



The role of mineral dust in cloud droplet formation processes: Insights from the Central Asian dust belt subregion

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Abstract.

In this article, the contribution of aged desert dust, contaminated with hygroscopic material, to droplet formation in liquid-water-containing tropospheric clouds in Central Asia is investigated. The study is based on long-term lidar observations of aerosol and cloud properties at Dushanbe (38.6°N, 68.9°E; 864 m height), Tajikistan. The continuous measurements, performed
5 between June 2019 and July 2023, cover three summer and three winter seasons in terms of cloud condensation nucleus concentration (CCNC) of the dust and non-dust (anthropogenic) CCN reservoirs and cloud droplet number concentration (CDNC) and effective radius of the water droplets in the base (lowest 100 m) of liquid-dominated cloud layers. The CDNC contains information about the actually activated CCN fraction of the total CCN reservoir. The data analysis reveals that aged, hygroscopic dust particles are able to sensitively influence or even control CDNC levels during the dust season (April to
10 October) and that the impact is relatively weak during the main winter months (December to February) when the dust particle concentration is partly very low and aerosol pollution from anthropogenic sources dominates. Case studies are presented to check the consistency in the retrieval products and to show the complexity of aerosol-cloud interaction in a region, probably representative for large parts of Asia, with aerosol mixtures of fresh and aged dust, anthropogenic haze, rural background aerosol, and even wildfire smoke after long-range transport. Water uptake by dust particles, observed at high relative humidity
15 close to cloud base, is a clear sign for the presence of dust particles containing hygroscopic material.

1 Introduction

Mineral dust is a key aerosol component in the atmospheric system. Soil particles, especially desert dust particles, influence solar and terrestrial radiation fluxes and cloud and precipitation formation. Dust particles are omnipresent in the atmosphere



from the surface to the tropopause and from the tropics to high northern and southern latitudes (Proestakis et al., 2025a, b; Lian et al., 2022). In the free troposphere, dust particles can be transported over more than 10000 km without efficient removal (Haarig et al., 2017). The most prominent examples are the long-range transport of Saharan dust towards the Caribbean, South, Central, and North America (Karyampudi et al., 1999; Rittmeister et al., 2017) and Asian dust plumes crossing Eastern Asia and the Pacific towards North America (Sassen, 2002; Shen et al., 2024).

It is well accepted that dust particles are efficient ice-nucleating particles (INPs) (Niemand et al., 2012; DeMott et al., 2015; Kanji et al., 2017; Knopf and Alpert, 2023) and widely influence ice formation in the troposphere (Froyd et al., 2022). Our studies confirm the strong impact of dust on the evolution of mixed-phase and cirrus clouds (e.g., Ansmann et al., 2008, 2009, 2019; Seifert et al., 2010; Kanitz et al., 2011).

In contrast to the role of dust in ice nucleation processes, the importance of dust particles to serve as cloud condensation nuclei (CCN) is not well understood and subject to a controversial discussion. This is related to the fact that dust particles close to the emission sources (e.g., deserts) may show a hydrophobic behavior (Kaaden et al., 2009) so that only large dust particles with diameters $> 1\text{--}2\text{ }\mu\text{m}$ may be able to form liquid-water droplets (Kelly et al., 2007) via water vapor adsorption. This process results in a surface film of water with reduced activity and promotes the formation of cloud droplets, even in the case of freshly emitted and chemically unprocessed dust particles (Sorjamaa and Laaksonen, 2007; Kumar et al., 2009; Laaksonen et al., 2016; Karydis et al., 2011, 2017). During long-range transport, chemical aging, i.e., interaction with trace gases via heterogeneous chemical reactions with acids such as HNO_3 and HCl on the surfaces of the dust particles takes place (Gibson et al., 2007; Kelly et al., 2007; Tang et al., 2016). Furthermore, dust particles provide reaction sites for the SO_2 oxidation into H_2SO_4 . These processes result in a coating of dust particles with soluble material, which augments the hygroscopicity of dust particles and therefore their ability to act as CCN (Karydis et al., 2017). In addition, cloud processing, i.e., the accumulation of soluble material on the surfaces of the dust particles after evaporation of cloud droplets that contained scavenged dust particles (Levin et al., 1996; Wurzler et al., 2000) occur so that dust particles may become increasingly hygroscopic and thus show an increased droplet nucleation efficiency. Aged, contaminated dust particles can be regarded as excellent CCN because of the favorable combination of accumulated soluble material on the dust particles and their comparably large sizes (Feingold et al., 1999; Levin et al., 2005).

The increasing complexity of aerosol mixtures introduces an additional challenge, namely the definition and identification of dust events. In many dust-affected regions, especially in the Eastern Mediterranean, Middle East, and Central Asia, mineral dust is frequently mixed with anthropogenic particles and may occur in elevated layers that are decoupled from the surface (Rogozovsky et al., 2023, 2024). Rogozovsky et al. (2024) showed that more than half of the dust-containing days detected by lidar observations over the Eastern Mediterranean were not captured by widely used dust-identification methods, based on in-situ observations or passive remote sensing from ground or space. The study further demonstrated that many dust episodes exhibit pronounced anthropogenic contributions and therefore cannot be unambiguously classified as either pure dust or pollution events. These findings suggest that the definition of a dust event remains a critical issue in aerosol-cloud-interaction (ACI) studies and emphasize the need for vertically resolved observations.



In addition to mineral dust sources, Central Asia and Iran are affected by saline dust emissions originating from desiccating lakes such as the Aral Sea, Lake Urmia, and Hamoun Lake (Xi and Sokolik, 2016; Hofer et al., 2017, 2020b; Banks et al., 2022; Moghim and Ramezanpoor, 2019; Moghim and Goftari, 2024; Alizadeh et al., 2026). These sources may contribute by roughly 10% to the regional aerosol burden (Banks et al., 2022) and have therefore to be considered in ACI studies in Central Asia.

Liquid-water cloud formation in a dusty and polluted environment remains a complex process with many facets. In the northern hemisphere, especially in Asia and Africa, there is practically no region without strong pollution and dust in the lower to middle troposphere, so that mixtures of pure and contaminated dust and anthropogenic aerosols particles (and existing cloud droplets) compete for the available excess water vapor during updraft events and respective water supersaturation conditions in liquid-water cloud processes. The impact of the natural aerosol component dust must be included in the ACI debate (see, e.g., Jia and Quaas, 2023; Gryspeerdt et al., 2023) when making hypotheses about the future climate development in a cleaner world with strongly reduced anthropogenic pollution.

Meanwhile, a number of model-based studies have been conducted and show the impact of aged and contaminated dust particles on water cloud formation (see the reviews in the articles of Karydis et al. (2017) and Milousis et al. (2025)). However, model-based findings and conclusions need to be confirmed by observations, preferably by long-term observations over years in key region with strong dust burden such as the polluted Eastern Mediterranean (Mamouri et al., 2016; Rogozovsky et al., 2024, 2025, 2026; Chudnovsky et al., 2026), Middle East, and Central Asia (Hofer et al., 2017, 2020a) or western Africa (Tesche et al., 2009; Gebauer et al., 2026). Such long-term observations are missing. Several field campaigns have investigated the CCN activity of mineral dust (Roberts et al., 2006; Sullivan et al., 2007; Twohy et al., 2009; Yamashita et al., 2011). Particularly, Bègue et al. (2015), using satellite, airborne, and ground-based observation and modeling, demonstrated that long-range transported Saharan dust accumulated soluble material and exhibited enhanced CCN compared with freshly emitted dust.

Motivated by the lack of required long-term observations of dust-cloud interaction, we upgraded our remote sensing (lidar) capabilities at Dushanbe, Tajikistan (Hofer et al., 2017; Engelmann et al., 2019; Müller et al., 2026) in Central Asia to cover both, observations of cloud and aerosol microphysical properties to allow for detailed liquid-phase ACI studies. The most promising way to investigate the impact of given aerosol conditions on the nucleation of liquid-water droplets is the study of the relationship between observations of the CCN concentration (CCNC) below cloud base and of the cloud droplet number concentration (CDNC) just above cloud base. Such observations can be provided by ground-based lidars (Schmidt et al., 2014, 2015; Jimenez et al., 2020a, b). However, to permit robust conclusions on ACI, one needs continuously monitoring lidars, collecting cloud and aerosol profile data over months, seasons, and years. With respect to dust-related ACI features, polarization lidar is of advantage because it permits the separation of dust and non-dust CCNC fractions (Mamouri and Ansmann, 2016).

To meet these requirements, we equipped the Dushanbe Polly (PORTabLe Lidar sYstem) lidar instrument with the recently introduced dual-field-of-view (dual-FOV) polarization lidar technique (Jimenez et al., 2020a, b, 2025). The dual-FOV lidar allows us to characterize droplet populations in the lowest 100 m of liquid-water and mixed-phase clouds in terms of cloud droplet number concentration (CDNC), effective, i.e., surface-area-weighted mean radius of the droplets, liquid water content,



and cloud extinction coefficient. Together with the lidar-derived aerosol products as presented by (Hofer et al., 2017, 2020a, b) we are now able to study the impact of dust and continental anthropogenic haze particles on liquid-droplet formation in detail on a long-term basis (Lipken, 2024).

2 Data and methods

The entire data analysis consists of four major steps. First, aerosol optical properties are retrieved from the polarization Raman lidar observations at Dushanbe and converted into dust and non-dust extinction profiles. Second, these extinction profiles are transformed into estimates of dust and continental CCN concentrations using the POLIPHON (Polarization Lidar Photometer Networking) approach (Mamouri and Ansmann, 2016; Ansmann et al., 2026), while cloud droplet number concentrations are independently retrieved from the dual-FOV polarization lidar measurements (Jimenez et al., 2020a). Details to these first two steps are given in Sect. 2.3 and 2.4. Third, three representative case studies are used to evaluate the physical consistency of the retrieval framework under contrasting aerosol conditions. These case studies are presented in Sect. 3 and highlight ACI processes in a region with complex mixtures of natural and anthropogenic particles. Finally, the validated methodology is applied to the complete 2019–2023 data set comprising more than 1600 cloud events to investigate seasonal aerosol–cloud interactions and the long-term variability of CCN activation over Central Asia and to quantify the contribution of mineral dust to cloud droplet formation. The Dushanbe lidar data analysis and parts of the ACI study were presented first in the master thesis of Lipken (2024).

2.1 Study area

Tajikistan is located in the central part of the so-called dust belt (Fig. 1) and is frequently affected by severe dust storms during the summer half year. Strong contributions from residential heating and fossil fuel combustion in winter control the air quality and aerosol characteristics from November to March (Fomba et al., 2024; Müller et al., 2026). Main source regions for particles in the main aerosol layer which frequently reaches up to 3–4 km height above ground, are Iran and Central Asia, followed by the Middle East region, with minor contributions from Europe and Northern Africa (Lipken, 2024).

2.2 Supporting observations

2.2.1 PollyXT

In the framework of the pilot study CADEX (Central Asian Dust EXperiment), a joint German-Tajik project between the Leibniz Institute of Tropospheric Research (TROPOS) and the Physical Technical Institute of the National Academy of Sciences of Tajikistan, we performed combined aerosol lidar and photometer observations at Dushanbe (38.6°N, 68.9°E; 864 m height), Tajikistan, from 17 March 2015 to 31 August 2016 (Hofer et al., 2017, 2020a, b). Goal of the pilot study was to obtain a representative height-resolved data set of optical, microphysical, and cloud-process-relevant properties of Central Asian



Figure 1. Geographical map with the names of a number of desert (in black). The eastern part of the dust belt (enclosed in green dashed boundaries) extending from western and northern Africa to the Gobi desert in China over 10000 km is shown. The lidar field site at Dushanbe, Tajikistan, is indicated in red (adapted from Hofer et al., 2017). Main Central Asian dust sources are the deserts east of the Caspian Sea.

aerosol mixtures of fresh and aged dust, salt aerosol from dried lakes, anthropogenic haze and continental rural background soil particles.

Motivated by the successful CADEX campaign, we broadened our activities and designed a new project to establish two permanent dust monitoring lidar stations, one in the Eastern Mediterranean (Cyprus) and one in Central Asia (Tajikistan). Financed by the German government, we built two new PollyXT (PORTable Lidar sYstem with eXTended capabilities) instruments (Engelmann et al., 2016; Hofer et al., 2017; Jimenez et al., 2020b) for the existing lidar/photometer stations at Limassol, Cyprus (Mamouri et al., 2013; Nisantzi et al., 2014), and Dushanbe, in the framework of the PoLiCyTa (PollyXT Lidars for Cyprus and Tajikistan) project (Engelmann et al., 2019). The measurements were started in Tajikistan in June 2019 (Müller et al., 2026) and in Cyprus in October 2020 (Mamouri et al., 2023). Both lidars permit dual-FOV polarization lidar observations. The lidar monitoring project is ongoing and meanwhile covers almost six years of observations at Dushanbe (Müller et al., 2026). The Dushanbe PollyXT belongs to the lidar network PollyNET (Baars et al., 2016), which is part of the European Aerosol Research Lidar Network (EARLINET) (Pappalardo et al., 2014) organized within the Aerosols, Clouds and Trace gases Research InfraStructure (ACTRIS) project (Laj et al., 2024).

Laser pulses are transmitted at 355, 532, and 1064 nm. Overall, 13 receiver channels are available to collect elastically backscattered photons at the three laser wavelengths, the cross-polarized signal components at 355 and 532 nm, the vibrational Raman signals of nitrogen at 387 and 607 nm, and of water vapor at 407 nm. Two receiver units, equipped with two different FOVs of 1 and 2 mrad, provide near-range (150 m to about 6–7 km height) and far-range (800 m to 30 km) aerosol information and allow the cloud microphysical retrieval in clouds with base heights from 500 m to about 6 km. All signals are measured with photoncounting photomultipliers and stored with 30 s temporal resolution. Vertical resolution is 7.5 m.

The PollyXT instrument is designed for automated continuous profiling of aerosols and clouds. Well-defined breaks are, however, necessary on a daily basis to apply different automated calibration and alignment procedures such as the $\Delta 90^\circ$ -



calibration (Freudenthaler, 2016) to check the performance of the full instrumental setup. The measured lidar data are uploaded and processed automatically (Baars et al., 2016, 2017; Yin et al., 2020; Klamt et al., 2024; Hofer et al., 2026) to be displayed
140 as “quicklooks” on the PollyNET homepage (PollyNET, 2026).

2.2.2 AERONET

An Aerosol Robotic Network (AERONET) sunphotometer station (Holben et al., 1998) is operated at Dushanbe since 2010. The AERONET photometer provides AOT observations at 8 wavelengths from 340 to 1640 nm and sky radiance measurements at 440, 675, 870, and 1020 nm at cloud-free conditions from shortly after sun rise to shortly before sunset (AERONET, 2026).
145 The temporal resolution varies between 3 and 15 minutes. We make use of the measured 500 nm AOT and of the spectral slope (Ångström exponent) of the AOT in the case study section 3.

2.2.3 MODIS MAIAC AOT at 1 km spatial resolution

In Sect. 3, we also show aerosol optical thickness (AOT) fields, obtained from the MODIS (MODerate Resolution Imaging Spectroradiometer) MAIAC (Multi-Angle Implementation of Atmospheric Correction) Level-2 gridded aerosol product
150 MCD19A2 (MAIAC, 2026). MCD19A2 is part of the MODIS Collection 6 MAIAC data suite and provides daily AOT retrievals over land at 1 km spatial resolution from observations acquired by the Terra and Aqua satellites (Lyapustin et al., 2018). MAIAC retrieves AOT using a time-series analysis of MODIS observations and simultaneous retrieval of surface bidirectional reflectance and aerosol properties. The performance of the high-resolution MAIAC AOT product has been evaluated in previous studies (Chudnovsky et al., 2013; Rogozovsky et al., 2021, 2023, 2025, 2026).

155 2.3 Retrieval methods

2.3.1 Aerosol optical properties

Well-established data analysis procedures are applied to the combined Raman and polarization lidar measurements to determine height profiles of the particle backscatter coefficient β_p and the respective dust and non-dust backscatter contributions, β_d and β_{nd} , respectively (Teschke et al., 2009; Mamouri and Ansmann, 2014; Hofer et al., 2017). In this study, we exclusively used the
160 optical properties measured at 532 nm. The backscatter coefficients were multiplied by typical dust and anthropogenic-aerosol extinction-to-backscatter ratios (lidar ratios) of $S_d = 40$ sr and $S_{nd} = 50$ sr at 532 nm wavelengths (Hofer et al., 2020b; Haarig et al., 2025), respectively, to obtain particle extinction coefficients for the dust and non-dust fractions:

$$\sigma_d = S_d \times \beta_d, \quad (1)$$

$$\sigma_{nd} = S_{nd} \times \beta_{nd}. \quad (2)$$

165 The optical properties are influenced by water uptake by hygroscopic particles at high relative humidity (RH) of $>75\%$ (Haarig et al., 2025). The Raman lidar option of the Dushanbe PollyXT enables us to determine height profiles of the water-vapor-to-dry-air mixing ratio and, in combination with temperature profiles taken from ECMWF (European Centre for



Medium-Range Weather Forecasts) data sets (ECMWF, 2026), the retrieval of relative humidity profiles (Dai et al., 2018). An example of measured profiles of aerosol optical properties and RH, showing dust water-uptake effects, is presented in Sect. 3.

170 2.3.2 Retrieval of cloud condensation nucleus concentration (CCNC)

The POLIPHON method (Mamouri and Ansmann, 2016; Ansmann et al., 2026) is applied to compute estimates of dust-related and non-dust-related CCN contributions to the CCN reservoir. We assume that non-dust particles are anthropogenic aerosol particles. By using the conversion factors $c_{100,d} = 2.5 \text{ Mm cm}^{-3}$ for dust particles and $c_{50,c} = 13.7 \text{ Mm cm}^{-3}$ for the continental fine-mode aerosol (Ansmann et al., 2026), the extinction coefficients σ_d and σ_{nd} are converted into particle number concentrations:

$$n_{100,d} = c_{100,d} \times \sigma_d, \quad (3)$$

$$n_{50,c} = c_{50,c} \times \sigma_{nd}. \quad (4)$$

The obtained number concentration $n_{50,c}$ considers anthropogenic particles with dry-particle radius $>50 \text{ nm}$. The dust particle number concentration $n_{100,d}$ considers all dust particles with dry-particle radius $>100 \text{ nm}$. $n_{50,c}$ and $n_{100,d}$ are interpreted as the CCN reservoirs for continental anthropogenic aerosol particles ($N_{CCN,c}$) and dust particles ($N_{CCN,d}$), respectively, available for droplet nucleation in the cloud base region at a water supersaturation of 0.2%. The total CCN reservoir is finally given by:

$$N_{CCN,t} = n_{100,d} + n_{50,c}. \quad (5)$$

It is assumed that all particles with dry-particle radius greater than the so-called critical radius of about 50 nm for anthropogenic aerosol particles (Schmale et al., 2018) and $>100 \text{ nm}$ (dust particles) can be activated to form water droplets when the water supersaturation level in a lofted air parcel is 0.2%. The motivation to assume a specific supersaturation value of 0.2% in the CCNC estimation is guided by the following arguments. The majority of the cloud observations at Dushanbe were conducted in altocumulus and stratocumulus cloud decks that formed in the upper part or at the top of the main aerosol layer (frequently coinciding with the daytime convective boundary layer or nighttime residual layer).

After the establishment of a closed cloud layer with an optical depth usually exceeding 2–3, the further cloud evolution is driven by radiative cooling of the cloud top region. The cooling initiates a vertical circulation system with regularly occurring downdrafts and updrafts (Shupe et al., 2008; Ansmann et al., 2009; Morrison et al., 2012; Jimenez et al., 2025). Updraft velocities at cloud base are typically $\leq 1 \text{ m s}^{-1}$ (Korolev and Mazin, 2003; Ansmann et al., 2009; Schmidt et al., 2014, 2015; Yang et al., 2019; Jimenez et al., 2020b, 2025). For such low updraft speeds the water supersaturation in the lofted air parcels entering the cloud layer from below is typically of the order of 0.2% or less (Korolev and Mazin, 2003; Siebert and Shaw, 2017; Yang et al., 2019). At these supersaturation conditions, new droplets can form and existing droplets can grow (Korolev and Mazin, 2003). Entrainment of dry air, droplet collision and coagulation processes control the finally established droplet size distribution. In downdrafts, reaching heights below the cloud layer, droplets evaporate. These cloud evolution features



were observed in polar and tropical altocumulus fields (Ansmann et al., 2009; Jimenez et al., 2025) and we assume that these
200 specific features also hold for midlatitude stratiform cloud layers.

Lipken (2024) analyzed the atmospheric stability conditions and related potential water supersaturation conditions over
Dushanbe as a function of season. The study was based on GDAS (Global Data Assimilation System, GDAS, 2026) mete-
orological data. The atmospheric stability parameter (Brunt-Väisälä-Frequency) showed highest values in winter, indicating
stable layering conditions with a weak potential for strong convective motions. Compared to the dust season (April to Oc-
205 tober), updrafts are thus expected to be, on average, much weaker in winter and, correspondingly, supersaturation levels in
droplet formation events are, on average, lower in winter (0.05%–0.1%) than in summer (around 0.2%). This expectation is in
line with simulations of Karydis et al. (2017). Sapkal et al. (2025) shows that the number of nucleated droplets and thus the
number of activatable CCN roughly decreases by a factor of 2 and 4 when the supersaturation decreases from 0.2% to 0.1%
and 0.05%, respectively. The critical (minimal) radius of activatable particles increases with decreasing water supersaturation
210 with the consequence that our CCN reservoir estimates (Eqs. 3–5 for 0.2% supersaturation) are too large by a factor of 2–4
when the given water supersaturation is of the order of 0.05% to 0.1%.

2.3.3 Retrieval of cloud droplet number concentration (CDNC)

The CDNC retrieval method is based on the dual-FOV polarization lidar observations and outlined in detail in Jimenez et al.
(2020a). Briefly summarized, this lidar technique enables us to determine multiple scattering by droplets in the liquid-water-
215 dominated cloud base zone from measurements of the volume linear depolarization ratio (at 532 nm) with two different FOVs.
The depolarization ratio is defined as the ratio of the cross-polarized to the co-polarized elastic backscatter coefficient. The
notation “cross” and “co” is used to indicate the plane of polarization orthogonal and parallel to the plane of linear polarization
of the transmitted laser pulses. From the difference in the light-depolarization strength, measured with two photomultipliers
(channels) for the two different FOVs, cloud microphysical properties such as effective droplet radius R_{eff} , the liquid water
220 content, and the cloud extinction coefficient σ_{cld} within a layer 50–100 m above cloud base, i.e., for a height of about 75 m
above cloud base can be derived in a very straightforward way. Finally, CDNC, traditionally denoted as N_d , is given by
(Schmidt et al., 2014; Jimenez et al., 2020a)

$$N_d = \frac{\sigma_{\text{cld}}}{2\pi k R_{\text{eff}}^2}. \quad (6)$$

For typical cloud droplet size distribution (modified gamma functions) in stratiform cloud layers, $k = 0.75 \pm 0.15$ (Miles et al.,
225 2000) has been found. With the relative uncertainty of 15% for R_{eff} , 20% for σ_{cld} , and 20% for k , the relative uncertainty of
 N_d is 30–35%. All measurement and retrieval products, used in the present study, are listed in Table 1. In addition, typical
relative errors caused by signal noise and required input parameters in the data analysis are given in the Table 1.



Table 1. Overview of Polly observational products, used in this study, and characteristic (typical) relative uncertainties in the determined and retrieved properties. r denotes aerosol particle radius.

Aerosol optical properties	Uncertainty	
Backscatter coefficient [$\text{Mm}^{-1} \text{sr}^{-1}$]	$\beta_p(532 \text{ nm})$	$\leq 10\%$
Extinction coefficient [Mm^{-1}]	$\sigma_p(532 \text{ nm})$	20%
Depolarization ratio	$\delta_p(532 \text{ nm})$	$\leq 10\%$
Cloud-relevant aerosol properties		
Dust CCN conc. ($r > 100 \text{ nm}$) [cm^{-3}]	$N_{\text{CCN,d}}$	35%
Non-dust CCN conc. ($r > 50 \text{ nm}$) [cm^{-3}]	$N_{\text{CCN,c}}$	50%
Total (dust and non-dust) CCN conc. [cm^{-3}]	$N_{\text{CCN,t}}$	50%
Droplet properties (liquid water clouds)		
Droplet number concentration [cm^{-3}]	N_d	35%
Droplet effective radius [μm]	R_{eff}	15%
Water vapor profiles		
Water-vapor-to-dry-air mixing ratio [g kg^{-1}]		$\leq 10\%$
Relative humidity [%]		10–20%

2.3.4 Aerosol-cloud-interaction (ACI) metrics

The link between the total CCNC (Eq. 5) and CDNC (Eq. 6) is frequently described by the susceptibility $s_{\text{ACI,liq}}$ (Feingold et al., 2001; Gryspeerd et al., 2023; Choudhury et al., 2026), defined as

$$s_{\text{ACI,liq}} = \frac{d \ln N_d}{d \ln N_{\text{CCN,t}}} \quad (7)$$

The susceptibility expresses the relative change of the droplet number concentration with a relative change in the respective aerosol loading. Reutter et al. (2009) distinguished three different regimes of CCN activation and cloud droplet formation, an updraft-limited regime, an aerosol-limited regime, and an aerosol- and updraft-sensitive regime (transitional regime). Updraft-limited conditions are preferably given when updraft velocities at cloud base are low so that the water supersaturation is low ($\leq 0.2\%$) and when, in addition, the ratio $N_d/N_{\text{CCN,t}} < 0.2$. The total aerosol particle number concentration $N_{\text{CCN,t}}$ considers all aerosol particles per cubic centimeter with diameters $> 10 \text{ nm}$. In this regime, N_d is directly proportional to the updraft velocity and practically independent of $N_{\text{CCN,t}}$, represented in our approach by the $N_{\text{CCN,t}}$ estimate, so that the susceptibility $s_{\text{ACI,liq}}$ is close to zero. In the aerosol-limited regime, the water supersaturation is usually $> 0.5\%$ and the fraction $N_d/N_{\text{CCN,t}}$ (in our approach $N_d/N_{\text{CCN,t}}$) of activated particles is high, ≈ 0.9 . In this regime, N_d is directly proportional to $N_{\text{CCN,t}}$ (or in our approach to the $N_{\text{CCN,t}}$ estimate) and the susceptibility $s_{\text{ACI,liq}}$ is close to one. According to these definitions, the



Dushanbe observations during the dust season belong to the transient regime, while updraft-limited conditions clearly prevail during the winter months.

2.4 Automated processing of the CCNC and CDNC database

245 The entire multiyear data analysis leading to the results discussed in Sect. 4 was automated. All in all, we analyzed 1602 cloud fields and the related aerosol environmental conditions (Lipken, 2024). In the case of multiple-layered cloud systems, only the lowest layer was considered. The analyzed cloud layers developed preferably in the dusty and polluted main aerosol layer. Examples are shown in Sect. 3.1–3.3. As mentioned in Sect. 2.3.1, the main aerosol layer often coincides with the convective boundary layer at daytime or the stable residual layer at nighttime. The analysis of the 1602 cloud fields was based on 45724
250 individual lidar observations, each averaged over 3 minutes. Each data set consists of the CDNC, cloud extinction coefficient, the liquid water content and droplet effective radius at 75 m above cloud base. Cloud layers were counted as individual events when they occurred at different heights. Cloud layers at the same height level were counted as individual events, when they were separated by >3600 s. The majority of cloud events showed lengths from 10 minutes up to 3–4 hours. The average (median) temporal length of the observed 1602 individual coherent cloud fields was about 1 h. In the discussion of the relationship
255 between $N_{CCN,d}$, $N_{CCN,t}$, and N_d in Sect. 4, the cloud-event mean N_d values are used, i.e., we computed the N_d mean value for each individual cloud field.

With a few exceptions (e.g., wildfire smoke events), the main aerosol types were dust and anthropogenic aerosol particles. To obtain robust estimates for $N_{CCN,d}$ and $N_{CCN,c}$ for each analyzed cloud event, we determined vertically and temporally mean dust and non-dust backscatter coefficients for the height range from 500 m above ground to 300 m below cloud base
260 with 3-minute temporal resolution. To obtain representative estimates of $N_{CCN,d}$ and $N_{CCN,t}$ for each individual cloud event, used in the multiyear data analysis in Sect. 4, we analyzed all aerosol profiles within ± 24 h around the center time of the cloud occurrence period. The selection of this quite large time window was motivated by the goal to make use of the robust aerosol Raman lidar method, that can be easily automated to obtain aerosol backscatter height profiles without the need of critical input parameters which may produce unwanted significant uncertainty. However, Raman lidar observations are only available
265 at nighttime. With the 48-h time window, we are able to obtain aerosol information before and after each individual cloud event. From the vertically and temporally averaged backscatter coefficients, we calculated the respective $N_{CCN,d}$, $N_{CCN,c}$, and $N_{CCN,t}$ values by using Eqs. (1)–(5).

We restricted the considered backscatter data to a height range up to 300 m below the cloud base to avoid a potential impact of water uptake by the hygroscopic dust and haze particles on the determination of the (dry-particle) dust and non-dust aerosol
270 fractions. For distances >300 m below cloud base, RH was usually below 75%. The contaminated dust particles retain their characteristic irregular shape as long as $RH < 75\%$ and cause a depolarization ratio of around 0.31. At these quasi dry aerosol conditions, the separation of non-dust and dust backscatter fractions is well applicable when using the depolarization-ratio method (Tesche et al., 2009; Mamouri and Ansmann, 2014; Hofer et al., 2017) as our field observation and long term studies in highly polluted and dusty environments such as Cabo Verde during the winter season (Tesche et al., 2009) and at Dushanbe in
275 Tajikistan throughout the year (Hofer et al., 2020a, b) confirm. An accurate separation of the dust and non-dust fractions is even

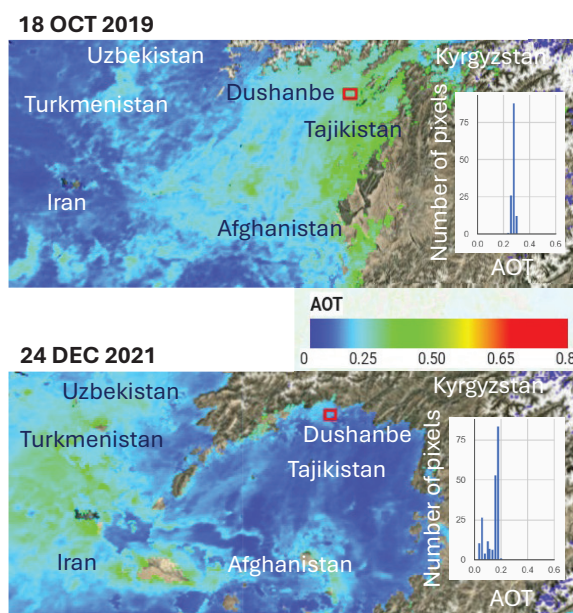


Figure 2. Aerosol conditions in Central Asia on 18 October 2019 and 24 December 2021 in terms of the 470 nm AOT retrieved from MODIS observations (6:30–9:30 UTC). The AOT histograms consider all MODIS-MAIAC results (for 1-km-by-1-km pixels within the red Dushanbe rectangle). Gray colors indicate cloud fields.

possible in cases with aged and internally mixed dust particles. This conclusion is supported by sun photometer observations at the lidar stations, i.e., by the good agreement between sun-photometer-derived fine-mode and coarse-mode AOTs and the respective lidar-derived non-dust and dust AOTs (Tesche et al., 2009; Hofer et al., 2020a). Complex mixtures of dust and anthropogenic pollution can be regarded as an external mixture of contaminated dust and anthropogenic aerosol particles in the case of the Dushanbe aerosol.

3 Case studies: evaluation of the retrieval framework

To evaluate the physical consistency of the retrieval framework, we examine three contrasting cases. The selected observations illustrate the complex nature of cloud condensation nuclei (CCN) activation in a region characterized by varying mixtures of natural and anthropogenic particles. We first analyze a dust-dominated aerosol-cloud interaction (ACI) scenario (Sect. 3.1), followed by an anthropogenic-dominated case (Sect. 3.2). Finally, Sect. 3.3 presents a special case of cloud formation within a long-range transported Canadian wildfire smoke plume.

Figure 2 illustrates the aerosol conditions around local noon on 17–18 October 2019 (Case 1: dust scenario) and 24 December 2021 (Case 2: near-pristine/anthropogenic scenario). These cases represent the characteristic extremes of the regional aerosol regime at Dushanbe: the spring-to-autumn dust season (April–October) and the winter pollution period (December–February)

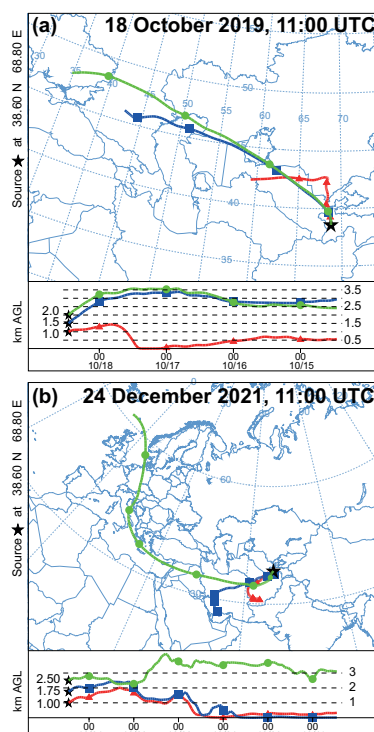


Figure 3. (a) Four-day HYSPLIT backward trajectories arriving at Dushanbe in 1000 m (red lines), 1500 m (blue lines), and 2000 m height above ground (green lines) on 18 October 2019, 11:00 UTC and (b) six-day backward trajectories arriving at Dushanbe in 1000 m (red lines), 1750 m (blue lines), and 2500 m height (green lines) on 24 December 2021, 11:00 UTC.

(Müller et al., 2026). Maps of the 470 nm aerosol optical thickness (AOT), derived from MODIS via the MAIAC algorithm (Fig. 2) show differences between the two regimes. Mineral dust produced AOT values of about 0.3–0.4 over Dushanbe on 17–18 October 2019. In contrast, wintertime anthropogenic haze on 24 December 2021 resulted in much lower AOTs, ranging from 0.05 to 0.15. The MAIAC AOT retrievals for the Dushanbe region (red box, Fig. 2) agree well with ground-based AERONET 500 nm AOT measurements of 0.30–0.35 on 17 October, 0.35–0.40 on 18 October 2019, and < 0.1 on 24 December 2021.

Cloud fields formed primarily at the top of the main aerosol layer, which extended up to 2.5–3.0 km in height.

Air mass history analyses further explain these contrasting optical properties. HYSPLIT (Hybrid Single Particle Lagrangian Integrated Trajectory) backward trajectories (Stein et al., 2015; Rolph et al., 2017) in Fig. 3a show that prior to arriving over Dushanbe on 18 October 2019, air masses traversed major Central Asian dust sources between the Caspian Sea and Tajikistan (including the Aralkum Desert), picking up mineral dust and regional pollution. Conversely, the trajectories for 24 December 2021 in Fig. 3b reflect weak stagnant winds, preventing dust and favoring accumulation of anthropogenic pollution particles during the slow travel of air masses across Iran, Afghanistan, and southern Tajikistan.

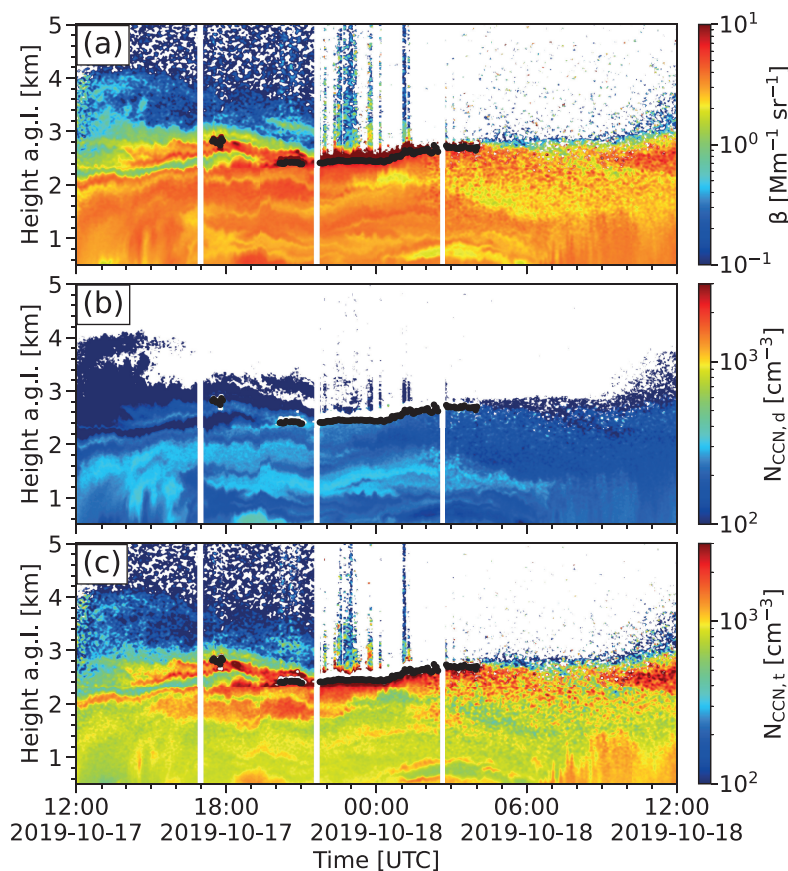


Figure 4. Cloud formation at the top of a 2.5–3 km thick dust-dominated aerosol layer observed over Dushanbe on 17–18 October 2019. Black symbols indicate the base height of shallow cloud fields. Panels a, b, and c show the aerosol layer in terms of the particle backscatter coefficient $\beta = \beta_d + \beta_{nd}$, the CCNC estimates for the dust CCN reservoir ($N_{CCN,d}$) and the total (dust + haze) CCN reservoir ($N_{CCN,t}$), respectively. White columns indicate automated 10-minute lidar alignment and calibration periods. Vertical resolution is 7.5 m, temporal resolution 180 s.

3.1 Dust event: 17-18 October 2019

Figure 4 shows the aerosol layering conditions on 17–18 October 2019 in terms of particle backscatter and CCNC estimates for the dust CCN reservoir and the total (dust + non-dust) CCN reservoir. Cloud evolution occurred in a dust-rich environment at the top of the main aerosol layer that reached to 2.5–3 km height above ground (Lipken, 2024). The dust layer was not well mixed during the night-time hours (14:00 UTC to 5:00 UTC, 19:00 to 10:00 local time) and showed many layering structures. An opaque cloud layer started to form at 2.5 km height around 20:00 UTC (1:00 local time) on 17 October 2019 and could then be observed for 8 hours. Water uptake by the hygroscopic dust and anthropogenic aerosol particles is visible in Fig. 4c at heights above about 1.8 km (orange to red colors below the cloud base). Because a trustworthy separation of dust and haze

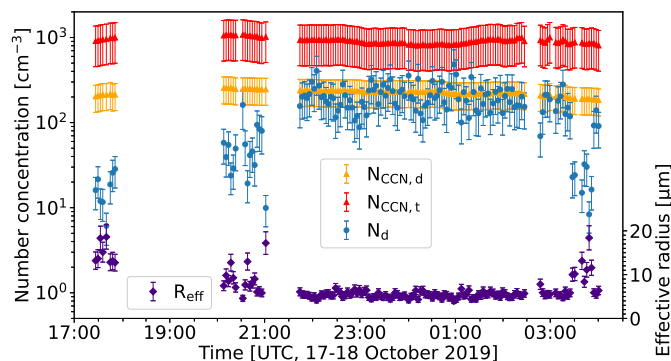


Figure 5. Cloud droplet number concentrations N_d (blue circles, indicating the actually activated CCN fraction) in comparison with the estimated concentrations of the potentially activatable dust CCN ($N_{CCN,d}$, orange, dust CCN reservoir) and total CCN ($N_{CCN,t}$, red, total CCN reservoir). The effective radius of the droplets R_{eff} (dark purple diamonds) is shown in addition. Uncertainties in the retrieval products according to Table 1 are indicated as error bars. White columns around 21:30 UTC and 2:45 UTC indicate periods of automated daily routine alignment and calibration procedures. A closed cloud deck was present from 21.30 UTC on 17 October to 4:00 UTC on 18 October 2019.

particles is no longer possible when $RH > 80\%$, the CCN reservoir values in Fig. 4b and c are only reliable up to about 1.6 km height before 22:00 UTC and up to 2.1 km height after 23:00 UTC. Taking water uptake into account, the dust-related number concentration $N_{CCN,d}$, available for droplet formation, was about $150\text{--}350\text{ cm}^{-3}$ (Fig. 4b), and $500\text{--}1000\text{ cm}^{-3}$ in case of $N_{CCN,t}$.

3.1.1 CDNC vs CCNC

Figure 5 compares the time series of cloud droplet number concentration (N_d) and droplet effective radius R_{eff} against the moving-window CCN reservoir estimates for the dust fraction ($N_{CCN,d}$) and the total particle ensemble ($N_{CCN,t}$). The effective radius is the droplet-cross-section-weighted mean radius and good proxy for the mean size of the droplets. The N_d and R_{eff} values are computed with 3 minute resolution, whereas the CCNC estimates $N_{CCN,d}$ and $N_{CCN,t}$, also given with 3 minute resolution, are 48 h mean values, i.e., calculated with a gliding temporal window lengths of 48 h and column averages (from 500 m above ground to 300 m below cloud base). The $N_{CCN,d}$ and $N_{CCN,t}$ values in Fig. 5 show the overall, and usually slowly varying aerosol conditions in the main aerosol layer rather than the actual aerosol situation below a detected cloud base.

After the establishment of a closed cloud deck at 21:30 UTC, the CDNC time series coincided well with the dust reservoir CCNC time series. As a reminder, all dust particles with radius $> 100\text{ nm}$ are assumed to be able to serve as CCN and to contribute to the dust CCN reservoir. During the stable cloud phase from 21:30 UTC to 3:30 UTC, the droplet size distribution, expressed by the droplet effective radius R_{eff} of around $5\text{ }\mu\text{m}$, showed a rather constant behavior. This may indicate a balance between droplet collision and coalescence, new droplet formation, and droplet evaporation. Droplet effective radii $< 10\text{ }\mu\text{m}$ indicate small droplets, i.e., typical sizes for non-precipitating clouds (Rosenfeld et al., 2014; Jimenez et al., 2025).



The variation of the 3-minute-mean N_d values in Fig. 5 reflects the impact of vertical velocity. During downdraft periods, new droplets are not formed and existing droplets may evaporate so that CDNC shows values around the minimum, i.e., values below 150 cm^{-3} . In contrast, during well organized updrafts, strong droplet nucleation in the lofted air parcels takes place, indicated by N_d values clearly exceeding $200\text{--}250 \text{ cm}^{-3}$. The water supersaturation level may be even considerably higher than 0.2% in well organized updrafts so that more droplets are formed than indicated by the CCNC reservoir values (assuming 0.2% water supersaturation). The large N_d values above 250 cm^{-3} are indicative for the actually activated number of CCN. All in all, the good agreement between the dust CCNC and CDNC in Fig. 5 suggests a strong, if not controlling, contribution of dust particles to the droplet nucleation processes during the long-lasting cloud event observation from 21:30 to 3:30 UTC.

3.1.2 Dust CCN: two critical questions

Evaluating the consistency of our N_{CCN} and N_d retrieval framework relies on two critical questions:

1 What is the minimum size of dust particles involved in the droplet nucleation processes? This question is discussed in this section 3.1.2.

2 Did the Dushanbe dust particles undergo chemical aging and show water-uptake effects at high RH? These aspects are illuminated in Sect. 3.1.3 and 3.1.4, based on Fig. 6 (for contaminated dust) and Fig.7 (for pristine, hydrophobic dust).

Kelly et al. (2007) and Koehler et al. (2009) investigated the ability of dust particles to nucleate water droplets. Koehler et al. (2009) pointed out that their laboratory results were consistent with the calculations and simulations of Kelly et al. (2007). In the case of insoluble, hydrophobic dust particles, only large particles with dry-particle radius $>500 \text{ nm}$ are able to initiate droplet nucleation at a water supersaturation of 0.2%. Koehler et al. (2009) analyzed the cloud-relevant behavior of dry and wet-generated Arizona Test Dust (ATD-wet, ATD-dry), a milled product originating from Arizona sands, of wet-generated Saharan dust after long-range transport, collected on the Canary Island of Lanzarote (CID-wet), and of dry-generated Saharan dust (collected in Egypt) in their laboratory studies. Concerning the generation process we refer to the explanation in the article of Koehler et al. (2009). The wet-generated particles contained some hygroscopic material on the surface. The dry-generated particles were assumed to be almost free of any hygroscopic material. Most natural airborne mineral dust particles generated from sources with low soluble contents likely will exhibit CCN activity more consistent with the less-hygroscopic dust particles. The observed CCN activity of the less hygroscopic dust types shows that particles with dry-particle radius >100 to 130 nm can readily serve as CCN at low water supersaturations of 0.2%. For ATD-wet and CID-wet, representing the more-hygroscopic dust types, the laboratory investigations showed minimum radii of the order of $50\text{--}60 \text{ nm}$.

Haarig et al. (2019) compared lidar-derived dust CCNC with airborne in situ observations of CCNC in the Saharan air layer (SAL) over Barbados during SALTRACE (Saharan Aerosol Long-Range Transport and Aerosol-Cloud-Interaction Experiment) (Weinzierl et al., 2017) in July 2013, about 4000 km west of Africa, and found good agreement with the in situ observations when using $n_{100,d}$ as proxy for the dust-related CCNC. The dust fraction in the SAL was $>90\text{--}95\%$ (Haarig et al., 2017; Groß et al., 2015) and the $\text{RH} < 50\%$ during the airborne and lidar observations. We assume that the dust particles in the SAL were chemically processed during the long-range transport over more than 5000 km from the Saharan dust

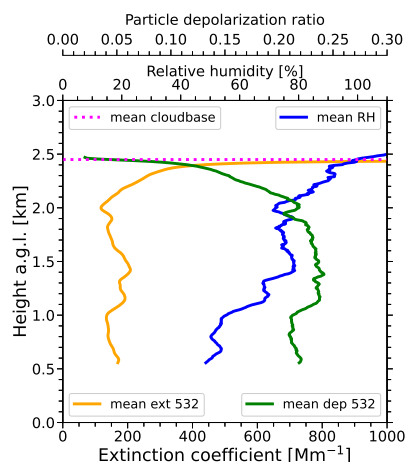


Figure 6. 532 nm light-extinction coefficient (orange profile, Eq. 1), particle depolarization ratio (green profile), and RH (blue profile). One-hour mean lidar profiles (23:00–24:00 UTC, 17 October 2019) are shown. Water uptake by aerosol particles is visible at heights above 2.1 km, when RH exceeded 75–80%. The strong decrease of the particle depolarization ratio above 2.1 km heights indicates a strongly increasing fraction of spherical dust particles. Relative uncertainties are of the order 20% (extinction profile), <10% (particle depolarization ratio), and 10–20% (relative humidity).

sources and after crossing populated regions of western central Africa towards the Caribbean and may have partly been involved in cloud processes on the way across the Atlantic. The use of recently updated extinction-to- $n_{100,d}$ conversion factors (Ansmann et al., 2026) further improved the agreement between the lidar-derived and measured dust CCNC values (personal communication, Moritz Haarig).

365 If we assume, that even dust particles with radius from 60–100 nm are hygroscopic and activatable, $N_{CCN,d}$ would be a factor of 5–6 larger than the $N_{CCN,d}$ considering dust particles with radius >100 nm as assumed in Fig. 5. If only dust particles with radius > 250 nm would be able to serve as CCN, then the impact of dust on droplet formation would be almost unimportant and the anthropogenic-aerosol particles would dominate droplet nucleation. $n_{250,d}$ is about 12–13 times lower than $n_{100,d}$ (Ansmann et al., 2026).

370 3.1.3 Hygroscopic dust

Regarding the second question — are the dust particles at Dushanbe contaminated as a result of chemical aging and cloud processing? — we analyzed simultaneously measured particle depolarization-ratio and RH profiles. The result for the 17–18 October case is shown in Fig. 6. A one-hour mean height profile of the particle extinction coefficient together with the respective RH profile (from lidar in combination with ECMWF temperatures, see Sect. 2.3.1), and the profile of the particle
375 depolarization ratio are presented. RH increased almost monotonically and reached 100% at cloud base. The particle extinction coefficient significantly increased when RH exceeded the 80% level from values of 140–150 Mm^{-1} at 80% to values of 220–250 Mm^{-1} at 90%. The strong increase of the particle extinction coefficient at heights above 2 km may be used as a first



indication that the aerosol mixture consisted of contaminated hygroscopic dust particles and anthropogenic aerosol particles. A more important indication for the presence of hygroscopic dust particles is the strong decrease of the particle depolarization ratio at heights >2 km with increasing RH in the 80–100% RH range. The particle depolarization ratio was close to 0.24 at 1200–1500 m height. This is indicative for a dust contribution to particle backscattering of about 80% (Tesche et al., 2011). The strong decrease from 0.22–0.23 at RH of 75% (at 1900 m height) to 0.15 at RH of 90% to 92% at 2300 m height can only be explained by a change of the dust-related depolarization ratio from 0.31 (for dry dust particles) to about 0.2 for wet dust particles with less irregular shape characteristics. In the analysis of changing optical properties with changing RH levels, we assumed hygroscopic dust as well as hygroscopic anthropogenic aerosol particles. The dust particles became obviously rounded and partly lost their irregular shape properties as a result of water uptake which then leads to a significant reduction of light depolarization during the backscattering process. This feature is the most unambiguous hint that the dust particles were hygroscopic. And this indication is an important prerequisite for our hypothesis that even small dust particles with radius of 100 nm can serve as CCN so that $N_{CCN,d}$, as used here, is an appropriate proxy for the CCN reservoir of contaminated dust particles.

3.1.4 Hydrophobic dust

Figure 7 shows a lidar observation of hydrophobic dust particles. The case is presented to provide a strong contrast to Fig. 6, i.e., an additional argument that Fig. 6 shows optical properties of contaminated, hygroscopic dust particles. The measurements in Fig. 7 were performed at Ouarzazate in the Atlas mountains in southeastern Morocco, close to Saharan dust sources, during SAMUM (Saharan Mineral Dust Experiment) 2006 (Heintzenberg, 2009; Freudenthaler et al., 2009; Ansmann et al., 2011) in May 2006. Two lidars were involved. The Munich SAMUM lidar (Freudenthaler et al., 2009) measured height profiles of the particle linear depolarization ratio (PDR) at 532 nm and the Leipzig lidar of TROPOS measured backscatter and extinction profiles and multiple wavelengths including 532 nm. The Leipzig lidar observations on 13 May 2006 were already discussed in Ansmann et al. (2008). As can be seen, the extinction coefficient of the fresh dust particles showed no dependence on relative humidity at all, even for $RH > 90\%$. RH monotonically increased from 10% at ground in 1150 m height above sea level to almost 100% at the top of the 4–5 km deep dust layer over the lidar station far away from big cities and industrialized regions. The conversion of the dust extinction coefficient into CCN values yield $20\text{--}25\text{ cm}^{-3}$, when considering dust particles with radius >250 nm in the CCNC estimation. The 532 nm particle linear depolarization ratio (PDR in % Fig. 7) was very high with values around 35% (or 0.35) up to heights of 4.5 km. This is a clear indication of a strong contribution to light depolarization by freshly emitted large dust particles. PDR slightly decreased to values around 31% at heights in the uppermost and humid part of the dust layer. This decrease may indicate water vapor adsorption by the pristine dust particles so that their shape characteristics slightly changed. However, it is also possible that the amount of very large dust particles decreased with height so that PDR showed more typical dust depolarization values of around 31% higher up. A third option is that a small fraction of hygroscopic (non dust) particles were present and caused a decrease of PDR at the dust layer top by an increasing impact on non-dust particle backscattering with increasing water uptake.

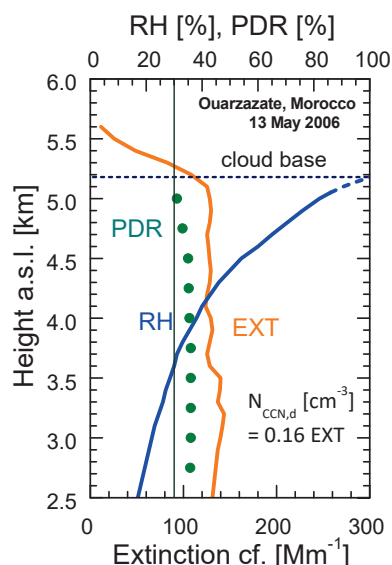


Figure 7. Light-extinction coefficient (EXT, orange, Leipzig lidar) and particle depolarization ratio (PDR, Munich lidar, green circles) of hydrophobic dust vs relative humidity (RH, blue, SAMUM radiosonde) in a well-mixed dust layer over southeastern Morocco close to Saharan dust sources. A cloud layer developed shortly after the shown lidar observation at Ouarzazate (30.9°N, 6.9°W, 1150 m height above sea level, a.s.l.) in the Moroccan Atlas mountains on 13 May 2006, 20:08-22:08 UTC (Ansmann et al., 2008). Extrapolation of the relative humidity to the cloud base is indicated by the dashed blue line. The vertical line indicates a PDR=0.3. Extinction values of 120-140 Mm^{-1} indicate dust CCNC levels of 20-22 cm^{-3} (assuming that only dust particles with radius >250 nm are potentially activatable) (Ansmann et al., 2026).

Kaaden et al. (2009) conducted ground-based in situ aerosol observations at Tinfou, Morocco, about 100 km southeast of Ouarzazate in the Saharan desert (close to the border to Algeria) during the SAMUM campaign and found that all particles in the diameter classes centered at 720, 840, and 960 nm were generally hydrophobic. During dust storms the fraction of hydrophobic particles significantly increased in the diameter classes centered at 150, 250 and 350 nm as well. Only the fine-
415 mode classes (centered at diameters of 30, 50 and 80 nm) were dominated (with number fractions $>90\%$) by hygroscopic particles, most probably by anthropogenic particles from nearby villages and touristic activities.

All in all, the height-independent particle extinction coefficient and almost height-independent PDR in Fig. 7 and the strong changes of the optical properties with height in Fig. 6 provide a convincing basis for the argumentation that the Dushanbe dust is significantly different from pristine, freshly emitted dust, and obviously to a large extent contaminated with hygroscopic
420 material as a result of chemical aging and cloud processing during long-range transport.

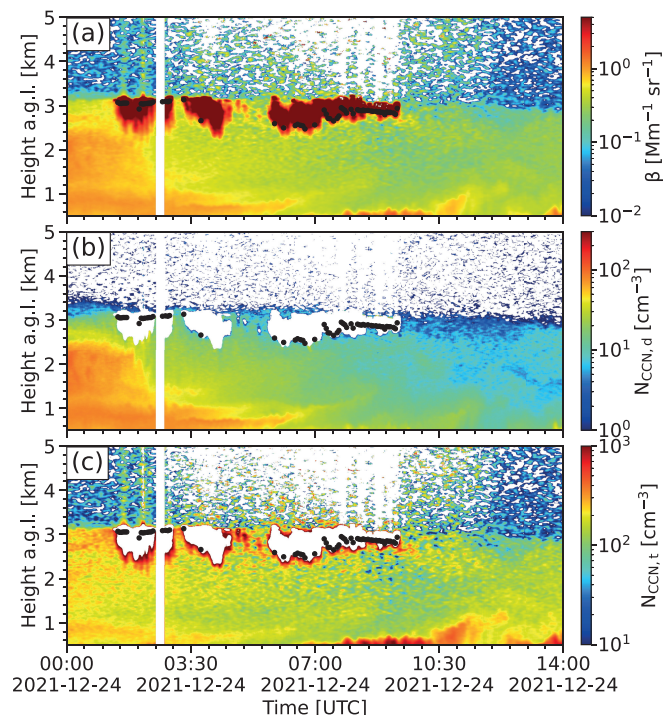


Figure 8. Same as Fig. 4, except for 24 December 2021. CCNC information is not given for the white areas (cloud layers, ice virga). Black symbols indicate the base height of the liquid-dominated cloud layer.

3.2 Near pristine/anthropogenic aerosol scenario: 24 December 2021

In contrast to the October dust case, the observation on 24 December 2021 represents a comparably clean aerosol scenario with a minimum impact of dust on water droplet nucleation (see Fig. 2 and 3). Figure 8 shows the lidar observations on this day. A dusty air mass (visible at heights below 2.5 km until 07:00 UTC) left the Dushanbe area so that cloud formation occurred in almost dust-free air at heights around 3 km. Cloud top temperatures were -18 to -20°C so that immersion freezing was initiated and ice virga developed.

A stable cloud deck was observed from 07:30 to 09:15 UTC (12:30 to 14:15 local time). To that time, the dust CCN reservoir showed low values of less than 20 cm^{-3} in the upper part of the aerosol layer, in which the cloud layers developed (Fig. 8b), whereas the haze-related CCN reservoir contained 100 – 300 particles per cm^3 according to Fig. 8c. Weak ice formation was detected to that time. It should be mentioned here, that ice nucleation does not influence the retrieval of CDNC when using the dual-FOV lidar technique (Jimenez et al., 2025).

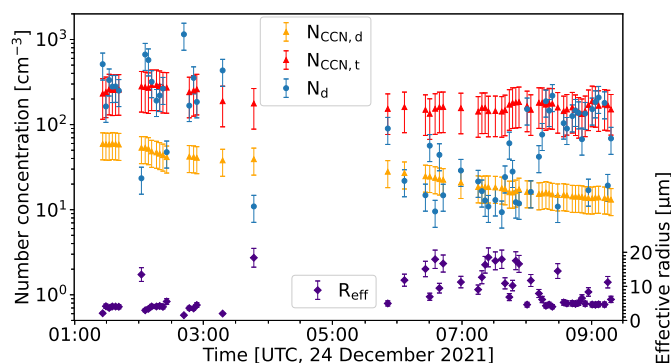


Figure 9. Same as Fig. 5, except for 24 December 2021.

3.2.1 CDNC vs CCNC

Figure 9 shows time series of the particle number concentrations for the dust and total CCN reservoirs, of CDNC, and of the droplet effective radius for the time period from 01:00–10:00 UTC on 24 December 2021. During times with well developed, stable cloud structures (indicated by R_{eff} around 5 μm), the N_d values were mainly between 100 and 400 cm^{-3} and close to the total CCN reservoir values (for the assumed supersaturation of 0.2%). By comparing the $N_{\text{CCN},t}$ and $N_{\text{CCN},d}$ time series, it can be seen that the total CCN reservoir $N_{\text{CCN},t}$ was dominated by the anthropogenic aerosol particle fraction in the case of well-developed cloud fields (01:00–03:00 UTC, 08:30–09:30 UTC). In the absence of a significant dust fraction, especially after 8:00 UTC, the small anthropogenic haze particles widely controlled the formation of droplets. The dust load was obviously too low to suppress the activation of pollution particles.

3.2.2 CDNC vs CCNC profiles

Figure 10 shows vertical CCNC profiles before and after the cloudy period from 01:30–09:30 UTC. The given 1-h mean $N_{\text{CCN},d}$ and $N_{\text{CCN},t}$ profiles for the 13:17–14:17 UTC time interval (dotted curves) represent well the aerosol conditions during the formation of the cloud layer observed from 08:12–09:19 UTC, for which the mean N_d and one-standard deviation are given. Fig. 10 corroborates that anthropogenic aerosol particles controlled droplet nucleation in the afternoon (after 8 UTC) on 24 December 2021. The consistency in the observations with similar values for N_d and $N_{\text{CCN},t}$ suggest that the water supersaturation was about 0.2%. For such a supersaturation level the number concentration $n_{50,c}$ (Eq. 4) is an appropriate proxy for $N_{\text{CCN},c}$ as often shown in the literature (Schmale et al., 2018; Düsing et al., 2018).

3.3 Wildfire smoke event: 1 June 2023

To test our retrieval framework under extreme aerosol conditions, we analyze a long-range Canadian wildfire smoke event on 1 June 2023. A deep aerosol layer was present over Dushanbe with cloud layers at 4–4.5 km height as shown in Fig. 11. According to the HYSPLIT backward trajectories (not shown), wildfire smoke from North America crossed the Atlantic and

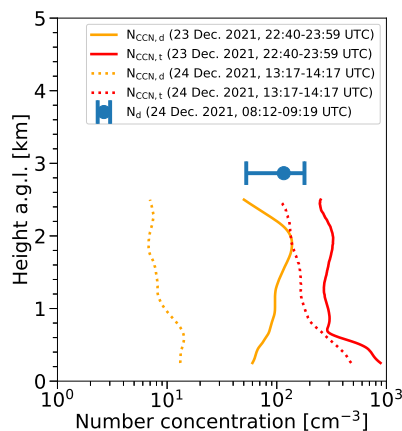


Figure 10. Dust CCNC (orange, dust CCN reservoir) and total CCNC (red, total CCN reservoir) profiles, observed before (solid lines) and after (dotted lines) the cloudy period, lasting from 01:30–09:30 UTC on 24 December 2021 (Fig. 9). Only CCNC profile segments for heights with $RH < 75\%$ (i.e., for dry particles) are shown. The mean CDNC (blue circle, assigned to the height of 75 m above mean cloud base) and corresponding one-standard deviation (horizontal bar) are computed from 22 3-minute mean CDNC values shown in Fig. 9 for the cloudy period from 08:12 UTC to 09:19 UTC. The CDNC mean value (blue) is close to the total CCN reservoir value for the clean period (dotted orange and red profiles). The uncertainties in the retrieval products are 35% (dust CCNC), 50% (total CCNC), and 35% (CDNC).

Europe and reached Central Asia at heights above 3 km after about two weeks of traveling. Aged wildfire smoke particles consist to more than 95% of organic carbon and contain 2–3% of black carbon. The particles are liquid in the lower troposphere and glassy in the upper troposphere and lower stratosphere (Knopf et al., 2018; Ansmann et al., 2021; Ohneiser et al., 2023).
 455 and glassy in the upper troposphere and lower stratosphere (Knopf et al., 2018; Ansmann et al., 2021; Ohneiser et al., 2023). The clouds at 4 and 4.5 km height above ground developed at temperatures of 265–268 K. Figure 12 provides insight into the contribution of the dust and smoke CCN to the total CCN reservoir. Here we applied the extinction-to-CCN conversion factors for anthropogenic aerosol particles (red profile segment), aged wildfire smoke (green profile segment), and mineral dust (orange profile) to the respective mean lidar backscatter coefficient profiles for the time period from 22:40–24:00 UTC (Ansmann et al.,
 460 2026), and assumed a 532 nm smoke lidar ratio of 70 sr in the computation of the smoke extinction coefficients (Haarig et al., 2018).

The good agreement between the cloud mean N_d value for the cloud layer occurring from 20:30 to 21:25 UTC (see Fig. 11) and the smoke CCNC profile values ($N_{CCN,s}$) in Fig. 12 suggest that our CDNC and CCNC retrieval framework worked very well and that the cloud droplets mostly formed on smoke particles. It should be emphasized here that aged smoke particles form
 465 an accumulation mode with a mode radius of 200–300 nm (Haarig et al., 2018; Ansmann et al., 2021). Aged smoke particles are hygroscopic and compared to anthropogenic fine-mode aerosol particles large and thus favorable CCN.

In summary, we can conclude from the three case studies that our CCNC and CDNC retrieval framework provides trustworthy results. Thus, the multiyear Dushanbe N_d , $N_{CCN,d}$ and $N_{CCN,t}$ data sets reflect well the complex aerosol and cloud properties and liquid-phase-related ACI conditions in Central Asia. Based on these data, the role of Central Asian dust on cloud
 470 droplet nucleation processes was investigated (Lipken, 2024). The main results are discussed in the next section.

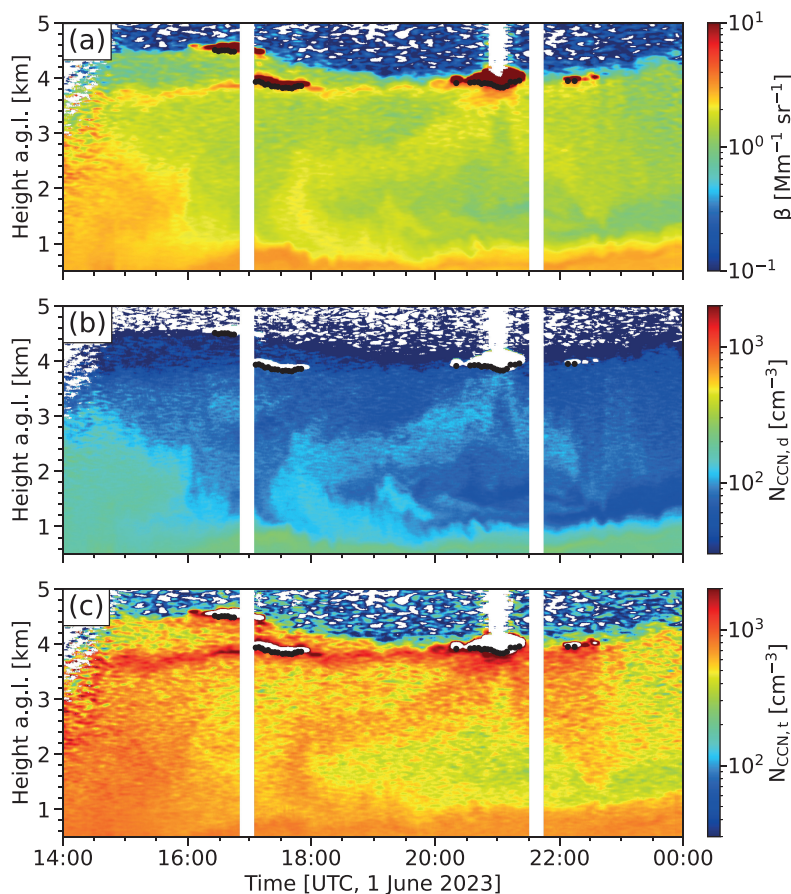


Figure 11. Same as Fig. 4, except for 1 June 2023. Black symbols indicate the base height of the liquid-dominated cloud layers. In c, we simply assumed that the non-dust CCN fraction in the entire aerosol layer consisted of anthropogenic aerosol particles.

4 Multiyear observations of aerosol-cloud interaction

This section presents the main findings from the multi-year observations. As outlined in Sect. 2.4, 1602 cloud events, ranging from short-term cloud layers lasting 10 minutes to persistent cloud decks lasting more than 10 hours, were analyzed. The observations spanned the summer dust seasons of 2019, 2022, and 2023, and the winter periods of 2019–2020, 2021–2022, and 2022–2023. The automated computation of the Dushanbe CDNC and CCNC data records is described in Sect. 2.3.2 to 2.4.

4.1 Seasonal variability of CCNC and CDNC

Figure 13 shows the time series of CDNC, i.e., of 1602 cloud-event mean N_d values (retrieved for the height of 75 m above cloud base) together with two time series of CCNC estimates characterizing the dust CCN reservoir $N_{CCN,d}$ and the total (dust

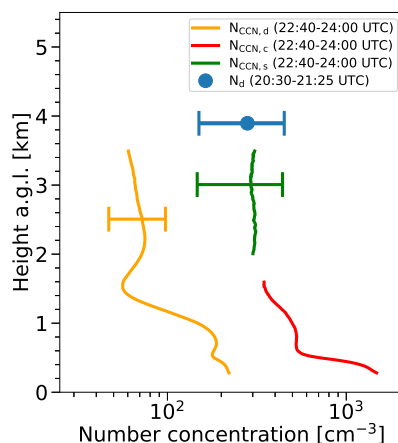


Figure 12. Profiles of dust CCNC (orange, $N_{CCN,d}$, dust CCN reservoir), wildfire smoke CCNC (green, $N_{CCN,s}$, smoke CCN reservoir), and continental haze CCNC (red, $N_{CCN,c}$, anthropogenic aerosol CCN reservoir), observed on 1 June 2023, 22:40–24:00 UTC, i.e., after the cloud period from 20:30–21:25 UTC for which the mean CDNC (blue, N_d) and the range of 3-minute mean CDNC values (horizontal bar) are shown. The N_d value is assigned to 75 m above the mean base height of the analyzed cloud layer. One-standard-deviation CCNC uncertainties for dust and smoke are shown as error bars.

and continental haze) CCN reservoir $N_{CCN,t}$. The N_d values can be used as a reasonable proxy for the fraction of the total
480 CCN reservoir $N_{CCN,t}$ actually activated.

Figure 13 shows a strong correlation between the dust CCN reservoir ($N_{CCN,d}$) and CDNC (N_d). The question is: Does the correlation indicate a sensitive, or even controlling influence of dust particles on droplet nucleation and this generally during all seasons of the year? Before we can answer this question, we need, in a first step, to explain the reasons for the observed seasonal cycles of N_d , $N_{CCN,d}$, and $N_{CCN,t}$.

485 4.1.1 Seasonal cycle of CDNC

The seasonal cycle of N_d mainly reflects the seasonally varying updraft conditions and the corresponding seasonal cycle of water supersaturation levels in the cloud base region. Weaker updrafts in winter lead to lower supersaturation values, on average, than in summer. This produces the corresponding seasonal pattern in Fig. 13, with lower N_d in winter and higher N_d in summer. In summer, the supersaturation was most probably around 0.2% as discussed in Sect. 2.3.2. In winter, the
490 supersaturation values are assumed to be frequently in the range from 0–0.1%. The assumed seasonal cycle of the water supersaturation is in line with global, season-resolved simulations of supersaturation conditions (Karydis et al., 2017). As pointed out in Sect. 2.3.2, the number of nucleated droplets roughly decreases by a factor of 2 and 4 when the supersaturation decreases from 0.2% to 0.1% and 0.05%, respectively (Sapkal et al., 2025). We may conclude, that the CDNC was widely controlled by updraft and respective water supersaturation conditions and only modulated by different winter-to-summer dust
495 conditions.

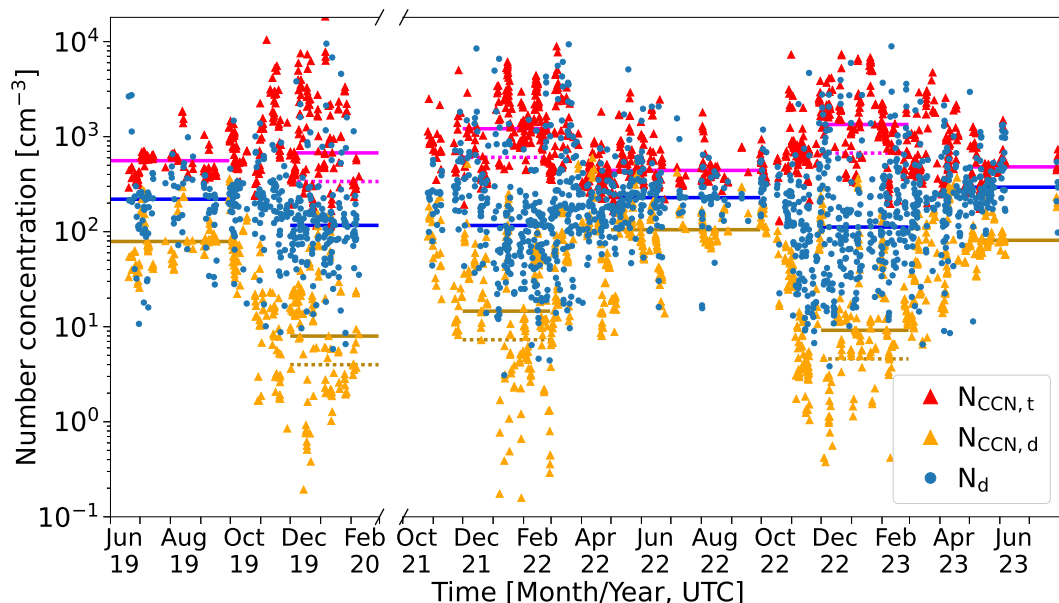


Figure 13. Long-term observations of cloud-event mean droplet number concentration at 75 m above cloud base (N_d , blue circles, indicating the actually activated CCN fraction) together with number concentration estimates for the dust CCN reservoir ($N_{CCN,d}$, orange triangles) and the total (dust + non-dust) CCN reservoir ($N_{CCN,t}$, red triangles), assuming a water supersaturation of 0.2% (see text for more explanations). Three summer and three winter seasons are covered by the continuous lidar observations at Dushanbe from June 2019 to July 2023. The horizontal lines show median values for the 5-month period from May to September and the winter season (December to February), for 0.2% (solid lines) and 0.1% (dotted lines, winter) supersaturation. A long break from February 2020 to October 2021 was caused by a broken laser that could not be repaired during the Covid-19 pandemic.

4.1.2 Seasonal cycle of the dust CCN reservoir

Several reasons are responsible for the pronounced seasonal cycle in the dust CCNC values. First of all, dust outbreak events are much less frequent during the winter season than during the summer half year (Hofer et al., 2020a; Kok et al., 2021; Nobakht et al., 2021; Fomba et al., 2024; Müller et al., 2026) as a result of low wind speeds close to the surface in the Middle East and Central Asia in winter. In addition, parts of Central Asia are covered by a snow deck during the winter months. As can be seen, the dust CCNC values decreased from values, mostly between 50 cm^{-3} and 500 cm^{-3} in the summer period of 2022 (April to October 2022) to values from $5\text{--}30 \text{ cm}^{-3}$ during the winter season of 2022–2023 (December 2022 to February 2023). However, occasionally, dust events with $N_{CCN,d}$ clearly above 50 cm^{-3} were observed in winter.

The pronounced seasonal cycle of $N_{CCN,d}$ is partly influenced by the summer-to-winter variations of the water supersaturation which is, as mentioned above, expected to be around 0.2% in summer and much lower with values of 0.05% to 0.1% during the winter months. We show all individual $N_{CCN,d}$ estimates and corresponding median values (solid lines) for a constant supersaturation of 0.2% (for all seasons) in Fig. 13. To indicate the reduction of the $N_{CCN,d}$ and $N_{CCN,t}$ values for a



supersaturation of 0.1% we plotted the resulting median values as dotted horizontal lines. This decrease of the water supersaturation from 0.2% in summer to around 0.1% or lower in winter makes the seasonal contrast in the $N_{CCN,d}$ time series even stronger and must be added to the strong drop in dust emissions in October-November of each year.

4.1.3 Seasonal cycle of the total CCN reservoir

During the winter months anthropogenic aerosol particles ($N_{CCN,e}$) clearly dominate the total CCN reservoir. The seasonal variations in the $N_{CCN,t}$ time series in Fig. 13 is to a large extent a function of the variation of the vertical depth of the haze layer. i.e., of the main aerosol layer. The vertical extent is much lower in winter than in summer (Hofer et al., 2020a) and correspondingly the $N_{CCN,t}$ winter values are much higher than the $N_{CCN,t}$ summer values. If we assume, in addition, that emissions, originating from traffic and industrial activities, are widely constant throughout the year and, in winter, a contribution by house heating (including residential wood consumption) (Fomba et al., 2024; Müller et al., 2026) must be added, the haze-related CCNC values increase by a factor of roughly 2–3 when the winter haze layer height decreases by a factor of 2 from summer to winter.

After discussing of the reasons for the seasonal cycles of the different time series in Fig. 13, we can return to our initial question concerning the impact of dust on the evolution of the liquid phase in water-containing clouds and what the remarkable correlation between $N_{CCN,d}$ and N_d tells us about the impact of dust on droplet nucleation.

4.2 Dust contribution to CCN activation

In Fig. 14, the relative contribution of dust particles to droplet formation over Dushanbe is shown. The same $N_{CCN,d}$ and N_d data as in Fig. 13 are used. Most $N_{CCN,d}/N_d$ values were found between 0.1 and 1. They indicate that about 10–100% of the activated CCN were dust particles. These numbers are in good agreement with respective simulations showing dust contributions to droplet nucleation of 5% to 40% (Karydis et al., 2017). Figure 14 also shows many rather low $N_{CCN,d}/N_d$ values, often clearly below 0.1, especially during the winter months. Then, the contribution of dust particles to droplet formation is not relevant.

A considerable number of cloud events show $N_{CCN,d}/N_d$ values from 1–10 and even up to 30. Ratios clearly exceeding 1 are typically linked to dust outbreaks or dust storms when the dust number concentration strongly increases but the meteorological conditions and thus the supersaturation levels remain widely unchanged (updraft-limited regime, see Sect. 2.3.4). However, we can not exclude that in several cases the CCN reservoir (at the relevant heights close to cloud base) was overestimated because $N_{CCN,d}$ is calculated as mean value of all backscatter coefficients observed between 500 m above ground to 300 m below cloud base. In case of intense dust storms, extremely large dust concentrations are often observed in the lowermost 1.5 km of the atmosphere. If cloud fields develop at heights of, e.g., >3 km, dust column CCNC estimates, dominated by the dust particles far below the cloud layers, are much too high and do not reflect the true CCNC conditions just below the base of the cloud fields. Furthermore, the CCNC computation includes lidar data, collected within 48 hours, while the cloud events may last for two hours or less. This is also a source for sometimes strong discrepancy between N_d and $N_{CCN,d}$, in both directions towards dust-dominated and dust-free scenarios.

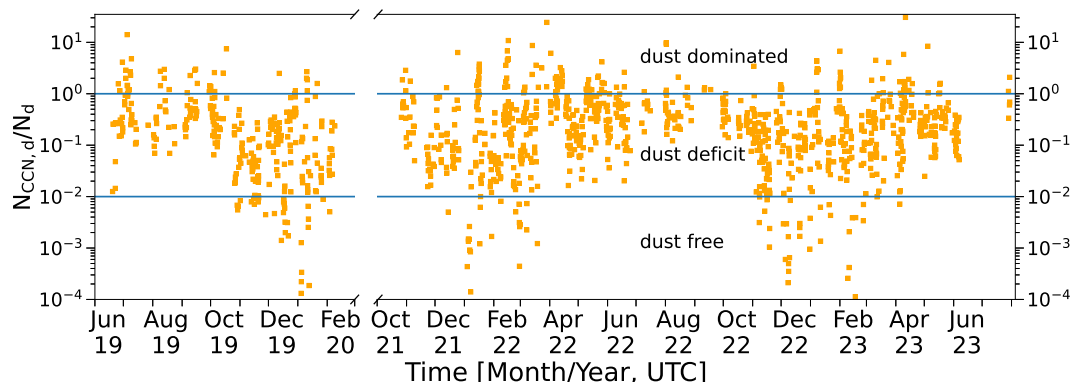


Figure 14. Ratio of the number concentration of activatable dust particles ($N_{CCN,d}$, dust CCN reservoir) to the observed droplet number concentration (N_d , indicative for the actually occurring CCN concentration). Values shown in Fig. 13 are used. Dust dominated and dust-free conditions are defined by $N_{CCN,d}/N_d \geq 1$ and ≤ 0.01 , respectively, indicated by horizontal lines.

Complementary to Fig. 14, Table 2 provides additional information about the contribution of the dust CCN reservoir to droplet formation. As can be seen, the majority of $N_{CCN,d}/N_d$ values accumulated between 0.1 and 1.0 (73% in summer, 47% in winter) with values >0.5 in 60% and 20% out of all cases in summer and winter, respectively. Regarding the total CCNC estimates (not shown), the ratio $N_{CCN,t}/N_d$ is typically in the range of 1 to 10 in summer and between 1 to 100 in winter.

Table 2. Frequency of occurrence of $N_{CCN,d}/N_d$ values shown in Fig. 14 during the summer period from April to October 2022 (dust season in 2022) and during the winter season from December 2022 to February 2023.

Values	Apr – Oct	Dec – Feb
>1	20%	10%
0.5–1	39%	10%
0.1–0.5	34%	37%
<0.1	7%	43%

545 Regarding the size of activatable dust particles, able to serve as CCN, the following remarks may be useful. As already discussed in Sect. 3.1, if we assume, that even dust particles with radius from 60–100 nm are hygroscopic and activatable, the $N_{CCN,d}/N_d$ ratio would shift from, e.g., 0.5–5 in Fig. 14 to 3–30, because the reservoir considering all dust particles with radius > 60 nm is a factor of 5–6 larger than the reservoir considering dust particles with radius >100 nm (Ansmann et al., 2026). These numbers corroborate the strong potential of hygroscopic dust CCN to influence cloud evolution and liquid droplet formation processes over Dushanbe during the dust season. If only dust particles with radius > 250 nm would be able to serve as CCN, as indicated by a large fraction of hydrophobic particles measured in situ at a desert site in Morocco (Kaaden et al.,

550



2009), the $N_{CCN,d}/N_d$ ratio would shift from 0.5–5 to about 0.05–0.5. Then, the impact of dust on droplet formation would be almost insignificant, and significant during very dusty days only.

4.3 Aerosol-cloud interaction susceptibility

555 With respect to the definitions of the updraft-limited, aerosol-limited, and transitional regimes in Sect. 2.3.4, all winter observations belong to the updraft-limited regime, and the summer observations widely belong to the transitional regime. We thus expected low values for the susceptibility $s_{ACI,liq}$ (Eq. 7). Lipken (2024) studied aerosol-cloud interaction based on monthly mean N_d and $N_{CCN,t}$ values. Months with only a few liquid water cloud events were excluded. Based on the remaining 24 monthly mean CCNC and CDNC pairs the resulting susceptibility $s_{ACI,liq}$ was 0.325 with an uncertainty of 30–40%. When
560 plotting $\log N_{CCN,t}$ versus $\log N_d$ on the basis of all individual CCNC and CDNC pairs, a clear relationship between the CCNC estimates and CDNC was not visible in the data clouds. ACI aspects will be further discussed in Sect. 5.

4.4 Cloud statistics

Cloud statistical results are presented in Fig. 15. As mentioned, the majority of the CDNC values was observed in altocumulus and stratocumulus cloud decks that formed in the upper part or at the top of the main dust and haze aerosol layer. The cloud
565 fields showed temporal lengths (lidar crossing time periods) of 10 minutes up to 3–4 hours in most cases. In few exceptions, cloud systems could be observed for more than 4 hours and up to 10 hours.

As can be seen in Fig. 15a and b, CDNC values accumulated between 100 and 350 cm^{-3} during the summer half year (spring to autumn) and 50 and 200 cm^{-3} during the winter season. The size of the droplets is characterized by the effective radius, i.e., the geometrical-cross-section-weighted mean droplet radius. The typically observed R_{eff} values ranged from 4–6 μm in
570 summer and 5–11 μm in winter. These R_{eff} numbers are indicative for non-precipitating cloud layers (Rosenfeld et al., 2014).

Cloud base heights in Fig. 15c reflect the vertical extend of the main aerosol layers which often coincide with the daytime boundary layer or nighttime residual layer. Most clouds formed in the top region of the near-surface aerosol layer. The lower cloud base temperatures in winter (on average around 265 K in Fig. 15e) compared to summer temperatures (around 271 K), and the correspondingly lower saturation water-vapor-to-dry-air mixing ratios in winter (2–4 g kg^{-1}) in Fig. 15f, compared to
575 the summer season (4–6 g kg^{-1}), may have contributed to the lower CDNC values in winter because of the lower amount of available water vapor for droplet formation in the presence of cloud droplets acting as a sink for water vapor as well.

As mentioned, the observed relatively high summer N_d values in Fig. 15a (median value close to 220 cm^{-3}) and low N_d values in winter (median value of around 100 cm^{-3}) reflect the changing water supersaturation with values most probably around 0.2% in summer and around 0.1% (or even less) in winter. The large scatter in the winter N_d values in Fig. 13 are
580 indicative for the large range of occurring supersaturation levels from almost 0% to 0.2% in winter.

Figure 15d contains the statistics to cloud extinction coefficients at 532 nm in the cloud base region. Extinction coefficients accumulate between 15 and 30 km^{-1} . Assuming vertical homogeneous extinction conditions and cloud vertical depths of mostly between 300 and 500 m, cloud optical depths of the analyzed clouds ranged from about 5–15.

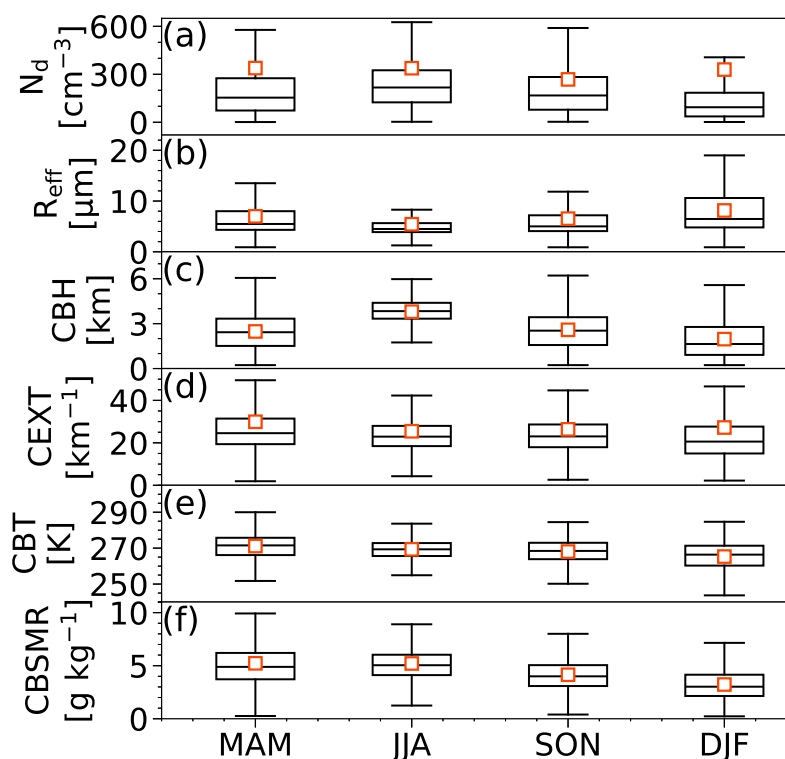


Figure 15. Cloud statistics based on 45724 individual lidar observations in 1602 liquid-water-dominated cloud fields observed over Dushanbe from June 2019 to July 2023. Seasonally resolved results are shown for (a) the cloud droplet number concentration N_d , (b) droplet effective radius R_{eff} , (c) cloud base height CBH above ground level (a.g.l.), and (d) cloud extinction coefficient CEXT (532 nm). (e) Cloud base temperature CBT and (f) respective saturation water-vapor-to-dry-air mixing ratio CBSMR are taken from meteorological data fields (ECMWF, 2026). Each box spans from the first quartile to the third quartile. The whiskers extend 1.5 times the interquartile range below and above the boxes, the mean values by red open squares, and the median values by black horizontal lines. MAM, JJA, SON, and DJF indicate spring, summer, autumn, and winter season, respectively.

5 Discussion

585 The lidar observations at Dushanbe of dust particle depolarization-ratio and RH profiles revealed that the strong decrease of the particle depolarization ratio with RH (for $RH > 80\%$) can only be explained by assuming hygroscopic dust particles that became rounded by water uptake. Chemical aging and cloud processing, as discussed in the introduction, are assumed to be responsible for the accumulation of soluble material on the surface of dust particles during long-range transport so that dust particles become increasingly hygroscopic during atmospheric transport and thus show an increased droplet nucleation efficiency with

590 lifetime (Karydis et al., 2017). Aged, contaminated dust particles can be regarded as excellent CCN because of the favorable combination of accumulated soluble material on the dust particles and their comparably large sizes (Feingold et al., 1999; Levin et al., 2005). Furthermore, Milousis et al. (2025) hypothesized that the large hygroscopic dust particles may coagulate



with the smaller anthropogenic aerosol particles below cloud base and may reduce in this way the number of anthropogenic CCN and thus the contribution of fine-mode aerosol to droplet nucleation processes. Manktelow et al. (2010) mentioned that the uptake of secondary aerosol material by dust particles and the respective reduction of concentrations of aerosol precursor gases could influence existing anthropogenic particle number and CCN concentrations (CCNC) by reduced growth and new particle formation. According to the studies of Lee et al. (2009), the total CCNC decreased by up to 10–20% in dusty regions by these processes.

From the comparison of lidar observations with respective airborne in situ measurements of dust CCNC profiles in the summertime SAL over Barbados during the SALTRACE campaign, we concluded that $n_{100,d}$ is an appropriate proxy for the dust CCNC ($N_{CCN,d}$ Haarig et al., 2019). The use of updated extinction-to- $n_{100,d}$ conversion factors (Ansmann et al., 2026) further improves the agreement between the lidar-derived and directly measured dust CCNC values.

Our multiyear Dushanbe observation of $N_{CCN,d}$, $N_{CCN,c}$, $N_{CCN,t}$, and N_d suggest that aged, hygroscopic dust significantly influences droplet nucleation processes, especially during the summer dust season. The ratio $N_{CCN,d}/N_d$, used as an indicator for the impact of the actually activated dust CCN fraction, accumulated between 0.1 and 1 in summer. This finding is in good agreement with respective simulated values of 0.05–0.4 (Karydis et al., 2017).

The Dushanbe CDNC statistics seem to represent well Central Asian cloud conditions. The CDNC values, mostly in the range from almost zero to 650 cm^{-3} agree well with observations of $30\text{--}700\text{ cm}^{-3}$ in western China (Qinghai Province, 2001–2004, Zhao et al., 2006; Karydis et al., 2011). The global simulations of Karydis et al. (2017) revealed CDNC values of $700\text{--}1000\text{ cm}^{-3}$ for the Central Asian region and western China during the winter season (December to February) and $900\text{--}1100\text{ cm}^{-3}$ for these regions during the summer season (June to August). The simulated larger CDNC values are due to the neglect of collision-coagulation and entrainment processes in the simulations (Karydis et al., 2011, 2017).

Finally, we have to comment on recent findings published by Choudhury et al. (2026). These authors concluded from their global ACI study with focus on the potential impact of dust particles on droplet formation that mineral dust has a negligible impact on cloud droplet numbers at all. The susceptibility $s_{ACI,liq}$ was close to zero for aerosol scenarios with a dust fraction close to 1. The study of Choudhury et al. (2026) was based on global observations with spaceborne lidar and radar and was restricted to the Oceans and continental regions, dominated by pollution, without a strong desert impact. The dust belt region and further regions with significant dust sources had to be excluded in their study because of the lack of good and useful observations. However, this lack of data leads to the question regarding the representativeness of their findings arises. Karydis et al. (2011, 2017) pointed out that the contribution of fresh and aged dust to cloud droplet formation is expected to be strongest in regions with significant dust sources. Our observations in the Central Asian dust belt subregion are in line with these findings of Karydis et al. (2017). As mentioned, we found $s_{ACI,liq}=0.325$ with an uncertainty of 30–40% based on the Dushanbe aerosol mixtures. Choudhury et al. (2026) obtained globally averaged values of $s_{ACI,liq}$ of 0.3–0.35 for a dust fraction <0.3 , of around 0.25 for a dust fraction of 0.3–0.6, about 0.1–0.2 for a dust fraction of 0.6–0.9, and <0.05 for a dust fractions of >0.9 .



625 6 Conclusions

We studied the impact of fresh and aged dust on cloud evolution and microphysical properties in Central Asia. The upgrade of the Dushanbe Polly lidar by implementing the dual-FOV polarization lidar technique was a big and unique step forward in our research strategy and capability to characterize aerosol-cloud-interaction in one of the key regions of climate change with very complex aerosol characteristics. The combination of the POLIPHON lidar method, used to obtain CCN reservoir estimates for the dust and non-dust aerosol fractions, in combination with the dual-FOV lidar technique opened the door to detailed ACI studies at Dushanbe.

Caused by changing meteorological conditions from summer (dust season) to winter with comparably low dust emissions, we observed a clear seasonal cycle in the observed CDNC as well as in the estimated dust and total CCN reservoir time series. When comparing the dust CCNC values (for 0.2% water supersaturation) and the CDNC values, which are roughly representative for the actually activated part of the CCN reservoir, we found a clear and relevant impact of dust on droplet nucleation processes during the summer half year (0.1–1 relative contribution to the activated CCN). A less pronounced impact during the winter season (December to February) was observed because of the much lower number of dust events and even the absence of dust on many winter days, while the anthropogenic aerosol sources continuously produce particles throughout the year. A specific feature of the dust burden in Central Asia is the unknown salt dust fraction. More research to quantify this influence is required.

The impact of fresh and aged, contaminated dust will remain probably at a similar impact level during the dust seasons in future. This needs to be considered in respective simulations of future climate conditions in many regions around the globe, especially in Africa, western North America, and Asia.

7 Data availability

Polly lidar observations (level 0 data, measured signals) are in the PollyNet database (PollyNET, 2026). All the analysis products are available at TROPOS upon request (polly@tropos.de). Lidar profiles from the ACTRIS station in Dushanbe are available via ACTRIS services at <https://data.earlinet.org/> (last access: 16 June 2026) as well.

Backward trajectory analysis has been performed by air mass transport computation with the NOAA (National Oceanic and Atmospheric Administration) HYSPLIT (HYbrid Single-Particle Lagrangian Integrated Trajectory) model http://ready.arl.noaa.gov/HYSPLIT_traj.php (HYSPLIT, 2026).

The Collection 6 MAIAC AOT data from the Terra and Aqua MODIS sensors were downloaded from the MODIS MAIAC data base (MAIAC, 2026).



8 Author contributions

CJ, AA, FL, JH, and AC designed the paper. AA wrote the manuscript and performed an extended literature study. The data analysis was performed by FL, CJ, JH, VF, HB, and AC. The Dushanbe lidar station was operated by RE, DA, JH, SFA and SKK. CJ, AA, AC, FL, JH, PS and UW were involved in the elaboration of the final version of the manuscript.

9 Competing interests

The contact author has declared that none of the authors has any competing interests.

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