Aerosol optical depths and sky brightness in the IR spectral region in arid zones of the globe

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The role of different factors, influencing the determination of aerosol scattering optical depth from observations of the sky brightness in the near-IR spectral region, is analyzed. For this purpose, we use the measurements of the optical depth and sky brightness at three sites; these sites have an arid aerosol and are presented in AERONET tables. The application of the so-called difference method allowed us to exclude the factor of surface reflection from this solution technique. The effect of variations of solar zenith angle and forward peak of the aerosol scattering phase function on the retrieval of the aerosol optical depth is estimated.

Because of a widespread damage of vegetation due to vigorous anthropogenic activity and growing number of forest fires the arid aerosol is becoming more and more typical for many regions of the globe. In this regard, the radiation studies of arid aerosol have become more and more urgent, especially against the background of currently observed global climate change. In the present paper, the experimental AERONET data[1] are used to study the relationship between aerosol optical depth τ_a and sky brightness $B(\Psi)$ observed in solar almucantar in the near-IR spectral range. Here, Ψ is the azimuth angle counted off the plane of the solar vertical (angle $\Psi = 0°$ corresponds to the direction toward the Sun). The data of τ_a and $B(\Psi)$ observations in two spectral regions at the wavelengths $\lambda_1 = 0.87$ μm and $\lambda_2 = 1.02$ μm are obtained using CIMEL sun photometers, employed by NASA for ground-based optical monitoring of atmospheric aerosol. We analyze τ_a and $B(\Psi)$ measurements at three sites: in Mongolia (Dalangazad), USA (New Mexico), and Arabian Peninsula (Solar Village).

For the visible spectral range, we have developed two methods of determination of aerosol scattering optical depth from sky brightness observations, the difference[2] and an integral one.[3] The methods are quite simple and can be used for analysis of vast observational material. However, in practical application of the second method it is necessary to specify surface albedo q; whose experimental values are unavailable in AERONET tables. Figure 1 presents the results measured in the near-IR on summertime surface albedo in different arid zones of the Earth.[4–6] As seen, the absolute values of surface albedo (1) can be quite large (up to 0.5) and (2) they strongly vary as a function of surface type. The same can also be said about directional reflection coefficients of light-colored clays, sands, and saline soils in semiarid and arid zones of Western Kazakhstan.[7] Therefore, in the absence of experimental data on q for the region where observation site is located, it is reasonable to use the difference method of determination of optical scattering depths from sky brightness, i.e., the method in which the role of albedo is automatically cancelled.[2] The present paper discusses the role of other factors, determining the relationship between the brightness and spectral atmospheric transmission in the IR spectral region.

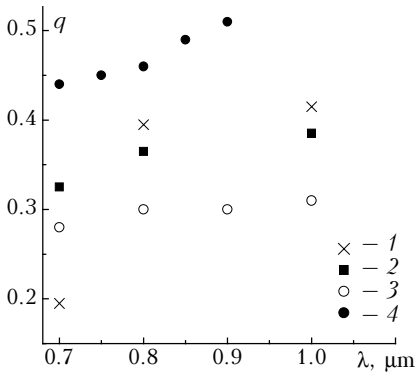

Fig. 1. Spectral behavior of albedo: motley grass steppes (curve *1*, Ref. 4); semiarid regions and northern deserts (curve *2*, Ref. 4); Gobi sand (curve *3*, Ref. 5); and white river sand (curve *4*, Ref. 6).

Initially, the azimuth distributions $B(\Psi)$ were recalculated to absolute scattering phase functions of brightness according to the following formula[4]:

$$f(\varphi) = \frac{B(\varphi)}{E_0 e^{-\tau m} m}, \qquad (1)$$

where E_0 is the spectral solar constant; m is the atmospheric mass along the direction toward the Sun; and τ is the total optical depth which includes molecular τ_m and aerosol τ_a scattering and absorption components. The scattering angle φ is related to azimuth Ψ via known formula

$$\cos\varphi = \cos^2 Z_0 + \sin^2 Z_0 \cos\Psi, \qquad (2)$$

where Z_0 is the solar zenith angle.

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The following remark is in order. The aerosol optical depths τ_a, presented in AERONET tables, are defined as the differences between total depths τ, determined according to Bouguer's method, and molecular scattering components $\tau_{m.s}$, and so they automatically include molecular $\tau_{m.a}$ and aerosol $\tau_{a.a}$ absorption components. These depths, especially $\tau_{m.a}$, seem to be low in the chosen spectral regions,[8] though this requires careful consideration.[9,10] At the same time, summing of τ_a to $\tau_{m.a}$ to retrieve the integrated extinction depth τ is a procedure required for further calculations of $f(\varphi)$ according to formula (1). We have calculated the values of $\tau_{m.s}$ taking into account the atmospheric pressure at each observation site.

In calculating $f(\varphi)$, we used E_0 averaged over the literature data[11, 12] taking into account the Earth–Sun distance. Also, we averaged the brightness at points symmetrical about the plane of solar vertical, with angles Ψ and $360° - \Psi$ for each angular distribution. Further processing was imposed only on those distributions for which $B(\Psi)$ and $B(360° - \Psi)$ differed, respectively, at $\Psi \geq 10°$ and $(360° - \Psi) \leq 350°$ by no more than 10%, and provided that the angular interval $\delta\Psi$ of systematic deviations did not exceed 30°.

In the region near solar aureole, i.e., for $\Psi < 10°$, use of $B(\Psi)$ and $B(360° - \Psi)$ had less stringent requirements: deviations could reach a factor of two. Most probably, this is not because real aerosol inhomogeneities existed in the atmosphere near the direction toward the Sun, but rather because the photometers were imprecisely pointing in directions to the left and to the right of the center of solar limb by symmetrical small angular distances, such as $\Psi = 2$ and 358° (in this case, the Ψ and φ angles practically coincide). Since the further analysis of observation data uses the integrals

$$\Delta_1 = 2\pi \int_0^{\pi/2} f(\varphi)\sin\varphi \, d\varphi \qquad (3)$$

and

$$\Delta_2 = 2\pi \int_{\pi/2}^{\pi} f(\varphi)\sin\varphi \, d\varphi, \qquad (4)$$

then, because of smallness of the angular interval $0 \leq \delta\varphi \leq 10°$, and especially because of weighting function $\sin\varphi$ involved in the integration, the errors in Δ_1 determination do not exceed 2%,[13] provided, of course, that the Δ_1 calculations use the mean values $[f(\varphi) + f(360° - \varphi)]/2$ for small scattering angles.

As argued in Ref. 14, in the visible spectral range the difference of integrals

$$\Delta = \Delta_1 - \Delta_2 \qquad (5)$$

is practically uniquely related to the aerosol scattering optical depth $\tau_{a.s}$. Smerkalov[4] has further suggested that such a relationship may also exist in the IR spectral range. Assuming the absence of aerosol absorption, i.e., letting $\tau_a = \tau_{a.s}$, he arrived at the following formula:

$$\tau_a = -0.02 + \sqrt{0.04 + \Delta/2}. \qquad (6)$$

Derivation of this formula is based on numerical solution of equation of radiative transfer in the atmosphere. According to Ref. 4, it must bear a universal character over a wide wavelength range from UV to IR region of the spectrum. In this paper, we study this based on the data of observations of atmospheric transmission and sky brightness at the above-mentioned sites with a predominantly arid aerosol.

Figure 2 compares aerosol optical depths τ_a, determined by Bouguer's method, with those, obtained using formula (6) and denoted as τ_a^*. The latter values are divided into two groups, corresponding to air mass intervals $2 \leq m \leq 3$ and $3 \leq m \leq 4.3$. Note that for $m > 4.3$ no observation data are presented in AERONET tables at all. The straight line in Fig. 2 is oriented at an angle of 45° with respect to the x-axis. From the figure it follows that there is a systematic discrepancy between compared aerosol optical depths, $\tau_a > \tau_a^*$, and this disparity increases with the decreasing m. Let us now discuss possible causes of this.

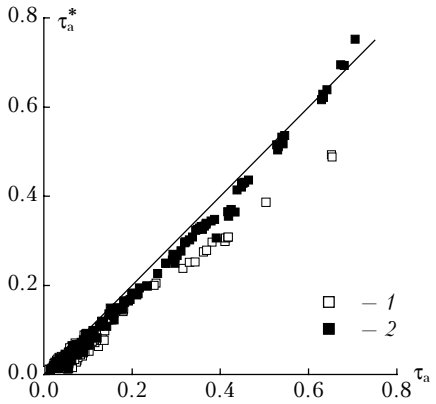

Fig. 2. Calculated τ_a^* versus τ_a: m = 2–3 (*1*); and m = 3–4.3 (*2*).

First and, probably, the main cause is as follows. Formula (6) has been derived for solar zenith angles $Z_0 = 76.9$ and 80°, corresponding to atmospheric masses $m > 4.4$, for which sky brightness measurements by CIMEL photometers were not performed. According to our previous results,[2,15] the difference of integrals Δ in the visible spectral range, undoubtedly, depends on Z_0. Most probably, this dependence is also true for the IR range. Therefore, a rigorous analysis of observation data must be based on results of solution of radiative transfer equation for those solar zenith angles for which sky brightness measurements have been performed.

The second cause of the discrepancy between τ_a and τ_a^* may be the neglect (in the final analysis) of the effect of absorption by air molecules and aerosol. As was shown earlier, the absolute scattering phase function of brightness $f(\varphi)$ weakly depends on the atmospheric absorptivity.[15] Therefore, τ_a^* determined via integrals (3) and (4) with the help of formula (6) will, actually, be the aerosol optical depth due to scattering.

Since the depth τ_a in the AERONET tables appears as the difference between τ and $\tau_{m.s}$, the $\tau_{m.a}$ and $\tau_{a.a}$ components are already included in it. If these latter are significant, the inequality $\tau_a > \tau_a^*$ naturally arises.

Finally, the third cause of the discrepancy between τ_a and τ_a^* may be the difference between model aerosol scattering phase function, which was used in calculations of the intensity of scattered radiation, from real scattering phase functions, which took place in observations of sky brightness in arid zones. The simplest hypothesis testing may be as follows. Since the problem of determination of scattering optical depth is, in fact, related to determination of the fluxes scattered into the forward and backward hemispheres,[2,4] let us compare their observed asymmetry coefficients

$$\Gamma = \frac{\int_0^{\pi/2} f(\varphi)\sin\varphi\, d\varphi}{\int_{\pi/2}^{\pi} f(\varphi)\sin\varphi\, d\varphi} \qquad (7)$$

with the value obtained from model calculations. Figure 3 shows Γ as a function of τ_a according to data of observations around $\lambda = 1.02$ μm. The presence of a maximum in the given family of the points undoubtedly indicates that multiple scattering affects the sky brightness and, hence, the value of Γ. Also included in the array of observation data is a model-calculated point Γ (Ref. 4). It well fits to the general ensemble. A totally analogous pattern also takes place for $\lambda = 0.87$ μm (not shown). This indicates that the shape of model-derived aerosol scattering phase function[4] is in a satisfactory correspondence with the observations, at least within the framework of the problem solved here. Hence, the differences between τ_a and τ_a^*, caused by natural variations of asymmetry of aerosol scattering phase function, most probably, will be random rather than systematic.

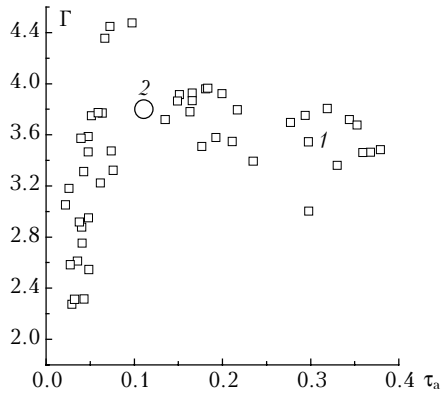

Fig. 3. Plot of Γ versus τ_a: observation data (*1*) and calculated data from Ref. 4 (*2*).

Let us derive empirical formulas relating τ_a^* and Δ in IR spectral range for above-indicated intervals of atmospheric masses. For the first group, when $2 \le m \le 3$, the formula has the form

$$\tau_a^* = -0.42\Delta^2 + 1.1\Delta + 0.01, \qquad (8)$$

while for the second group, i.e., when $3 \le m \le 4.3$, it is:

$$\tau_a^* = -0.14\Delta^4 + 0.61\Delta^3 - 1.03\Delta^2 + 1.1\Delta + 0.01. \qquad (9)$$

The goodness of this polynomial representation of relationship between aerosol optical depth and sky brightness is seen from Fig. 4; it compares the results of determination of aerosol optical depths, found according to formulas (8) and (9) from sky brightness observations and denoted as τ_a^*, with data of measurements of aerosol depth according to Bouguer's method τ_a for both wavelengths in two atmospheric mass intervals simultaneously.

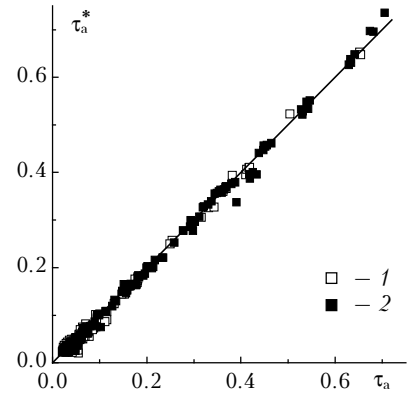

Fig. 4. Plot of τ_a^* versus τ_a: $m = 2$–3 (*1*); and $m = 3$–4.3 (*2*).

It is well known that the relative error in τ_a measurements grows with decreasing atmospheric turbidity. Therefore, the root-mean-square deviations of τ_a^* from τ_a were calculated separately for the cases of high ($\tau_a < 0.1$) and usual ($\tau_a \ge 0.1$) atmospheric transmission, and then for entire interval of τ_a variations. They are presented in the Table.

Root-mean-square deviations of τ_a^* from τ_a, %

λ, μm	$\tau_a < 0.1$	$\tau_a \ge 0.1$	All τ_a
0.87	14	6	11
1.02	21	5	15

From results presented above we can conclude the following. Observations of sky brightness (absolute scattering phase functions $f(\varphi)$) in arid regions of the globe for atmospheric masses $4.3 \ge m \ge 2$ make it possible to determine with formulas (8) and (9) the aerosol optical depths in the IR spectral range for different levels of atmospheric turbidity without parallel measurements of the surface albedo. For different levels of atmospheric turbidity, independent of the shape of aerosol scattering phase function, the aerosol optical depths agree with τ_a determined by Bouguer's method to within the root-mean-square error ranging from 5 to 21%. This last accuracy is evidently insufficient for solution of basic problem of determination of aerosol absorption depth $\tau_{a.a}$ because of the smallness of this component.[16] Therefore, in derivation of final formulas it is necessary to take

into account the effect of the scattering phase function on fluxes scattered into the forward and backward hemispheres. This can be done using radiative transfer theory for different aerosol models and is planned to be performed in the nearest future.

References

1. http://aeronet.gsfc.nasa.gov.
2. T.B. Zhuravleva, V.E. Pavlov, and V.V. Pashnev, Atmos. Oceanic Opt. **16**, No. 4, 347–351 (2003).
3. T.B. Zhuravleva, A.S. Shestukhin, V.E. Pavlov, and V.V. Pashnev, Atmos. Oceanic Opt. **16**, Nos. 5–6, 417–423 (2003).
4. V.V. Smerkalov, *Applied Atmospheric Optics* (Gidrometeoizdat, St. Petersburg, 1997), 334 pp.
5. http://www.aeoliandust.com/jp/project/d6/d6aoki.html.
6. V.K. Kagan and K.Ya. Kondratyev, *Fundamentals of Information Theory of Vision in the Atmosphere* (Gidrometeoizdat, Leningrad, 1968), 168 pp.
7. Zh.D. Zhalmukhamedova and V.E. Pavlov, Issled. Zemli iz Kosmosa, No. 3, 22–25 (2003).
8. L.A. Egorova and V.E. Pavlov, Atm. Opt. **2**, No. 1, 19–22 (1989).
9. D.M. Kabanov, "*Experimental studies of atmospheric aerosol optical depth over ocean and continent*," Cand. Phys.-Math. Sci. Dissert., Tomsk (2002), 209 pp.
10. S.M. Sakerin, *"Passive sensing of optical characteristics of the atmosphere over ocean and continent,"* Doct. Phys.-Math. Sci. Dissert., Tomsk (1998), 376 pp.
11. D. Neckel and H. Labs, Solar Phys. **90**, No. 2, 205–258 (1984).
12. E.A. Makarova, A.V. Kharitonov, and T.B. Kazachevskaya, *Solar Radiation Flux* (Nauka, Moscow, 1991), 400 pp.
13. T.Z. Muldashev, V.E. Pavlov, and Ya.A. Teifel, Izv. Akad. Nauk SSSR, Fiz. Atmos. Okeana **27**, No. 8, 831–841 (1991).
14. G.Sh. Livshits and V.E. Pavlov, in: *Atmospheric Optics* (Nauka, Moscow, 1968), pp. 59–62.
15. V.E. Pavlov, T.B. Zhuravleva, V.V. Pashnev, and A.S. Shestukhin, Vestn. KazNU, No. 4(32), 34–41 (2002).
16. O. Dubovik, B. Holben, T. Eck, A. Smirnov, Y. Kaufman, M. King, D. Tanre, and I. Slutsker, J. Atmos. Sci. **59**, 590–608 (2002).