The standard CZCS algorithm: found to be inappropriate for the atmosphere over the Baltic Sea*

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Baltic Sea
Aerosol radiance
Water-leaving radiance
Case 1 Waters
Case 2 Waters

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Abstract

In the standard CZCS atmospheric correction method the aerosol radiance is derived on the assumption that the water-leaving radiance at 670 nm is zero. This assumption, justified for water whose reflectance is determined solely by absorption (Case 1 Waters), gives rise to errors where the reflectance of the water is significantly influenced by scattering (Case 2 Waters), *e.g.* the Baltic Sea. The values of the aerosol radiance are too high and those of water-leaving radiance too low in comparison with the experimental ones. The relative error for the aerosol radiances (normalised to their values at 670 nm) decreases with wavelength from 60% at 443 nm to 13% at 550 nm.

1. Introduction

The total radiance L_S received by a sensor is determined by the water-leaving component L_W and the atmospheric radiance – the sum of the Rayleigh L_R and aerosol L_A components. Assuming that the Rayleigh radiance is known (Gordon, 1978, 1981; Robinson, 1985; Sturm, 1981), the other components exert the main influence on the total radiance at the sensor.

The aim of this paper is to show that the standard CZCS algorithm used for processing the global data set is not appropriate for the Baltic Sea. In the standard two-channel switching algorithm by Gordon *et al.* (1983), $L_W(670)$ is assumed to be zero. The same initial conditions also appear

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in an iterative algorithm by Smith and Wilson (1981) and a three-channel algorithm by Clark (1981). L_A is determined on the basis of this assumption. The principal difficulty, however, lies in the fact that this assumption is valid for the open ocean (Case 1 Waters) but not for turbid coastal waters (Case 2 Waters).

In the present paper the spectral values of the aerosol radiance and the aerosol optical thickness for the atmosphere over the Baltic Sea were determined on the basis of simultaneous measurements of the direct and diffuse irradiances at visible wavelengths: 400, 443, 490, 520, 550, 620 and 670 nm (Olszewski *et al.*, 1995a,b; Kuśmierczyk-Michulec and Darecki, 1996). The spectral values of the water-leaving radiance (Darecki *et al.*, 1994) were measured by means of a Multiwavelength Environmental Radiometer at visible wavelengths. On the basis of this data set the consequences of replacing the experimental values of the aerosol optical thickness and the water-leaving radiance for Baltic Sea conditions by the constants fully justified for Case 1 Waters (Morel and Prieur, 1977) are discussed.

2. The atmospheric radiance

The atmospheric radiance is the sum of the Rayleigh component $L_R(\lambda)$ and the aerosol radiance $L_A(\lambda)$. The Rayleigh radiance is given by Gordon *et al.* (1988):

$$L_R(\lambda) = \frac{\tau_R(\lambda) p_R(\theta, \theta_s) F_s'(\lambda) m}{4\pi}, \tag{1}$$

where the Rayleigh optical thickness $\tau_R(\lambda)$ is defined by Van Stokkom and Guzzi (1984), m – the path length through the atmosphere, according to Kasten (1966), is expressed by

$$m = [\cos(\theta) + 0.15(93.885 - \theta)^{-1.259}]^{-1}, \tag{1a}$$

$p_R(\theta, \theta_s, \lambda)$ is taken from Gordon (1988), and the dependences θ_s and θ represent the two paths through the atmosphere: the first is the path from the Sun to the Earth, the second the path from the Earth to the satellite.

Values of the mean extraterrestrial irradiance F_s were taken from Neckel and Labs (1984) and corrected for the Earth–Sun distance. Following Gordon *et al.* (1983), the effects of the absorbent gases in the atmosphere (within the CZCS bands only ozone absorption is of importance) were determined by assuming two trips of the extraterrestrial irradiance through the atmosphere. Then $F_s'(\lambda)$ was expressed by (Gregg *et al.*, 1993)

$$F_s'(\lambda) = F_s(\lambda) T_O(\lambda, \theta_s) T_O(\lambda, \theta), \tag{2}$$

where T_O represents the transmittance function for ozone absorption. The aerosol radiance $L_A(\lambda)$ may be obtained from Gordon and Castaño (1989):

$$L_A(\lambda) = \frac{\omega_A(\lambda)\tau_A(\lambda)p_A(\theta,\theta_s)F_s'(\lambda)m}{4\pi}, \tag{3}$$

where $\omega_A(\lambda)$ is the single scattering albedo of the aerosol, and $p_A(\theta,\theta_s)$ is a factor to account for the probability of scattering to the spacecraft for three different paths from the Sun (Gordon and Castaño, 1989).

3. The aerosol optical thickness

The aerosol optical thickness of the atmosphere over the Baltic Sea τ_A was determined on the basis of simultaneous measurements (Olszewski *et al.*, 1995b) of the total $E_{tot}(\lambda)$ and diffuse $E_{dif}(\lambda)$ solar spectral irradiances at seven spectral channels across the visible spectrum (400 nm to 670 nm). All data were collected on cloudless days in 1993 and 1994 over the southern Baltic Sea. The aerosol optical thickness $\tau_A(\lambda)$ was obtained from the expression

$$\tau_A(\lambda) = m^{-1}\ln T_A(\lambda)^{-1}, \tag{4}$$

where the transmittance function for the aerosol extinction T_A is defined as

$$T_A(\lambda) = \frac{E_{tot}(\lambda) - E_{dif}(\lambda)}{F_s(\lambda)T_R(\lambda)T_O(\lambda)\cos(\theta_s)}; \tag{5}$$

T_R is the Rayleigh scattering transmittance function (Van Stokkom and Guzzi, 1984). The transmittance function for ozone absorption T_O was based on the mean daily amounts of atmospheric ozone measured by means of a Dobson spectrophotometer at the Belsk Geophysical Observatory (*Atmospheric ozone ...*, 1996).

As shown by Kuśmierczyk-Michulec and Darecki (1996), the application of empirical orthogonal functions (EOF) has enabled all spectra of the aerosol optical thickness to be defined by a single general equation that taken only the first mode $h_1(\lambda_j)$ into account:

$$\tau_{Ai}(\lambda_j) = h_1(\lambda_j)\beta_{i1} + <\tau_A(\lambda_j)> \quad j = 1, ..., 7;\ \ i = 1, ..., 66, \tag{6}$$

where $<\tau_A(\lambda_j)>$ is the mean spectrum (see Fig. 1). The subscript i symbolises each successive measurement, the subscript j corresponds to the number of spectral channels. The amplitudes β_{i1} determine the temporal variability of $\tau_A(\lambda)$. In this paper $\beta_{\max}$ and $\beta_{\min}$ – the maximum and minimum amplitude of τ_A defining the maximum and the minimum spectrum respectively – are used instead of β_{i1} (see Fig. 1).

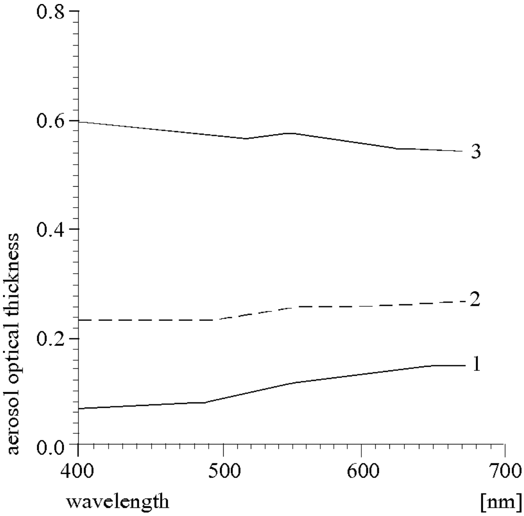

Fig. 1. The spectrum of the aerosol optical thickness recorded over the Baltic Sea in 1994: 1 – minimum ($\beta_{\min}$ = –0.3789), 2 – mean, 3 – maximum ($\beta_{\max}$ = 0.8696). The details concerning the aerosol optical thickness over the Baltic Sea are given by Kuśmierczyk-Michulec and Darecki (1996)

4. Results

In the standard CZCS atmospheric correction method (Gordon *et al.*, 1983; Gordon and Castaño, 1989) the aerosol radiance $L_A(\lambda)$ is related to $L_A(670)$ by the expression

$$L_A(\lambda) = \frac{F_s(\lambda)T_O(\lambda)}{F_s(670)T_O(670)} \left(\frac{\lambda}{670}\right)^{B(\lambda)} L_A(670), \tag{7}$$

where the values of $B(\lambda)$ are assumed to be 0.12, 0.00 and 0.00 for 443, 520 and 550 nm respectively, which are typical marine haze values (McClain and Yeh, 1994). The question arises whether this algorithm might be applied to the atmosphere over the Baltic Sea. To answer this question some comparisons need to made.

Let $L_A(\lambda)^{\mathrm{Case\,2}}$ be the aerosol component calculated on the basis of the experimental values of $\tau_A(\lambda)$ for the atmosphere over the Baltic Sea (see eqs. (2)–(5)). The experimental values of the aerosol optical thickness are included in the range shown in Fig. 1 between the maximum and minimum spectrum (maximum and minimum amplitude of the aerosol optical thickness).

Let us assume that the discrepancies between $L_A(\lambda)^{\mathrm{Case\,2}}$ and $L_A(\lambda)^{\mathrm{Case\,1}}$ (eq. (6)) are the 'error' ϵ given by

$$\epsilon = \frac{\frac{L_A(\lambda)}{L_A(670)}^{\text{Case 1}} - \frac{L_A(\lambda)}{L_A(670)}^{\text{Case 2}}}{\frac{L_A(\lambda)}{L_A(670)}^{\text{Case 2}}}. \tag{8a}$$

Taking into account the relations

$$\frac{L_A(\lambda)}{L_A(670)}^{\text{Case 1}} \sim \left(\frac{\lambda}{670}\right)^{B(\lambda)} \tag{8b}$$

and

$$\frac{L_A(\lambda)}{L_A(670)}^{\text{Case 2}} \sim \frac{\tau_A(\lambda)}{\tau_A(670)}, \tag{8c}$$

the 'error ϵ' may be rewritten as

$$\epsilon \cong \frac{\left(\frac{\lambda}{670}\right)^{B(\lambda)} - \frac{\tau_A(\lambda)}{\tau_A(670)}}{\frac{\tau_A(\lambda)}{\tau_A(670)}}, \tag{8}$$

where $\lambda = 443$, 520, 550 nm. The ϵ values obtained for the CZCS bands are listed in Tab. 1.

Table 1. The arithmetic estimation of the error within the CZCS bands

	443 nm	520 nm	550 nm
ϵ_{ad}	0.606	0.397	0.134
$\sigma_{\epsilon_{ad}}$	0.789	0.41	0.168
$\epsilon_{\max}$	1.404	0.808	0.302
$\epsilon_{\min}$	–0.193	–0.013	–0.034

where ϵ_{ad} is the mean value (see eq. (9)), and $\epsilon_{\max}$ and $\epsilon_{\min}$ are the respective maximum and minimum errors determined by the limits of the standard deviation $\sigma_{\epsilon_{ad}}$.

In order to determine the consequences of the assumption $L_W(670) = 0$, the relation of the water-leaving radiance to the total radiance was found:

$$U_W = \frac{L_W(\lambda)}{L_S(\lambda)}, \tag{9}$$

where the total radiance at the sensor is given by

$$L_S = L_A + L_R + T' L_W, \tag{10a}$$

where T', the diffuse transmittance, is given by Gordon *et al.* (1983) and Gordon and Castaño (1989), and determines the quantity of the water-leaving radiance received by the sensor.

The values of the water-leaving radiance for the Baltic Sea are taken from Darecki *et al.* (1994). All the data were collected in 1993 (from April to September). The mean value of L_W was calculated only for cloudless days (only 105 of the 273 radiance spectra data were selected). The maximum and minimum values of the water-leaving radiance were analysed (see Fig. 2).

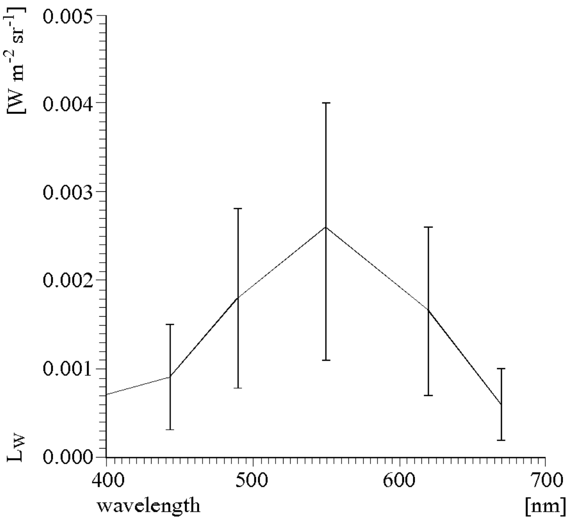

Fig. 2. The mean value of the water-leaving radiance for the Baltic Sea, based on experimental data in Darecki *et al.* (1994). All data were collected in 1993 (from April to September)

The contribution U_W of the water-leaving radiance L_W to the total radiance L_S is presented in Tab. 2. The relative dispersion ΔU_W was defined as

$$\Delta U_W = \pm \frac{U_W(L_W^{\max}, \beta_{\min}) - U_W(L_W^{\min}, \beta_{\max})}{U_W(L_W^{\max}, \beta_{\min}) + U_W(L_W^{\min}, \beta_{\max})}, \tag{10}$$

where $L_W^{\max}$ and $L_W^{\min}$ are the respective maximum and minimum values of the water-leaving radiance, determined by the limits of the standard deviation (see Fig. 2), and $\beta_{\max}$, $\beta_{\min}$ are the respective maximum and minimum amplitudes of the aerosol optical thickness (see Fig. 1).

Evidently (Tab. 2), the contribution U_W of the water-leaving radiance to the total radiance at the sensor within the CZCS bands is greatest for the maximum values of L_W and minimum values of the amplitude $\beta_{\min}$. However, the minimum contribution is recorded for $L_W^{\min}$ and $\beta_{\max}$.

The relative dispersions ΔU_W at wavelengths 443 nm and 670 nm are similar, although the contribution U_W of the water-leaving radiance to the total radiance at $\lambda = 670$ nm is much higher than at $\lambda = 443$ nm (see Tab. 2). This can be explained by the stabilising effect of the Rayleigh radiance. The component L_R decreases with wavelength increase, its values at $\lambda = 670$ nm are therefore about 10 times smaller than at $\lambda = 400$ nm. The total radiance $L_S(670)$ is thus more sensitive to variations in the components $L_A(670)$ and $L_W(670)$.

Table 2. The contribution U_W of the water-leaving radiance to the total radiance at the sensor within the CZCS bands

λ [nm]	$L_W^{\max}$ [W m^{-2}sr^{-1}]	$L_W^{\min}$ [W m^{-2}sr^{-1}]	$\frac{U_W(L_W^{\max}, \beta_{\max})}{U_W(L_W^{\min}, \beta_{\max})}$	$\frac{U_W(L_W^{\max}, \beta_{\min})}{U_W(L_W^{\min}, \beta_{\min})}$	ΔU_W [%]
400	0.0013	0.0001	$\frac{1.86}{0.15}$	$\frac{1.9}{0.15}$	± 85.7
443	0.0015	0.0003	$\frac{2.6}{0.53}$	$\frac{2.67}{0.54}$	± 66.9
490	0.0028	0.0008	$\frac{6.74}{2.02}$	$\frac{7.0}{2.10}$	± 55.3
550	0.0040	0.0011	$\frac{14.07}{4.27}$	$\frac{14.81}{4.52}$	± 55.2
620	0.0026	0.0007	$\frac{15.76}{4.75}$	$\frac{16.93}{5.15}$	± 56.2
670	0.0010	0.0002	$\frac{9.46}{2.04}$	$\frac{10.39}{2.26}$	± 67.2

5. Conclusions

The direct application of the standard CZCS atmospheric correction algorithm to the atmosphere over the Baltic Sea gives rise to a number of discrepancies. The most important conclusions are:

- The values of the aerosol radiance estimated on the basis of the CZCS algorithm are much higher than the experimental ones. The relative error (see eq. (9)) for the aerosol radiances (normalised to their values at 670 nm) decrease with wavelength from 60% at 443 nm to 13% at 550 nm (see Tab. 1).

- The assumption that the water-leaving radiance $L_W(670) = 0$, fully justified for most low-pigment Case 1 Waters, is invalid for Baltic Sea conditions. Moreover, the relation of the water-leaving radiance to the total radiance at 670 nm depends on the amplitude of the aerosol optical thickness and varies from 2% to 10% (see Tab. 2).

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