Combined use of satellite and surface observations to infer the imaginary part of refractive index of Saharan dust

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[1] We present a method for the retrieval of the imaginary part of refractive index of desert dust aerosol at near UV wavelengths. The method uses observations of radiances at 331 and 360 nm by the Total Ozone Mapping Spectrometer, and aerosol optical depth measurements by the Aerosol Robotic Network. The derived values of imaginary part of refractive index of Saharan dust aerosol at 360 nm are significantly lower than previously reported values. The average retrieved values vary between 0.0054 and 0.0066 for different geographical locations. The results reported in this work are in good agreement with the results of several recent investigations. *INDEX TERMS:* 0305 Atmospheric Composition and Structure: Aerosols and particles (0345, 4801); 0669 Electromagnetics: Scattering and diffraction; 0933 Exploration Geophysics: Remote sensing. **Citation:** Sinyuk, A., O. Torres, and O. Dubovik, Combined use of satellite and surface observations to infer the imaginary part of refractive index of Saharan dust, *Geophys. Res. Lett.*, *30*(2), 1081, doi:10.1029/2002GL016189, 2003.

1. Introduction

[2] Absorption and scattering of solar radiation by dust affects the sign of aerosol radiative forcing [*Sokolik and Toon*, 1999]. The correct modeling of dust radiative forcing as well as the development of aerosol remote sensing techniques require precise knowledge of the optical properties of mineral dust in a wide spectral range.

[3] *Patterson et al.* [1977] used Saharan dust samples and laboratory measurements to produce one of the most widely used data sets of imaginary part of refractive index (k) of mineral dust in the range 300–700 nm. This data set combined with typical aerosol particle size distribution (PSD), indicates rather high dust absorption [*Sokolik and Toon*, 1999]. Recent remote sensing observations [*Kaufman et al.*, 2001; *Dubovik et al.*, 2002] conclude that mineral dust is much less absorbing in the visible part of spectrum than previously assumed. *Dubovik et al.* [2002] derived a value of k of 0.0025 at 440 nm which is significantly smaller than $k \approx 0.01$ reported by *Patterson et al.* [1977]. Thus, the inconsistency between laboratory measurements and remote sensing results for dust absorption in the visible casts uncertainty on the accuracy of k at shorter wavelengths.

[4] The issue of imaginary part of dust refractive index in the near UV has recently been addressed [*Colarco et al.*, 2002]. The derived values of k over the ocean were found to be less than those predicted by laboratory measurements. The aim of this work is to perform additional independent studies on dust absorption in the near-UV considering oceanic and continental locations and extending analysis to different times of the year.

[5] In this paper we present a retrieval method to infer the imaginary part of the refractive index of Saharan dust aerosols in the near UV. The method combines satellite observations of radiances at two near UV channels by the Total Ozone Mapping Spectrometer (TOMS), and ground based observations of aerosol optical depth by the Aerosol Robotic Network (AERONET). Since the retrieval of aerosol properties in the UV is sensitive to the location of the aerosol layer [*Torres et al.*, 1998], the retrieval method discussed here also derives the height of the aerosol layer. Retrieved values of k at 360 nm are compared to laboratory measurements and to results from other remote sensing techniques.

2. Retrieval Method

[6] The method retrieves imaginary part of refractive index of dust aerosols at 360 nm along with aerosol layer height above the ground. We used TOMS observations of the top of atmosphere radiances at 331 nm and 360 nm over Cape Verde (16°N, 22°W) and Dakar (14°N, 16°W) islands influenced by Saharan dust transported over the Atlantic Ocean and over Bidibahn (14°N, 2°W, in Burkina Faso) and Bondoukoui (11°N, 3°W, in Burkina Faso) sites located near the Saharan desert. These observations were combined with the independent collocated measurements of aerosol optical depth at 440 nm provided by AERONET [*Holben et al.*, 1998].

[7] The retrieval algorithm relies on several aerosol model assumptions. The aerosol PSD used in our approach is based on seven years of AERONET retrievals of atmospheric column size distribution of aerosols at Cape Verde [*Dubovik et al.*, 2002]. These data suggest a bi-modal lognormal PSD with parameters, which are functions of aerosol optical depth at 1020 nm. To take into account this dependence of PSD on aerosol loading, we employed three different PSD's corresponding to different values of aerosol optical depth at 440 nm. The relative spectral dependence of imaginary part of dust refractive index in the range 331–440 nm was assumed to be the same obtained by *Patterson et al.* [1977]. Table 1 summarizes the parameters of the selected PSDs. Although spherical aerosol particles were assumed, a sensitivity analysis to quantify the error due to this assumption was carried out.

[8] Our retrieval approach uses precalculated lookup tables of upwelling radiances at 331 and 360 nm for a number of aerosol laden atmosphere models of varying aerosol layer height (Z), optical depth (τ_{440}), and imaginary part of the refractive index at 360 nm (k_{360}). In producing the tables the following ranges of variability of parameters were used: $0.5 \le Z \le 5$ (km), $0.58 \le \tau_{440} \le 2.18$, $0 \le k_{360} \le 0.0158$.

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Table 1. Aerosol Models Used in Retrieval Method

Parameters of PSD for τ440 = 0.58	Fine mode	Coarse mode
σ_N (width)	1.72	1.79
R_N (modal radius), μm	0.05	0.69
N (number concentration), μm-2	15.4	0.07
Parameters of PSD for τ440 = 1.11		
σ_N (width)	1.8	1.7
R_N (modal radius), μm	0.04	0.82
N (number concentration), μm-2	26.5	0.11
Parameters of PSD for τ440 = 2.18		
σ_N (width)	1.99	1.54
R_N (modal radius), μm	0.03	1.1
N (number concentration), μm-2	70.6	0.144

[9] To model aerosol vertical distribution we used a Gaussian profile with maximum aerosol concentration at height Z and width equal 1 km. The assumption of a single dust layer over the east cost of North Africa (not too far from the Saharan dust sources) is based on the results of the Lidar In-space Technology (LITE) experiment [*Winker et al.*, 1996]. The range of variability for aerosol optical depth was selected to represent a rather large aerosol loading in order to ensure sufficient sensitivity of the measurements to the retrieved parameters. For each of the aerosol models in Table 1 the imaginary part of refractive index varies from zero (no absorption) to the value reported by *Patterson et al.* [1977], (strong absorption). To describe the real part of dust refractive index, the value 1.55 obtained by *Patterson et al.* [1977] was used.

[10] The relationship between the 331/360 nm radiance ratio and the 360 nm radiance measured by TOMS for a known aerosol optical depth constitutes the basis of the retrieval method as illustrated in Figure 1.

[11] In each case, the relationship depicted in Figure 1 was obtained for the value of aerosol optical depth measured by AERONET. The TOMS measurements were then used to determine k_{360} and Z by using the PSD associated with the AERONET measured optical depth (τ_{440}).

[12] TOMS and AERONET measurements were collocated in space by restricting the positions of TOMS pixels to be within 1° latitude by 1° longitude box centered at the AERONET site. The time of satellite overpass, in turn, was restricted to be within 30 minutes between two successive measurements. Finally, an average value was obtained using retrieval results from all the pixels within the selected area.

[13] Because of the large TOMS footprint (40 × 40 km^2), commonly used cloud masking methods based on spatial homogeneity cannot be applied. We have, instead, developed a parametric method of minimizing sub-pixel cloud contamination effects based on the Aerosol Index (AI), and the Lambert Equivalent Reflectivity of pixel (R_p). These two TOMS derived quantities are combined in the expression

$$r = \frac{R_P - R_S}{AI}, \tag{1}$$

where R_s is the actual near UV surface reflectivity, which is available from a monthly climatological database [*Herman and Celarier*, 1997]. The above expression takes advantage of the large sensitivity of the AI to the presence of mineral aerosols. Since the near-UV surface albedo is practically wavelength independent, the R_p for a pure molecular atmosphere is also spectrally independent and the resulting AI is close to zero. Clouds produce a spectrally independent increase in R_p, and, therefore, yield also near-zero AI values. The presence of aerosols produces a spectrally dependent R_p, resulting in positive AI's for absorbing aerosols, and negative AI's for small size non-absorbing aerosols [*Torres et al.*, 1998].

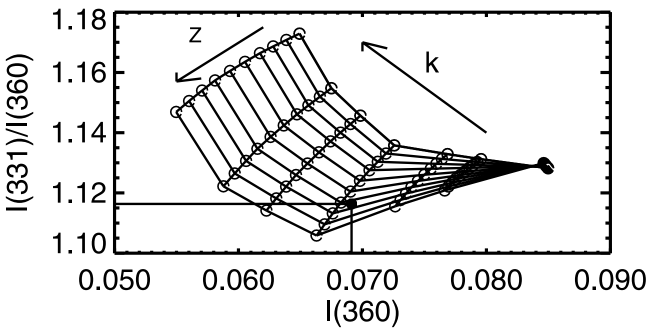

Figure 1. Ratio of the top of atmosphere radiances at 331/360 nm as function of 360 nm radiance for fixed value of aerosol optical depth. The solid circle represents the retrieval of k_{360} and Z associated with a hypothetical set of satellite measurements. The aerosol layer height and imaginary part of refractive index increase in the direction of the arrows.

[14] Figure 2 shows the relationship between the ratio r and the difference R_p-R_s, and illustrates the basic idea of using the ratio r as an identifier of cloud contaminated pixels. The observed relationship allows the separation of the data into distinct branches, each corresponding to a constant value of AI and increasing R_p. The increase in R_p while AI remains constant is, by definition, spectrally independent and, therefore, indicative of cloud presence. The points along every branch indicate different levels of sub-pixel cloud contamination with cloud amount increasing with both r and R_p. Negative values of r are produced by either non-absorbing aerosols (i.e. negative AI), or strongly absorbing aerosols that yield positive AI, and slightly negative values of the R_p-R_s difference. By imposing a restriction on the range of variability of the ratio r, the effect

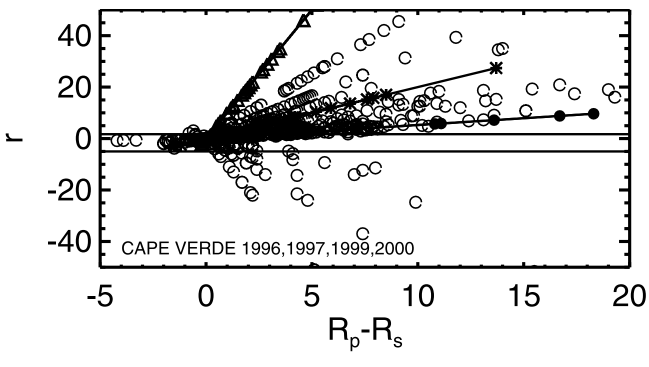

Figure 2. The ratio r as a function of the difference R_p-R_s, for TOMS observations at Cape Verde. The observed branches are associated with different values of the Aerosol Index: 0.1 (triangles), 0.5 (asterisks), 1.9 (solid circles). Open circles indicate intermediate AI values. The two lines of constant values of r, indicate the range of variability for minimum cloud contamination (see text for details).

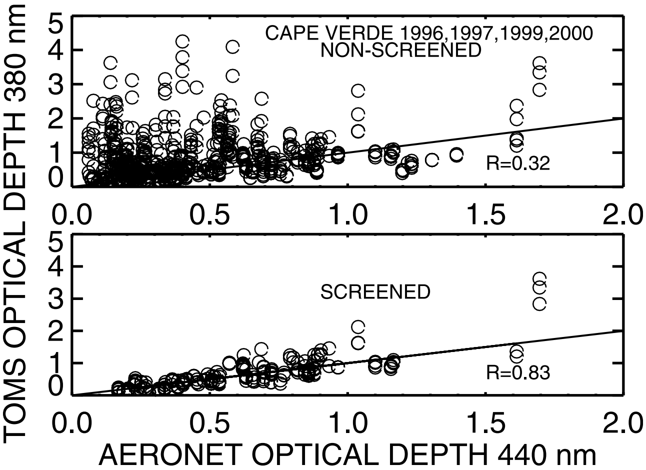

Figure 3. Comparison between TOMS optical depth retrievals and AERONET measurements. Solid lines represent the one-to-one relationship between two optical depths. R is the correlation coefficient between two data sets.

of sub-pixel cloud contamination is reduced. The allowed range of variability of the parameter r, was chosen based on the comparison of aerosol optical depth derived from TOMS observations [*Torres et al.*, 2002] to AERONET optical depth measurements under cloud-free conditions [*Smirnov et al.*, 2000]. The range $-5 < r < 1.7$ for positive values of AI was determined by selecting the conditions that produced the best comparison of the two datasets as illustrated in Figure 3.

3. Sensitivity Analysis

[15] The main error sources in retrieval results are the uncertainties in radiometric calibration, and assumptions used in the forward modeling. The radiometric calibration of the TOMS sensor at 331 and 360 nm is known to be accurate within 1% [*Jaross et al.*, 1998], which yields errors of less than 5% in the derived refractive index. Forward modeling was done with a Gauss-Seidel vector radiative transfer code [*Herman and Browning*, 1965], that accounts accurately for gas absorption, molecular and aerosol multiple scattering, and polarization effects. Small corrections were applied for ozone absorption at 331 nm, and for Raman scattering effects at both wavelengths. Retrieval errors associated with the assumptions of spherical particles and the width of the dust layer were examined. In addition, an analysis on the variability of optical depth within the area of collocation was also carried out.

Table 2. Average Retrieval Result From This Analysis Compared to Other Methods

SITE	TOMS + AERONET (This work)	AERONET	TRANSPORT MODEL + TOMS
Cape Verde, Dakar	0.0054 (C, D) (0.004–0.0065)	0.0046 (C) (0.0036–0.0056)	0.004 (C) 0.005 (D)
Bidibahn, Bondoukoui	0.0066 (0.0038–0.01)	N. A	N. A.

The AERONET value was reduced to 360 nm by linearly extrapolating the 440 and 670 nm values. Numbers in parentheses represent the range of variability.

[16] To evaluate the sensitivity to the assumption of spherical particles, we have used a mixture of randomly oriented polydisperse spheroids, which provides a more realistic modeling of optical properties of non-spherical desert dust particles [*Mishchenko et al.*, 1997].

[17] Synthetic data were simulated assuming the particles to be spheroids with an aspect ratio (the ratio of axes) equal to 1.8 and k_{360} = 0.006, for the actual satellite viewing geometry over Cape Verde. A similar aerosol model for non-spherical particles was used in sensitivity studies by *Dubovik et al.* [2000]. The T-matrix code [*Mishchenko and Travis*, 1994] was used to calculate light scattering by randomly oriented spheroids with the same volume size distribution as for spherical particles. The results of our numerical tests show that the retrieval error in k is minimum in the vicinity of the scattering angle 150° and could be as large as 60% at 180°. Within the range $140° \leq \vartheta_{sca} \leq 160°$ the resulting maximum error in k_{360} is about 20%, while a maximum 10% error in aerosol layer height is obtained. Retrieval results outside the above range of scattering angles were rejected.

[18] We also evaluated the uncertainty of the retrieval due to the spatial variability of the aerosol optical depth within the area of collocation. To do this, we examined the change in

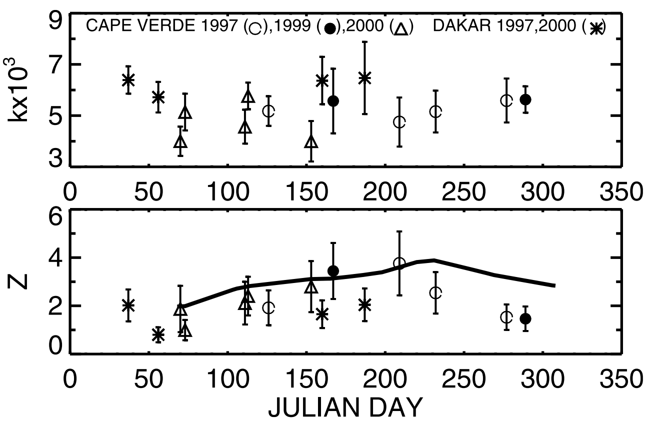

Figure 4. Retrieval results for imaginary part of refractive index and aerosol layer height for Cape Verde and Dakar.

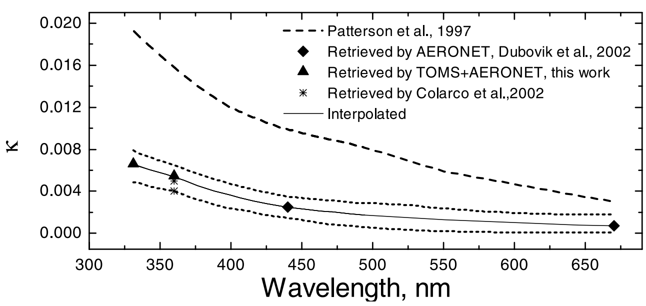

Figure 5. The solid line represents the spectral dependence of k of Saharan dust obtained by means of linear interpolation using 331, 360, 440, and 670 nm values. The upper and lower limits (dotted lines) represent the observed range of variability. The dashed line shows the *Patterson et al.* [1977] data.

optical depth at the AERONET sites, during a 5-hr interval, which is the time required for an air parcel to travel within the collocation area at a typical wind speed of 5 m-sec^{-1}. We found that on most days, the spatial variability was less than 5% with only two days exceeding 10%. The sensitivity analysis of the assumption on the width of the dust layer shows that a 1 km uncertainty in the width produce errors of about 5% in the retrieved imaginary part of refractive index and about 20% in the derived aerosol layer height.

4. Retrieval Results

[19] We applied the described retrieval method to TOMS measurements over four selected AERONET sites on several days in 1997, 1999 and 2000 (see Figure 4). The error bars were obtained as the root mean square of the retrieval errors due to spatial variability, calibration, dust particles non-sphericity, and uncertainty in the width of aerosol vertical distribution. Maximum errors of about 24% in refractive index and 51% in aerosol layer height were obtained.

[20] Table 2 presents a comparison of the retrieval results in this analysis with those of AERONET for Cape Verde [*Dubovik et al.*, 2002] and the results of *Colarco et al.* [2002] for Cape Verde and Dakar. The results of all the three approaches are in good agreement with each other.

[21] The values of k_{360} reported here are about one third of the laboratory measurements of *Patterson et al.* [1977]. Since the size of Saharan dust aerosols is known to be insensitive to relative humidity effects [*Li-Jones et al.*, 1998], the discrepancy between the results of the remote sensing analysis and those of the laboratory measurements cannot be explained in terms of differences in optical properties associated with the hypothetical water content of the ambient aerosol.

[22] The retrieved values of aerosol layer height are consistent with the predictions of dust transport model [*Ginoux et al.*, 2001] for Cape Verde and Dakar. The solid line at the Figure 4 represents the climatological values of aerosol layer height for Cape Verde produced by using this model.

[23] Using the obtained average values of k_{360} and k_{331} for Cape Verde and Dakar, and the values of k_{440} and k_{670} derived from AERONET we have inferred the values of k in the range 331–670 nm by means of linear interpolation in logarithmic scale as shown in Figure 5.

5. Discussion and Conclusions

[24] We have presented the method for simultaneous retrieval of imaginary part of refractive index of Saharan dust and aerosol layer height above the ground. The approach uses TOMS measurements of top of atmosphere radiances at 331 nm and 360 nm and independent information on aerosol optical depth at 440 nm provided by AERONET. The method was applied to the combination of TOMS and AERONET measurements over four AERONET sites, which are dust-dominated locations. The retrieved values of imaginary part of refractive index of dust aerosols are found to be lower than the value reported by *Patterson et al.* [1977].

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