

Ceilometer Retrieval of the Boundary Layer Vertical Aerosol Extinction Structure

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ABSTRACT

The CT25K ceilometer is a general-purpose cloud height sensor employing lidar technology for detection of clouds. In this paper it is shown that it can also be used to retrieve aerosol optical properties in the boundary layer. The authors present a comparison of the CT25K instrument with the aerosol lidar system and discuss its good overall agreement for both the range-corrected signals and the retrieved extinction coefficient profiles. The CT25K aerosol profiling is mostly limited to the boundary layer, but it is capable of detecting events in the lower atmosphere such as mineral dust events between 1 and 3 km. Assumptions needed for the estimation of the aerosol extinction profiles are discussed. It is shown that, when a significant part of the aerosol layer is in the boundary layer, knowledge of the aerosol optical depth from a sun photometer allows inversion of the lidar signal. In other cases, surface observations of the aerosol optical properties are used. It is demonstrated that additional information from a nephelometer and aethalometer allows definition of the lidar ratio. Extinction retrievals based on spherical and randomly oriented spheroid assumptions are performed. It is shown, by comparison with the field measurements during the United Arab Emirates Unified Aerosol Experiment, that an assumption about specific particle shape is important for the extinction profile inversions. The authors indicate that this limitation of detection is a result of the relatively small sensitivity of this instrument in comparison to more sophisticated aerosol lidars. However, in many cases this does not play a significant role because globally only about 20% of the aerosol optical depth is above the boundary layer.

1. Introduction

The role of atmospheric aerosols in modifying the radiation budget of the earth-atmosphere climate system is being increasingly understood and recognized

(Hansen et al. 1997; Haywood et al. 1999; Ramanathan et al. 2001). There are still large uncertainties of the aerosol radiative forcing on regional scales (Houghton et al. 2001) because of the lack of sufficient knowledge of aerosols' optical, physical, and chemical properties and their large spatial and temporal variability.

Much of the recent work has been devoted to reducing aerosol forcing uncertainties by using global circulation models (Chin et al. 2002; Takemura et al. 2002) and transport models (Collins et al. 2001). Establish-

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TABLE 1. List of instrumentation used in this study.

Experiments and instruments		
UAE ²	SAWA	Wavelength (nm)
CT25K ceilometer	CT25K ceilometer	905
	AEROSOL FUB lidar	355, 532, 1064
AERONET Cimel	AERONET Cimel	380, 440, 500, 675, 870, 936, 1020
	Microtops (ozonometer, aerosol)	305, 312, 320, 380, 440, 500, 675, 870, 936, 1020
TSI nephelometer		450, 550, 700
AE-30 aethalometer		370, 470, 520, 590, 660, 880, 950
Radio soundings	Radio soundings	

ment of observational networks such as the Aerosol Robotic Network (AERONET; Holben et al. 2001), European Aerosol Research Lidar Network (EARLINET), Micropulse Lidar Network (MPLNET; Welton et al. 2001), and Regional East Atmospheric Lidar Mesonet (REALM) dedicated to monitoring aerosol properties and vertical distribution supported by satellite observations (Wielicki et al. 1996) resulted in fast progress in this field.

The vertical distribution and composition of aerosols and their optical properties are needed as input to radiative transfer models allowing for determination of aerosol radiative forcing. Although lidars are good tools for mapping the spatial and temporal distribution of the atmospheric aerosol, it is not straightforward to obtain quantitative estimates of atmospheric aerosol concentration or aerosol extinction. Major difficulties come from uncertainties in the aerosol extinction-to-backscatter ratio (called lidar ratio) and the lidar calibration constant. The lidar ratio depends on the aerosol phase function in the backscatter direction and the single-scattering albedo (SSA). Both parameters can be measured at the surface, but their vertical variability may significantly increase the uncertainties of aerosol optical properties derived from lidar measurements. Other sources of lidar (or ceilometer) uncertainties are related to laser signal and receiver characteristics. There are calibration techniques being developed (Zhang and Hu 1997; O'Connor et al. 2004) that partially rely on additional information and atmospheric conditions.

In this paper we discuss the use of ceilometers to measure aerosol optical properties. These instruments are relatively robust and inexpensive and are widely deployed at airports to measure cloud bases. Section 2 describes observational sites and instruments used in this study. Section 3 provides technical information about the CT25K ceilometer, overlap, and water vapor corrections. In section 4 we compare the range-corrected lidar and ceilometer signals based on the Saharan Aerosol over Warsaw (SAWA) campaign in

2005. In section 5 we describe an algorithm to derive the lidar ratio based on the nephelometer and aethalometer observations and present comparison of inversion methods used to obtain vertical profiles of aerosol extinction coefficient.

2. Description of observation sites and instrumentations

The experimental part of this study is based on two campaigns: United Arab Emirates (UAE) Unified Aerosol Experiment (UAE²; Remiszewska et al. 2007; Markowicz et al. 2008) and SAWA (Markowicz et al. 2006). The Vaisala CT25K laser ceilometer was deployed during the UAE² and SAWA campaigns and measured the backscattered light. Table 1 describes some of the instruments used during these two campaigns.

a. United Arab Emirates Unified Aerosol Experiment

UAE² took place in August and September of 2004. During UAE² the aerosol optical, physical, and chemical properties and meteorological parameters were measured using the Mobile Atmospheric Aerosol and Radiation Characterization (MAARCO) station. MAARCO is a shipping container that was modified to function as an easily shipped laboratory. The observational station was located in the northeastern part of the UAE at 24.700°N, 54.659°E, about 10 m above the sea level close to the shore. During UAE² the aerosol absorption coefficient at the surface was obtained from the AE-30 aethalometer produced by the Magee Scientific Company (Hansen et al. 1996; Allen et al. 1999). Values measured at seven wavelengths (370, 470, 520, 590, 660, 880, and 950 nm) were corrected for scattering artifact, the deposit spot size, the AE-30 flow rate, and the manufacturer's calibration (Remiszewska et al. 2007). Measurements of aerosol scattering and hemispheric backscattering coefficients were made with two

integrating nephelometers (Model 3563, TSI Inc.; Anderson et al. 1996) at three wavelengths (450, 550, and 700 nm); one operated at near-atmospheric conditions and the other at constant relative humidity of about 35%. Radio soundings were collected at the Abu Dhabi airport about 30 km from MAARCO.

b. SAWA

SAWA took place at the University of Warsaw Institute of Geophysics in Warsaw, Poland, in April and May of 2005. This place is about 45 km from the Belsk geophysics observatory, where an AERONET station is located. One of the goals of this campaign was to estimate the aerosol forcing of mineral Saharan dust over central parts of Europe during spring. During SAWA we used an aerosol lidar developed at the Free University of Berlin (FUB). It is a multiwavelength backscattering lidar based on a solid-state 10-Hz Nd:YAG laser (Big Sky Laser CFR 200). The laser delivers pulses of 10-ns duration and energies of 60, 50, and 30 mJ at 1064, 532, and 355 nm, respectively. The FUB lidar has a multiaxial design, in which each of the beams is emitted separately and the respective positions between the beam and the telescope's axes can be independently adjusted. Such a solution allows researchers to compensate for differences in the overlap functions resulting from the wavelength-dependent beam divergence (Stelmaszczyk et al. 2005). The receiving telescope's field of view (FOV) is typically set to 1.5 mrad during a daylight operation. It can be increased by opening the iris placed in the focal plane of the telescope. In this way the overlap range can be significantly reduced; however, the compromise between the incomplete overlap and the background light noise must always be attained. The large telescope's FOV effects the multiple scattering, which cannot be simply neglected, and it will affect the retrieved backscatter and extinction coefficients. In this study we neglected the multiple scattering effects.

To detect the returning signals, two photomultiplier tubes (Hamamatsu R7400-U04 and R7400-U02) are used for the UV and visible (VIS) and a silicon avalanche photodiode (EG&G C30954/5E) for the IR. The vertical resolution for this device is 7.5 m and results from the bandwidth of a transient recorder (20 MHz, Licel). The intensity at 532 nm is additionally measured in two perpendicular polarizations, one being parallel and the other perpendicular to the laser light. To increase the signal-to-noise ratio (SNR), the averaging over typically 1000 laser shots is required. This corresponds to approximately 1 min, 40 s of acquisition time for a single aerosol profile.

Two sun photometers were used: Cimel and Microtops. The Cimel sun photometers are part of the AERONET sun photometer network (Holben et al. 2001) and are used to measure the direct and diffuse solar radiation in eight spectral bands (340, 380, 440, 500, 675, 870, 936, and 1020 nm). The spectral aerosol optical thickness (AOT) and the total water vapor are calculated (Plana-Fattori et al. 1998) on the basis of the direct observation. Scans of diffuse sky radiation provide information for the single-scattering albedo, the scattering phase function, and the aerosol size distribution retrievals (Dubovik and King 2000). The Microtops sun photometer is used to measure the AOT at 380, 440, 500, 675, 870, and 1020 nm; the total columnar water vapor; and ozone (Morys et al. 2001). Radio soundings from the Legionowo station, which is 35 km from the Institute of Geophysics, were also included in the analysis.

3. The CT25K ceilometer

a. General information

The CT25K ceilometer is a general-purpose cloud height sensor employing lidar technology for detection of clouds (O'Connor et al. 2004), but it also allows researchers to determine the mixing height (Eresmaa et al. 2006). The CT25K provides reliable determination of cloud height up to 7.5 km with vertical resolution of 15 or 30 m (Table 2). The transmitter system is based on the indium gallium arsenide (InGaAs) laser diode at 905 ± 5 nm, its repetition rate is 5.57 kHz, and the divergence of light beam is ± 0.53 mrad. The CT25K receiver consists of two identical silicon avalanche photodiodes (APDs), which are sensitive at around 905 nm. The first APD receives the actual signal and sends the pulses current through a transimpedance amplifier with selectable gain. The second one compensates most of the optical cross-talk appearing in the lens system by means of a half-bridge connection. The divergence of the receiver's FOV is only 0.66 mrad. Narrow light beam and receiver FOV reduce the effects of the multiple scattering.

Averaging time of the acquisition system can be set between 15 and 120 s. During measurement campaigns we set vertical resolution to 15 m and time averaging was set to 15 s.

Built with the unique single-lens design, the CT25K provides good accuracy starting from close to zero height. With several caveats it is possible to use the ceilometer data to retrieve aerosol backscatter at 905 nm. For ambient temperature below 20°C the CT25K laser diode is heated and kept at 20°–25°C. The maxi-

TABLE 2. The CT25K Vaisala ceilometers and FUB lidar technical information.

		CT25K	FUB lidar
General information			
Vertical resolution		15 m	7.5 m
Acquisition (averaging) time		15–120 s	100 s
Measurement range		15–7500 m	7.5–10 000 m
Mean overlap range		Up to 550 m	Up to 350 m
Transmitter			
Laser source		InGaAs laser diode	Nd:YAG (Big Sky Laser, CFR 200)
Center wavelength		905 ± 5 nm at 25°C	1064, 532, 355 nm
Repetition rate		5.57 kHz	10 Hz
Nominal pulse properties at full range measurement	Energy	$1.6 \mu\text{J} \pm 20\%$ (factory adjustment)	60, 50, 30 mJ in 1064, 532, 355 nm
	Peak power	16 W typical	—
	Width, 50%	100 ns typical	8 ns
Beam divergence		± 0.53 mrad edge ± 0.75 mrad diagonal	Full divergence < 1 mrad
Receiver			
Detector		Silicon APD	Photomultiplier R7400-P04 and R7400-U02 for 355 and 532 nm, avalanche photodiode for 1064 nm
Detection type		Detection, 8 bits analog-to-digital (A/D) converter with 3-MHz transient recorder	Photon counting at 355- and 532-nm analog detection, 12 bits A/D converter with 20-MHz transient recorder
Interface filter		Center wavelength 908 nm	Center wavelength: 1064, 532, 355 nm respectively; full width at half maximum = 1 nm for all wavelengths
50% passband		35 nm at 890–925	—
FOV divergence		± 0.66 mrad	Typical 1.5 mrad
Optical system			
Optical system focal length		377 mm	800 mm
Effective lens/mirror diameter		145 mm	200 mm
Lens/mirror transmittance/reflectance		96% typical	$> 90\%$ for all wavelengths

energy emitted by the laser is at 905 ± 5 nm (at 25°C). This wavelength is temperature sensitive and changes at a rate of about 0.25 nm K^{-1} . Unfortunately, this is a critical parameter for water vapor effects, which depend significantly on emitted laser wavelength (see section 3c).

Rapid decrease of the ceilometer signal with height is observed because of water vapor absorption for large specific humidities. The CT25K reports a two-way attenuation coefficient, which is defined by the range-corrected signal

$$S = C\beta T_r^2, \quad (1)$$

where C is the instrument constant (lidar constant); β is a sum of the Rayleigh and the aerosol backscatter coefficient (backscatter coefficient for the water vapor is negligible); and T_r is the atmospheric transmittance, which can be written as

$$T_r = T_R T_A T_{\text{H}_2\text{O}}, \quad (2)$$

where T_R is the Rayleigh transmittance, T_A is the aerosol transmittance, and $T_{\text{H}_2\text{O}}$ is the water vapor transmittance.

b. Overlap correction

The overlap function describes the loss of signal strength in the near field due to the optical design of the instrument. Signals at ranges greater than the overlap range are not affected by this effect. The CT25K ceilometer is based on enhanced single-lens optics, which, in comparison to the traditional biaxial design, has a small overlap range. This provides for reliable detection of the lowest clouds and aerosols and reduces the effect of multiple scattering.

The overlap correction may be determined from horizontal measurements of a horizontally homoge-

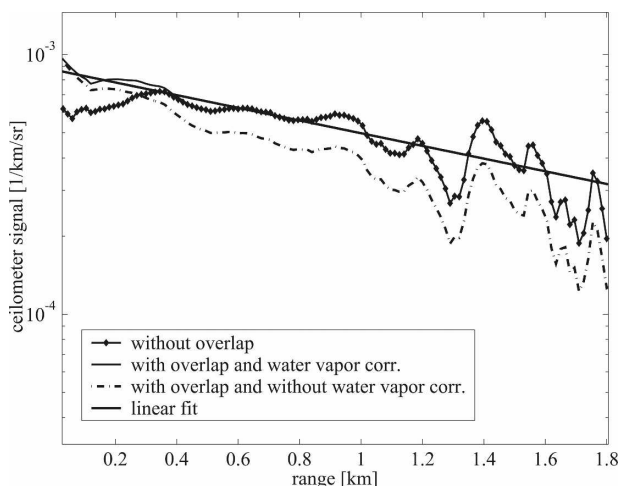


FIG. 1. Logarithm of the CT25K ceilometer range-corrected signal (two-way attenuation coefficient) as a function of range for the horizontal scan performed on 20 May 2005 (SAWA). Signals without (dotted line) and with overlap (dash line) are plotted. The solid line corresponds to the range-corrected signal with water vapor correction, and the solid straight line is a linear fit.

neous atmosphere. The deviation from the expected attenuated backscattered return and actual signal is a measure of the overlap function (Berkoff et al. 2003; Welton and Campbell 2002). In the case of the CT25K ceilometer, the overlap appears up to about 450–550 m.

The reported two-way attenuation coefficient by the CT25K system includes overlap correction, but we performed several horizontal measurements to check the reduction of the signal close to the observational system. Above the overlap range the logarithm of lidar signal is a linear function of range. The slope of this function equals twice the extinction coefficient:

$$\ln S(z) = \begin{cases} \ln(C\beta) - 2\sigma z + \ln O(z) & z < z_{\text{Overlap}} \\ \ln(C\beta) - 2\sigma z & z \geq z_{\text{Overlap}} \end{cases}, \quad (3)$$

where $O(z)$ is the overlap function. Knowledge of the backscatter coefficient at the surface allows us to estimate the lidar calibration constant C .

Reduction of a lidar signal due to the near-field problem appears up to 450 m (Fig. 1) but is compensated by the manufacturer's overlap function. Close to the instrument the signal is slightly stronger than expected. The solid line corresponds to the two-way attenuation (range corrected) signal with correction for the water vapor absorption and manufacturer overlap function. Differences between the original signal without water vapor correction and the signal corrected for the water vapor absorption are significant. For the water vapor

absorption-corrected signal, the slope is smaller and the extinction coefficient is about 0.28 km^{-1} , while without water vapor absorption it reaches 0.40 km^{-1} . About 1.2 km from the instrument the SNR decreases and reduces the CT25K range of aerosol detection. At this distance the significant fluctuation suggests horizontal inhomogeneity, but this is an artifact of SNR.

We performed several horizontal measurements (under different meteorological conditions including different ambient temperatures) to check the stability of the overlap function. We found that the variability of this correction function reaches 20% at 60 m, 15% at 95 m, 4% at 200 m, and 2% at 300 m from the ceilometer. Above 300 m changes of the overlap function are negligible. The main reason of the time variability of the overlap function is the small FOV of the CT25K ceilometer (0.5 mrad). For the ambient temperatures larger than 25°C the instruments are not temperature stabilized, and this may be the reason for the overlap correction changes. However, horizontal inhomogeneity in aerosol properties may also be a contributing factor to our estimates.

c. The CT25K signal correction due to the water vapor absorption

The electromagnetic radiation emitted by the CT25K laser diode ($905 \pm 5 \text{ nm}$) is absorbed by the water vapor. Transmission at 905 nm for the total water vapor amount of 1 g cm^{-2} is about 0.91 and decreases to 0.82 for 3 g cm^{-2} . The ceilometer signal must be corrected for water vapor absorption because of this significant reduction of transmittance. The radio sounding data (which are available usually twice a day) are used to estimate the profile of specific humidity, but this profile is scaled by the sun photometer total water vapor retrieval. This method allows us to take into account the temporal variability of the water vapor in the lower troposphere. Radiative transfer calculations are performed to determine transmission as a function of the water vapor content. In this study we use version 4.1 of the moderate resolution atmospheric transmission (MODTRAN) modeling code (Berk et al. 1998), which we run in the transmittance mode. We use the correlated- k option for the band model and use 1 cm^{-1} wavelength resolution. The water vapor absorption depends on the wavelength. For this reason, the spectral shape of the laser emitted energy needs to be known. We used the monolithic miniature spectrometer (MMS1-UV-VIS) Zeiss spectrometer to determine the emission spectrum of the CT25K laser diode. Spectral accuracy of this spectrometer is 0.3 nm, and temperature drift is about 0.02 nm K^{-1} .

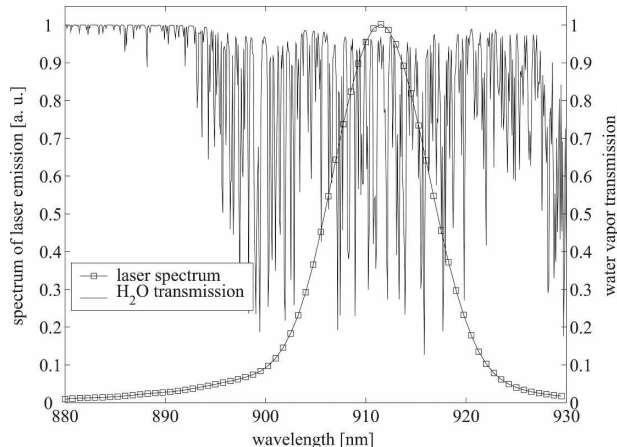


FIG. 2. The normalized spectrum of diode laser emission (solid line with squares) measured for the laser temperature of 28°C. The water vapor transmission (solid line) is plotted for the total water vapor content of 2 g cm^{-2} .

Figure 2 shows the normalized spectral dependence of laser energy (solid line with squares) and variability of transmission due to the water vapor absorption (solid line) as a function of wavelength. Emitted radiation is relatively wide and has a half width of about 10 nm. The mean water vapor transmission $\langle T_r \rangle$ can be calculated from the emission-weighted integral

$$\langle T_r \rangle = \frac{\int T_r(\lambda) E(\lambda) d\lambda}{\int E(\lambda) d\lambda}, \quad (4)$$

where $E(\lambda)$ is spectral radiance emitted by the laser diode and T_r is calculated atmospheric transmission due to the water vapor effect.

Temperature drift of the laser diode emission spectrum plays an important role. The CT25K laser diode temperature is kept at 20°–25°C, but for higher temperatures it is not stabilized. There is a 0.25 nm K^{-1} wavelength shift of laser emission. Figure 3 shows the variability of water vapor transmission for different water vapor contents as a function of the peak wavelength. Significant changes of the water vapor transmission appear only for large water vapor contents. Variability of the water vapor transmission for the total water vapor content smaller than 2 g cm^{-2} is small. Thus, the peak wavelength dependence can only be significant for tropical atmospheres.

The water vapor modification leads to a standard lidar equation where the right-hand term depends on aerosol and molecular optical properties only:

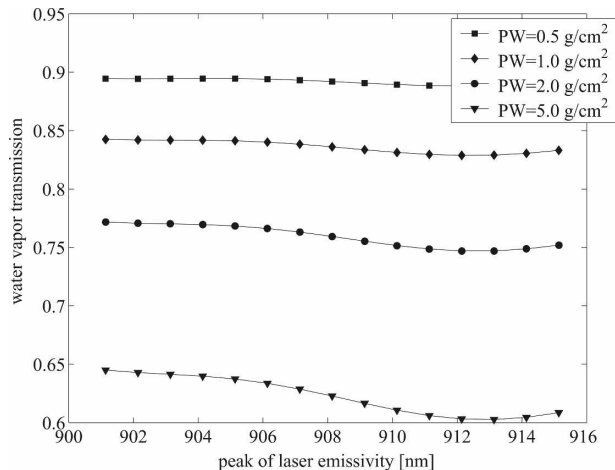


FIG. 3. Calculated water vapor transmission as a function of the peak emissivity wavelength for several total water vapor contents [precipitable water (PW; g cm^{-2})].

$$S(z) = \frac{S_{\text{raw}}(z)}{T_{\text{H}_2\text{O}}^2(z)} = C\beta(z)T^2(z), \quad (5)$$

where $S_{\text{raw}}(z)$ is a raw ceilometer signal and $S(z)$ is a ceilometer after the water vapor correction. Figure 4 shows the difference between vertical profiles of the CT25K original range-corrected signals (solid lines) and the signals corrected for the water vapor absorption (dotted). In the first case (Fig. 4a) the water vapor correction in the boundary layer reaches 30% because of the high value of specific humidity. In the second case (Fig. 4b) the specific humidity in the boundary layer is significantly smaller and this correction is also small. However, in the mineral dust layers, which are observed between 1.5 and 3 km, applying the water vapor correction in Eq. (5) leads to an increase of the ceilometer range signal by about 30%–40%.

4. Comparison of the range-corrected signals between the CT25K ceilometer and the aerosol lidar

In this section we compare the range-corrected signals of the ceilometer and the FUB lidar. Figure 5 shows temporal variability of these signals measured during SAWA on 1 April 2005. The magnitude of the range-corrected signals is different because of different instrument calibrations C . The diurnal evolution of the boundary layer measured by the ceilometer is consistent with observations made with the FUB lidar. The top of the boundary layer height is similar in both cases.

Figure 6 shows the range-corrected signals and the FUB lidar total (aerosol plus molecular) depolarization for several hours with significant dust above the bound-

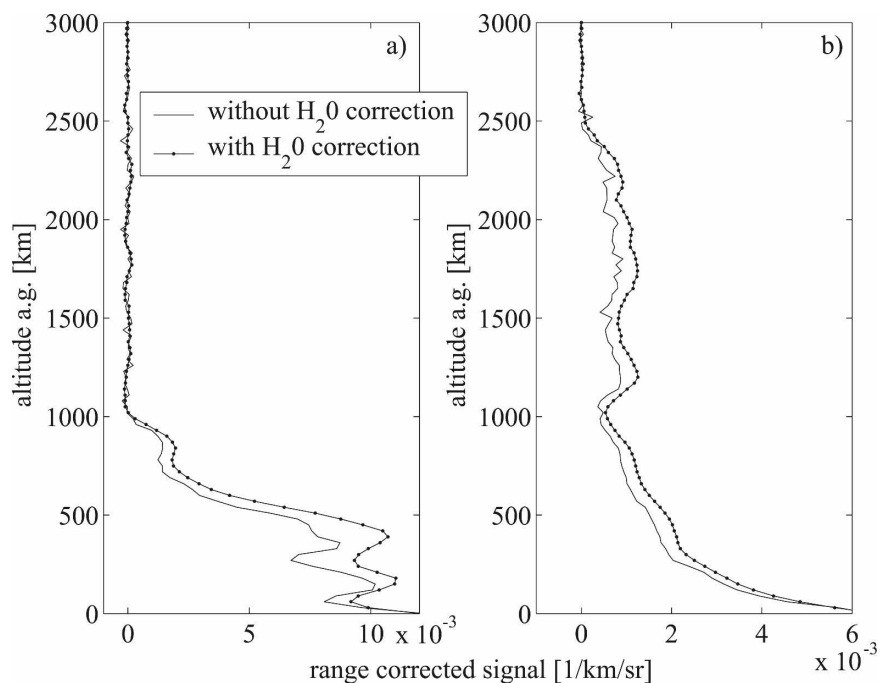


FIG. 4. Comparison of the range-corrected signal with (dots) and without (solid line) water vapor correction for two different atmospheric conditions: (a) 0320 and (b) 2230 UTC 13 Sep 2004 (UAE²). The boundary layer specific humidity is higher in (a) and so is the water vapor correction.

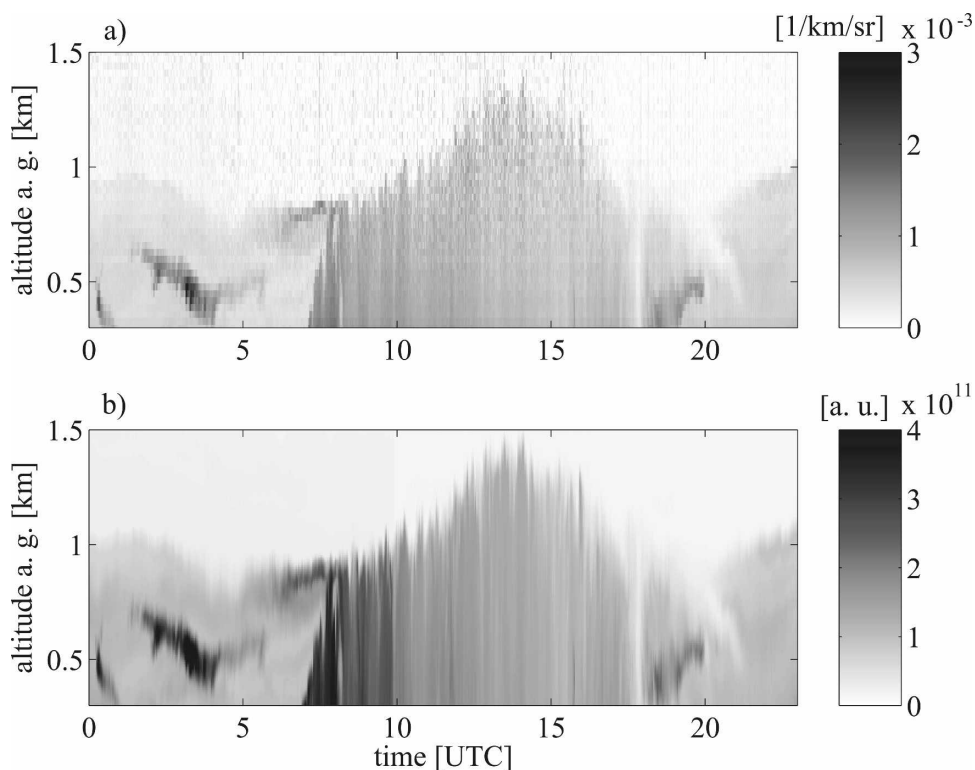


FIG. 5. The (a) ceilometer and (b) FUB lidar range-corrected signal during clear-sky conditions (SAWA; 1 Apr 2005). The FUB lidar signal is measured at 1064 nm, whereas the ceilometer is measured at 905 nm.

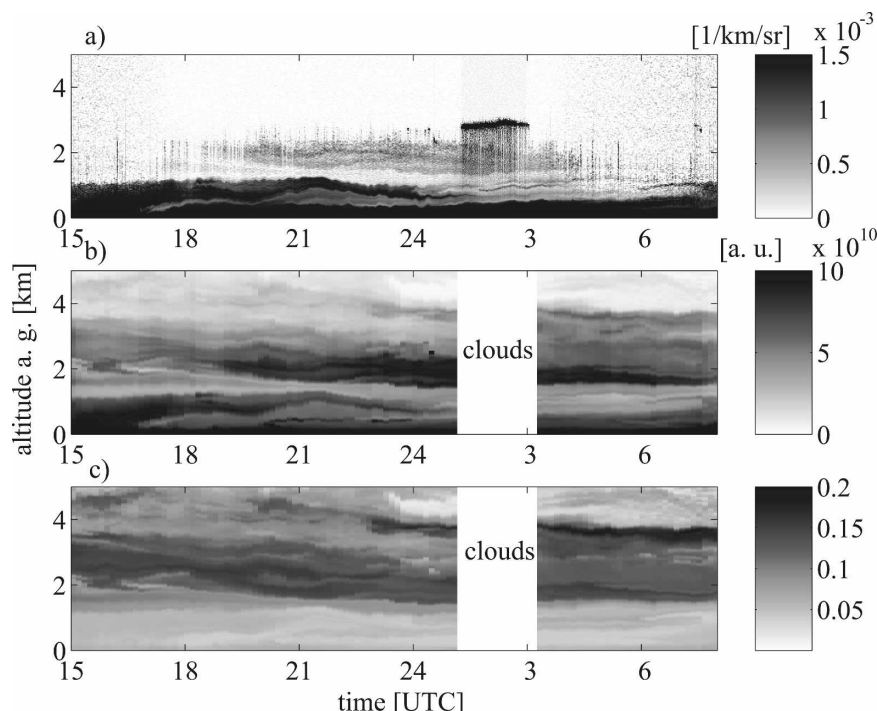


FIG. 6. The (a) ceilometers, (b) FUB lidar range-corrected signal, and (c) FUB lidar total depolarization at 532 nm during a dust event for 13–14 Apr 2005 (SAWA). The FUB lidar signal corresponds to 1064 nm, and the ceilometer signal corresponds to 905 nm.

ary layer. Relatively large values of the total depolarization above 1.5 km (Fig. 6c) are indicative of non-spherical particles transported from desert regions. The source of this transport was confirmed by the backtrajectories analysis. The ceilometer detects this aerosol plume, but the magnitude of returned signal is significantly smaller than in the FUB lidar case. The ceilometer SNR is reduced because of signal strength and water vapor absorption in the atmosphere. Thus, for a large distance, only a layer with significant optical thickness (usually a cloud) can be detected by this instrument. The dust layer is detected by the ceilometer during the nighttime because the solar radiation significantly increases the instrument's noise.

One of the important parameters related to lidar retrievals is the SNR. In this study, we use the method described by Xie and Zhou (2005), which allows us to calculate SNR based on the range-corrected signal. Figure 7 shows a comparison of the SNR as a function of the altitude for the CT25K ceilometer and the FUB lidar during the clear-sky conditions. The solid and dotted lines correspond to the FUB lidar signal during the day and at night, respectively. The dotted line and open circles correspond to the same times but for the CT25K. For the ceilometer, the SNR is significantly smaller in comparison to the FUB lidar. The SNR for the ceilometer decreases

with altitude and at 1.5–2.5 km is about 10, which is a limiting value of detection. The SNR can be improved by longer averaging times. Results in Fig. 7 were obtained for the time window of 200 s applied to both signals.

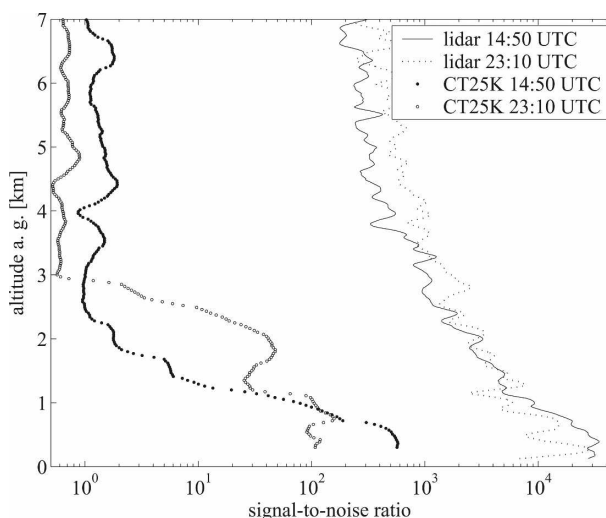


FIG. 7. The SNR for the FUB lidar (solid and dotted line) and the CT25K ceilometer (open and dotted circles) as a function of altitude on 13 Sep 2004. The SNRs were calculated based on the range-corrected signals (after overlap correction) averaged over 200-s intervals.

TABLE 3. Description of the lidar inversion methods.

Method	Required information	Assumptions
Backward modified Klett ^a	AOT at 905 nm, molecular extinction and backscatter coefficients	Lidar ratio constant with altitude, top of the aerosol layer (1.5–2 km)
Forward Klett ^b	Lidar ratio at 905 nm, aerosol extinction near the surface	Lidar ratio constant with altitude, spherical or spheroid particles
Forward Porter ^c	Lidar ratio at 905 nm, lidar calibration value, aerosol phase function at 180° near the surface	Aerosol phase function at 180° constant with altitude, spherical or spheroidal particles

$$^a \beta_A(z-1) = \frac{S(z-1)\Psi(z-1, z)}{\frac{S(z)}{\beta_R(z) + \beta_A(z)} + \frac{\Delta z}{R_A} [S(z) + S(z-1)\Psi(z-1, z)]} - \beta_R(z-1).$$

$$^b \beta_A(z) = \frac{S(z)}{\frac{S(z-1)\Psi(z-1, z)}{\beta_R(z-1) + \beta_A(z-1)} - \frac{\Delta z}{R_A} [S(z) + S(z-1)\Psi(z-1, z)]} - \beta_R(z)$$

$$R_A = \frac{\int \beta_A(z) dz}{\text{AOT}} \quad \text{adjusted value of inverse lidar ratio.}$$

$$^c \sigma_{As}(z) = \frac{1}{P_A} \left\{ \frac{4\pi S(z)}{C[T_M(z - \Delta z)T_A(z - \Delta z)]^2} - P_M \sigma_{Ms} \right\}$$

where β_A and β_M are aerosol and molecular backscatter coefficients, $S(z)$ = range-corrected signal, Δz = vertical layer, R_A and R_M are the inverse of lidar ratio for aerosol and molecular scattering, AOT = aerosol optical thickness, σ_{As} and σ_{Ms} = aerosol and molecular scattering coefficients, P_A and P_M are aerosol and molecular scattering phase functions at 180°, and Ψ is defined by equation

$$\Psi(z-1, z) = \exp \left\{ \left(\frac{1}{R_A} - \frac{1}{R_R} \right) [\beta_R(z) + \beta_R(z-1)] \Delta z \right\}.$$

5. Inversion algorithms for the ceilometer observation data

To obtain vertical profiles of aerosol extinction and backscatter we use three different retrieval methods. These require additional information about the aerosol optical properties such as lidar ratio (ratio of particle extinction to backscatter coefficient) or AOT. In the Klett technique (Klett 1981) the lidar ratio is assumed to be known. Modification of this method allows application of observations of the aerosol optical depth to avoid the lidar ratio assumption. In both cases the lidar ratio is constant with altitude, but in the second case the lidar ratio is calculated as an adjustment of the integrated extinction coefficient to the total AOT. Welton et al. (2002) discussed that using additional information, such as aerosol optical depth in Klett's technique, reduces errors in derived extinction profiles to $\pm 0.005 \text{ km}^{-1}$. However, assumptions about constant-with-altitude lidar ratio may lead to uncertainties (Sasano et al. 1985).

The small range of ceilometer aerosol detection leads to complications when using the modified Klett's retrieval algorithm. The AOT measured by the sun photometers is defined for the whole atmospheric column,

while the ceilometer detects aerosol mostly in the boundary layer. The total AOT, which should be applied in the inversion computation, is difficult to estimate and depends on the meteorological situation. In addition, there is an assumption in the algorithm that at some level the total backscatter equals the molecular backscatter in the absence of aerosols. In the case of the CT25K we assume that this level is at about 1.5–2 km in height where the aerosol backscatter coefficient is still significant in most atmospheric conditions.

The last method used in this study is described by Porter et al. (2000). Their forward-stepping algorithm requires information about the single scattering albedo and the backward phase function or the lidar ratio. In addition, the lidar calibration coefficient C (see Table 3) is also required. This quantity is usually unknown but can be estimated on the basis of the horizontal measurements. Porter et al. (2000) showed that using incorrect values of the lidar calibration value leads to large errors in the derived aerosol scattering coefficient, and these errors increase with distance. The lidar calibration can be done only during horizontally uniform meteorological conditions when extinction is independent of the distance.

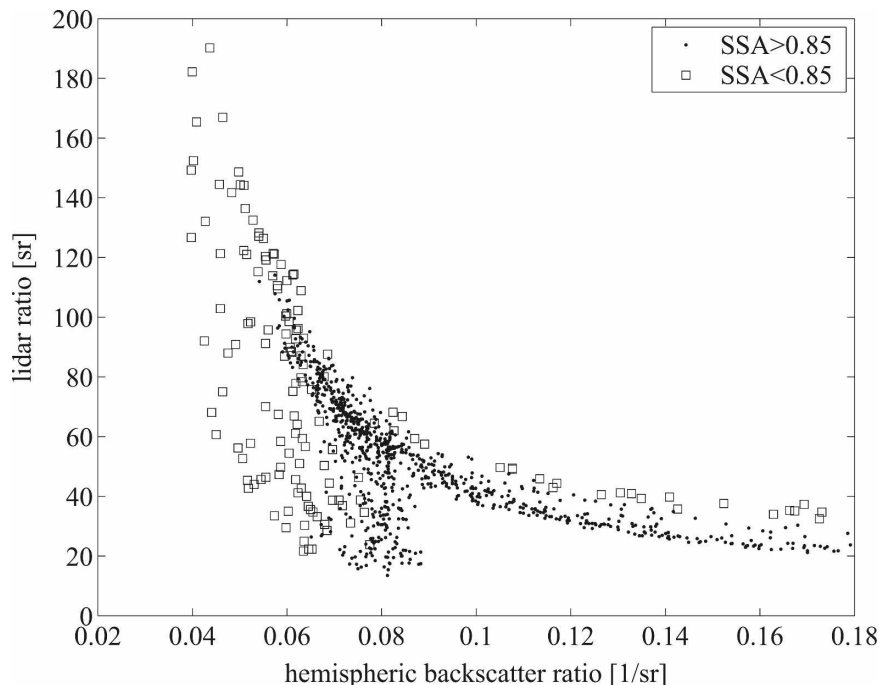


FIG. 8. The lidar ratio as a function of the hemispheric (nephelometer) backscatter to the total scattering coefficient ratio at 550 nm. Closed circles are for $SSA > 0.85$ and open squares are for $SSA < 0.85$; spherical particles are assumed here.

a. Additional aerosol optical information

The backward scattering can be estimated assuming the Henyey–Greenstein relationship:

$$P_{HG}(\cos\Theta) = \frac{1 - g^2}{(1 + g^2 - 2g \cos\Theta)^{3/2}}, \quad (6)$$

or performing more detailed calculations base on the Lorenz–Mie solution (Bohren and Huffman 1983). Assuming asymmetry parameter $g = 0.6$ and $g = 0.7$ and the single scattering albedo $\omega = 1$, the lidar ratio

$$LR = \frac{4\pi\omega}{P_{HG}(\pi)} \quad (7)$$

is equal to 80 and 121 sr, respectively. More detailed calculations indicate that the lidar ratio for the same asymmetry parameter can be significantly different, for example, because of refractive index changes. In addition, the asymmetry parameter is not measured directly and can only be estimated, usually with significant uncertainties, on the basis of the nephelometer observations (Andrews et al. 2006). In section 5b we will estimate the lidar ratio based on the surface observations.

b. Retrieval of the lidar ratio based on the surface observation

Determination of the lidar ratio from the integrating nephelometer is not straightforward because the mea-

sured hemispheric backscatter coefficient does not correlate well with the backscattering coefficient. Figure 8 shows a relationship between the lidar ratio and the backscatter ratio (ratio of the hemispheric backscatter to scattering coefficient at 180°) for two SSA ranges. These results are derived on the basis of the Lorenz–Mie solution to scattering problems on spherical particles for different refractive index and for the lognormal aerosol size distributions. It can be seen that estimation of the lidar ratio based on the backscatter fraction, especially for large particles (small backscatter fraction), is not a function of the SSA only. Generally speaking, the lidar ratio is a function of the refractive index, the aerosol size distribution, and particle shape. For this reason, we will retrieve the aerosol size distribution and refractive index, and estimate the lidar ratio.

To retrieve sizes and refractive index we use scattering, hemispheric backscatter, and absorption coefficients at three wavelengths (450, 550, and 700 nm). The scalar cost function (Rodgers 2000)

$$J = [(\mathbf{y} - F(\hat{\mathbf{x}}))^T \mathbf{S}_y^{-1} (\mathbf{y} - F(\hat{\mathbf{x}})) + (\hat{\mathbf{x}} - \mathbf{x}_a)^T \mathbf{S}_a^{-1} (\hat{\mathbf{x}} - \mathbf{x}_a)] \quad (8)$$

is minimized, where F denotes forward model; $\mathbf{y}$ is a measurement vector; $\hat{\mathbf{x}}$ are retrieved parameters; $\mathbf{x}_a$ represents the best guess of retrieval parameters based

of a priori information; and $\mathbf{S}_y$ and $\mathbf{S}_a$ are covariance matrices for uncertainties of measurements and model disparity and uncertainties of a priori information, respectively. The solution can be found by Newton iteration method:

$$\hat{\mathbf{x}}_i = \hat{\mathbf{x}}_{i-1} + \mathbf{S}\{\mathbf{K}^T \mathbf{S}_y^{-1} [\mathbf{y} - F(\hat{\mathbf{x}}_{i-1})] + \mathbf{S}_a^{-1} (\mathbf{x}_a - \hat{\mathbf{x}}_{i-1})\}, \quad (9)$$

where $\hat{\mathbf{x}}_i$ represents retrieval parameters for i th iteration, $\mathbf{K}$ is a weighting function whose elements are defined by

$$\mathbf{K}_{ij} = \frac{\partial F_i(\hat{\mathbf{x}})}{\partial \hat{\mathbf{x}}_j}, \quad (10)$$

and $\mathbf{S}$ denotes

$$\mathbf{S} = (\mathbf{S}_a^{-1} + \mathbf{K}^T \mathbf{S}_y^{-1} \mathbf{K})^{-1}. \quad (11)$$

The forward model is described by

$$\mathbf{y} = F(\mathbf{x}, \mathbf{b}) + \varepsilon, \quad (12)$$

where $\mathbf{y}$ and $\mathbf{x}$ represent vectors of observation and retrieved parameters, $\mathbf{b}$ represents a vector of constant model input parameters, and ε is the error between model and observation. This forward model is based on either the Lorenz–Mie solution of the scattering problem for the spherical particles or the T matrix (Mishchenko and Travis 1998; Mishchenko et al. 2000) solution for randomly oriented spheroidal homogeneous particles. In addition, we consider only an internal mixture of the aerosol particles, and the aerosol size distribution is described by two lognormal probability distribution functions. From the forward model we obtain measured quantities—scattering, hemispheric backscatter, and absorption coefficients:

$$\mathbf{y}_i = F(\mathbf{x}_i, \mathbf{b}) + \varepsilon = \pi \int r^2 n(r) Q_i(\mathbf{x}, m) dr + \varepsilon, \quad (13)$$

where $\mathbf{y}_i$ is a vector of the measured quantities; $n(r)$ is an aerosol size distribution; and Q_i is a cross-section efficiency for the scattering, the hemispheric backscattering, and the absorption ($i = \text{scat}, \text{bscat}, \text{abs}$) obtained from the forward model.

The measurement vector $\mathbf{y}_i$ includes nine elements: scattering, hemispheric backscatter, and absorption coefficients at three wavelengths (450, 550, and 700 nm). The retrieved vector contains a number of particles in fine and coarse modes, a mode radius of fine and coarse particles, and an imaginary part of the refractive index. For spheroids we retrieve the aspect ratio defined as the ratio of spheroid axes. We assume a constant value of the real part of the refractive index equal to 1.52, which is a typical value for the mineral dust. The covariance matrices $\mathbf{S}_y$ and $\mathbf{S}_a$ are defined on the basis of

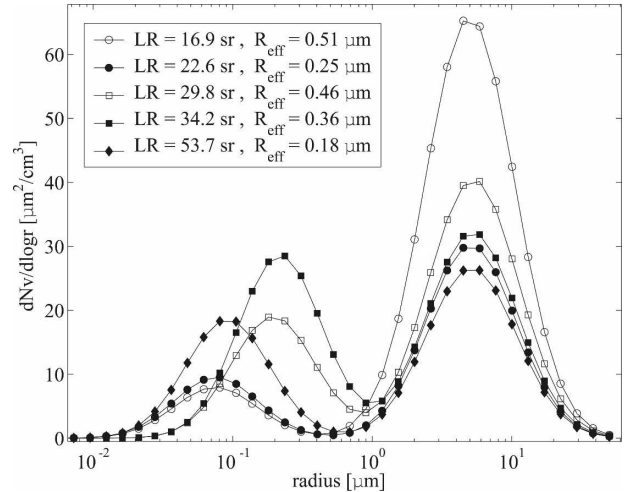


FIG. 9. Retrieved aerosol size distribution ($dN/d \ln r$) and lidar ratios for several effective radii R_{eff} (μm).

observations and a priori (climatological) information about errors. The Optical Properties of Aerosols and Clouds—Global Aerosol Data Set (OPAC–GADS; Hess et al. 1998) includes global aerosol climatology, and we used it to estimate the covariance matrices.

When the size distribution, the effective radius (ratio of third to second moment of size distribution), and the refractive index are retrieved we can calculate the lidar ratio at 905 nm. Figure 9 shows several aerosol size distribution retrievals and lidar ratios. The relationship between the shape of the aerosol size distribution and the lidar ratio is complicated because of refractive index influence. The correlation coefficient between the aerosol effective radius and the lidar ratio obtained from the retrieval is small (-0.34).

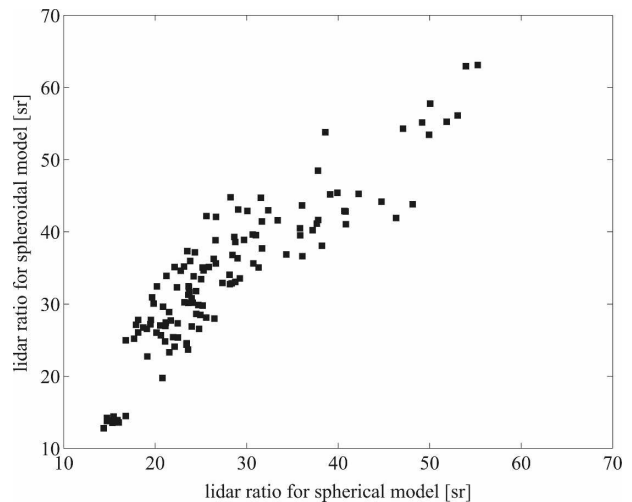


FIG. 10. Comparison of the lidar ratio for spherical and spheroidal particles.

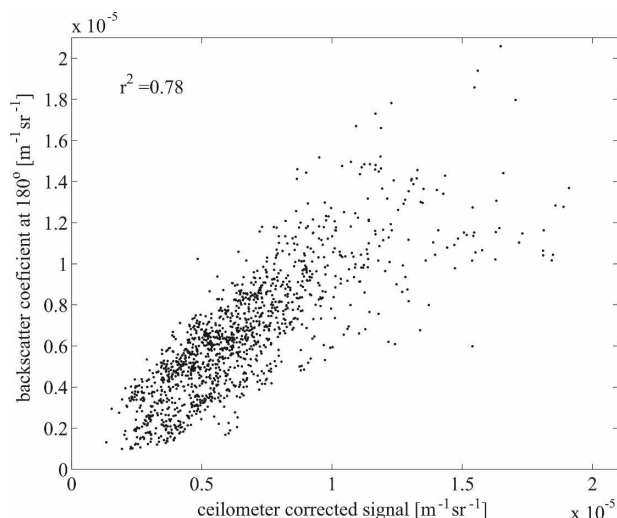


FIG. 11. Comparison of the backscatter coefficients (at 180°) retrieved from the nephelometer and aethalometer surface observations and compared to the CT25K ceilometer signal at 30 m above the surface. For the CT25K, the lidar constant can be determined from this correlation.

Figure 10 presents the correlation between the lidar ratio for spherical and spheroidal particles. Differences between these quantities are caused by different scattering properties of the spherical and nonspherical particles. The lidar ratio for the spheroidal model is larger

than that for the equivalent (by volume) spheres. This result is consistent with that obtained by Mishchenko et al. (1997) and Mattis et al. (2002).

c. Comparison of aerosol optical properties from remote and in situ observations

Figure 11 presents the correlation between the backscatter coefficient at 180° obtained from the surface observations and from the CT25K measurements. The surface backscatter is derived using the retrieval procedure that combines data from the nephelometer and the aethalometer with an assumption about spherical particles. Ceilometer backscatter coefficients were estimated at 30 m above the surface (the first data point) from the CT25K signal. The ceilometer backscatter coefficient [see Eq. (1)] depends on the accuracy of lidar constant and overlap correction. The correlation coefficient of the backscatter coefficient at 180° based on these two different methods is significant (0.78), but the root-mean-square (rms) difference is relatively large ($2.1 \times 10^{-6} \text{ m}^{-1} \text{ sr}^{-1}$). The difference between these quantities is due to retrieval and observation errors. The retrieval uncertainties of the lidar ratio are between 10% and 15%. Observational errors for the scattering and the hemispheric backscattering coefficient are about 10% and error in the estimation of the ab-

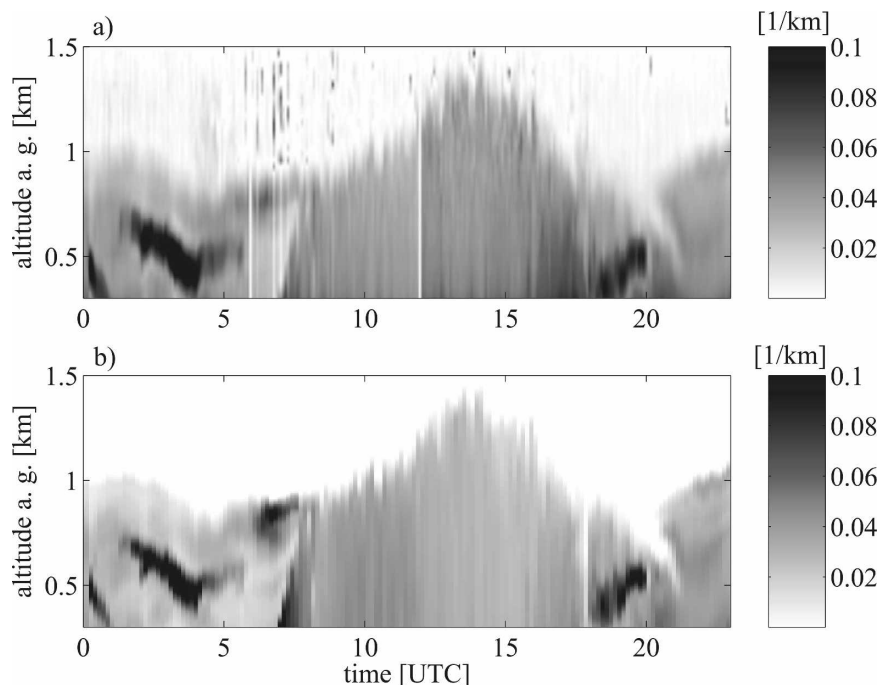


FIG. 12. Diurnal variability of retrieved aerosol extinction coefficients based on the modified backward Klett's algorithm. (a) The ceilometer retrievals and (b) aerosol FUB lidar observation performed during SAVA on 1 Apr 2005. The FUB lidar signal is 1064 nm, and the ceilometer is 905 nm.

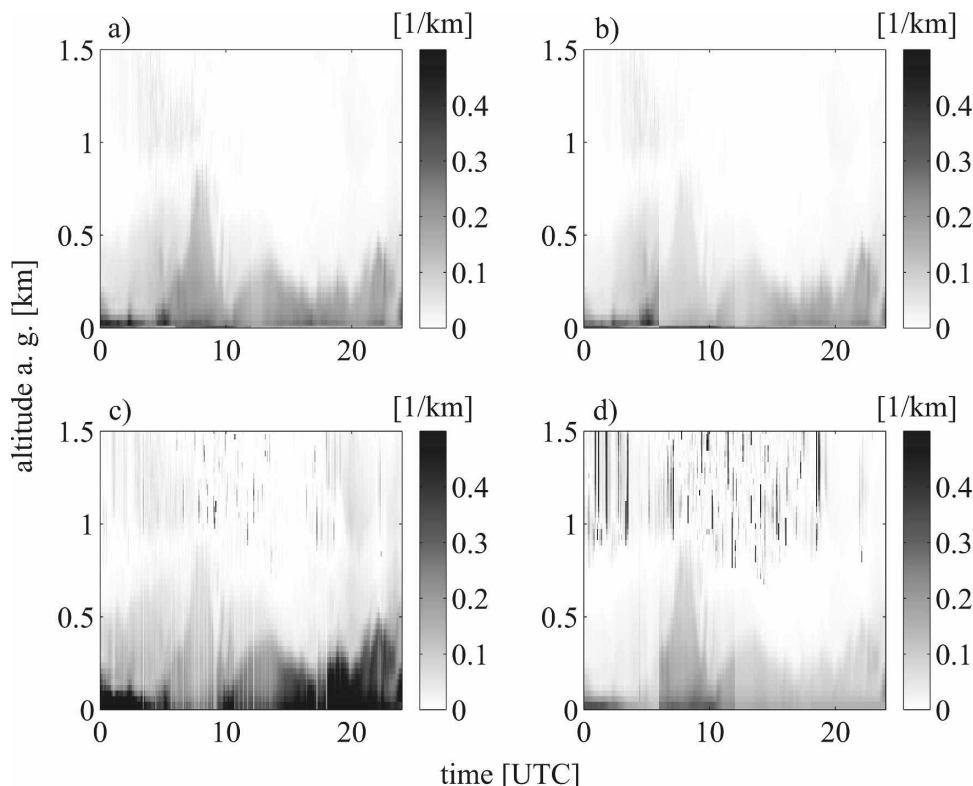


FIG. 13. Comparison of the diurnal (UAE²; 9 Sep 2004) variability of the aerosol extinction coefficient obtained by (a) Porter's spheroidal algorithm, (b) Porter's spherical algorithm, (c) Klett's backward algorithm (50% of the AOT between 0 and 1.5 km), and (d) forward Klett's algorithm for a constant lidar ratio.

sorption coefficient is about 12% (Remiszewska et al. 2007).

Significant correlation between surface observations and those measured by the CT25K backscatter is consistent with the study of Muenkel et al. (2004). Based on one year of data they show that the CT25K backscatter coefficient at 30 m is strongly correlated ($r = 0.83$) with the PM₁₀ (particulate matter particles of 10- μ m diameter or less) concentrations.

6. Comparison of the aerosol extinction coefficient obtained from a different approach

In this section we discuss ceilometer and FUB lidar inversions based on data from two field projects: SAWA and UAE². In Fig. 12a temporal variability of the aerosol extinction coefficients profiles retrieved from the CT25K ceilometer is presented and the FUB lidar inversion is shown in Fig. 12b. These inversions are based on a modified backward Klett algorithm. The structure of the aerosol layer is well captured by both instruments but some differences are also apparent: the ceilometer's SNR is smaller, and extinction based on

the CT25K ceilometer is enhanced close to the surface and reduced in the upper boundary layer in comparison to the FUB lidar results.

Figure 13 shows a comparison of the diurnal variability of the retrieved extinction coefficients based on different algorithms for the UAE² data. Figure 13a is for Porter's scheme with an assumption of spheroidal particles, Fig. 13b is for Porter's and a spherical model, Fig. 13c is for the modified backward Klett method with 50% of the total AOT between the ground and 1.5 km, and Fig. 13d is for the forward Klett method with the constant lidar ratio. The main features of the aerosol layer structure are similar. The differences between Figs. 13a and 13b are due to the lidar ratio values. For spheroidal particles the lidar ratio is larger than for the spherical particles, and an increase of the lidar ratio corresponds to a larger value of the retrieved extinction coefficients. Figure 13c presents calculations based on the modified Klett algorithm with the assumption that 50% of the total AOT is between the surface and 1.5 km. Apparently for this case the extinction is too large, especially close to the surface—probably because less than 50% of AOT is between the ground and 1.5 km

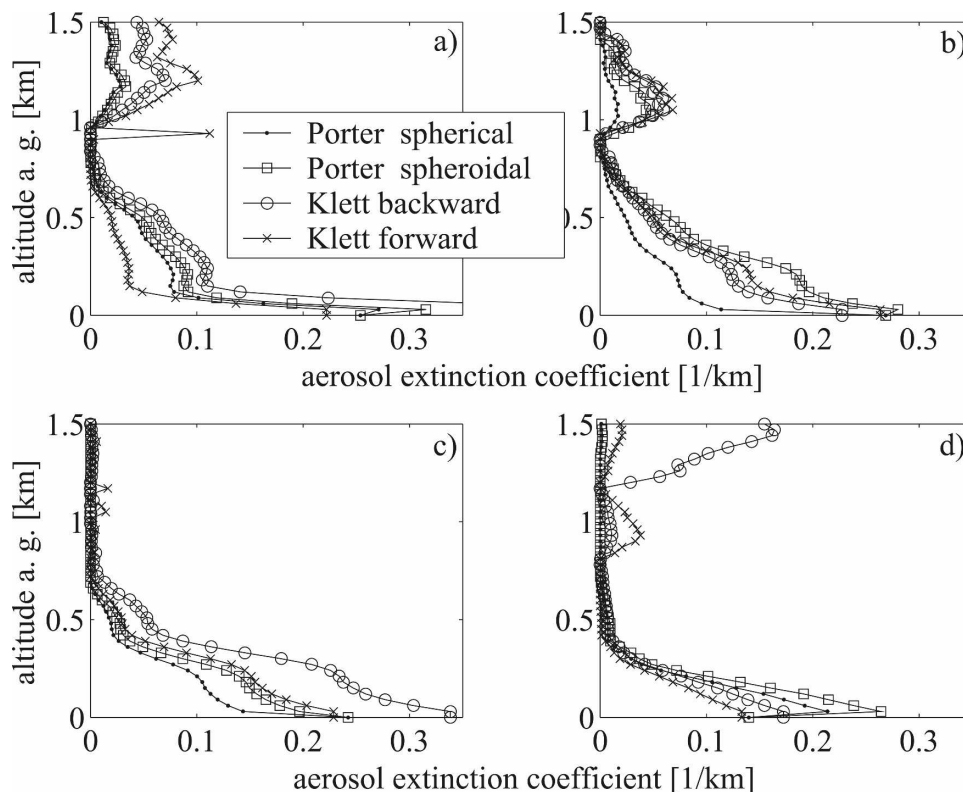


FIG. 14. Comparison of vertical profiles (UAE^2 ; 9 Sep 2004) of the aerosol extinction retrieved by Porter's algorithm for spheroidal model (closed circles) and for spherical model (open squares), modified Klett's backward algorithm with 50% of the AOT in the boundary layer (open circles), and forward Klett's algorithm (stars); (a)–(d) for different times.

because of the dust present above the boundary layer. In this case an assumption that at 1.5 km the total backscatter equals the molecular backscatter is not satisfied.

The forward Klett algorithm with a constant lidar ratio gives noticeably better results but produces large noise above 1 km.

Differences between retrieved extinction profiles are shown in Fig. 14. Open squares correspond to Porter-spheroids method, closed circles represent Porter-spherical method, open circles define the backward modified Klett method with 50% of the AOT in the boundary layer between 0–1.5 km, and stars represent the forward Klett method with constant lidar ratio and spherical particles. Figures 14a–d show the results for 0300, 0600, 1200, and 1800 UTC. It can be seen that the vertical structure depends on the method. All methods, with the exception of the backward Klett's, start with the same extinction coefficient defined close to the surface. Subsequently, the extinction coefficient is determined by the lidar ratio or the scattering phase function. Differences of the extinction determined by Porter's method for spheroidal and spherical particles are consistent with differences of the lidar ratio. The stan-

dard backward Klett method with an assumed lidar ratio turned out to be unstable for the ceilometer data. Several simulations performed using this method (not shown in this study) produced large noise and unrealistic profiles of the extinction coefficient.

Part of the AOT measured by photometric methods is above the ceilometer range of aerosol detection. Assuming a starting level at about 1.5 km, the total backscatter coefficient is not equal to the molecular backscatter coefficient. Figure 15 shows aerosol extinction profiles obtained from the modified backward Klett algorithm for several assumed fractions of the AOT. Triangles and the solid line correspond to results calculated on the basis of the Porter's spheroid and spherical models, respectively. For the AOT fraction of about 25%, the aerosol extinction profiles based on Klett's and Porter's methods are close to each other. Thus, in this case, a significant part of the aerosol is not detected by the ceilometer. For the modified Klett algorithm, having 100% of the AOT in the boundary layer leads to a large overestimation of the aerosol extinction coefficients. Based on this analysis we conclude that the modified Klett technique (Welton et al. 2002), which

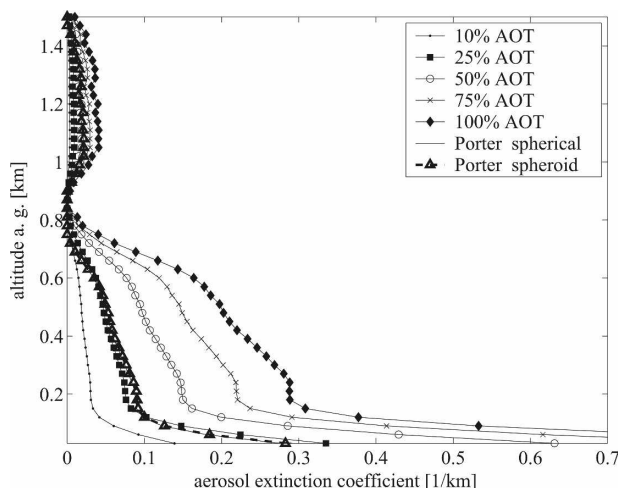


FIG. 15. Retrieved aerosol extinction coefficients from backward Klett's algorithm with different fraction of measured AOT (UAE²; 9 Sep 2004). Dotted triangles and the dotted line correspond to the forward Porter's algorithm for spheroid and spherical particles, respectively.

utilizes the total AOT, cannot be used for the CT25K extinction retrieval.

7. Summary and conclusions

Use of the CT25K ceilometer to measure aerosol extinction profile is limited to the boundary layer because of its small SNR. In the upper atmosphere, the ceilometer is able to detect only clouds or densely polluted air layers. Significant uncertainties of the aerosol measurements are due to the water vapor absorption. The CT25K laser emits its peak energy at about 905 nm where moderate water vapor absorption bands are located. The peak wavelength shifts with laser temperature changes. We show that this effect is small and can be neglected except for the case of very humid atmosphere. We also present a water vapor-corrected ceilometer signal and notice that water vapor reduces the transmission up to 20%–30% depending on the humidity. Correction for the water vapor absorption decreased uncertainties of the retrieved aerosol extinction coefficients. Precise information about the variability of the water vapor distribution in the lower atmosphere is usually unknown, which complicates the calculation of this correction. However, Fig. 16 shows that 5% and 20% uncertainty in the assumed specific humidity leads to 1%–2% and 5%–6% errors in the aerosol extinction coefficients at 1.5 km above the ceilometer.

We found that overlap correction for the CT25K needs to be applied to first 450–550 m above the surface. Our horizontal measurements performed during

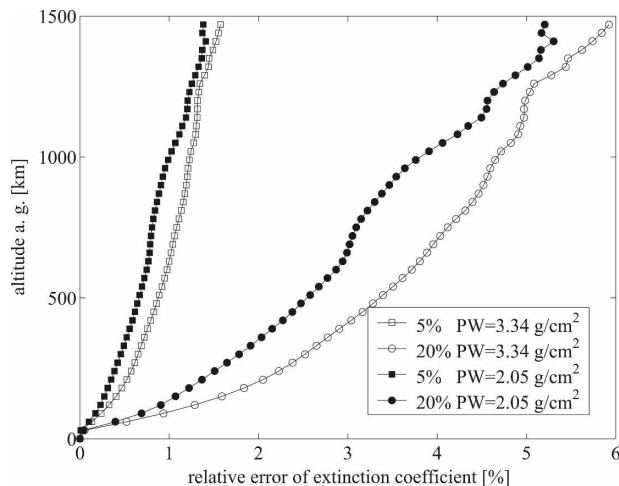


FIG. 16. Relative error of the retrieved aerosol extinction as a function of the range for different uncertainties is assumed for the specific humidity. Dotted squares and circles correspond to errors in the specific humidity of 5% and 20% for the total water vapor content of 3.34 g cm⁻². The opened squares and circles are the same but for a more drier atmosphere (the total water vapor content 2.05 g cm⁻²).

nearly horizontally uniform atmospheric conditions shows that obtained the overlap correction is consistent with manufacturer correction function.

We also present several methods to retrieve the extinction structure of the atmosphere. In all retrieval methods we assume a constant (with altitude) lidar ratio or scattering phase function. In this study we limit the extinction retrieval to the boundary layer where the lidar ratio is fairly stable with altitude.

The most promising methodologies are based on additional surface observations of extinction (absorption and scattering) with additional closure obtained by the retrieval of size distribution and refractive index and subsequent estimation of the lidar ratio. To this end, we extended Porter's method to both spherical and randomly oriented spheroidal particles. Our sensitivity studies show that uncertainties of calculated lidar ratios vary between 5% and 20%, which corresponds to 7% and 25% of the aerosol extinction uncertainties at 1.5 km from the ceilometer.

In Porter's method we have to determine the lidar calibration constant. This quantity was estimated from horizontal measurements. We also compare Klett's and Porter's forward algorithms with Klett's backward algorithm based on observed total aerosol optical depth and the assumption that there is a layer in the atmosphere that is above the aerosol layers. Because the ceilometer signal decreases significantly with height we made several calculations with varying AOT fraction. Vertical profiles of the aerosol extinction coefficient

are sensitive to the AOT fraction, and Klett's backward algorithm can be used when the aerosol layer is located mostly within the boundary layer.

The CT25K is able to detect the dust layer up to 3.5 km. We found a good agreement between the ceilometer and aerosol lidar range-corrected signals in the boundary layer. The correlation between the top of the boundary layer heights and the structure of aerosol layer measured by these instruments is very good. Reduction of the range of aerosol detection in comparison to the standard lidar system leads to complications with retrieval of the aerosol extinction profiles. Our study indicates how it is possible to use ceilometers to study the boundary layer aerosol structure and derive quantitative information about the extinction profiles.

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