeta_{r_{eff,ci,C2}}$ of 2.1% are obtained. The analysis shows, that deviations between SMART-Albedometer and MODIS in the retrieved cloud properties are only slightly enhanced by the non-linearity in the retrieval algorithm. Additionally, cloud properties derived from the MODIS cloud product (blue) are also shown in Fig. 6.1c and 6.1d, where $\eta$ with the subscript of "L" describes the respective mean standard deviation along the selected time series.

**Figure 6.1:** Time series of cirrus $\tau$ (a) and $r_{\text{eff}}$ (b) retrieved from SMART-Albedometer (black) and MODIS (red) using combination 1 (C1). The dark shaded area describes retrieval uncertainties. $\eta_S$ (SMART-Albedometer) and $\eta_M$ (MODIS) represent the mean standard deviation along time series. (c) and (d) are the respective properties retrieved using combination 2 (C2). Cloud properties derived from the MODIS cloud product (MYD06_L2), $\tau_L$ and $r_{\text{eff,L,2130}}$, are shown in blue (only in panel c and d) with the corresponding $\eta_L$.

Cirrus properties retrieved using combinations C1 and C2 are compared to the MODIS cloud product (combination C4). Along the selected time series, all combinations show that $\tau_{ci}$ is homogeneous as indicated by the small standard deviation $\sigma_{\tau_{ci}} < 1$. However,
it is found that $\tau_{\text{ci},L,C4}$ derived from the MODIS cloud product significantly overestimates $\tau_{\text{ci},C2}$ (see Fig. 6.1c). The absolute deviation between the mean value $\tau_{\text{ci},L,C4}$ and $\tau_{\text{ci},C2}$ is found up to 4.7 (160% relative difference). For the MODIS cloud product, the retrieval is always performed with the assumption of a single cloud layer even if a multilayer condition is detected (Platnick et al., 2017). Omitting the low liquid water cloud consequently results in a significant overestimation of the retrieved $\tau_{\text{ci}}$. Including a low liquid water cloud in the radiance ratio retrieval as applied to SMART-Albedometer and MODIS, more realistic $\tau_{\text{ci}}$ are obtained. Furthermore, small differences between $\tau_{\text{ci},C1}$ and $\tau_{\text{ci},C2}$ are found. For a cirrus cloud where the particle size decreases towards the cloud top, it is expected that $r_{\text{eff},C1} > r_{\text{eff},C2}$. Due to the remaining coupling between $\tau$ and $r_{\text{eff}}$ (non-orthogonal radiance lookup tables), these differences propagate into the retrieved $\tau$, and lead to $\tau_{\text{ci},C1} > \tau_{\text{ci},C2}$.

![Figure 6.2](image)

**Figure 6.2**: Same as Fig. 6.1 but for the DCC case.

The results from all approaches show that the mean $r_{\text{eff,ci},C1} > r_{\text{eff,ci},C2} > r_{\text{eff,ci},C4}$. It should be noted, that due to omitting the underlying liquid water cloud $r_{\text{eff,ci},C4}$ underestimates the actual value. The difference between $r_{\text{eff,C1}}$ and $r_{\text{eff,C2}}$ results from the different $w_{\text{in}}$ as discussed in Sec. 5.3.2, which makes $r_{\text{eff,C1}} > r_{\text{eff,C2}}$ for a cirrus with decreasing particle size towards the cloud top. Additionally, the results show that the standard deviation $\sigma_{r_{\text{eff,ci},C1}} > \sigma_{r_{\text{eff,ci},C2}} > \sigma_{r_{\text{eff,ci},C4}}$. This indicates, that the horizontal variability of ice crystals is higher.
Comparison of cloud optical thickness and particle effective radius

in lower cloud layers, while close to the cloud top the ice crystals are distributed more homogeneously along the flight legs. Smaller ice particles with low sedimenting velocity remain at the higher altitudes, while larger ice particles with faster sedimenting velocity drop into the cloud layers below. This sedimentation is horizontally inhomogeneous due to the variability of the vertical wind velocity and leads to a size sorting and the observed horizontal variability of the particle sizes. The analysis shows, that the uncertainty $\delta r_{eff,ci,C1} > \delta r_{eff,ci,C2}$. This confirms, that retrievals of $r_{eff}$ using a wavelength with a smaller absorption by cloud particles will result in a larger uncertainty. Additionally, it is found that increasing $\tau$ and $r_{eff}$ has a positive correlation with increasing $\delta \tau$ and $\delta r_{eff}$, which is due to decreasing sensitivity in the radiance lookup tables for larger $\tau$ and $r_{eff}$.

Time series of DCC optical thickness and effective radius retrieved using $C_1$, $r_{eff,dcc,C1}$ and $\tau_{dcc,C1}$ are shown in Fig. 6.2a and 6.2b, respectively. A $\zeta_{\tau_{dcc,C1}}$ of 1.2 % and a $\zeta_{r_{eff,dcc,C1}}$ of 6.2 % are obtained between SMART-Albedometer and MODIS retrievals. Compared to the cirrus case, the larger horizontal variability indicates a strong evolution of microphysical properties in the deeper layer of DCC. Fig. 6.2c and Fig. 6.2d show time series of DCC optical thickness and effective radius retrieved using $C_2$, $r_{eff,dcc,C2}$ and $\tau_{dcc,C2}$. A $\zeta_{\tau_{dcc,C2}}$ of 3.6 % and a $\zeta_{r_{eff,dcc,C2}}$ of 4.6 % are obtained in this case. In addition to the fast cloud evolution, larger 3-D radiative effects are likely influencing the observations, which can enhance the deviations of retrieved cloud properties. The cloud properties derived from the MODIS cloud product (blue) are also presented in Fig. 6.2c and 6.2d. In this case (over land), the MODIS cloud product algorithm uses $C_3$. The high values of standard deviation $\sigma_{\tau_{dcc}}$ from approach $C_1$, $C_2$, and $C_3$, which are up to 10.1, indicate that $\tau_{dcc}$ is heterogeneous except in the anvil region. The DCC anvil is observed between 17:56:00 - 17:56:20 UTC, which is characterized by relatively smaller $\tau$ between 8 - 15. Later, $\tau_{dcc}$ increases sharply corresponding to the DCC core and decreases again towards the cloud edge. The mean value $\tau_{eff,dcc,C1} > \tau_{eff,dcc,C2}$ indicates decreasing particle size towards the cloud top. It is found, that $\tau_{eff,dcc,C3}$ is larger than $\tau_{eff,dcc,C2}$ corresponding to the different assumptions of the ice crystal habit of plate (SMART-Albedometer and MODIS retrievals) and aggregated columns (MODIS cloud product). Given that $\sigma_{\tau_{eff,dcc,C1}} > \sigma_{\tau_{eff,dcc,C2}}$ and $\sigma_{\tau_{eff,dcc,C2}} < \sigma_{\tau_{eff,dcc,C3}}$, this illustrates that the particle sizes are more homogeneous in the level of $r_{eff,dcc,C2}$ compared to the level of $r_{eff,dcc,C1}$ and $r_{eff,dcc,C3}$.

6.2 Comparison of particle effective radius between retrievals and in situ measurements

The retrieved and in situ $r_{eff}$ are compared for the cirrus case. Here, the terminology of $r_{eff}(z)$ is used to describe the particle effective radius sampled at a specific vertical layer $z$, while the retrieved $r_{eff}$ represents a bulk property of the entire cloud as discussed in Sec. 5.3.2. CCP provides $r_{eff}(z)$ at 1 Hz temporal resolution. Further, the data are averaged to
derive \( r_{\text{eff}}(z) \) with a vertical resolution of 65 m. Fig. 6.3a shows that CCP detected a cirrus between 10.7 and 12 km with the mean values (solid line) ranging between 3 - 30 \( \mu m \). The grey area illustrates the estimated uncertainties of the in situ data. The smallest particles with \( r_{\text{eff}} = 3.1 \mu m \) are found at the cloud base \( z_b = 10.7 \) km and grow rapidly up to 30.2 \( \mu m \) at \( z = 10.8 \) km. Later, \( r_{\text{eff}} \) decreases reaching a value of 8.4 \( \mu m \) at the cloud top \( z_b = 12 \) km.

![Figure 6.3](image)

**Figure 6.3**: (a) Profile of effective radius \( r_{\text{eff}}(z) \) derived from in situ CCP (solid line) with the corresponding uncertainties (grey area). (b) Comparison of the in situ \( r_{\text{eff},w}^* \) and the mean value of \( r_{\text{eff}} \) retrieved from SMART-Albedometer and MODIS using \( \lambda \) between 1240 nm - 3700 nm. Horizontal error bars represent the standard deviation of \( r_{\text{eff}} \), while vertical error bars are the uncertainty of \( z_{w}^* \). (c) Scatter plots between the in situ \( r_{\text{eff},w}^* \) and the mean value of retrieved \( r_{\text{eff}} \). The dashed line is the one-to-one line. The labels at each data point describe the wavelengths used to retrieve the \( r_{\text{eff}} \).

To compare retrieved and in situ \( r_{\text{eff}} \), the vertical weighting function \( w_m \) has to be considered. A direct comparison between \( r_{\text{eff}} \) and \( r_{\text{eff}}(z) \) at a single layer is inappropriate because both are defined differently. Note that the \( w_m \) in this study is calculated in terms of \( \tau \) increasing from the cloud top towards the cloud base. Therefore, the conversion of geometrical altitude and optical thickness \( \tau(z) \) needs to be specified. For this purpose,
IWC(z) measured by WARAN and $r_{\text{eff}}(z)$ derived from CCP are converted into a profile of the extinction coefficient $b_{\text{ext}}(z)$ following the scheme introduced by Fu and Liou (1993) and Wang et al. (2009):

\[ b_{\text{ext}}(z) \approx \text{IWC}(z) \cdot \left[ a + \frac{b}{r_{\text{eff}}(z)} \right], \tag{6.1} \]

where $a = -6.656 \times 10^{-3}$, $b = 3.686$. $b_{\text{ext}}(z)$ is in the unit of $m^{-1}$, IWC(z) in $g \, m^{-3}$, and $r_{\text{eff}}(z)$ in $\mu m$. The extinction profile is used to calculate $\tau(z)$ by integrating $b_{\text{ext}}(z)$ from the cloud top to the altitude level $z$. Using $\tau(z)$, $r_{\text{eff}}(z)$ can be converted into $r_{\text{eff}}(\tau)$. To calculate the $w_m$, the cloud is divided into 20 layers, where each cloud layer is assigned to a $r_{\text{eff}}(\tau)$. Finally, the $r_{\text{eff}}(\tau)$ is convoluted with the $w_m$ to calculate the in situ weighting-estimate $r_{\text{eff},w}$ given by Eq. 5.6 to allow a comparison with the retrieved $r_{\text{eff}}$. Similarly, the weighting altitude $z^*_w$, which characterizes the altitude of corresponding weighting estimate $r_{\text{eff},w}$ and retrieved $r_{\text{eff}}$ is calculated by Eq. 5.8. Due to different degree of absorption at each wavelength in the near-infrared spectrum, $z^*_w$ will vary accordingly, as shown by the results in Sec. 5.3.5. The stronger the absorption by cloud particles in the wavelength, the higher the $z^*_w$ (closer to the cloud top). Conversely, when the absorption is smaller, $z^*_w$ will be located at a lower altitude (closer to the cloud base).

**Table 6.1**: The mean standard deviation of $r_{\text{eff}}$ from in situ (CCP), retrievals (SMART-Albedometer and MODIS), and MODIS cloud product (MYD06_L2) for different near-infrared wavelengths between 1240 nm - 3700 nm. The wavelengths have been sorted in order that the degree of absorption by cloud particles increases to the right.

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The comparison of $r_{\text{eff},w}$ and the mean value of retrieved $r_{\text{eff}}$ is presented in Fig. 6.3b by symbols. Horizontal error bars represent the standard deviation of $r_{\text{eff}}$. Vertical error bars indicate the estimated uncertainty of the $z^*_w$ with a value of 40 m. This value is defined as the standard deviation of $z^*_w$ by varying ice crystal habits in the forward simulations. Additionally, the $r_{\text{eff}}$ retrieved using SMART-Albedometer radiance measurements at $\lambda = 1500$ nm, 1550 nm, and 1700 nm, and also MODIS radiances centered at $\lambda = 2130$ nm, and 3700 nm (band 20) are applied in this comparison. The retrieval and the calculation of $w_m$ for $\lambda = 3700$ nm are performed by considering both solar and thermal radiation. Using these additional wavelengths allows to enhance the vertical resolution of retrieved $r_{\text{eff}}$. 

Fig. 6.3b shows, that in situ $r_{eff,w}^*$ and retrieved $r_{eff}$ agree within the standard deviation for all altitudes and reproduce the decrease of particle size towards the cloud top. However, it is obvious that although retrievals of $r_{eff}$ using multi near-infrared wavelengths result in particle sizes from different cloud altitudes, this retrieval technique only provides information of particle size at the upper layers. This is because the retrieved $r_{eff}$ represents a vertically weighted value, where the upper layers are weighted at most.

Table 6.1 summarizes the mean standard deviation $\eta$ of in situ $r_{eff,w}^*$ and retrieved $r_{eff}$ from SMART-Albedometer and MODIS, and $z_w^*$ for near-infrared wavelengths between 1240 nm - 3700 nm. Additionally, MODIS cloud products (MYD06_L2), $r_{eff,L,2130}$ and $r_{eff,L,3700}$ are included in the table for the comparison. To quantify the agreement between in situ $r_{eff,w}^*$ and retrieved $r_{eff}$, the normalized mean absolute deviation $\zeta$ is calculated. The deviations of in situ $r_{eff,w}^*$ and SMART-Albedometer $r_{eff}$ range between $\zeta = 3.2\%$ ($\lambda = 1500$ nm) and $\zeta = 10.3\%$ ($\lambda = 1550$ nm). Between $r_{eff,w}^*$ and MODIS $r_{eff}$, the $\zeta$ results in a value between 1.5\% for $\lambda = 3700$ nm and 9.1\% for $\lambda = 1640$ nm. Overall, the values of $\zeta$ are in the range between 1.5 – 10.3\% and agree within the horizontal standard deviation, as shown in Fig. 6.3b.

The $r_{eff}$ derived from the MODIS cloud product are obviously affected by the low liquid water cloud, which is not included in the algorithm of MODIS operational retrieval. Therefore, a $\zeta$ of 47.5\% and 19.3\% is obtained for $r_{eff,L,2130}$ and $r_{eff,L,3700}$ respectively. The deviation for $r_{eff,L,2130}$ is smaller because the absorption by the ice crystals at $\lambda = 3700$ nm is very strong. Consequently, the first top layers will dominate the absorption, and therefore reduce the effect of the underlying liquid water cloud. Fig. 6.3c shows scatter plots of in situ $r_{eff,w}^*$ and $r_{eff}$ retrieved from SMART-Albedometer (black triangles) and MODIS (red dots), while the dashed line represents the one-to-one line. There is a good agreement between in situ $r_{eff,w}^*$ and retrieved $r_{eff}$ with a correlation coefficient $R^2$ of 0.82. The variability of particle size distributions, the uncertainties of deriving $r_{eff}$ from the in situ measurements, the presence of liquid water cloud below cirrus, and the uncertainties caused by the choice of ice crystal shapes for the retrievals are considered as the main contributor to address the discrepancies between in situ $r_{eff,w}^*$ and retrieved $r_{eff}$. 
7 Retrieval of the vertical profile of particle effective radius

This Chapter belongs to the fourth part of the thesis. To begin with, the novelty of the new retrieval technique is introduced in Sec. 7.1. This will mainly elaborate limitations of the conventional retrieval technique as that applied in the previous Chapters, and further to motivate of how measurements of passive remote sensing can be used to obtain the full vertical profile of \( r_{eff} \). Sec. 7.2 focuses on the detailed methodology of the new retrieval technique and the analysis of Shannon information content. Additionally, the analysis of the forward model uncertainty is presented. Eventually, the result and validation of the retrieved profile are given in Sec. 7.3.

7.1 Basic ideas

The previous Chapters have shown that using wavelengths with different degree of absorption results in \( r_{eff} \) from different cloud altitudes due to different vertical penetration depths. However, it is unclear whether retrievals of \( r_{eff} \) using spectral measurements are capable to reconstruct the full vertical profile of \( r_{eff} \). Particularly, the results in Fig. 6.3 clearly show that retrievals using multiple wavelengths of SMART-Albedo meter and MODIS with different absorption characteristics only provide information on particle sizes located at the upper layers, close to the cloud top. The spectrally computed \( w_m \) in Fig. 5.11 indicates that deeper in the lower layers, there is seemingly only little information afforded by the spectral measurements. To respond such a question, thus, it is required to plot the weighting estimate \( r^*_{eff} \) obtained from spectral measurements alongside the vertical profile of \( r_{eff} \) assumed in the forward simulation. For this purpose, a typical profile of cirrus (cloud A in Fig. 5.10a) is applied. Detailed specifications of the profile are summarized in Table 5.2. The \( r^*_{eff} \) is calculated using Eq. 5.6 for the spectral range between 1000 and 2000 nm. To illustrate the altitude of the \( r^*_{eff} \), the weighting altitude \( z^*_w \) is utilized, which is calculated by Eq. 5.8.
7. Retrieval of the vertical profile of particle effective radius

Fig. 7.1 shows the resulting $r_{\text{eff}}^*$ (color-coded) calculated for $\lambda = 1000-2000$ nm alongside the original profile of $r_{\text{eff}}$ (black line). Here, the color code also illustrate the degree of absorption with respect to the wavelength. Smaller values of $r_{\text{eff}}^*$ correspond to those calculated using higher absorbing wavelengths (lower penetration depth; higher $z_w^*$). In turn, the lower the absorption at the wavelength (higher penetration depth; lower $z_w^*$), the larger the $r_{\text{eff}}^*$. This is because particle sizes are generally larger at the lower layers. Although this approach is in line with the theory, which also fit well with the original profile, only particle sizes at the upper layers can be obtained. Given that the cloud extends from 10 to 12 km altitude, the resulting $r_{\text{eff}}^*$ only covers particle sizes located at altitudes between 11.0 and 11.4 km (20% of the vertical extent). With respect to $\tau$, this corresponds to $\tau$ between 0.75 and 1.5 for cloud A with $\tau_c = 3$.

![Figure 7.1](https://via.placeholder.com/150)

**Figure 7.1:** The weighting estimate of effective radius $r_{\text{eff}}^*$ computed for $\lambda = 1000-2000$ nm. The black line represents the vertical profile of $r_{\text{eff}}$ assumed in the forward simulation. The color code represents $r_{\text{eff}}^*$ according to the degree of absorption at the wavelength.

From those results, it can be concluded that by using the conventional technique applied to the measurements of reflected solar radiation (passive remote sensing), it is impossible to obtain the full vertical profile of $r_{\text{eff}}$. This is because the peak of $w_m$ is not distributed evenly throughout the cloud vertical extent. To do so, this technique needs to be extended by putting some constraints on the shape of the vertical profile of $r_{\text{eff}}$ with respect to a vertical coordinate (King and Vaughan, 2012). This then allows portions of the profile for which the measurement contains significant information to be used to derive the parameters determining the profile throughout. The important vertical coordinate specifying $r_{\text{eff}}$ is
τ, which is defined from the cloud top. If small contributions such as gaseous absorption, molecular scattering and surface reflection are ignored, the vertical extent and position of a cloud in geometric space make no contribution to the resulting upward radiance.

7.2 Retrieval methodology and information content

7.2.1 Bayesian optimal estimation retrieval

Lookup table approaches may be the most convenient for operational retrievals. However, when more retrieval parameters and wavelengths are introduced, the effort of computing enough lookup tables to retrieve the cloud properties becomes inefficient and impractical. Given these facts, a Bayesian optimal estimation is utilized in the new retrieval technique. This technique has shown an elegant framework in retrieving the cloud profile, which allows to analyze the information content of spectral measurement, as well as to perform post analyses. Using this approach, the output can be directly linked to the retrieval uncertainties. Bayesian optimal estimation uses probability density functions (PDFs) to link the measurement vector space \( y \) to the state vector space \( x \) along with their uncertainties (hereafter bold variables are vectors or matrices, unless otherwise stated). This method allows to find a solution that is most likely to be consistent with the measurements and any given a priori knowledge, within their uncertainties. In the Bayesian theory, the PDF of the state being true retrieved state given a set of measurements (retrievals) corresponds to the PDF of the a priori weighted by the PDF of the measurements knowing the state (forward model):

\[
P(x|y) = \frac{P(y|x) P_a(x)}{P(y)}. \tag{7.1}
\]

The inverse problem is set out as follows; \( y \) is the result of the mapping of \( x \) into measurement space via a forward model \( F(x) \) so that:

\[
y = F(x) + \epsilon, \tag{7.2}
\]

where \( \epsilon \) describes the errors from the measurements and the forward simulation. \( y \) is a vector consisting of the upward radiance measured above the cloud at different wavelengths \( (I_{\lambda_1}^+, I_{\lambda_2}^+, \ldots, I_{\lambda_n}^+) \). \( F(x) \) calculates \( y \) for a given \( x \). In this application, \( x \) is comprised of three cloud parameters to be retrieved: total optical thickness (\( \tau_c \)), particle effective radius at the cloud base (\( r_{\text{eff,b}} \)), and at the cloud top (\( r_{\text{eff,t}} \)). The forward simulation and the setup of cloud vertical profiles have been described in Sec. 5.3.1. The upward radiance is calculated for the entire cloud layer with \( \tau_c \), not for each cloud layer. In the retrieval scheme, Eq. 7.2 should be inverted to infer \( x \) from a given set of measurements \( y \) considering \( \epsilon \). In
order to perform the inverse mapping, however, it is necessary to calculate the Jacobian matrix $K$. It is calculated by perturbing the components of the state vector one-by-one to find the resulting change in the forward simulation. The response in the upward radiance due to the changes of $\tau_c$, $r_{\text{eff},b}$, and $r_{\text{eff},t}$ within the range of expected values is non-linear. Therefore, $K$ is a function of the state vector as follows:

\[
K = \begin{bmatrix}
\frac{\partial I_1}{\partial \tau_c} & \frac{\partial I_1}{\partial r_{\text{eff},b}} & \frac{\partial I_1}{\partial r_{\text{eff},t}} \\
\frac{\partial I_2}{\partial \tau_c} & \frac{\partial I_2}{\partial r_{\text{eff},b}} & \frac{\partial I_2}{\partial r_{\text{eff},t}} \\
\vdots & \vdots & \vdots \\
\frac{\partial I_k}{\partial \tau_c} & \frac{\partial I_k}{\partial r_{\text{eff},b}} & \frac{\partial I_k}{\partial r_{\text{eff},t}}
\end{bmatrix}
\]

where $I_1, I_2, \ldots, I_k$ are the upward radiance measurements and $\tau_c, r_{\text{eff},b}, r_{\text{eff},t}$ are the parameters of the state vector.

The optimal estimation method requires the a priori state $x_a$. This vector corresponds to the prior knowledge of the state vector, such as before the measurements have been performed. In order to take different sources of uncertainties into account, each $x_a$ and $y$ are assigned to the corresponding error covariance matrices, $S_a$ and $S_y$, respectively. Another error covariance matrix that has to be considered is $S_f$. This matrix represents uncertainties attached to intrinsic parameters and non-retrieved parameters assumed in the forward simulation. The two matrices $S_a$ and $S_y$ has to be summed up to form the total error covariance matrix $S_e$. For simplification, an assumption that each component of the measurements and the state vectors are independent from each other, is used (Sourdeval et al., 2013; King and Vaughan, 2012). Consequently, the off diagonal elements of matrices $S_a$, $S_y$, and $S_f$ have null values, as shown by Eq. 7.4 and 7.5.

\[
S_a = \begin{bmatrix}
\sigma_r^2 & 0 & 0 \\
0 & \sigma_{r_{\text{eff},b}}^2 & 0 \\
0 & 0 & \sigma_{r_{\text{eff},t}}^2
\end{bmatrix}
\]
The diagonal components of $S_y$, i.e. $\sigma_{y\lambda}^2$, are obtained from the measurement uncertainties of SMART-Albedometer centered at $\lambda$. Due to a non-linearity in the system, an iterative approach with a first guess $x_0$ is required. Once $x_0$ been specified, the Levenberg-Marquardt iterative formula (Levenberg, 1944; Marquardt, 1963) is used to calculate the next guess $x_{i+1}$. Climatology and model forecast data often provide a convenient a priori $x_a$ but they are not the only source. In this study, the synergetic in situ measurements give advantages to define $x_a$ along with its uncertainties. $x_a$ can be also used as the first guess, but following this approach is not mandatory. The iterative formula is given by:

$$x_{i+1} = x_i + \left[(1 + \gamma)S_a^{-1} + K_i^T S_\epsilon^{-1} K_i\right]^{-1} \{K_i^T S_\epsilon^{-1} [y - F(x_i)] - S_a^{-1} [x_i - x_a]\}$$

(7.6)

where $\gamma$ is defined as a positive damping parameter (e.g., Rodgers, 2000; Fletcher, 1971), which needs to be evaluated in each step according to the changes of the cost function $\chi$. Under the assumption of Gaussian distribution, $\chi$ of the system is given by Eq. 7.7. As a general strategy of retrieval techniques, $\chi$ has to be minimized in order to find the optimal solution. From Eq. 7.7, it can be analyzed that there are contributions of the measurement and the a priori. Given these facts, $\chi$ goes to a minimum value when the forward model is close to the measurement, or when the state introduces a high proximity to the a priori, or it is also possible by the combination of both conditions.

$$\chi = [y - F(x)]^T S_\epsilon^{-1} [y - F(x)] + [x - x_a]^T S_a^{-1} [x - x_a]$$

(7.7)

The convergence analysis is needed to determine the correct criteria to stop the iteration. Rodgers (2000) concluded that it is not necessary to continue the iteration until there is no change in the solution at machine precision. The solution is expected to differ from the true maximum probability state by a quantity, which is noticeably small (smaller than the error of the solution), or the difference between the measurement and the forward simulation can be explained by the uncertainties. Following Iwabuchi et al. (2014) and Sourdeval et al. (2015), the converged solution is obtained when $\chi$ yields a value that is lower than the length of the measurement vector.
Fletcher (1971) found that Marquardt’s strategy had drawbacks. Thus, he proposed a strategy of updating $\gamma$ based on the ratio of the change of $\chi$ to that computed with the linear approximation to the forward model. This ratio will be unity if the linear approximation is satisfactory, and negative if $\chi$ has increased rather than decreased. The aim is to find a value of $\gamma$, which restricts the new value of $x$ to lie within linear range of the previous estimate (so-called trust region). The strategy is summarized as following:

- If the ratio is greater than 0.75, reduce $\gamma$.
- If the ratio is less than 0.25, increase $\gamma$.
- Otherwise make no change.
- If $\gamma$ is less than some critical value, use zero.

The numbers of 0.75 and 0.25 are found by experiments. Fletcher (1971) suggested a factor of two for reducing $\gamma$, and a factor between 2 and 10 for increasing it. By using this update algorithm, on average, less than ten iterations of evaluating the forward simulation and $K$ are sufficient to obtain the optimal solution. In this regard, the optimal estimation retrieval profoundly shows a better efficiency than the lookup table technique since there is no need to calculate the whole combinations of the possible solution. Once the solution has been obtained, the covariance matrix of the solution $\hat{S}$ can be calculated by:

\[
\hat{S} = \left[\hat{K}^T S^{-1} \hat{K} + S_a^{-1}\right]^{-1},
\]

\[
= S_a - S_a\hat{K}^T \left[S_a + \hat{K} S_a \hat{K}^T\right]^{-1} \hat{K} S_a,
\]

where $\hat{K}$ is defined as the Jacobian matrix of the solution state. The square root of the diagonal elements of $\hat{S}$ gives an estimate of the retrieval uncertainties ascribed to each retrieval parameter.

### 7.2.2 Shannon information content and wavelength selection

In a Bayesian approach, the knowledge of the state of the system before a measurement is made is expressed as a prior PDF $P(x)$. After performing the retrieval the knowledge of the state is expressed by a posterior PDF $P(x|y)$. The Shannon information content (Shannon and Weaver, 1949) describes the gain in information from making a measurement, which is given by the difference between the entropy of prior PDF $S[P(x)]$ and the entropy of posterior PDF $S[P(x|y)]$. Based on Rodgers (2000), the information content of a measurement can be evaluated either in the state or in the measurement space. In the basis of the state space, the information content $H$ is expressed by:
\[ H = S [P(x)] - S [P(x|y)], \]
\[ = \frac{1}{2} \ln |S_a| - \frac{1}{2} \ln |\hat{S}|, \]
\[ = \frac{1}{2} \ln |\hat{S}^{-1}S_a|, \] (7.9)

where \( \hat{S} \) represents the posterior error covariance matrix given by Eq. 7.8. With the entropy defined as the natural logarithm of the total number of states, \( H \) provides the information in units of bits, which implies that the measurements allow to distinguish between \( e^H \) states within the prior state space. Substituting \( \hat{S} \) into the second expression of Eq. 7.9 gives the following formula:

\[ H = \frac{1}{2} \ln |(K^T S^{-1} K + S^{-1}_a) S_a|, \]
\[ = \frac{1}{2} \ln |S_a^\frac{1}{2} K^T S^{-1}_e K S_a^\frac{1}{2} + I|, \]
\[ = \frac{1}{2} \ln |\tilde{K}^T \tilde{K} + I|, \] (7.10)

where \( \tilde{K} = S_a^\frac{1}{2} K S_a^\frac{1}{2} \) is defined as the the modified Jacobian matrix and \( I \) represents an identity or singular matrix. To apply Eq. 7.10 to a single wavelength, it is needed to take the row of \( K \) corresponding to the wavelength number so each wavelength has an associated error variance \( \sigma^2 \). Considering that \( \tilde{K}^T \tilde{K} \) has non-zero eigenvalues, Rodgers (2000) rewrote the information content in Eq. 7.10 as:

\[ H = \frac{1}{2} \sum_{i=1}^{n} \ln (1 + \Gamma^2_i), \] (7.11)

where \( \Gamma \) represents the singular value of \( \tilde{K} \). The singular value can be obtained by the singular value decomposition (SVD). It is similar to the eigenvalue decomposition, but can also be applied for non-square matrices. Decomposing \( \tilde{K} \) gives advantages for analyzing wavelengths that bring the most information pertaining to each retrieval parameters (so-called partial information content). It is useful to select the optimal combination of wavelengths in the retrieval. An entropy of zero represents a completely defined solution. Since the information content is a relative measure, a high value corresponds to knowledge gained. Conversely, a low value describes a loss of information. When calculating the information content, please keep in mind that the information gained by making a measurement is in dependence on the definition of the a priori state.
Fig. 7.2 shows the partial information content of $\tau_c$, $r_{\text{eff},b}$, and $r_{\text{eff},t}$ in the logarithmic scale base 10. Here, the information content is calculated above the a priori knowledge of a cirrus cloud with $\tau_c = 3$, $r_{\text{eff},b} = 40 \, \mu m$, and $r_{\text{eff},t} = 10 \, \mu m$. It is computed for wavelengths corresponding to the SMART-Albedometer. To represent the condition during the cirrus case, the forward simulations are performed with $\theta_0$ of 36° and the standard atmospheric profile of mid-latitude summer (Anderson et al., 1986). In addition to the information content, the co-albedo of GHM by Baum et al. (2014) with $r_{\text{eff}} = 10$ and 20 $\mu m$ is shown in Fig. 7.2b. For $\tau_c$, the values of $H$ vary in the range between 0.8 and 1.7 bits, where the peak is found at the scattering wavelengths (co-albedo $\approx 0$). These high values indicate high information for retrieving $\tau_c$. This result is in line with the sensitivity test conducted in Sec. 5.1.3 although the computation was performed by assuming a vertically homogeneous cloud. Thus, the assumption on the vertical profile of $r_{\text{eff}}$ does not give a significant impact in the retrieval of $\tau_c$.

![Figure 7.2](image)

**Figure 7.2:** The partial information content in term of $\tau_c$, $r_{\text{eff},b}$, and $r_{\text{eff},t}$, calculated above the a priori knowledge of a cloud state with $\tau_c = 3$, $r_{\text{eff},b} = 40 \, \mu m$, and $r_{\text{eff},t} = 10 \, \mu m$ (a). The co-albedo of GHM by Baum et al. (2014) with $r_{\text{eff}} = 10$ and 40 $\mu m$ is shown in (b).

Unlike the analysis of $\tau_c$ (already well known), the analysis on wavelengths that provide sufficient information for retrieving $r_{\text{eff},b}$ and $r_{\text{eff},t}$ of ice clouds has never been done so far. In Fig. 7.2a, the values of $H$ for $r_{\text{eff},t}$ differ in the range between 0.02 and 2.1 bits, where the peak is found at the wavelengths around 1500 nm. This is the range, the absorption by cloud particles is considerably high, as shown by enhancing the co-albedo. Thus, they contain most information for retrieving $r_{\text{eff},t}$. In term of $r_{\text{eff},b}$, the analysis is more complex. When looking at the result in Fig. 5.11, we may think that the retrieval of $r_{\text{eff},b}$ requires a wavelength with a minimum co-albedo. This allows photons to penetrate deeper in the
cloud. Nevertheless, the result in Fig. 7.2b proves that this hypothesis is invalid. The values of $H$ for $r_{eff,b}$ are obtained in the range between 0.01 and 1.1 bits, where the maximum is found at the wavelengths around 1550 nm. This is the region where the absorption is small enough, so that a significant portion of photons can penetrate to the lower levels of the cloud, but not too small so that the upward radiance becomes invariant to the changes of $r_{eff,b}$. Hence, it should be kept in mind that the retrieval of $r_{eff,b}$ requires a wavelength with a sufficient absorption, neither too high nor too small.

While it is useful to determine the wavelengths that contain information on different retrieval parameters, it is not known how much information is repeated in the different channels and whether a retrieval combining many wavelengths is advantageous rather than selecting just a few. King and Vaughan (2012) and Wang et al. (2016) have concluded that using many wavelengths reduces the retrieval uncertainties. However, using too many wavelengths can also introduce challenges due to measurement uncertainties ascribed to each wavelength. This may conduce a conflict in the retrieval and result in a spurious solution. Moreover, the enhancement of time expenses has to be considered. Thus, in this study, only a sufficient number of measurements that contain a sufficient information content. Furthermore, measurements in gaseous and molecular absorption, e.g., water vapor, $O_2$, and $CO_2$ must be ignored. In total, five wavelengths are selected according to those criteria. The five wavelengths consist of $\lambda = 645, 1500, 1510, 1550, \text{and } 1560$ nm.

$$\mathcal{R}_{1500} = \frac{I^\uparrow_{1500}}{I^\uparrow_{645}}, \mathcal{R}_{1510} = \frac{I^\uparrow_{1510}}{I^\uparrow_{645}}, \mathcal{R}_{1550} = \frac{I^\uparrow_{1550}}{I^\uparrow_{645}}, \text{and } \mathcal{R}_{1560} = \frac{I^\uparrow_{1560}}{I^\uparrow_{645}} \quad (7.12)$$

The measurement at $\lambda = 645$ nm is selected due to its high sensitivity on $\tau_c$. At $\lambda = 1500$ and 1510 nm, there are large amount of information pertaining to $r_{eff,t}$. While at $\lambda = 1550$ and 1560 nm, there is enhancing sensitivity in term of $r_{eff,b}$. Considering the radiance ratio technique applied in this study, all the measurements at $\lambda = 1500, 1510, 1550, \text{and } 1560$ nm are normalized to the measurement at $\lambda = 645$ nm, as shown by Eq. 7.12. In the retrieval, $I^\uparrow_{645}$ is used most to retrieve $\tau_c$, $\mathcal{R}_{1500}$ and $\mathcal{R}_{1510}$ are for $r_{eff,t}$, whereas $\mathcal{R}_{1550}$ and $\mathcal{R}_{1560}$ are for $r_{eff,b}$.

### 7.2.3 Estimation of the forward model uncertainties

The error covariance matrix in Eq. 7.5 denotes that the uncertainty of the measurement state space $S_e$ does not only account the measurement uncertainty, but also should consider the uncertainty of the forward simulation. Surprisingly, L’Ecuyer et al. (2006) and Coddington et al. (2012) found that the contribution of the uncertainty of the forward simulation to $S_e$ is larger than the measurement uncertainty itself. In this study, parameters pondered to contribute to the uncertainty of the forward simulation are summarized in Table 7.1.
Table 7.1: Parameters and data sources that are pondered for estimating the uncertainty of the forward simulation. The molecular absorption and extraterrestrial radiation represent the intrinsic parameters, while the others represent the non-retrieved parameters.

<table>
<thead>
<tr>
<th>Parameter</th>
<th>Data source</th>
</tr>
</thead>
<tbody>
<tr>
<td>Molecular absorption</td>
<td>(1) Retran coarse, (2) Retran medium, (3) Retran fine by Gasteiger et al. (2014), and (4) Lowtran by Ricchiazzi et al. (1998).</td>
</tr>
<tr>
<td>Surface albedo</td>
<td>Five times measured in-flight by the SMART-Albedometer.</td>
</tr>
<tr>
<td>Surface wind speed</td>
<td>Three times of surface wind speed were obtained from dropsonde devices released around the location of the cirrus: (1) 3 m s(^{-1}), (2) 5 m s(^{-1}), and (3) 7 m s(^{-1}).</td>
</tr>
<tr>
<td>Atmospheric profile</td>
<td>Three times of atmospheric profiles (temperature and humidity) were acquired from the dropsonde devices, which are adjusted to the standard atmospheric profile of mid-latitude summer by Anderson et al. (1986).</td>
</tr>
<tr>
<td>Cloud geometrical altitude</td>
<td>The cloud top is varied between 11 and 12 km. While for the cloud base, the altitude is set between 9 and 10 km according to the measurements of WALES.</td>
</tr>
<tr>
<td>Aerosol haze</td>
<td>(1) Maritime aerosol and (2) Rural aerosol by Shettle (1989).</td>
</tr>
</tbody>
</table>

For estimating the uncertainty, forward simulations are performed by perturbing each parameter with different data sources enlisted in Table 7.1 to calculate the spectral upward radiances. For this purpose, a typical profile of cirrus (cloud A in Fig. 5.10a) is applied. Data sources of the intrinsic parameters (molecular absorption and extraterrestrial spectral irradiance) and the aerosol haze are directly taken from the LibRadtran library (Mayer, 2005; Emde et al., 2016). Data of atmospheric, relative humidity, and wind speed profiles are obtained from the dropsonde devices that were released three times in the area close to the location of the cirrus. Furthermore, the profiles of temperature and relative humidity are adjusted to the standard atmospheric profile of mid-latitude summer (Anderson et al., 1986). The information of surface wind speed are inferred from the measured profiles of wind speed at the sea level altitude. Data of cloud geometrical altitudes (cloud top and base) are provided by the water vapor and lidar experiment in space (WALES) aboard of HALO (Wirth et al., 2009). Along the cirrus section, the cloud top altitudes were located between 11 and 12 km altitude, while the cloud base altitudes were between 9 and 10 km altitude. The surface albedo was measured five times by the SMART-Albedometer using a technique described in Wendisch et al. (2004).
7.2. Retrieval methodology and information content

Figure 7.3: The standard deviation of upward radiance in the spectral range between 400 and 2000 nm. The calculation is carried out by perturbing each parameter with different data sources enlisted in Table 7.1.

Fig. 7.3 shows the resulting standard deviation $\sigma$ of spectral upward radiances. Overall, the values of $\sigma$ are in the range between $10^{-6}$ and $10^{-2}$ W m$^{-2}$ nm$^{-1}$ sr$^{-1}$, indicating a considerably spectral dependence. At a wavelength that is sensitive to a specific parameter, the value will increase correspondingly, and vice versa. In this study, the approach of L’Ecuyer et al. (2006) is adapted to compute the uncertainty to define $S_\sigma$. Firstly, the fractional uncertainty $\sigma_f$ (%) has to be calculated from the ratio between the standard deviation and the mean value of upward radiance. Furthermore, the resulting $\sigma_f$ from each of the seven parameters are combined to calculate the total effective fractional uncertainty $\sigma_t$ (%) at each selected wavelength. If it is assumed that each source is uncorrelated, the combined uncertainty from all sources is given by the square root of the sum of the squares of each component as follows:

$$\sigma_t = \sqrt{\sum_{i=1}^{n} \sigma_{f,i}^2}.$$ 

(7.13)

In Eq. 7.13, the index $i$ expresses the number of parameters involved (in this study, $i = 7$). Table 7.2 summarizes the values of $\sigma_f$ and $\sigma_t$ calculated at $\lambda = 645$ nm, 1500 nm, 1510 nm, 1550 nm, and 1560 nm, according to the wavelengths applied in the retrieval. The values of $\sigma_f$ vary between 0.1 and 2.8 % depending on the parameter and the wavelength. In general, the uncertainties in the visible range is smaller than those in the near-infrared range, in agreement with L’Ecuyer et al. (2006). From the seven analyzed parameters, it is found that changes on the cloud geometrical altitudes only result in small $\sigma_f$ that range...
between 0.1 and 0.2%. This means that the assumption of the cloud geometrical altitudes are not sensitive to the upward radiances at the five wavelengths, and therefore, they do not contribute largely to the retrieval uncertainties. Apart from that, it is surprising that the intrinsic parameters introduce the largest uncertainties. The maximum value of \( \sigma_t \) in term of the molecular absorption is found to be 2.8% at \( \lambda = 1500 \) nm, while for the extraterrestrial spectral irradiance is about 1.2% at \( \lambda = 645 \) nm. Given that they contribute most to the total uncertainty, the intrinsic parameters has to be selected carefully.

**Table 7.2:** The fractional uncertainty \( \sigma_t \) due to intrinsic and non-retrieved parameters assumed in the forward simulation. The last line represents the total effective fractional uncertainty \( \sigma_t \).

<table>
<thead>
<tr>
<th>Parameter</th>
<th>Fractional uncertainty ( \sigma_t ) (%)</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>645 nm</td>
</tr>
<tr>
<td>Molecular absorption</td>
<td>0.19</td>
</tr>
<tr>
<td>Extraterrestrial irradiance</td>
<td>1.22</td>
</tr>
<tr>
<td>Surface albedo</td>
<td>0.89</td>
</tr>
<tr>
<td>Surface wind speed</td>
<td>0.61</td>
</tr>
<tr>
<td>Atmospheric profile</td>
<td>0.20</td>
</tr>
<tr>
<td>Cloud geometrical altitude</td>
<td>0.04</td>
</tr>
<tr>
<td>Aerosol haze</td>
<td>0.08</td>
</tr>
<tr>
<td>Total uncertainty ( \sigma_t ) (%)</td>
<td>1.66</td>
</tr>
</tbody>
</table>

The results in Table 7.2 imply that the resulting uncertainty has a spectral dependence. The largest uncertainty at \( \lambda = 645 \) nm is caused by the extraterrestrial irradiance. The molecular absorption contributes most to the uncertainty at \( \lambda = 1500 \) nm and 1510 nm, while the surface albedo results in the highest uncertainty at \( \lambda = 1550 \) nm and 1560 nm. This findings also emphasize that assuming a spectrally flat uncertainty is inappropriate because the uncertainty significantly varies with wavelengths depending. By applying Eq. 7.13 to calculate \( \sigma_t \), values of 1.7% for \( \lambda = 645 \) nm, 3.7% for \( \lambda = 1500 \) nm, 3.3% for \( \lambda = 1510 \) nm, 3.2% for \( \lambda = 1550 \) nm, and 2.8% for \( \lambda = 1560 \) nm are obtained. Each of those values are further summed up with the measurement uncertainties in order to determine the error covariance matrix \( S_e \) given by Eq. 7.5.

### 7.3 Result and validation

The cirrus case analyzed here stems from that analyzed in Sec. 6.1. However, the time series are extended from 13:55:30 to 13:58:00 UTC to analyze the full section of the cirrus.
This also helps to assess the capability of the new retrieval technique, particularly at the cloud edges, where $\tau_c$ is typically thinner. In the forward simulations, the geometrical cloud altitudes (cloud top and base) have to be defined. In this case, these information are provided by the measurements of WALES, which agree nicely with those measured by the in situ measurements. The new retrieval scheme is applied to the measurements of SMART-Albedometer to obtain $\tau_c$, $r_{\text{eff},b}$, and $r_{\text{eff},t}$. Once those three parameters have been obtained, the vertical profile can be reconstructed by the parameterization technique described in Sec. 5.3.1. Please note that in this study, the shape parameter $k$ is fixed to 3. This value is determined according to the observation.

![Figure 7.4: Time series of the vertical profile of $r_{\text{eff}}$ as a function of $\tau$. The color code indicates the magnitude of $r_{\text{eff}}$. There are white areas indexed as I, II, and III, which represent a cloud-free region, dark current measurement, and retrieval failure, respectively.](image)

Fig. 7.4 shows the time series of the the profile of $r_{\text{eff}}$ as a function of $\tau$. Note that $\tau$ is defined as 0 at the cloud top and increases linearly toward the cloud base. Along the analyzed cloud section, the values of $\tau_c$ vary from 1.2 to 5, yielding very similar results with the retrieval assuming a vertically homogeneous cloud, as shown in Sec. 6.1. The values of $r_{\text{eff},t}$ at the cloud top are relatively homogeneous that range between 4.7 and 14 $\mu$m. At the cloud base, the values of $r_{\text{eff},b}$ are considerably heterogeneous (between 15 and 40 $\mu$m). In general, the $r_{\text{eff}}$ shows a sharp increase at the cloud base (only in a very small scale) and then, it monotonically decreases toward the cloud top. Obviously, this feature is governed by the parameterization applied to obtain the full vertical profile.

There are several white areas indexed as "I", "II", and "III". The index 'I' represents when no clouds were measured or when the value of $\tau_c$ is less than 1.2. This is the number set in the retrieval algorithm as the restriction of $\tau_c$ for being retrieved. Considering the underlying liquid water cloud, the retrieval uncertainties my significantly enhanced, particularly for
retrieval of the vertical profile of particle effective radius \( r_{\text{eff,b}} \), when the cirrus is not sufficiently thick. Therefore, the results for \( \tau_c \) less than 1.2 are discarded in the analysis. The index "II" describes when the shutter was closed for dark current measurements. Those are required for the in-flight sensor calibration.

**Figure 7.5:** The comparison of the vertical profile of \( r_{\text{eff}} \) measured by the in situ CCP (black) and retrieved from SMART-Albedometer measurements (red). The shaded areas indicate the corresponding uncertainties. For the retrieved profile, the original uncertainties are obtained at the cloud top and the cloud base. The values in between are linearly interpolated.

The index "III" denotes a retrieval failure that can be classified as either a total or a partial failure. In this application, the retrieval can introduce a total failure if all components of the measurement state space can not be reconciled by the forward simulation considering the uncertainties. This kind of failure is easier to be recognized. Also, it is common in cases of the conventional retrieval. Apart from that, the retrieval can result in a partial failure if the three parameters are retrieved incompletely. For this application, sometimes \( r_{\text{eff,b}} \) can not be retrieved because the horizontal variability of \( r_{\text{eff}} \) at the lower layers are too high. Thus, the forward simulation given the assumed vertical profile can not reproduce the similar magnitude of measured upward radiance due to the variability of \( r_{\text{eff}} \) at the lower layers. Nevertheless, in this particular case, the retrieved \( \tau_c \) and \( r_{\text{eff,t}} \) are still valid. The retrieval of \( \tau_c \) is weakly influenced by the profile of \( r_{\text{eff}} \). Whereas, due to strong absorption at \( \lambda = 1500 \text{ nm} \) and \( 1510 \text{ nm} \), only the first layers of the cloud contribute most to the absorption that are already sufficient for inferring \( r_{\text{eff,t}} \). The upper layers of cirrus are typically composed of small and more homogeneous particles that fit much better to the assumed profile. However, due to size sorting and sedimentation processes in cirrus, larger
7.3. Result and validation

particles that are less buoyant drop resulting in mixture particle sizes in the lower layers. When the horizontal variability is too high, it is very likely that the partial failure can happen.

For the validation, the retrieved profile is averaged over the same domain, as that analyzed in Sec. 6.1 (from 13:56:20 to 13:57:35 UTC). To allow this comparison, the retrieved profile at each measurement point is normalized to the geometrical altitudes measured by the WALES. Fig. 7.5 shows the comparison of the retrieved (red line) and the in situ (black line) profiles of \( r_{\text{eff}} \). The shaded areas described the respective uncertainties. For the retrieved profile, the uncertainties are originally given by the uncertainty of \( r_{\text{eff},t} \) at the cloud top and \( r_{\text{eff},b} \) at the cloud base only, while the values in between are interpolated linearly. From the result, it can be seen that the uncertainty toward the cloud base becomes larger which is attributed to decreasing sensitivity on the retrieval of \( r_{\text{eff},b} \). A sensitivity study based on the information content clearly shows a loss of information, when it is calculated for a cloud state with higher \( \tau_c \). Sourdeval et al. (2013) used a limit of 0.5 to analyze measurements that bring enough information on parameters to be retrieved. For the case analyzed in this study, this corresponds to \( \tau_c = 10 \) that is defined as the limit of \( \tau_c \) allowing reliable retrievals of the vertical profile of \( r_{\text{eff}} \). This number may differ for liquid water clouds due to differences in the single scattering properties between ice crystals and liquid water droplets.

The result in Fig. 7.5 confirms the applicability of the new retrieval technique to reconstruct the vertical profile of \( r_{\text{eff}} \). The retrieved profile also reproduces the typical feature of particles sizes in cirrus. The comparison with the in situ profile yields a considerably good agreement, within the uncertainties. The relative deviation between the mean value of both profiles is found to be about 5\% at the cloud top. It increases toward the cloud base with the values of up to 15\%. The deviation between the retrieved and the measured profile is caused by the measurement uncertainties, the choice of ice crystal habit in the forward simulation, the variability of the size distributions, and the possible impacts of the underlying liquid water cloud below the cirrus. Given the fact the the \( k \) parameter determining the shape of the profile is fixed in this application, putting \( k \) as an independent parameter to be retrieved along with the other three parameters will help to improve the result. It is conceivable that by increasing \( k \), the particle sizes at the lower layers close to the cloud base will increase. As the consequence, this has to be compensated by smaller particle sizes at the upper layers. Therefore, a better agreement with the in situ data is expected. Overall, the result from the new retrieval technique significantly improve the result from the conventional technique, which only provides \( r_{\text{eff}} \) located at the upper layers. Using the extended technique, the full vertical profile of \( r_{\text{eff}} \) can be obtained, which fits well with the profile measured by the CCP.
8 Summary and conclusion

Within this PhD thesis, optical thickness $\tau$ and particle effective radius $r_{\text{eff}}$ of cirrus and deep convective clouds (DCCs) are investigated using measurements from airborne and satellite passive remote sensing. Different wavelength combinations are employed to obtain the cloud properties and further to study the vertical structure of $r_{\text{eff}}$. For validation, retrieved values of $r_{\text{eff}}$ are compared with concurrent airborne in situ measurements for the cirrus case. Numerous sensitivity studies based on radiative transfer simulations have been carried out to study the retrieval uncertainties, the vertical photon transport in the cloud, the feasibility of retrieving the vertical profile of $r_{\text{eff}}$ based on the measurements of passive remote sensing. In the end, a new retrieval technique based on measurements of passive remote sensing is developed to obtain the full vertical profile of $r_{\text{eff}}$. The main conclusions of these studies will be summarized within the next sections.

8.1 Airborne campaigns

The ML-CIRRUS and the ACRIDICON-CHUVA campaigns have been conducted to study cirrus clouds over Europe and tropical DCCs over the Amazon rainforest using HALO research aircraft. Within these campaigns, HALO was equipped with a comprehensive set of remote sensing and in situ instruments (Wendisch et al., 2016; Voigt et al., 2017). Spectral upward radiances reflected by cirrus and DCCs were measured by the SMART-Albedometer, while in situ cloud probes were employed to sample microphysical properties throughout the vertical extent of the clouds. In particular flights, HALO performed measurements above clouds, which were closely collocated with overpasses of the MODIS-Aqua satellite. This unique setup was applied in order to validate the quality of primary remote sensing measurements and further to assess the retrieval algorithm and products.

8.2 Comparison of upward radiance

Accurate solar radiation measurements are necessary to retrieve high-quality cloud products such as the cloud optical thickness $\tau$ and particle effective radius $r_{\text{eff}}$. Small measurement uncertainties propagate in the retrieval and amplify the retrieval uncertainties.
To assess the measurement quality, the spectral upward radiances measured by SMART-Albedometer and MODIS are compared for two cloud cases, a cirrus above liquid water clouds and a DCC topped by an anvil cirrus. Several filters have been applied to select these two cloud cases. To minimize the effects caused by the cloud evolution, the time shift between SMART-Albedometer and MODIS is limited. The analysis for cirrus clouds shows that only measurements within a time shift of 500 s can be directly compared. For DCCs, the time shift is strictly restricted to 300 s pondering the fast cloud evolution. Only straight flight legs with altitude changes of less than 50 m and pitch and roll angles less than 3° are considered in the analysis. Further, measurements at the cloud edges are discarded to avoid biases due to sharp changes of upward radiance and higher 3-D radiative effects. For the cirrus case, the upward radiance comparisons yield values of normalized mean absolute deviation between 0.2 % and 7.7 %, while for the DCC case, the values lie between 1.5 % and 8.3 %. The deviations are generally larger for the DCC case due to remaining influences from the cloud evolution. Additionally, stronger 3-D radiative effects should be considered in this case. It is obtained that the upward radiance at $\lambda = 1240$ nm measured by MODIS is systematically larger by 15 % compared to that measured by SMART-Albedometer. Because of issues on the detectors of MODIS band 6, the upward radiance of MODIS at $\lambda = 1640$ nm is retrieved based on the measurement of MODIS band 7 ($\lambda = 2130$ nm). For ice clouds, this approach can be applied assuming that ice crystals have similar optical properties at the two bands resulting similar characteristics on the reflected radiation. The applicability of the retrieval has been tested by comparing the retrieved values with the measurements extracted from the remaining detectors of MODIS band 6.

8.3 Retrieval of cloud optical thickness and particle effective radius

Due to the multilayer condition, a liquid water cloud layer is considered in the forward simulation of the cirrus case. The properties of the liquid water cloud are estimated by comparing measured and simulated spectral upward radiances, particularly in the absorption bands of water vapor and oxygen-A. An optimal combination of cirrus and liquid $\tau$ results in a good fit in the absorption bands. Assuming an overestimated value of liquid $\tau$, thereby an underestimated value of cirrus $\tau$, reduces the upward radiance in the absorption bands because a larger amount of radiation is transmitted by the cirrus and further absorbed by water vapor and oxygen-A below the cirrus. An opposite condition is obtained when the liquid $\tau$ is underestimated, thus the cirrus $\tau$ is overestimated. However, please note that the estimation of liquid $\tau$ requires a precise knowledge on the geometrical altitudes (cloud top and base). Similarly, wavelengths with high absorption by cloud particles are applied to estimate the liquid $r_{\text{eff}}$. 
8.3. Retrieval of cloud optical thickness and particle effective radius

In the retrieval, the cloud phase index \( I_p \) is applied to distinguish the cloud thermodynamic phase. The retrieval can be run either in the ice or liquid water mode, depending on the information given by \( I_p \). Due to different absorption characteristics between ice and liquid water particles at \( \lambda = 1550 \) and \( 1700 \) nm, measurements of upward radiances yield a positive slope (\( I_p > 0 \)) for ice clouds and a negative slope (\( I_p < 0 \)) for liquid water clouds. However, the presence of low liquid water clouds might bias the resulting \( I_p \) if the cirrus is not sufficiently thick (\( \tau < 2 \)), or when the measurements take place over an ocean surface. In these conditions, the information is misleading since \( I_p \) will result in a negative value, indicating a liquid water cloud.

Based on the measurements of upward radiance, a radiance ratio retrieval is applied to retrieve \( \tau \) and \( r_{\text{eff}} \). Two combinations, \( C1 (I_{645}^\uparrow \) and \( \Re_{1240} = I_{1240}^\uparrow / I_{645}^\uparrow \) ) and \( C2 (I_{645}^\uparrow \) and \( \Re_{1640} = I_{1640}^\uparrow / I_{645}^\uparrow \) ), are applied in the retrieval algorithm. Due to different particle absorption at \( \lambda = 1240 \) and \( 1640 \) nm, the retrievals will result in \( r_{\text{eff}} \) from different cloud altitudes. Analyses on the vertical weighting function have shown that for \( \lambda = 1240 \) nm, the lower cloud layers are more weighted compared to those for \( \lambda = 1640 \) nm. In this way, retrievals using \( C1 (\lambda = 1240 \) nm) will result in \( r_{\text{eff}} \) from a lower altitude. Because the upper layers are more weighted, retrievals using \( C2 (\lambda = 1640 \) nm) produce \( r_{\text{eff}} \) from a higher altitude. Given that cirrus particle sizes generally decrease with increasing altitude, retrievals of \( r_{\text{eff}} \) using \( C1 \) result in larger values of \( r_{\text{eff}} \) than using \( C2 \). For liquid water clouds with increasing particle sizes toward the cloud top, an opposite result is expected. To some degree, retrievals using those two combinations give a snapshot of the vertical variation of \( r_{\text{eff}} \) in the cloud. The vertical weighting function clearly shows that each cloud layer contributes the absorption imprinted in the upward radiance, where the weighting depends on the cloud profile itself and the applied wavelength. Thus, it must be kept in mind that \( r_{\text{eff}} \) retrieved using this technique does not represent a particle size at a single cloud layer. Instead, it represents a bulk property of the entire cloud layer.

Possible biases resulting from the assumption of the vertical profile of \( r_{\text{eff}} \) in the retrieval is investigated. A systematic deviation is found between assuming a vertically homogeneous and a realistic cloud profiles in the retrieval. For ice clouds with decreasing particle sizes towards the cloud top, retrievals assuming a vertically homogeneous cloud result in an underestimation of \( r_{\text{eff}} \) by 1 µm. The deviation increases when the retrieval is performed using a less absorbing wavelength (e.g., \( \lambda = 1240 \) nm) because the lower layers contribute more strongly to the absorption. For a homogeneous profile, the weighting at the lower layers is smaller, while it is larger at the upper layers. However, for this type of retrieval, the assumption on the vertical profile of \( r_{\text{eff}} \) does not give a significant impact, since both profiles produce a similar total absorption at the applied wavelength.

With helps of the vertical weighting function, the vertical penetration depth of the radiation scattered into the view of the remote sensing instruments is quantified. Here, the penetration depth is defined as the distance between the cloud top and the location of weighting estimate of particle effective radius \( r_{\text{eff}}^* \). The penetration depth is definable two
terms, either by the optical thickness $\tau_w$ or the geometrical thickness $h_w$. The penetration depth at $\lambda = 1000, 1240, 1500, 1550, 1640, 2130, and 3700$ nm is analyzed. While it largely depends on $\tau$, only at $\lambda = 3700$, the influence of cloud geometrical thickness $h$ slightly enhances. For the other wavelengths, changes of $h$ do not significantly alter the vertical weighting function, as long as the profile of $\tau$ remains unchanged. The penetration depth decreases with increasing $\tau$. For large $\tau$, the weighting at the upper layers is strengthened, reducing the amount of radiation transmitted to the lower layers reduces. Thus, in this condition, the penetration depth is smaller. Compared to ice clouds, the penetration depth in liquid water clouds is generally higher due to smaller absorption given by liquid water droplets at the same particle size. Only at $\lambda = 3700$ nm, the penetration depth of liquid water clouds is lower due to higher absorption by liquid water droplets. In addition to $\tau$, the penetration depth is highly affected by the solar zenith angle $\theta_0$. Increasing $\theta_0$ results in a lower penetration depth since the photons interact more strongly with the upper layers, reducing the contributions of the lower layers to the absorption.

Retrieval uncertainties due to the occurrence of multilayer condition and the assumptions of ice crystal habit and surface albedo are quantified. The presence of liquid water clouds below cirrus leads to an overestimation of the retrieved cirrus $\tau$, when the low cloud is not considered in the forward simulation. This is because the upward radiance measured by the sensor located above cirrus is presumed to be reflected by the cirrus only, neglecting the role of the underlying cloud. An overestimation of liquid $\tau$ will artificially decrease the retrieved cirrus $\tau$ because the low liquid cloud contributes largely to the upward radiance, and vice versa. Given these facts, the liquid $\tau$ must be determined correctly since a wrong assumption almost directly propagates to the retrieved cirrus $\tau$. The presence of low liquid water clouds also influences the retrieved cirrus $r_{\text{eff}}$, particularly when the cirrus layer is not sufficiently thick ($\tau < 5$). In this condition, multiple reflections between the two clouds alter the vertical weighting function at the lower layers of the cirrus. Consequently, for cirrus with decreasing particle size towards the cloud top, the retrieved $r_{\text{eff}}$ becomes larger. The impact vanishes when the cirrus layer is sufficiently thick with $\tau > 5$.

The influence of surface albedo $\rho$ on the retrieval of $r_{\text{eff}}$ is analyzed by the changes of the vertical weighting function. It is found that its influence is similar with that given by the underlying liquid water cloud. Assuming a higher value of $\rho$ enhances the weighting at the lower layers due to the multiple reflections between the surface and the cloud. For higher $\rho$, the retrieved $r_{\text{eff}}$ is located at a lower altitude, and vice versa for lower $\rho$. Apart from that, this study has shown that the impact of ice crystal habit on the retrieval uncertainties is considerably high. In general, assuming a habit with a lower asymmetry parameter $g$ results in a smaller $\tau$ and a larger $r_{\text{eff}}$. The backward scattering is enhanced for a larger $g$, resulting in a higher upward radiance measured by the sensor. For the range of $\tau$ (1-8) and $r_{\text{eff}}$ (10-45 $\mu$m) analyzed in this study, assuming general habit mixtures (GHM) in the retrieval while in reality the habit is either aggregated plates or aggregated columns, results in uncertainties of up to 30% for $\tau$ and 49% for $r_{\text{eff}}$. 


8.4 Comparison of cloud optical thickness and particle effective radius

The cloud properties retrieved by SMART-Albedometer and MODIS are compared for the two cloud cases. For the cirrus case, the normalized mean absolute deviation yields a value of up to 1.2% for $\tau$ and 2.1% for $r_{\text{eff}}$. The deviations are slightly larger than those found in the comparison of upward radiance. This shows that the errors are only slightly amplified by the non-linearity in the retrieval algorithm. The cirrus $\tau$ derived from the MODIS cloud product overestimate the retrieval results by a factor of 1.6 because the MODIS cloud product algorithm only assume a single cloud layer. For the DCC case, the deviation is in the range of 3.6% for $\tau$ and 6.2% for $r_{\text{eff}}$. In this case, the fast cloud evolution and larger 3-D radiative effects are the major issue. The values of $r_{\text{eff}}$ derived from the MODIS cloud product are systematically larger due to the selection of ice crystal habit. While the retrievals implements the habit of aggregated plates, MODIS cloud products are retrieved based on aggregated columns. For both cloud cases, it is found that the particle sizes decrease toward the cloud top. While the particle sizes are more homogeneous in the upper layers, a higher horizontal variability is observed in the lower layers due to the sedimentation processes. Indeed, using $\lambda = 1240$ nm is useful to infer $r_{\text{eff}}$ from lower altitudes. However, the retrieved properties are associated with higher uncertainties due to lowering sensitivity to the changes of $r_{\text{eff}}$. Using C1, the resulting uncertainties are found to be about 12% for $\tau$, while they can be as high as 60% for $r_{\text{eff}}$. The uncertainties are lower for C2, which are about 10% for $\tau$ and 8% for $r_{\text{eff}}$.

For the cirrus case, retrieved $r_{\text{eff}}$ are compared to in situ measurements. A vertical weighting technique is applied to allow the comparison of retrieved and in situ $r_{\text{eff}}$. Using additional near-infrared wavelengths of SMART-Albedometer and MODIS increases the information on particle sizes extracted from the spectral measurements and the vertical resolution of retrieved $r_{\text{eff}}$. The normalized mean absolute deviation between retrieved and in situ $r_{\text{eff}}$ ranges between 1.5% and 10.3%, which falls within the standard deviation. A robust correlation coefficient is obtained with a value of 0.82. The best agreement is obtained for $\lambda = 3700$ nm due to its high absorption by cloud particles. Therefore, only the first layers, which are likely more homogeneous, contribute to the retrieved $r_{\text{eff}}$. For $r_{\text{eff}}$ at the lower layers, in general, the deviation increases corresponding to enhancing particle inhomogeneities. The values of $r_{\text{eff}}$ derived from the MODIS cloud product underestimate the in situ data with deviations between 19% and 49% due to the assumption of a single cloud layer. In general, the variability of particle size distributions, the uncertainties of deriving $r_{\text{eff}}$ from the in situ measurements, the remaining influences by the presence of liquid water clouds below cirrus, and the uncertainties by the unconstrained choice of ice crystal habit in the retrieval are identified as the major contributors, which can reveal the discrepancies between retrieved and in situ $r_{\text{eff}}$. The values of $r_{\text{eff}}$ derived from spectral measurements represent the feature of particle sizes in ice clouds, which decrease with
increasing altitude. However, the retrieved $r_{\text{eff}}$ only captures particle sizes at the cloud top and mid-layers. At the lower layers, it is hardly possible to be retrieved because of the limitation of the penetration depth given by this approach.

### 8.5 Retrieving the vertical profile of cirrus properties

This study has shown that by utilizing the conventional retrieval technique, it is impossible to obtain the full vertical profile of $r_{\text{eff}}$, even if spectral measurements have been deployed. This is due to the fact that the retrieved $r_{\text{eff}}$ represents a vertically weighted value, where the cloud top layers are weighted at most. To infer the full vertical profile, it is necessary to extend the method by putting some constrains on the shape of the profile of $r_{\text{eff}}$ with respect to a vertical coordinate, such as $\tau$. This allows portions of the profile for which the measurement contains significant information to be utilized to derive the parameters determining the profile. In the new retrieval approach, a Bayesian optimal estimation is applied to obtain the vertical profile of $r_{\text{eff}}$ as a function of $\tau$. This technique has shown an elegant framework in retrieving the vertical profile, which also gives advantages to analyze the information content of spectral measurement, as well as to quantify the retrieval uncertainties. Within this new approach, the total optical thickness $\tau_c$, the particle effective radius at the cloud top $r_{\text{eff},t}$, and the cloud base $r_{\text{eff},b}$ are seek to be retrieved. By knowing those parameters, the vertical profile of $r_{\text{eff}}$ can be reconstructed. Prior to the retrieval, a sufficient wavelength combination needs to be determined. For this purpose, the Shannon information content is utilized to identify wavelengths, which add the most information pertaining to each retrieval parameters. In addition to the measurement, uncertainties due to intrinsic and non-retrieved parameters assumed in the forward simulation are taken into account. Surprisingly, the analysis shows that this type of uncertainty contributes more strongly to the retrieval uncertainties.

The new technique is applied to obtain the profile of $r_{\text{eff}}$ as a function of $\tau$ for the cirrus case. Subsequently, the retrieved profile is compared to the corresponding profile measured by the CCP. The comparison yields a good agreement with a deviation of about 5\% at the cloud top, increasing toward the cloud base with values of up to 15\%, within uncertainties. The retrieved profile also reproduces a typical feature of particle sizes in cirrus with increasing particle inhomogeneity towards the cloud base. The deviation at the lower layers is generally larger due to this issue. Additionally, the potential error sources are due to the uncertainties of deriving $r_{\text{eff}}$ from the in situ measurements and the remaining influences by the presence of liquid water clouds below cirrus. In this retrieval, the parameter determining the shape of the vertical profile of $r_{\text{eff}}$ is fixed. Putting this parameter as an independent parameter to be retrieved convincingly shows a potential to improve the result. Eventually, this approach shows a significant improvement in obtaining the vertical profile of cirrus properties, which should be considered in the future generation of the retrieval technique based on measurements of passive remote sensing.
A Matlab code to generate the cloud profiles

The following MATLAB function, namely "iceprof.m", will generate two files that are needed to specify the cloud in the radiative transfer simulation. The first file consists three columns: geometrical altitude (km), IWC (g m$^{-3}$), and $r_{\text{eff}}$ (µm). The profile of $r_{\text{eff}}$ here is in accordance with that described in Sec. 5.3.1. The second file has one column that specifies $\tau$ of each layer. All data are presented in the descending order from the cloud top. Dividing the cloud into many layers might theoretically better to increase the vertical resolution. However, due to a limitation in the radiative transfer and abrupt decreasing sensitivity for $d\tau$ less than 0.15, the optimum number of layer is found to be about 20 (for $\tau_c \approx 3$). This number can be increased of up to 30 for larger values of $\tau_c$.

Figure A.1: (a) illustrates the parameterization applied to develop the profile of $r_{\text{eff}}$, while (b) is for IWC.

Fig. A.1 illustrates how the parameterization is applied for generating the profiles of $r_{\text{eff}}$ (a) and IWC (b). Developing the profile of IWC is relatively more straight forward since it is assumed to decrease linearly with altitudes. For $r_{\text{eff}}$, the profile is developed with helps of a geometric transformation, namely the reflection over the line $y = mx + q$. In this way (see Fig. A.1a), the aim is to mirror the exponential decrease (blue line) over the linear decrease (red line). The equation of the geometric transformation is given by Eq. A.1.
\[
\begin{bmatrix}
  r_{\text{eff},e} \\
  z
\end{bmatrix} = \begin{bmatrix}
  \cos 2\alpha & \sin 2\alpha \\
  \sin 2\alpha & -\cos 2\alpha
\end{bmatrix} \times \begin{bmatrix}
  r_{\text{eff},e} \\
  z - q
\end{bmatrix} + \begin{bmatrix}
  0 \\
  q
\end{bmatrix} 
\] (A.1)

The first term on the right side of Eq. A.1 yields the reflection matrix with an angle of \(2\alpha\). In this equation, \(r_{\text{eff},e}\) denotes the exponential decrease of \(r_{\text{eff}}\). The rotation angle \(\alpha\), which represents the gradient of the linear decrease of \(r_{\text{eff}}\), is obtained by:

\[
\alpha = \arctan (m),
\] (A.2)

where

\[
m = \frac{h}{r_{\text{eff},t} - r_{\text{eff},b}}. 
\] (A.3)

Here, \(h\) represents the cloud geometrical thickness. Additionally, the variable \(q\) in Eq. A.1, which specifies the intercept, is obtained by:

\[
q = m \cdot r_{\text{eff},b}.
\] (A.4)

As the result (see the black line in Eq. A.1a), \(r_{\text{eff}}\) will initially increase up to a certain point and turns to decrease towards the cloud top.
% A function to setup the profile of cirrus
% Written by: trismono_candra.krisna@uni-leipzig.de
% version: V26.02.2018

% Input parameters:
% tc = total cloud optical thickness
% rb = particle effective radius at cloud base (\unit{\mu}m)
% rt = particle effective radius at cloud top (\unit{\mu}m)
% iwcb = ice water content at cloud base (g/m^3)
% iwct = ice water content at cloud top (g/m^3)
% zt = cloud top altitude (m)
% zb = cloud base altitude (m)
% n = number of cloud layer (n = 20 is ideal for the radiative transfer)
% k = shape parameter (for cirrus, k is typically 3–5)

% Command:
% [prof, tau] = prof(tc, rb, rt, iwcb, iwct, zb, zt, n_layer, k)

% Example:
% [prof, tau] = prof(2, 40, 10, 0.1, 0.03, 10000, 12000, 20, 3)

"prof" will result in a cloud profile as the input in Libradtran:
% z(km) IWC(g/m^3) reff(\unit{\mu}m)
% "tau" generates the optical thickness profile from cloud top

function [prof, tau] = iceprof(tc, rb, rt, iwcb, iwct, zb, zt, n_layer, k)

% basic parameters
n_layer = n_layer + 1 ; % number of cloud layer
h = abs(zt-zb) ; % geometrical thickness
z = linspace(0,h,n_layer) ; % altitude of each layer
A. Matlab code to generate the cloud profiles

```matlab
% 72
a0 = rt + rb ;
a1 = rt ^k ;
a2 = rt ^k - rb ^k ;

% exponential decrease reff
re_exp = a0 - ( a1 - a2 . * z . / h ) . ^ ( 1 / k ) ;

% linear decrease reff
re_linear = linspace(rb, rt, n_layer) ;

% parameters of line
p1 = [ rb , 0 ] ; % coordinate at the cloud base
p2 = [ rt , h ] ; % coordinate at the cloud top
m = ( p2 ( 2 ) - p1 ( 2 ) ) / ( p2 ( 1 ) - p1 ( 1 ) ) ; % slope
q = p1 ( 2 ) - m * p1 ( 1 ) ; % intercept

% rotation angle
alpha = atand(m) ;

% reflection over line y = mx+q
for i = 1 : numel(re_exp)
    var(:,i) = [ cosd(2*alpha) sind(2*alpha); sind(2*alpha) - cosd(2*alpha) ] * [ re_exp(i); z(i)-q ] + [0;q] ;
end

% setup output parameters and adding IWC
re = ( var ( 1 , :) ' ) ; % reff result
zl = ( var ( 2 , :) + zb ) ' . / 1000 ; % reff altitude (km)
iwc = linspace(iwcb,iwct,n_layer)' ; % ice water content (g/m^3)
prof = flipud([zl, iwc, re]) ; % make descending order

% tau profile
tau = flipud(linspace(tc, 0 , n_layer)) ' ; % \tau profile
end
```
B Impact of the vertical profile of particle effective radius on the spectral downward irradiance

Fig. B.1 shows the resulting spectral downward irradiance $F_{\lambda}^{\downarrow}$ at the Earth’s surface ($z = 0$) for assuming a realistic (black) and a vertically homogeneous (red) profiles. In this example, $\tau$ is fixed to 3. The simulations are conducted by assuming $\theta_0 = 36^\circ$ and $\rho = 0.05$ (spectrally flat). Panels (a) and (b) represents resulting $F_{\lambda}^{\downarrow}$ in the solar ($\lambda = 300-4000$ nm) and terrestrial ($\lambda = 4000-90000$ nm) ranges. Assuming different profiles of $r_{\text{eff}}$ introduces discrepancies, particularly in the absorption wavelengths by cloud particles (e.g., at $\lambda = 900-1150$ nm, 1200-1350 nm, 1450-1750 nm, 2000-2400 nm, 8000-15000 nm). These will consequently lead to discrepancies in the resulting cloud forcing, as shown in Fig. 1.4.

Figure B.1: Comparison of spectral downward irradiance for assuming different profiles: realistic profile (black) and vertically homogeneous profile (red). (a) and (b) are the results for the solar ($\lambda = 300-4000$ nm) and terrestrial ($\lambda = 4000-100000$ nm) ranges, respectively.
Bibliography


<table>
<thead>
<tr>
<th>Author(s)</th>
<th>Title</th>
<th>Journal</th>
<th>Volume</th>
<th>Pages</th>
<th>Year</th>
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List of Symbols

<table>
<thead>
<tr>
<th>Symbol</th>
<th>Description</th>
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<tbody>
<tr>
<td>$\chi$</td>
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<table>
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<tr>
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</tr>
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</tr>
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</tr>
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</tr>
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</tr>
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<td>$RSR$</td>
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# List of Abbreviations

<table>
<thead>
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<th>Abbreviation</th>
<th>Description</th>
</tr>
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<tbody>
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<td>Advanced Very High Resolution Radiometer</td>
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<td>BRDF</td>
<td>Bidirectional Reflectance Distribution Function</td>
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<td>C</td>
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<td>Cloud-Aerosol Lidar and Infrared Pathfinder Satellite Observations</td>
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<td>FWHM</td>
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<td>General Habit Mixture</td>
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<td>Geostationary Operational Environmental Satellite</td>
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<td>High Altitude and Long Range Research Aircraft</td>
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<td>High-Performance Instrumented Airborne Platform for Environmental Research</td>
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<td>libRadtran</td>
<td>Library for radiative transfer</td>
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<td>Moderate Resolution Imaging Spectroradiometer</td>
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<td>Standard Bi-spectral Retrieval</td>
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<td>Tropopause Transitional Layer</td>
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<td>UTC</td>
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<td>Water Vapor Analyzer</td>
</tr>
<tr>
<td>WCB</td>
<td>Warm Conveyor Belt</td>
</tr>
</tbody>
</table>
## List of Figures

1.1 Global distribution of cirrus and deep convective clouds ........................................... 2  
1.2 Ice crystal shapes as a function of temperature .............................................................. 6  
1.3 Radiative forcing of ice and liquid water clouds ............................................................ 8  
1.4 Radiative forcing due to the vertical profile of effective radius ..................................... 13  
2.1 Geometry to define radiance and irradiance ................................................................. 18  
2.2 Downward solar irradiance and absorption bands ......................................................... 19  
2.3 Single scattering albedo different ice crystal habits and imaginary part of refractive index of ice ................................................................................................................ 21  
2.4 Scattering phase function of different ice crystal habits ................................................ 23  
3.1 HALO flight paths during ML-CIRRUS and ACRIDICON-CHUVA campaigns .......... 30  
3.2 Instrumentation setup of HALO during ML-CIRRUS and ACRIDICON-CHUVA campaigns .................................................................................................................. 33  
3.3 Footprint of SMART-Albedometer radiance inlet ......................................................... 34  
4.1 MODIS relative spectral response .................................................................................... 40  
4.2 Retrieval of MODIS radiance band 6 ................................................................................ 41  
4.3 Illustration of the binning method applied to SMART-Albedometer measurements ........................................................................................................................................... 41  
4.4 Data filter for selecting the case studies .......................................................................... 43  
4.5 Cloud fields and selected flight legs ................................................................................ 44  
4.6 Comparison of upward radiance for the cirrus case ....................................................... 45  
4.7 Comparison of upward radiance for the DCC case ....................................................... 46  
4.8 Comparison of spectral upward radiance for the two cases ........................................... 46  
5.1 Illustration for discriminating the properties of cirrus and low liquid clouds ............... 52  
5.2 Sensitivity test on the spectral upward radiance and the co-albedo of different particle sizes ...................................................................................................................................... 54  
5.3 Lookup tables for the cirrus and the DCC cases ............................................................ 55  
5.4 Uncertainties of cirrus retrievals due to the properties of low liquid cloud ................. 57  
5.5 Uncertainties of retrieved cirrus effective radius for higher optical thicknesses ........... 58  
5.6 Single scattering asymmetry parameter and the co-albedo of different ice habits ................................................................................................................................. 60  
5.7 Uncertainties of cirrus retrievals due to the choice of ice crystal habit ........................... 60
<table>
<thead>
<tr>
<th>Figure</th>
<th>Description</th>
<th>Page</th>
</tr>
</thead>
<tbody>
<tr>
<td>5.8</td>
<td>Cloud phase index of pure cirrus and multilayer cloud</td>
<td>63</td>
</tr>
<tr>
<td>5.9</td>
<td>Cloud phase index of pure cirrus and liquid water clouds</td>
<td>64</td>
</tr>
<tr>
<td>5.10</td>
<td>Vertical weighting function of two idealized clouds</td>
<td>67</td>
</tr>
<tr>
<td>5.11</td>
<td>Spectral vertical weighting function and the co-albedo</td>
<td>68</td>
</tr>
<tr>
<td>5.12</td>
<td>Vertical weighting function due to changes of surface albedo</td>
<td>69</td>
</tr>
<tr>
<td>5.13</td>
<td>Spectral surface albedo derived from MODIS BRDF/Albedo product</td>
<td>70</td>
</tr>
<tr>
<td>5.14</td>
<td>Vertical weighting function due to the presence of low liquid clouds</td>
<td>72</td>
</tr>
<tr>
<td>5.15</td>
<td>Setup of the clouds used for quantifying penetration depths</td>
<td>74</td>
</tr>
<tr>
<td>5.16</td>
<td>Penetration depth of cirrus clouds in terms of optical and geometrical thicknesses</td>
<td>75</td>
</tr>
<tr>
<td>5.17</td>
<td>Comparison of penetration depths in cirrus and liquid water clouds</td>
<td>76</td>
</tr>
<tr>
<td>5.18</td>
<td>Vertical weighting function for different solar zenith angles</td>
<td>77</td>
</tr>
<tr>
<td>6.1</td>
<td>Comparison of retrieval results for the cirrus case</td>
<td>80</td>
</tr>
<tr>
<td>6.2</td>
<td>Comparison of retrieval results for the cirrus case</td>
<td>81</td>
</tr>
<tr>
<td>6.3</td>
<td>Comparison between remote sensing and in situ effective radius</td>
<td>83</td>
</tr>
<tr>
<td>7.1</td>
<td>Profile of the weighting estimate effective radius</td>
<td>88</td>
</tr>
<tr>
<td>7.2</td>
<td>Partial information content and co-albedo of different particle sizes</td>
<td>94</td>
</tr>
<tr>
<td>7.3</td>
<td>Standard deviation of spectral upward radiances given the uncertainty parameter</td>
<td>97</td>
</tr>
<tr>
<td>7.4</td>
<td>Retrieved profile of effective radius as a function of optical thickness</td>
<td>99</td>
</tr>
<tr>
<td>7.5</td>
<td>Comparison of the retrieved and the in situ profile of effective radius</td>
<td>100</td>
</tr>
<tr>
<td>A.1</td>
<td>Parameterization of cloud profiles</td>
<td>109</td>
</tr>
<tr>
<td>B.1</td>
<td>Spectral downward irradiance for assuming different profiles</td>
<td>113</td>
</tr>
</tbody>
</table>
List of Tables

3.1 Characteristics of the SMART-Albedometer spectrometers ............... 32
3.2 Uncertainties of radiance measurements .................................. 34
3.3 Specifications of MODIS bands .............................................. 36

4.1 Flight descriptions and atmospheric conditions ............................ 42
4.2 Statistics of the upward radiance comparison for the two cases .......... 47

5.1 Retrieval uncertainties due to different wavelength combinations and solar zenith angle .......................................................... 62
5.2 Specifications of vertically inhomogeneous clouds ....................... 66
5.3 Effective penetration depth for ice and liquid water clouds ............ 78

6.1 Statistic of remote sensing and in situ effective radius .................. 84

7.1 Parameters governing uncertainties in the forward simulation .......... 96
7.2 Statistic of uncertainties of the forward simulation ...................... 98
I, Trismono Candra Krisna, declare that this dissertation entitled, "Airborne Passive Remote Sensing of Optical Thickness and Particle Effective Radius of Cirrus and Deep Convective Clouds" and the work presented in it are my own. I confirm that:

- This work was done wholly or mainly while in candidature for a research degree at this University.

- Where any part of this thesis has previously been submitted for a degree or any other qualification at this University or any other institution, this has been clearly stated.

- Where I have consulted the published work of others, this is always clearly attributed.

- Where I have quoted from the work of others, the source is always given. With the exception of such quotations, this thesis is entirely my own work.

- I have acknowledged all main sources of help.

- Where the thesis is based on work done by myself jointly with others, I have made clear exactly what was done by others and what I have contributed myself.

Signed: Trismono Candra Krisna

Date: 23 January 2019
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