Empirical relationship between
aerosol scattering phase function and optical thickness
of atmosphere above the ocean

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ABSTRACT

This paper presents a statistical analysis of aerosol phase functions measured under marine conditions with the total atmospheric optical thickness. The result of the analysis is an empirical equation that expresses the aerosol phase function of scattering in the visible region through two tabulated empirical functions and a total aerosol optical thickness at 745 *nm*. The extrapolation of the atmospheric optical thickness to the visible spectrum is accomplished *via* an Ångsrom-type empirical equation that includes the wavelength of light and the aerosol optical thickness at 745 *nm*.

Keywords: marine aerosols, scattering, atmospheric optics, optical properties

1. INTRODUCTION

It is known that about 90% of upwelling light that reaches the upper level of clear atmosphere above the ocean is due to the scattering in the air and reflection from the water surface.[1-12] Such a large share of scattered light implies that the detailed knowledge of such optical atmospheric parameters, as an atmospheric optical thickness and a scattering phase function, is very important for optical algorithms of atmospheric correction. Variability of atmospheric optical parameters is largely due to the variations in content and properties of atmospheric aerosol. Experimental measurements of the aerosol optical thickness of atmosphere $\tau_A(\lambda)$ show that $\tau_0 \equiv \tau_A(\lambda_0)$ at some fixed wavelength λ_0 can vary several times on the distances of the order of several hundred kilometers.[13, 14] Though the value of τ_0 is less than the corresponding value above the earth surface, it is still higher than the value of Rayleigh optical thickness $\tau_R(\lambda_0)$. Consequently, any algorithm of atmospheric correction requires knowledge of the total aerosol optical thickness $\tau_A(\lambda)$ and of the aerosol scattering phase function $p_A(\theta)$ related to the whole atmospheric layer.

The algorithms for atmospheric correction of optical signatures reflected from the ocean [4–11] either include too many input parameters or perform inadequately under marine conditions. Some of those algorithms incorporate extensive Mie, LOWTRAN and MIDTRAN calculations. These calculations are unsuitable for the real-time processing of remotely measured optical signatures

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of the ocean-atmosphere system. The optical radiation reflected from the sea and modified by the atmospheric scattering can be expressed analytically through the optical thickness of the atmosphere, the atmospheric scattering phase function and the properties of the sea and sea-air interface. The knowledge of an empirical relationship between the atmospheric optical thickness and the scattering phase function together with some algorithm to derive these properties from optical measurements can greatly simplify the atmospheric correction procedure.

Methods of restoring $p_A(\theta)$ and $\tau_A(\lambda)$ from optical satellite measurements are either based on processing radiance coefficients measured at several viewing angles or depend upon utilizing statistical relationships between $p_A(\theta)$ and $\tau_A(\lambda)$ and τ_0 in visible part of the spectrum. These relationships may be represented as follows: [1]

$$p_A(\theta) = \langle p_A(\theta) \rangle + \left[B_\theta(p_A, \tau_0)/B(\tau_0, \tau_0) \right] \left(\tau_0 - \langle \tau_0 \rangle \right), \tag{1}$$

$$\tau_A(\lambda) = \langle \tau_A(\lambda) \rangle + \left[B_\lambda(\tau_A, \tau_0)/B(\tau_0, \tau_0) \right] \left(\tau_0 - \langle \tau_0 \rangle \right), \tag{2}$$

here the angular braces denote averaging over statistical ensemble of embraced parameters, and values B_θ, B_λ, and B denote cross- and auto-correlation matrices. Currently we do not have enough experimental data measured above the ocean to recreate relationships of types (1) and (2). Instead, we use currently available data to find relationships between the aerosol optical thickness in visible region and the aerosol scattering phase function with the aerosol optical thickness τ_0 of the atmosphere in the infrared part of the spectrum.

2. DEPENDENCE OF AEROSOL SCATTERING PHASE FUNCTION ON AEROSOL OPTICAL THICKNESS OF ATMOSPHERE IN NEAR INFRARED PART OF SPECTRUM

As a basis for the functional representation of an atmospheric aerosol scattering phase function above the ocean we take aerosol phase functions of Ref. [15] measured above the Atlantic ocean. The statistical analysis of these phase functions shows that all values $p_A(\theta_n)$ at all tabulated scattering angles $(\theta_n = 0, 10°, ..., 180°)$ are well correlated with the first Legendre polynomial decomposition coefficient x_1 of the phase function $p_A(\theta)$:

$$p_A(\theta) = A(\theta) + x_1 D(\theta), \quad x_1 = (3/2)\int_0^\pi p_A(\theta) \cos\theta \sin\theta \, d\theta. \tag{3}$$

The values of two empirical functions $A(\theta)$ and $D(\theta)$ are given in Table 1.

The phase function $p_A(\theta)$ is normalized according to:

$$0.5\int_0^\pi p_A(\theta)\sin\theta \, d\theta = 1, \tag{4}$$

and functions $A(\theta)$ and $D(\theta)$ satisfy the following conditions:

$$\int_0^\pi A(\theta)\sin\theta\, d\theta = 2, \quad \int_0^\pi D(\theta)\sin\theta\, d\theta = 0. \tag{5}$$

By integrating Eq. (3) over the solid angle of the backward hemisphere we obtain the following relationship between a backscattering probability on aerosol particles and parameter x_1:

$$B_A = 0.5\int_{\pi/2}^\pi p_A(\theta)\sin\theta\, d\theta \cong 0.5 - 0.2\, x_1. \tag{6}$$

Equations (3)-(6) describe aerosol scattering phase functions measured at the sea level. Additionally, according to the Ref. [16], the atmospheric optical parameters of the vertical atmospheric profile linearly correlate with the optical parameters on the sea level, which means that we can apply relationships (3)-(6) to the whole atmospheric thickness.

Comparing Eq. (3) with Eq. (1) we have:

$$x_1 = k_0 + k_1\tau_0. \tag{7}$$

In Ref. [17] it is shown that when $\tau_0 \to 0$ the phase function $p_A(\theta)$ transforms into the phase function similar to the Rayleigh one, *i. e.* $B_A \to 0.5$. It means that $k_0 \to 0$ when $x_1 \to 0$. Consequently,

$$p_A(\theta) = A(\theta) + k_1\tau_0\, D(\theta) , \tag{8}$$

$$B_A = 0.5 - 0.2\, k_1\, \tau_0. \tag{9}$$

The coefficient k_1 in Eqs. (8) and (9) is unknown on this stage. Using experimental data published in Ref. [17], we can establish connection between aerosol backscattering probability B_A and the horizontal visibility range S (in m):

$$B_A = 0.23\, S^{0.55}, \quad r^2 = 0.77, \tag{10}$$

here and elsewhere in this paper r^2 denotes a correlation coefficient.

<u>Table 1.</u> Empirical functions $A(\theta)$ and $D(\theta)$

θ, degrees	A	D	r^2	θ, degrees	A	D	r^2
0	2.07	5.59	0.992	100	0.74	-0.30	0.999
10	1.95	4.53	0.989	110	0.77	-0.32	0.998
20	1.45	2.66	0.991	120	0.80	-0.33	0.998
30	1.30	1.29	0.991	130	0.94	-0.40	0.998
40	1.24	0.45	0.987	140	1.11	-0.45	0.998
50	1.05	0.08	0.715	150	1.37	-0.55	0.997
60	0.95	-0.13	0.929	160	1.71	-0.65	0.995
70	0.80	-0.19	0.971	170	2.34	-0.88	0.992
80	0.74	-0.24	0.994	180	2.38	-0.85	0.994
90	0.67	-0.24	0.990				

Another study, [18] based on more than fifteen hundred measurements, proposes a correlation relationship between $\tau(550)$ and S. Taking into account that $\tau(550) = \tau_R(550) + \tau_A(550)$ $\cong 0.1 + \tau_A(550)$, where τ_R is a Rayleigh optical thickness, and accepting, according to the Ref. [4], the relationship $\tau_A(550) \cong \tau_A(740)(740/550)^{0.93}$, we have the following empirical formula:

$$S \cong 400(0.8 - 3\tau_0). \tag{11}$$

Eliminating S from Eq. (11) with the use of Eq. (10), we have the following equation that connects backscattering probability on aerosol particles B_A with the aerosol optical thickness of atmosphere:

$$B_A = 0.23\left[400(0.8 - 3\tau_0)\right]^{0.55} \cong 0.5(1 - 2\tau_0). \tag{12}$$

Comparing Eq. (12) with Eq. (9) we have: $k_1 \cong 5$.

Let us estimate variations of k_1 for the atmosphere above the ocean. For the typical continental atmosphere, $B_A \cong 0.13$ at $\tau_0 = \tau_A(740) = 0.2$.[4] Using Eq. (9) we have $k_1 \cong 9.0$. Since the clear atmosphere above the ocean typically is more transparent than the atmosphere above land masses and since marine aerosol scattering phase function is less elongated than the continental one, the value of 9.0 may be considered as the estimate of k_1 from above, *i.e.*,

$$k_1 \leq 9. \tag{13}$$

Let us find an estimate for k_1 from below. G. Livshitz proposes the following relationship between the infrared aerosol optical thickness τ_0 and integral probabilities of forward F_μ and backward B_μ scattering: [2]

$$\tau_0 = C\left(F_\mu - B_\mu\right) \equiv C\left(1 - 2B_\mu\right), \tag{14}$$

here F_μ and B_μ are computed using the radiance indicatrix $\mu(\theta)$ defined by Ref. [19]:

$$F_\mu = 0.5 \int_0^{\pi/2} \mu(\theta)\sin\theta\, d\theta, \quad B_\mu = 0.5 \int_{\pi/2}^{\pi} \mu(\theta)\sin\theta\, d\theta, \tag{15}$$

$$0.5 \int_0^{\pi} \mu(\theta)\sin\theta\, d\theta = F_\mu + B_\mu = 1. \tag{16}$$

The analysis of the large data sets, obtained in different locations and during the various seasons of a year, [20] shows that in the near infrared part of the spectrum (740-750 *nm*) Eq. (14) is valid with the values of $C = 1.26$ and $r^2 \geq 0.95$.

Radiance indicatrix $\mu(\theta)$, defined by Ref. [19], is connected to the single-scattering aerosol

phase function $p_A(\theta)$ and the Rayleigh scattering phase function on air molecules $p_R(\theta)$ by the following equation:

$$\mu(\theta) = (1 - f_2)\left[r\, p_R(\theta) + (1 - r)\, p_A(\theta)\right] + f_2\, p_2(\theta) \qquad (17)$$

here r is a share of Rayleigh scattering, $p_R(\theta)$ is a Rayleigh scattering phase functin, f_2 is a share of multiple scattering contribution, and $p_2(\theta)$ is an addition counting scattering orders higher than the second one. Since the phase function $p_2(\theta)$ in Eq. (17) is very close to the spherical one, we have:

$$\left.\begin{array}{l} F_\mu = (1 - f_2)\left[r/2 + (1 - r)(1 - B)\right] + f_2/2, \\ B_\mu = (1 - f_2)\left[r/2 + (1 - r)B\right] + f_2/2. \end{array}\right\} \qquad (18)$$

By using Eq. (18) with Eq. (14) and taking into account that $C = 1.26$, we have

$$\tau_0 = 1.26(1 - f_2)(1 - r)(1 - 2B_A), \ \text{ or } \ \tau_0 < 1.26(1 - 2B_A). \qquad (19)$$

Comparing Eqs. (19) and (13), we have the following range of variability for the coefficient k_1:

$$2 \le k_1 \le 9. \qquad (20)$$

In the next paragraph, by using radiance coefficients of the ocean-atmosphere system with the Eq. (8), we make an estimate for the spectral dependence of aerosol optical thickness $\tau_A(\lambda)$.

3. ESTIMATION OF AEROSOL OPTICAL THICKNESS USING SATELLITE-MEASURED RADIANCE OF THE OCEAN-ATMOSPHE SYSTEM

Let us adopt an atmospheric optical model of the studies. [4-7, 12] Measured at nadir radiance reflection coefficient of the ocean-atmosphere system is represented as

$$\rho = (\rho_R + \rho_A + \rho_S)T_{oz} \qquad (21)$$

here $T_{oz} = \exp\left[-\tau_{oz}(\lambda)(1 + m)\right]$ is a total transmittance by the absorbing ozone layer, $m = 1/\cos z_s$ is the atmospheric optical mass, z_s is the solar zenith angle. The first term in the right of Eq. (21) is the Rayleigh part of the radiance coefficient. It is represented as

$$\rho_R(\lambda) = \frac{m}{4}\left[1 + F(z_s)\right]p_R(z_s)\tau_R(\lambda), \qquad (22)$$

where p_R is the Rayleigh scattering phase function, and τ_R is the Rayleigh total optical thickness of atmosphere. They are represented as follows: [21]

$$p_R(\theta) = 0.7629 + 0.7113\cos^2\theta, \quad \tau_R(\lambda) = 1.545 \cdot 10^{10}\lambda^{-4.086}, \qquad (23)$$

here λ is a wavelength of light in *nm*, and

$$F(z_s) = R_F(0) + R_F(z_s), \tag{24}$$

where $R_F(\varphi)$ is a Fresnel reflection coefficient of light from the ocean surface. In this study we disregard the roughness of the water surface. This approximation is acceptable only when the solar glint is not detected by the receiver, *i.e.* for the solar zenith angles $z_s > 30°$.

Radiance reflection coefficient of an aerosol part of the atmosphere that takes into account an atmospheric haze both from the scattering of the direct and reflected light is expressed as follows:

$$\rho_A(\lambda) = \frac{m}{4}\left[p_A(180° - z_s) + F(z_s)\, p_A(z_s)\right] \tau_A(\lambda), \tag{25}$$

here $\tau_A(\lambda)$ is an aerosol atmospheric optical thickness, $p_A(\theta)$ is an aerosol phase function of scattering. The typical values for aerosol optical thickness $\tau_A(\lambda)$ and ozone optical thickness $\tau_{oz}(\lambda)$ for ozone content of 0.325 *atm.-cm* [4] are given in Table 2.

Radiance reflection coefficient of the sea that consider the light ascending from the sea depth is:

$$\rho_S = \rho_W T_\Sigma \tag{26}$$

here ρ_W is a seawater reflection coefficient, and T_Σ is the atmospheric transmittance of the light going from the Sun to the surface and from the surface to the receiver:

$$T_\Sigma = 0.25\left(1 + e^{-\tau_R}\right)\left(1 + e^{-\tau_R m}\right) - \left(1 - e^{-B_A \tau_A (1+m)}\right)e^{-\tau_R (1+m)}. \tag{27}$$

In the near infrared part of the spectrum ($\lambda_0 = 745\,nm$) the ocean can be regarded as fully absorbing medium and we can consider $\rho_w(\lambda_0) = 0$. In this case by using Eqs. (8), and (21)-(23) we can derive an equation that expresses an aerosol optical thickness $\tau_0 \equiv \tau_A(\lambda_0)$ through the radiance reflection coefficient of the ocean-atmospheric system $\rho_0 \equiv \rho_0(\lambda_0)$:

$$\tau_0 = \frac{2}{k_1}\left\{\frac{d(z_s)}{q(z_s)} - \sqrt{\left[\frac{d(z_s)}{q(z_s)}\right]^2 - \frac{2k_1}{m\,d(z_s)}\left[\frac{\rho_0}{T_{oz}(\lambda_0)} - \rho_R(\lambda_0)\right]}\right\}, \tag{28}$$

where the functions $d(z_s)$ and $q(z_s)$ are defined by the following equations:

Table 2.

λ, nm	440	480	520	560	600	640	675	700	745
τ_{oz}	0.0003	0.0071	0.0183	0.0390	0.0503	0.0285	0.0146	0.0083	0.0030
τ_A	0.1146	0.0964	0.0821	0.0708	0.0617	0.0542	0.0487	0.0453	0.0400

$$d(z_s) = -2\big[D(180° - z_s) + F(z_s)\,D(z_s)\big], \Big\}$$
$$q(z_s) = 0.5\big[A(180° - z_{s_s}) + F(z_s)\,A(z_s)\big]. \Big\} \qquad (29)$$

The numerical values of the functions $d(z_s)$ and $q(z_s)$ for a number of the solar zenith angles z_s are given in Table 3.

Let us represent the spectral dependence of aerosol optical thickness in the following form:

$$\tau_A(\lambda) = \tau_0 \left(\frac{\lambda_0}{\lambda}\right)^{\alpha(\lambda)}, \qquad (30)$$

where $\tau_0 \equiv \tau_A(\lambda_0)$ and $\lambda_0 = 745\,nm$. If $\alpha(\lambda)$ is an arbitrary function, this equation represents an arbitrary spectral dependence of an aerosol optical thickness. When $\alpha = const$, Eq. (30) coincides with the Ångstrom formula. [22]

In the visible part of the spectrum the portion of internal sea light ascending to the upper atmospheric level $\big(\sim \rho_w\, T_\Sigma\, T_{oz}\big)$ is about ten times smaller than the portion of light from the atmospheric haze. By neglecting the term $\rho_S\, T_{oz}$ in Eq. (21) we introduce an error of the order of 10%. This procedure is more precise than the use of the expression (1). By using Eq. (21) with $\rho_S = 0$ and Eq. (30) we have the following equations for restoring parameter $\alpha(\lambda)$ (Ångstrom parameter) from the radiance reflection coefficient of the ocean-atmospheric system measured from a satellite:

$$\alpha(\lambda) \cong \ln\big[\rho_\Sigma(\lambda)/\rho_\Sigma^0\big]\big/\ln(\lambda_0/\lambda), \qquad (31)$$

where

$$\rho_\Sigma(\lambda) = \rho(\lambda)/T_{oz}(\lambda) - \rho_R(\lambda), \quad \rho_\Sigma^0 \equiv \rho_\Sigma(\lambda_0). \qquad (32)$$

We used Eqs. (28) and (31) for the estimation of τ_0, $\alpha(440)$ and $\alpha(560)$ from the ocean-atmospheric reflection coefficient ρ measured from the satellite. [6] It was found that the Ångstrom parameter $\alpha(\lambda)$ can be represented with the following equation:

$$\alpha(\lambda) = \beta(\lambda)\,/\,\tau_0, \quad \beta(\lambda) \approx 0.08 \pm 0.03. \qquad (33)$$

This gives us the following approximate formula for the aerosol optical thickness of the atmosphere:

Table 3.

z_s, degrees	10	20	30	40	50	60	70	80
d	1.40	1.08	0.98	0.87	0.78	0.68	0.70	0.77
q	1.21	0.88	0.71	0.58	0.50	0.44	0.45	0.51

$$\tau_A(\lambda) \cong \tau_0 \left(\frac{\lambda_0}{\lambda} \right)^{\frac{0.08}{\tau_0}}, \tag{34}$$

here $\tau_0 \equiv \tau_A(\lambda_0)$, $\lambda_0 = 745\,nm$. Equation (34), Eq. (8) with $\lambda_0 = 745\,nm$ and Table 1 give us the complete one-parameter model of atmospheric marine aerosol optical properties.

4. CONCLUSION

This study demonstrates that the optical properties of aerosol atmosphere above the ocean in some cases may be determined by one parameter — the total aerosol optical thickness in the near infrared, $\tau_0 \equiv \tau_A(745nm)$. [24] The spectral dependence of the total aerosol atmospheric optical thickness above the ocean is represented by the equation:

$$\tau_A(\lambda) = \tau_0 \left(\frac{745}{\lambda} \right)^{\frac{0.08}{\tau_0}}, \quad 0.01 \le \tau_0 \le 0.1, \quad 400\,nm \le \lambda \le 750\,nm. \tag{35}$$

The aerosol scattering phase function p_A and the probability of backscattering on aerosol particles B_A are computed by the following equations:

$$p_A(\theta) = A(\theta) + 5\tau_0 D(\theta), \quad B_A = 0.5 - \tau_0, \quad 0.01 \le \tau_0 \le 0.1, \tag{36}$$

with the empirical functions $A(\theta)$ and $D(\theta)$ given in Table 1.

It is necessary to note, that Eqs. (35)-(36) are derived from measurements restricted in time and location and made by different researchers. These equations should be used with discretion. Alternative approaches can be found in Refs. [13, 14]

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