Bachelor Thesis

Cloud Condensation Nuclei from Satellite Lidar Observations

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Abstract

Aerosol-cloud interactions appear to be the dominant factor of uncertainty in quantifying atmospheric drivers of a changing climate due to anthropogenic perturbation. Aerosol optical properties have been widely used as a proxy to investigate the global distribution of Cloud Condensation Nuclei (CCN) number concentrations. Spaceborne lidar instruments are identified to give a realistic view of atmospheric aerosol abundance in terms of vertically resolved optical quantities. This work investigates the relationship between simulated CCN and aerosol extinction applying data from the European Monitoring Atmospheric Composition and Climate (MACC) project. The numerical approach reveals a reasonable correlation between the quantities on a global scale. However, certain factors are identified to impact the CCN-extinction correspondence and need to be included in parameterization schemes more accurately. In a second step, the simulation setup is combined with aerosol optical features that have been provided by the Cloud-Aerosol Lidar with Orthogonal Polarization aboard the CALIPSO satellite. A varying degree of global correlation emerges for different areas of the Earth’s surface. Certain factors (e.g. extinction-to-backscatter ratios) need to be considered for the derivation of global CCN fields from spaceborne lidar products. However, scattering ratios in an interval ranging from 2 to 3 are found to hold a strong (linear) connection to simulated CCN.
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1 Introduction

The latest assessment report (AR5) of the Intergovernmental Panel on Climate Change (IPCC) states that the human influence on the climate system is proven (Stocker et al., 2013). Changes in the concentration of several atmospheric constituents, such as greenhouse gases, reactive gases, and aerosols alter the Earth’s energy budget. The transformation of the natural environment due to human emissions, quantified as perturbation to Top of Atmosphere (TOA) radiative fluxes, is termed radiative forcing (Forster et al., 2007). Since the industrial revolution, an increasing atmospheric concentration of the greenhouse gas CO$_2$ appears to be the main driver for a warming climate. However, a larger burden of atmospheric aerosol caused by anthropogenic emissions is identified to counter the effect of greenhouse gases by cooling the climate system (Forster et al., 2007; Boucher et al., 2013).

1.1 Aerosol interactions

The modification of the net global radiation budget by aerosols appears in terms of different mechanisms: Aerosol forcing directly occurs via radiation scattering and absorption in the solar and thermal spectrum, and additionally by radiation emission at thermal wavelengths (Lohmann and Feichter, 2005).

In addition, suitable aerosols can be capable of initiating cloud droplet (or ice particle) formation at a given saturation ratio. These particles are established as Cloud Condensation Nuclei, hereafter abbreviated as CCN, in liquid droplet nucleation processes (e.g. Boucher et al., 2013). Aerosol-cloud interactions appear by transforming cloud characteristics and abundance of either warm clouds or ice/mixed-phase clouds, initiated from the liquid phase (Stier, 2016; Bellouin et al., 2013; Boucher et al., 2013).

There are different processes referred to as aerosol indirect effect: First, an increase in the amount of CCN-relevant aerosol particles in a cloud enhances its droplet number concentration. Consequently, in a given cloud with constant liquid water content, the size of aqueous droplets would decrease. Twomey (1959) concluded that this process results in an enhanced radiation reflection by the cloud (Lohmann and Feichter, 2005; Boucher et al., 2013; Twomey, 1959). This indirect process is called cloud albedo effect or Twomey effect. Second, a higher concentration of cloud droplets with smaller radii leads to the suppression of precipitation, which results in a prolonged cloud lifetime and increased atmospheric reflectivity (Lohmann and Feichter, 2005; Boucher et al., 2013; Albrecht, 1989). The correspondence of the enhancement of the aerosol number to the hydrological cycle of a cloud is known as cloud lifetime effect or Albrecht effect.

A semi-direct effect can be initiated by the absorption of solar radiation by aerosols, which may cause an evaporation of cloud particles due to the heating of
the aerosol layer. This process is particularly pronounced for absorbing aerosols such as Black Carbon ([Lohmann and Feichter 2005] [Boucher et al. 2013]).

1.2 Observation of CCN

As noted by the IPCC AR5, aerosol-cloud interactions emerge as the most uncertain factor in estimates of the anthropogenic perturbation ([Boucher et al.] 2013) [Stocker et al.] 2013). Thus, the investigation of global aerosol and cloud distribution and their various interactions is a fundamental step to advance the understanding of human impact on the Earth’s climate. This motivates an accurate global determination of spatiotemporal CCN abundance and the underlying processes of cloud-droplet formation (e.g. [Bellouin et al.] 2013). However, aerosol-cloud interactions occur locally and emerge from complex physical and chemical processes. This yields observational difficulties, so that researching atmospheric aerosol-cloud interactions is often fraught with uncertainties ([Stier] 2016) [Boucher et al. 2013].

Direct measurements of aerosol particles provide a realistic view of atmospheric CCN fields. In-situ techniques make use of counting mechanisms, registering the number of formed droplets in a chamber at adjusted air saturation ratio ([Liu and Li] 2014). However, direct measurements are sparse and cannot sufficiently represent CCN abundance on a global scale ([Shinozuka et al.] 2015) [Stier 2016].

Consequently, many studies rely on numerical modelling to mimic atmospheric particulates and quantify aerosol-cloud interactions. The use of models provides a spatially and temporally consistent data set. Aerosol models include the most important particles, considering relevant chemical and physical processes via parameterization schemes ([Stier] 2016). However, the global modelling of atmospheric constituents is challenging and simplifying assumptions appear to be unavoidable. Consequently, model outputs do not necessarily conform to observations ([Bellouin et al.] 2013).

Observational products from remote sensing provide an increasingly detailed view of aerosols and clouds ([Bellouin et al.] 2013) [Winker et al.] 2010). The exploitation of satellite sensors enables the daily retrieval of radiative properties that can constrain atmospheric aerosol on a global scale. However, spatial and temporal gaps in direct measuring techniques hamper a sufficient testing of the underlying assumptions so that no reliable global CCN estimation exists ([Stier] 2016) [Shinozuka et al.] 2015) [Mamouri and Ansmann 2016].

A large portion of prior studies of aerosol-cloud interactions uses the Aerosol Optical Depth (AOD) as substitute for CCN ([Shinozuka et al.] 2015). Per definition, light extinction throughout the whole atmospheric column contributes to the optical depth. Consequently, the CCN-AOD relationship is affected by the vertical distribution of the particles. In addition, the AOD is sensitive to the total aerosol surface area, which complicates the relation to CCN number concentrations. Another point
is that the composition (e.g., hygroscopicity) of aerosol particles affects the AOD and CCN properties differently, which is why the columnar quantity might not be a reliable proxy (Shinozuka et al., 2015). Recent studies identify vertical profiling in terms of aerosol optical properties as more appropriate for a large-scale CCN assessment.

1.3 Current state of research

The modelling study of Stier (2016) investigates the global link between aerosol optical properties and CCN. It is concluded that the AOD variability explains only 25% of CCN variance at cloud base for 71% of the globe. The correlation significantly improves when the local Aerosol Extinction Coefficient (AEC) replaces the columnar AOD as proxy. In addition, it is noted that the correlation further improves by making use of the local Ångström parameter (Deuzé et al., 2001) since it gives lower weight to large particles.

Liu and Li (2014) researches the role of aerosol optical properties to estimate surface CCN by accounting for certain influential factors. Data is acquired from different Atmospheric Radiation Measurement (ARM) Climate Research Facility sites for regions with certain aerosol background. The results closely match with demonstrations of Stier (2016). Liu and Li (2014) determines that CCN fields correlate reasonably well with the AOD for observed regions, but only in areas where large aerosol particles (sea salt and dust) are not dominant. Higher correlations are indicated between surface CCN and the surface aerosol scattering coefficient. A strong dependence of the CCN-scattering relationship on ambient relative humidity (RH) conditions is observed.

This points out an important factor for global CCN estimations. Even though high CCN-extinction/scattering correlations account for a high potential of vertically resolved optical quantities to derive CCN, the relationship is strongly affected by humidity growth: A hygroscopic particle that has been activated adds to light extinction with increasing humidity conditions due to water uptake without contributing to the CCN number concentration (e.g., Shinozuka et al., 2015; Quinn et al., 2008; Liu and Li, 2014).

A recent study of Shinozuka et al. (2015) further examines the link of dry-particle extinction coefficient and CCN vertical distributions. Using in-situ extinction data for dried particles, the humidity response is assumed to be negligible. The result indicates an improvement of the CCN-extinction relationship in regions of lower ambient relative humidity.

Based on previously mentioned correlation studies, the importance of investigating vertical structure of aerosol and clouds is evident. It is implied that if the vertical profile of aerosol scattering properties is known, CCN profiles may be estimated. Mamouri and Ansmann (2016) establish a way to derive profiles of cloud-relevant
particles from polarisation lidar products. The classification of different aerosol types (here: marine, continental and desert dust) is pointed out as major step since cloud relevance is strongly tied to size and chemical composition (e.g. [Quinn et al. 2008]). Particle number concentrations are estimated from aerosol extinction coefficients, assuming certain extinction-to-backscatter ratios. Finally, vertical profiles of CCN are concluded for set humidity conditions.

In summary, previous studies use aerosol radiative properties provided by remote sensing or in-situ methods as a constraint for CCN. Correlation studies have investigated the link between aerosol parameters and cloud microphysical properties. A general problem is identified by the fact that the AOD is a vertically integrated property. Recent studies illuminate the need to consider the vertical structure of the particles: Aerosols will contribute to light extinction regardless of altitude, whereas CCN relevant processes specifically take place at cloud base ([Shinozuka et al. 2015]). Consequently, stronger relationships are expected for vertically resolved aerosol optical quantities. This matches the findings from correlation studies: Vertically resolved aerosol radiative properties have a large capability to globally assess CCN. This implies a high potential of spaceborne lidar instruments, as they simultaneously assess height profiles of particle backscatter from which the local extinction coefficient can be derived. However, complicated relationships are expected even between CCN and aerosol extinction: Atmospheric aerosol particles vary in terms of size, shape and composition. Environmental conditions such as the ambient humidity also matter.

This work connects to previous correlation studies and examines the link between CCN and vertically resolved aerosol optical quantities. Simulated 3-dimensional CCN fields are related to the simulated local (model layer) aerosol extinction coefficient. The modelling allows consistently accessing the relationship between both quantities. In a second step, the modelled data is evaluated with products from the spaceborne Cloud-Aerosol Lidar with Orthogonal Polarization (CALIOP). This work does not provide a method of deriving CCN distributions from aerosol radiative properties, but investigates the correspondence of aerosol extinction/scattering and CCN.
2 Methods and Data

This work applies data from the aerosol analysis of the European Monitoring Atmospheric Composition and Climate (MACC) project. The simulated parameters are the CCN number concentration and the local AEC. Both quantities are based upon the same model output and represent properties of the main atmospheric aerosol types: mineral dust, sea salt, Black Carbon (BC), Organic Matter (OM), and sulphate. The parameters are mapped on a Gaussian grid with a horizontal resolution of $0.75^\circ \times 0.75^\circ$. The vertical resolution is set to 60 levels, extending from the surface up to 10 Pa. More information about the MACC aerosol analysis can be found in Morcrette et al. (2009), Benedetti et al. (2009), and Bellouin et al. (2013).

To further investigate the role of spaceborne lidar techniques for constraining global CCN abundance, the modelled quantities are related to observed aerosol optical properties obtained by the Cloud-Aerosol Lidar with Orthogonal Polarization (CALIOP) aboard the CALIPSO satellite. The analysis applies Scattering Ratios (SRs) categorized in certain intervals that are provided by the GCM-Oriented CALIPSO Cloud Product (GOCCP). The data is aggregated onto a global 2-degree latitude-longitude grid. The altitude is referenced to sea level. A detailed description of the CALIOP-GOCCP is reported in Chepfer et al. (2010).

2.1 MACC aerosol analysis

The Copernicus Atmosphere Monitoring Service (CAMS) provides information of atmospheric constituents in the form of the series of the European MACC projects. MACC is based upon the Integrated Forecasting System of the European Centre for Medium-Range Weather Forecasts (ECMWF) (Eskes et al., 2015). The modelling includes global abundance, transport, and sources and sinks of several atmospheric constituents which are identified as being relevant for a changing climate system.

The aerosol component of the ECMWF IFS (Morcrette et al., 2009) simulates the mass-mixing ratio of above-mentioned aerosol types. Mineral dust and sea salt are separated into three size classes. OM and BC are simulated with hydrophobic and hydrophilic modes, respectively. In total, 11 tracers are distinguished. Emissions are found through certain model parameters (e.g. near-surface wind field or soil moisture) and emission inventories (e.g. GFED or Global Fire Emission Database). Sinks include mechanisms such as wet and dry deposition (Benedetti et al., 2009, Bellouin et al., 2013).

For better agreement with observations, a data-assimilation module (Benedetti et al., 2009) combines the modelling with retrieved products from satellite instruments. The control variable is provided by the total aerosol mass-mixing ratio that shifts with respect to observations. The assimilated property is the AOD at a wavelength of 550 nm. It is provided by the Moderate-resolution Imaging Spectroradiometry...
Methods and Data

The AOD is derived based upon precomputed aerosol optical properties, namely mass extinction coefficient $\alpha$, single scattering albedo $\omega$, and asymmetry parameter $g$ (see Benedetti et al. 2009; Reddy et al. 2005 for further information).

2.1.1 Parameterization schemes

CCN

A fundamental requirement needed to derive microphysical properties of atmospheric aerosol is a detailed knowledge about the particle size distribution (Liu and Li 2014). Aerosol number concentrations are calculated from the MACC-II analysis and aerosol properties assumed in the IFS radiative transfer code. A detailed description of the computation is reported in Block et al. (2017).

The applied particle number distribution for an aerosol species $i$ appears in the following form:

$$N_i(r) = N_{a,i} \cdot \frac{1}{\sqrt{2\pi} \cdot \ln \sigma_{g,i}} \cdot \exp \left( -\frac{\ln^2 (r/r_{0,i})}{2 \ln^2 \sigma_{g,i}} \right).$$

The fixed parameters are established as count median radius $r_{0}$, and geometric standard deviation $\sigma_{g}$, which originate from Table 2 of Reddy et al. (2005). The size distributions are applied for each aerosol mode and are passed on to the CCN parameterization scheme. Note that only the hydrophobic aerosol modes have been considered to contribute to CCN. No dust components appear in the CCN parameterization setup (Block et al. 2017).

Cloud-droplet formation from an initial aerosol mainly depends on particle size, chemical composition, and ambient environmental conditions (Abdul-Razzak and Ghan 2000; Quinn et al. 2008). An estimation of atmospheric CCN-activity is found through Köhler theory. The derivation of CCN number concentrations is found in section 2 of Block et al. (2017) and corresponds to the elaborations in Pruppacher and Klett (1997).

Köhler theory establishes a relationship between the water vapour saturation ratio and the size of a solution droplet. Considering an equilibrium between water vapour and the droplet, the Köhler equation is given by

$$\frac{e_a}{e_{sat,w}} = \exp \left( \frac{A}{a} - \frac{B}{a^3} \right)$$

with $e_a$, the equilibrium water vapour pressure over the droplet of a given radius $a$, and $e_{sat,w}$, the water vapour pressure over a plane water surface. In order to simplify the expression, the formal definition of the power series for the exponential function
with a complex number $z$ is applied (see page 37 of Bronstein et al., 2013) with the approximation

$$\exp(z) = \sum_{k=0}^{\infty} z^k / k! = 1 + z + \frac{z^2}{2} + \frac{z^3}{6} + ... \approx 1 + z \quad \text{for } |z| \ll 1. \quad (3)$$

This simplifies the Köhler equation to the more convenient form

$$\frac{e_a}{e_{\text{sat},w}} = 1 + \frac{A}{a} - \frac{B}{a^3} \quad (4)$$

or

$$S = \frac{e_a}{e_{\text{sat},w}} - 1 = \frac{A}{a} - \frac{B}{a^3}, \quad (5)$$

with $S$, the supersaturation. Two effects appear to influence the equilibrium between solution droplet and water vapour. First, the water vapour pressure over a curved surface is higher compared to a flat-water surface. Second, a soluble aerosol material corresponds to a decreased water vapour pressure above the droplet surface. The curvature (Kelvin) effect and the solution (Raoult) effect are represented by the parameters $A$ and $B$, respectively. The equations characterising both parameters correspond to the expressions given in Pruppacher and Klett (1997) and Abdul-Razzak and Ghan (2000), and can be written as follows:

$$A = \frac{2M_w \sigma_{w/a}}{RT \rho_w} \quad \text{and} \quad B = \frac{3}{4\pi} \nu \phi \epsilon m_b M_w \quad (6)$$

With:

- $\sigma_{w/a}$ the surface tension of pure water
- $M_w$ and $\rho_w$ the molecular weight and density of pure water, respectively
- $R$ the universal gas constant
- $T$ the temperature ($K$)
- $M_s$ and $m_b$ the molecular weight and mass of the dissolved substance, respectively
- $\nu$ the number of ions the dissolved substance dissociates within water
- $\phi$ the osmotic coefficient
- $\epsilon$ the mass fraction of the aerosol material
Both curvature and solution effect are satisfied for a given droplet and yield the saturation ratio by means of the Köhler theory.

Equation (5) is now converted to a form from which CCN fields can be estimated with respect to the prescribed aerosol number concentration: The saturation ratio corresponds to the radius $r_s$ of a dry particle, that is contained in the Raoult term. By replacing the mass $m_s$ of the dry particle, and assuming a spherical shape, equation (6) for parameter $B$ can be transformed into the following expression:

$$B = \nu \phi \epsilon \frac{\rho_s}{\rho_w} \frac{M_w}{M_s} r_s^3$$

(7)

If the maximum of the saturation ratio is reached, the wet particle is termed as activated. Consequently, the aerosol is suitable for acting as CCN. Maximizing $S$ with respect to the equilibrium radius for a solution droplet $a$ yields the maximum (critical) supersaturation in the form

$$S_{\text{crit}} = \sqrt{\frac{4A^3}{27B}}$$

(8)

which corresponds to the critical droplet radius:

$$a_{\text{crit}} = \sqrt{\frac{3B}{A}}$$

(9)

If the actual supersaturation is larger than the critical supersaturation, the considered droplet will grow. A typical value of supersaturation in real cloud conditions ranges from 0.1 to 1.0 % (Mamouri and Ansmann, 2016). By re-defining the parameter $B$ as

$$B = \nu \phi \epsilon \frac{\rho_s}{\rho_w} \frac{M_w}{M_s}$$

(10)

and extracting the radius of the dry particle $r_s$, its critical value is obtained by equation (6) as a function of curvature and solution parameters, and the critical supersaturation $S_{\text{crit}}$:

$$r_{s,\text{crit}} = \frac{A}{3B^2} \left( \frac{2}{S_{\text{crit}}} \right)^{\frac{3}{2}}.$$ 

(11)

The total CCN concentration is derived with respect to the prescribed aerosol number concentration for each mode, considering equation (11) and supersaturations of 0.2, 0.4 and 1.0 %.

**Aerosol extinction coefficient**

Light extinction by atmospheric aerosol depends on particle properties such as composition, size, and shape (e.g. Waggoner et al., 1981), and the wavelength of the
incoming radiation. Mie theory is used to quantify the interaction of small atmospheric particles with incident radiation by assuming the aerosols as perfect spheres. Applying Mie theory, the extinction coefficient $\rho_e$ for a spherical particle with the radius $a$ can be expressed as

$$\rho_e = Q_e \pi a^2$$

with $Q_e$, the efficiency coefficient for extinction, and $\pi a^2$, the cross-sectional area of the aerosol. The extinction efficiency is a function of the so-called size parameter ($x$), and the index of refraction for a given medium. The size parameter of a spherical particle is defined as the ratio of its circumference to the wavelength of the incident radiation:

$$x = \frac{2\pi r}{\lambda}$$

For each tracer, aerosol material dictates the effective refractive index (see Table 2, for the considered modes). However, the refractive index of an aerosol particle might change due to water uptake. Therefore, aerosol water is included in the parameterization via volume-averaging the respective refractive indices of aerosol material and water. Consequently, aerosol extinction is additionally parameterized in terms of ambient RH conditions. The required RH fields are pre-calculated from the model parameters temperature, pressure, and specific humidity (Benedetti et al., 2009; Reddy et al., 2005).

The aerosol extinction coefficient is based upon the MACC aerosol mass-mixing ratio. Refractive indices and size parameters for each mode serve as input for look-up tables, leading to the extinction cross-section with respect to the considered wavelength (here: 532 nm). Finally, the aerosol extinction coefficient is derived by integrating the extinction cross-section over the size range, using a prescribed lognormal distribution (see Benedetti et al., 2009; Reddy et al., 2005).

### 2.1.2 CCN-extinction relationship

In summary, both the CCN number concentration and the local extinction coefficient are parameterized with respect to the mass-mixing ratio of the MACC aerosol analysis. CCN concentrations are related to particle size and chemical features of the respective aerosol mode at different reference supersaturations. Aerosol extinction coefficients are based upon size and refractive indices of the particles with respect to the wavelengths of interest. In addition, for hygroscopic tracers, ambient RH conditions in combination with water affinity are considered for aerosol optical behaviour.

Theoretically, aerosol extinction has a high potential to sufficiently constrain CCN. Given fixed ambient conditions and identical size, shape, and composition of aerosols, CCN concentrations at a certain supersaturation are linearly related
to aerosol extinction (Stier 2016). However, in real-world conditions, atmospheric particulates vary markedly in terms of their features (Lamb and Verlinde 2011). Consequently, even the CCN-extinction relationship is expected to be complex, since optical behaviour and CCN activity are related to the influential factors in their own specific way (Shinozuka et al. 2015).

**Humidity**

With enhanced ambient humidity, a hygroscopic aerosol particle might swell and additionally contribute to light extinction. As shown by Figure 1 of Reddy et al. (2005), light extinction is significantly larger for higher values of RH. Hence, changes in extinction ability are strongly tied to varying humidity conditions. Köhler theory shows an indirect proportional relation of $S$ to the critical radius for fixed $A$ and $B$ parameters. Hence, an increased supersaturation leads to the activation of smaller particles and in total to a higher activation rate of aerosols to CCN. Consequently, the relationship is theoretically affected by a changing humidity environment.

**Composition**

Aerosol composition in the CCN parameterization scheme is determined from many variables and is comprised of all aerosol material factors that go into parameter $B$ of the Köhler equation. In the framework of Mie theory, the refractive index is a material property and thus also refers to composition features. However, the influence of the refractive index only, is expected to be low as it varies little around that for water at visible wavelengths (Lamb and Verlinde 2011).

Recent studies conclude that the impact of aerosol composition might be simply represented by an aerosol hygroscopicity factor (Shinozuka et al. 2015; Liu and Li 2014; Petters and Kreidenweis 2007). Köhler theory states that the stronger the hygroscopicity of a particle, the higher its capability to initiate cloud droplet formation. In addition, in a humid environment the particle composition has a deep influence on optical features. If varying only the RH, hygroscopicity determines the swelling effect due to water uptake. As extinction is parameterized as function of RH, aerosol hygroscopicity is considered to impact the CCN-extinction relationship.

**Particle size**

Theoretically, CCN fields possess their number peak in the smaller size ranges. This may be verified as follows: In general, atmospheric aerosols are typically most numerous in smaller size ranges (Lamb and Verlinde 2011; Seinfeld and Pandis 2016). Köhler theory yields the number concentration of aerosols out of the spectrum that are capable of initiating cloud-droplet formation by deriving the minimal radius a particle can have to do so. This yields the critical radius in the form of equation (11) as function of the supersaturation, and the parameters $A$ and $B$. Even for a case,
Table 1: The molecular weight and mass of the substance of hygroscopic aerosol species that are considered in the CCN parameterization. Properties originate from Benedetti et al. (2009), Reddy et al. (2005) and references therein.

<table>
<thead>
<tr>
<th>Aerosol</th>
<th>$\rho_s$ [g/cm$^3$]</th>
<th>$M_w$ [g/mol]</th>
</tr>
</thead>
<tbody>
<tr>
<td>Sea salt small</td>
<td>2.60</td>
<td>58.443</td>
</tr>
<tr>
<td>Sea salt medium</td>
<td>2.60</td>
<td>58.443</td>
</tr>
<tr>
<td>Sea salt large</td>
<td>2.60</td>
<td>58.443</td>
</tr>
<tr>
<td>Sulphate</td>
<td>1.84</td>
<td>96.0631</td>
</tr>
<tr>
<td>Black Carbon</td>
<td>1.80</td>
<td>12.01</td>
</tr>
<tr>
<td>Organic Matter</td>
<td>1.76</td>
<td>180.00</td>
</tr>
</tbody>
</table>

where $A$ appears as maximal, and $B$ as minimal (which would yield the maximum possible critical radius), $r_{\text{crit}}$ happens to be found in the smaller size ranges. Table 1 shows all CCN-relevant particles (hygroscopic modes) with the respective molecular weight and mass of the dissolved substance, that go into the solution parameter of equation (5). Considering equation (7) of parameter $B$, the value appears minimal for the smallest aerosol density, and the largest molecular weight. Both cases are found for OM. By calculating the corresponding critical radius (with given constants and values taken from section 5 of Block et al. (2017), the critical radius at 0.2 % is approximately 70 nm. In summary, even the maximal possible critical radius of activation is found in smaller modes. Consequently, for higher supersaturations, the critical radius is found to be even smaller (for 0.4 %: 45 nm, for 1.0 %: 24 nm).

On the other hand, particles in fine modes have a weak influence on light extinction. This is evident by considering equation (12) of Mie theory: The product of extinction efficiency and cross-sectional area of the particle yields the extinction coefficient. The extinction efficiency is maximal when $a \approx \lambda$. Hence, for midvisible wavelengths, the extinction efficiency is maximal for large-sized particles. With increasing size parameter, the extinction efficiency does not drop to value for those size ranges, where the CCN number peak is expected (see Figure 2.39 of Lamb and Verlinde (2011)). That accounts for high aerosol extinction coefficients for large aerosols, where both particle size and extinction efficiency are high.

In summary, it is expected that the CCN-extinction correlation varies due to all introduced influential factors. Therefore, knowledge of typical aerosol size and composition is of key importance for taking in account the factors and for reducing uncertainties in the global assessment of CCN abundance through aerosol extinction.
2 Methods and Data

2.2 CALIOP

Data from CALIOP is expected to have high potential to improve the understanding of global aerosols-clouds interactions. The spaceborne lidar is an active instrument aboard the polar-orbiting CALIPSO satellite. It was launched in April 2006 as the first polarization lidar to provide global atmospheric measurements [Winker et al., 2010].

The basic principle of a spaceborne lidar to provide atmospheric profiles is explained as follows: A laser continuously emits light pulses towards the Earth’s surface. On its way, the radiation interacts with atmospheric constituents. A certain part of the signal is scattered back, at a 180-degree angle with respect to the incident beam, and returns to the instrument. The measurement of the time interval between emitting a laser impulse and detecting the respective return signals allows a vertical resolution of the backscatter, assuming the speed of light [Weitkamp, 2006].

The CALIOP mission provides a two-wavelengths lidar observation at 532 and 1064 nm, respectively. The laser emits linearly polarised light at both wavelengths. The backscatter return signal is detected through a telescope in different channels. Two channels separately analyse the parallel and cross-polarised part of the return radiation at 532 nm, with respect to the original state of polarisation [Winker et al., 2010]. The ratio of both signals is referred to as the depolarisation of the backscattered, linearly polarised laser light. Spherically shaped particles retain their polarisation properties, whereas irregular shaped aerosols (e.g. dust) yield a depolarisation. Therefore, the functionality of polarisation-lidar techniques enables the separation of certain atmospheric constituents through their shape [Mamouri and Ansmann, 2016].

2.2.1 Scattering Ratios and joint histogram

The GCM-Oriented CALIPSO Cloud Product (GOCCP) has been set up to provide estimates of the cloudiness simulated by general circulation models (GCMs). A deeper description of the lidar product is reported in Chepfer et al. (2010). In the atmosphere both molecules and particles (cloud and aerosol particles) contribute to atmospheric scattering [Omar et al., 2009]. The scattering ratio (SR) is introduced as the relation of the lidar-observed backscattering to the hypothetical backscattering if there were no particles in the atmosphere. This yields the correction of the molecular backscattering (Rayleigh regime). SR histograms are used to record the occurrence frequency of certain SRs for each vertical grid box at a given altitude and time. The so-called Contoured Frequency-by-Altitude Diagram (CFAD) then appears as a function of time (one timestep per day for polar satellites), latitude and longitude, altitude (40 equidistant levels), and a certain interval of SR (15 overall with a range from 0 to 100). The histogram allows the classification of certain
atmospheric layers, following the method of [Chepfer et al. (2010)]. Where the lidar SR signal is too high for a clear-sky situation \((0.01 \leq \text{SR} \leq 1.2)\), but is too low for a cloud layer \((\text{SR} > 5)\), the SRs may represent the abundance of aerosols (e.g., over the Sahara). Note that for the following analysis clouds are removed and only SRs in an interval ranging from 1 to 3 are considered.

### 2.3 Statistics

In order to investigate the link between MACC-CCN fields and aerosol optical properties, Pearson product-moment linear correlation coefficients are computed by means of the following form:

\[
    r = \frac{1}{n} \sum_{i=0}^{N-1} \left( \frac{x_i - \bar{x}}{\sigma_x} \right) \left( \frac{y_i - \bar{y}}{\sigma_y} \right)
\]

(14)

The parameters \(\bar{x}\) and \(\bar{y}\) are the average of the values \(x\) and \(y\), respectively, and \(\sigma_x\) and \(\sigma_y\) are the standard deviation of these values. CCN-relevant particle number concentrations have been adjusted to \(S = 0.2, 0.4, \text{ and } 1.0\%\), respectively. The particle extinction coefficient refers to the 532-nm CALIOP wavelength. The quantities are considered for certain months of the years 2015 and 2016 and for different pressure levels (1000, 850, 700, and 500 hPa). Temporal correlation coefficients are computed for each grid box from daily pairs. In addition, spatial correlations are performed for respective longitude-latitude fields and all timesteps.
3 Results

3.1 Comparison of simulated CCN and simulated AEC

The globally mapped MACC-simulated CCN abundance near surface shows maximum values over continental regions with the main sources for atmospheric aerosol (Figure 1, leftmost panels). The distribution of the maximal CCN concentration corresponds to the maximum values of the local AEC on the map of the 1000-hPa layer.

![Figure 1: Map of MACC-simulated CCN$_{0.2\%}$ in m$^{-3}$ (left panels) and local (model layer) AEC$_{532\text{ nm}}$ in m$^{-1}$ (middle panels), averaged for Jan, Mar, Jun and Oct of 2016. Right panels: the Pearson coefficient of linear correlation computed for each model grid box of daily pairs for model layers at 1000 hPa, 850 hPa, 700 hPa, and 500 hPa, respectively.]

The mapping of the Pearson coefficient for a linear relation reveals that CCN fields correlate reasonably well with the aerosol extinction for large areas of the
globe at corresponding pressure levels (Figure 1, rightmost panels). The global mean correlation coefficient is $r = 0.63$ for the surface-near layer and constantly increases with enhanced model level.

The vertically resolved CCN show a high correspondence to aerosol extinction throughout the troposphere (Figure 2). The correlation between the temporal mean vertical profiles of the quantities is statistically robust with a global mean Pearson coefficient $r$ ranging from 0.6 to 0.8.

Even though the globally averaged Pearson coefficient of the CCN and AEC fields reveals reasonable values, a statistically robust CCN-extinction correspondence does not continuously emerge for all areas of the globe, especially not for the surface near layer, as evident by Figure 1. Light extinction appears to be higher
Results

in relation to the respective CCN concentration for significant parts of the Sahara and the Arabian Peninsula, and for large areas of the Southern Ocean in a band of approximately 50°S to 70°S. These differences are supported by significantly down-grading correlations on the respective map. The Pearson coefficient dominates with negative values, indicating that CCN and AEC are anti-correlated at the prescribed regions of the Earth. On the other hand, for large parts of the continents (e.g. North/South America, and Europe) the correlation appears as continuously robust, which is evident by high values of positive Pearson coefficients.

By paying attention to higher altitudes of the global maps of Figure 1, an improving global mean correlation is notable: The Pearson coefficients of linear correlation adds up to global mean values of $r = 0.64$ at 850 hPa, and $r = 0.74$ at 700 hPa and 500 hPa, respectively. However, the predefined regions in the Southern Ocean and the Sahara reveal a different behaviour in terms of the changing CCN-extinction relationship in the vertical: The correlation consistently improves with enhancing model level above the ocean area (also evident in the rightmost panel of Figure 1). On the other hand, above the African continent the area of negative correlation shifts towards the Saharan dust outflow with rising altitude.

![Figure 3: Map of regions used in the analysis, referred to as Sahara (red), South America (black) and the Southern Ocean (blue).](image)

The concentration on individual regions of the Earth should further reveal reasons of a locally varying relationship between the CCN and AEC. The areas of interest are defined on the map of Figure 3 and account for parts of the Sahara, the Southern Ocean (in a band of 50°S to 70°S), and South America.
### 3 Results

Mean vertical profiles of CCN concentration and the extinction coefficient of the Southern Ocean reveal smaller values for both quantities in the lower parts of the atmosphere compared to the other regions chosen (Figure 4). In addition, the correlation coefficient is low at the surface, with \( r = 0.3 \), but consistently rises with altitude, which corresponds to the maps of the Pearson coefficient for that region in Figure 1. Up to a pressure of around 850 hPa, the CCN concentration in the Southern Ocean increases more distinctly with altitude and corresponds to light extinction in a better way. This is evident, since the correlation profile of the Southern Ocean starts to overlap with the continuously robust global mean profile of Pearson’s coefficient. The selected region of the African continent dominates by light extinction throughout the whole troposphere, compared to the other panels. In addition, the correlation is below 0.5 for pressure levels extending from the surface up to 500 hPa. The low correlation is supported by the negative values of the correlation coefficient for the respective region in the maps of Figure 1. The CCN-extinction relationship found for South America exceeds the global mean in the lower layers, and stays robust throughout the troposphere.

![Figure 4: Vertical profiles (averaged for Jan, Mar, Jun, and Oct of 2016) of MACC-simulated CCN_{0.2 %} (red; in m^{-3}) and the AEC_{532 nm} (blue; in m^{-1}) for the Sahara, the Southern Ocean and South America, respectively. Mean profiles of the temporal correlation coefficient for the regions of interest and for the whole globe (right panel).](image-url)
3 Results

<table>
<thead>
<tr>
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<td>0.642</td>
<td>0.638</td>
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<td>700 hPa</td>
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<td>500 hPa</td>
<td>0.738</td>
<td>0.736</td>
<td>0.725</td>
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</table>

Table 2: Coefficients of Pearson correlation between MACC-simulated AEC$_{532 \text{nm}}$ and CCN with respect to different values of supersaturation (0.2, 0.4, and 1.0 %) for different model levels.

For another comparison, different values of the supersaturation $S$ in the CCN parameterization scheme have been selected. Table 2 reveals a slightly degrading CCN-extinction relationship with higher values of supersaturations for the selected model layers, respectively. In addition, for each supersaturation, the correlation enhances with altitude until about 700 hPa.

3.2 Comparison of simulated AEC/CCN and CALIOP scattering ratios

Finally, retrievals from the spaceborne lidar CALIOP are applied for an evaluation with the simulated quantities from the MACC aerosol analysis. The midpoint of each of the 15 SR-bins of the joint histogram is used for calculating a scattering ratio per grid box.
Figure 5: Map of GOCCP-derived SRs and MACC-simulated local (model layer) AEC$_{532\ nm}$ (right), averaged for three months (Feb, Mar, Apr) of 2015. The quantities are determined for corresponding atmospheric layers at 1000 hPa, 850 hPa, 700 hPa, and 500 hPa, respectively. Correlation coefficients between corresponding latitude-longitude fields appear with $r_{1000\ hPa} = -0.056$, $r_{850\ hPa} = 0.289$, $r_{700\ hPa} = 0.313$, and $r_{500\ hPa} = -0.123$. 
3 Results

The global mapping of the GOCCP-derived scattering ratios reveals a variable degree of correspondence with the MACC-simulated extinction quantities for the atmospheric levels of interest (Figure 5). The Pearson coefficient of corresponding latitude-longitude fields shows the best result with a value of $r = 0.313$ at the 700-hPa layer. Even though the global mean Pearson coefficient is comparatively low, correspondences between the quantities are found in an area extending from West Africa eastward, across the Arabian Peninsula and southwest Asia into Central China. The same occurs for the 850-hPa layer, even though less significantly. For the atmospheric pressure level at 850 hPa and the near-surface level big parts of the Tropical Ocean stick out in terms of comparatively high SRs. The global correlation deteriorates to a Pearson coefficient of $r = 0.289$ at 850 hPa. For the lowest layer and for the upper troposphere (at 500 hPa) no distinct correspondence between simulated AECs and the GOCCP-derived SRs is evident, and the parameters even appear as slightly anti-correlated on a global average.

In the next step, the selected bins for the scattering ratios (SR between 1 and 3) have been converted to respective bins for the MACC-simulated AEC via deriving a middle extinction coefficient per interval (Figure 6). Considering the results from the modelling study, the extinction data can be related to CCN fields, assuming a linear relationship between the modelled quantities. This assumption includes some uncertainties, especially near the surface, where the global mean Pearson coefficient has been identified as lowest (see Figure 1). However, the scaling can serve as a rough orientation for the connection of the CALIOP product and global CCN abundance. Note that in the following description the expressions SR and AEC refer to the mean values of the chosen bins of the joint histogram only.
3 Results

Figure 6: Pairs of SR (y-axis) and MACC-derived AEC<sub>532 nm</sub> (x-axis, bottom) for each SR bin between 1 and 3, that are rescaled to CCN<sub>0.2%</sub> (x-axis, top). Pairs of values have been generated for the SR bins, via converting each bin midpoint to a respective AEC (taking into account all grid boxes and timesteps of the MACC-AEC data set). SR-AEC<sub>(CCN)</sub> pairs are formed for corresponding atmospheric pressure levels at (a) 1000 hPa, (b) 850 hPa, (c) 700 hPa, and (d) 500 hPa, respectively. Linear correlation coefficients are calculated for each atmospheric layer chosen.

$\text{r} = 0.956$

$\text{r} = 0.963$

$\text{r} = 0.885$

$\text{r} = 0.490$
3 Results

For the considered atmospheric layers a degrading linear regression coefficient $r$ is evident with increasing altitude, by comparing the different panels of Figure 6. Panel (a) and (b), representing the near-surface level and the 850-hPa layer, respectively, reveal a robust linear regression with coefficients $r$ higher than 0.95. For lower values of the SRs, the slope between successive data points is comparatively steep and flattens with increasing values. For values of the SRs of 2 and higher, the regression appears as almost perfectly linear, especially evident in panel (a) for the layer near surface. Panel (c) of the 700-hPa level features a similar shape in terms of the data point distribution. However, the slope between successive data points for smaller values of the compared quantities is slightly tilted towards negative, which becomes even more obvious for the 500-hPa pressure level. However, a robust linear connection emerges in an interval of approximately $2 \leq \text{SR} \leq 3$ for all vertical levels chosen.
4 Discussion

Globally mapped Pearson coefficients of linear correlation compare to the common understanding that CCN are well correlated to aerosol extinction (Figure 1). All pressure levels reveal a correlation coefficient \( r \) above 0.6. The vertical distribution of spatial correlation coefficients (Figure 2) further shows a consistently robust relationship throughout the whole troposphere. Note that the CCN peak (Figure 3) is around an atmospheric pressure of 700 hPa, as the aerosols are only being activated as long as \( S_{\text{max}} \) is larger than \( S_{\text{crit}} \) and the air parcel is assumed to rise adiabatically (Block et al., 2017).

Despite the robust global-mean correlation, atmospheric aerosol varies in terms of particle size, chemical composition and mixing state (Lamb and Verlinde, 2011; Quinn et al., 2008). It has been pointed out in section 2.1.2 that certain factors impact CCN and extinction differently and may have an impact on the CCN-extinction relationship. Consequently, errors are incurred in calculating the global mean correlation. For the lowest chosen pressure layer (1000 hPa), the correlation between CCN and particle extinction is robust for large areas of the globe. However, the frequent presence of dust and sea salt particles near the surface (where most aerosol sources are found) degrade the correlation. Outside the boundary layer speciation errors matter less as the aerosol fields are more homogeneous.

By looking deeper at specific features of these aerosol types, variations in the correlation may be better understood: Sea salt can dominate large parts of the planet’s oceanic regions with 50 to 70 % of aerosol masses (Boucher et al., 2013). The particles can be expected as particularly numerous in areas of the Southern Ocean (see Figure 2 of Bellouin et al., 2013 for marine aerosols). Sea salt aerosols are found in coarse and accumulation modes and are effective at light extinction. The salt aerosols are entirely treated as highly hygroscopic species in the parameterization scheme due to their chemical features. Both size and composition characteristics add up to a strong CCN activity in the parameterization via Köhler theory. However, for lower parts of the troposphere, CCN are less numerous compared to the corresponding particle extinction, which is better seen from the degraded correlation profile of the Southern Ocean (Figure 4, rightmost panel). This might be verified as follows: As pointed out by Stier (2016), ambient humidity conditions not only account for water uptake by atmospheric aerosols, but also particle removal via scavenging. The wet deposition parameterization schemes include in-cloud scavenging by large-scale and convective precipitation, and below-cloud washout when falling droplets entrain more aerosols to fall towards the surface (Bellouin et al., 2013). The wet (and dry) deposition schemes were adapted directly from the Laboratoire de Météorologie Dynamique (LMD) model (Benedetti et al., 2009) and are detailedly described in Reddy et al. (2005); Boucher et al. (2002). From Table 3 of Reddy et al. (2005) it is seen that most of the sea salt aerosols are emitted in larger size modes and the
removal is dominated (with 69%) by dry deposition due to the comparatively high gravitational force. However, 31% of the aerosol mass is removed by wet deposition and even 91% of aerosol masses in the smaller size bins. Since the humidity in Southern Ocean regions is high, the aerosol scavenging processes might be very efficient and account for decreased correlation coefficients below the 850-hPa pressure level.

Another region of interest is defined by the chosen area of the Sahara. The high extent of dust particles in the desert makes this part of the planet interesting for investigating the CCN-extinction relationship. Mineral dust is predominantly found in coarse and super-coarse modes, with a small accumulation mode (Boucher et al., 2013). The dust particles appear to be highly dominant by extinction due to their large size, but they are rather hydrophobic. In the CCN parameterization, dust is entirely considered to not contribute to CCN number concentrations, which accounts for a low correlation throughout the whole troposphere. Dust particles in the MACC aerosol analysis have been parameterized in terms of near-surface wind stress, effecting particle dispersion from the Earth’s surface. Depending on the weather conditions, dust may stay in the atmosphere for a long time and is transported farther from its sources, affecting the geographic pattern of aerosol impacts. That is supported by the maps of the Pearson coefficient (Figure 1), as the area of low correlation gets shifted westward towards the Saharan dust outflow into the Atlantic Ocean with enhanced model level.

The selected region for South America (as well as other continental regions) accounts for robust correlations throughout the troposphere (neglecting levels above 500 hPa). In those areas, anthropogenic aerosols (biomass burning) are presumably dominating atmospheric CCN fields. These particles are predominantly represented in Aitken, and accumulation ranges. The extinction is less distinct for those modes and seems to correspond to CCN in a better way.

As previously mentioned, not only particle properties influence the CCN-extinction relationship, but also the ambient environment, especially the humidity. A decreasing Pearson coefficient has been identified for higher supersaturations (see Table 2): The enhancement of $S$ leads to a higher particle activation rate of aerosols with smaller radii. Particles in smaller modes significantly add to CCN number concentrations at appropriate $S$ (number peak), but scarcely contribute to light extinction. However, the degradation of the Pearson coefficient is not significant, which is presumably since at 0.2% supersaturation, the number peak of the aerosol spectrum is already included in the CCN concentration (see 2.1.2 Particle size).

The results of the modelling study clearly prove that the considered influential properties of the quantities themselves, but also from the ambient environment, complicate the CCN-extinction correlation. However, a linear connection between modelled CCN and AEC can be considered as a good approach, especially for the
700-hPa and 500-hPa pressure level. However, from an aerosol-cloud perspective, CCN are most relevant at the cloud base altitude. Significant cloud particle formation processes particularly occur at altitudes around 700 hPa (which is at roughly 3 km). Consequently, the atmospheric level at 500 hPa matters less for describing the relationship between aerosol optical characteristics and cloud microphysical features, and improving the quantification of aerosol-cloud interactions.

By combining simulated quantities and spaceborne lidar products, high correlation coefficients for the surface layer as well as 850/700-hPa pressure levels are shown in Figure 6. However, the low SR bins (between 1 and 2) show very little variation for changes in modelled AEC/CCN. This might be explained as follows: The Mie regime for atmospheric radiation at 532 nm is applicable for particles with radii that are higher than approximately 0.01 µm (Petty, 2006). On the other hand, the extinction efficiency appears negligibly small for particle radii ranging from 0.01 to 0.1 µm (see Figure 2.39 of Lamb and Verlinde, 2011). Note that for radii below 0.01 µm the Rayleigh approximation is applicable, which describes radiation scattering by air molecules for the considered wavelength (Petty, 2006). The lidar SR was introduced as the ratio of the lidar backscatter signal to the molecular scattering signal. Consequently, for SR = 1, the lidar backscatter signal is due to scattering by molecules. As mentioned above, for radiation at 532 nm, the Rayleigh approximation is assumed only for scattering by air molecules. Hence, the interval of SR bins ranging from 1 to 2 can be understood as a “transition” zone where the Rayleigh approximation is not suitable anymore, but Mie theory reveals only slightly rising extinction efficiencies for increasing particle size. In summary, SRs only in an interval ranging from 2 to 3 can be successfully related to the simulated quantities.

The global mapping of the scattering ratio and the respective aerosol extinction data appears to be highly uncertain in terms of global mean correlations (Figure 5). Nevertheless, as already pointed out in the modelling part, globally averaged correlations involve large error sources in describing the relationship. Therefore, certain regions of the globe emerge where a linear connection might be appropriate. This is particularly obvious for the aerosol-cloud relevant layer at 700 hPa (and 850 hPa). The scattering ratios are strongly tied to aerosol extinction for the regions of the so-called “dust belt”, reaching from West Africa eastward into Central China. On the other hand, the global correlation seems to be mainly deteriorated by large areas of the Tropical Ocean. High SRs are evident along typical tropical cyclone paths, where a large extent of optically active marine aerosol is expected, as the particles are effectively being injected into the planetary boundary layer via sea spray (Boucher et al., 2013). A large error in the mean correlation might occur due to the fact that the extinction-to-backscatter ratio (lidar ratio) is not considered in the study. However, different aerosol backgrounds obey certain lidar ratios that
need to be included in the transformation algorithm from direct lidar products to extinction data.

Omar et al. (2009) establishes a way to derive aerosol extinction data from lidar-retrieved backscattering with assumed lidar ratios for each aerosol mode. The study applies the polarisation lidar technique of CALIOP to generate vertically resolved distributions of aerosol types and their respective optical characteristics. The lidar ratio of aerosol products appears as one variable for each aerosol mode. The 532 nm lidar ratio of clean marine aerosol is much smaller (20 sr) compared to e.g. dust (40 sr), or polluted dust (65 sr). Therefore, a lower level of aerosol extinction would be derived for typical marine-aerosol regions compared to other areas of the globe. This matches the panel for the MA CC-extinction, showing smaller values compared to high SRs in the Tropical Ocean. Hence, it can be expected that the global mean linear correlation will improve if specific lidar ratios are considered.
5 Conclusion

A commonly used quantity for estimating global CCN abundance is the AOD. However, errors emerge by integrating aerosol light extinction over the whole atmospheric column. CCN distributions might be more accurately derived from vertically resolved aerosol optical features.

The comparison of simulated CCN and AEC confirms that CCN are closely related to aerosol extinction in terms of a linear correlation. These results strongly agree with the modelling study of Stier (2016). If aerosols were consistent in terms of size, composition and humidity response, the aerosol extinction coefficient would yield a first-class approach for CCN. However, certain impacting factors provoke a variable degree of linear correlation. This has been verified by comparing different planetary regions with certain aerosol backgrounds, and by changing environmental conditions (humidity). The results match with the findings of e.g. Shinozuka et al. (2015), Liu and Li (2014) and imply that the accuracy of parameterization schemes can be further improved by taking in account those factors more accurately.

Even though numerical models possess a consistent field of global aerosol, parameterization schemes are approximations to the environmental conditions. The simulation of atmospheric constituents is associated with a degree of uncertainty due to assumptions in emission, transport and nonlinear physical processes involving aerosols. Consequently, the simulation might vary markedly from observations.

On a global scale, aerosol optical properties are easier to measure than CCN. The work of Winker et al. (2013) shows a high potential of spaceborne lidar techniques to represent tropospheric aerosol, considering retrievals from CALIOP. The derived extinction profiles appear to be suitable for providing a comprehensive view of global tropospheric aerosol abundance. Consequently, lidar observations from space are considered a reliable method in terms of aerosol vertical profiling, from which CCN can be estimated.

Relating the CALIOP products to the modelled quantities shows that the correlation is difficult on a global scale and needs the inclusion of certain assumptions (e.g. lidar ratios). However, for altitudes that are particularly interesting for aerosol-cloud interactions, the lidar products might sufficiently represent CCN abundance.

In summary, relating the products from spaceborne lidar techniques to CCN is implied to be a convenient solution for advancing the understanding of so far highly uncertain aerosol-cloud interactions.
References


References


Selbstständigkeitserklärung


Leipzig, 19.12.2017

Ort, Datum

Unterzeichner