Papers

The aerosol optical thickness over the Baltic Sea

OCEANOLOGIA, No. 38 (4)
pp. 423–435, 1996.
PL ISSN 0078–3234

Baltic Sea
Aerosol optical thickness
Empirical
orthogonal functions
Air temperature
Visible spectrum

Jolanta Kuśmierczyk-Michulec,
Mirosław Darecki
Institute of Oceanology,
Polish Academy of Sciences,
Sopot

Manuscript received June 6, 1996, in final form September 7, 1996.

Abstract

The results of applying empirical orthogonal functions (EOF) to the decomposition and approximation of spectra of the aerosol optical thickness are presented. The aerosol optical thickness is calculated on the basis of the experimental data of the direct and diffuse irradiances collected on cloudless days in 1993 and 1994 over the southern Baltic Sea. The main finding is that the amplitude of aerosol optical thickness in the visible spectrum is dependent on air temperature.

1. Introduction

The aerosol radiance $L_A(\lambda)$ is a key parameter in atmospheric correction models of remotely-sensed data in marine applications (Gordon, 1978, 1981; Sturm, 1981; Robinson, 1985). Appearing explicitly in the expression for $L_A(\lambda)$, the aerosol optical thickness $\tau_A(\lambda)$ is a physical quantity measuring the attenuation power of air molecules with respect to a specific wavelength of the incident light. According to Weller and Leiterer (1988) marine aerosols have local properties dependent on their origin and meteorological conditions. Therefore the aerosol optical thickness reflects nearly all daily and seasonal changes occurring in the atmosphere (for instance, the changes in the relative humidity or temperature). In order to analyse more precisely the spatial and temporal changeability of the aerosol optical thickness spectra, the method of empirical orthogonal functions (EOF) (Lorenz, 1956; Nielsen, 1979; Preisendorfer, 1988; Jankowski, 1994) was applied. The results show that all data can be defined by means of one general equation with air temperature as the only local variable. Such formulae may be very useful especially in atmospheric correction models, where the accurate determination of the aerosol radiance is very important. This article presents the preliminary results of the modelling of the aerosol optical thickness for the atmosphere over the Baltic Sea.

2. Methods of measurement

The aerosol optical thickness of the atmosphere over the Baltic Sea was determined on the basis of simultaneous measurements of the total $E_{tot}(\lambda)$ and the diffuse $E_{dif}(\lambda)$ solar spectral irradiances in seven spectral channels across the visible spectrum (for the range of wavelengths from 400 nm to 670 nm).

The details of this measurement method are presented in Olszewski *et al.* (1995b). The idea of this method is to remove cyclicly from the photodetector field of view successive parts of the horizon (one of them including all direct solar radiation). This is done by means of a strip diaphragm with fixed dimensions, rotating automatically around the optical axis of a cosine collector. The strip is equal in width to the diameter of the cosine collector; its bottom edge lies in the cosine collector plane, and its top edge at the zenith angle, which is no smaller than the minimal solar zenith angle during measurements. Even though the strip covers the Sun exactly, it will always cover some part of the horizon as well. The entire recording process is controlled by an appropriate computer program. The maximum relative error of the measurement does not exeed 7% (obviously, any decrease in the width of the strip will reduce this error) (Olszewski *et al.*, 1995b). Moreover, the main ideas of this method – continuous and automatic measurement, and the maximum stability of its parameters – are implemented under natural marine conditions, even on board a ship in motion.

3. The aerosol optical thickness

The aerosol optical thickness $\tau_a(\lambda)$ can be obtained from the following expression:

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

where m according to Kasten (1966) is expressed by

$$m = [\cos(\theta) + 0.15(93.885 - \theta)^{-1.259}]^{-1}, \tag{2}$$

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

$$T_A(\lambda) = \frac{E_{tot}(\lambda) - E_{dif}(\lambda)}{F_s(\lambda) T_R(\lambda) T_W(\lambda) T_O(\lambda) T_U(\lambda) \cos(\theta)}. \tag{3}$$

Values of the mean extraterrestrial irradiance $F_s(\lambda)$ are taken from Neckel and Labs (1984) and corrected for the Earth–Sun distance after Gordon *et al.* (1983):

$$F_s(\lambda) =< F_s(\lambda) > \left[1 + 0.0167 \cos \frac{2\pi(d-3)}{365}\right]^2, \tag{4}$$

where d is the sequential day of the year.

T_R, T_W, T_O and T_U are the respective transmittance functions for Rayleigh scattering, water vapour absorption, ozone absorption and uniformly mixed gas absorption. According to the Neckel and Labs spectrum for the seven wavelengths taken into account (400, 443, 490, 520, 550, 620 and 670 nm), the absorption of water vapour and uniformly mixed gas can be neglected. Then eq. (3) can be replaced by

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

In this paper the transmittance function for Rayleigh scattering was taken from Bird (1984). The transmittance function for ozone absorption was determined on the basis of the daily mean contents of atmospheric ozone, measured by means of a Dobson spectrophotometer at the Belsk Geophysical Observatory (*Atmospheric ozone* ..., 1996). The uncertainty resulting from the replacement of the hourly mean ozone contents by the daily mean values is $\Delta T_O = \pm 0.01$.

The aerosol optical thickness was calculated on the basis of eqs. (1), (2), (4), (5) and the empirical values of the total E_{tot} and the diffuse E_{dif} solar spectral irradiances collected on cloudless days of 1993 and 1994 over the southern Baltic Sea. The uncertainty of the total irradiance measurement resulting from the sensitivity of the detector to the temperature is $\Delta E_{tot} = \pm 2\%$ (Olszewski *et al.*, 1995a). The uncertainty of the diffuse irradiance measurement $\Delta D = \Delta(E_{dif}/E_{tot}) = \pm 0.05$ (Olszewski *et al.*, 1995b). The maximum measurement uncertainty of the aerosol optical thickness $\Delta\tau$ (Szydłowski, 1981) is ± 0.06.

The diagrammatic location of measurement stations is shown in Fig. 1. The measurements were carried out not only in the region of the open Baltic Sea but also off the coast of Poland. This set of data thus includes various cases of the aerosol optical thickness reflecting different hydrometeorological conditions (*i.e.* wind velocity, air humidity, temperature). Examples of various spectral types of $\tau_A(\lambda)$, obtained for the atmosphere over the Baltic Sea are shown in Figs. 2 and 3. The spatial (Fig. 2) and temporal (Fig. 3) variabilities are evident.

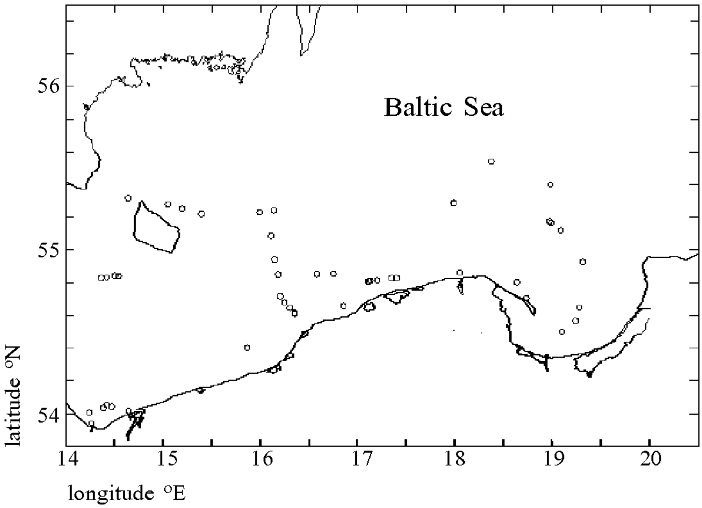

Fig. 1. Location of measurement stations

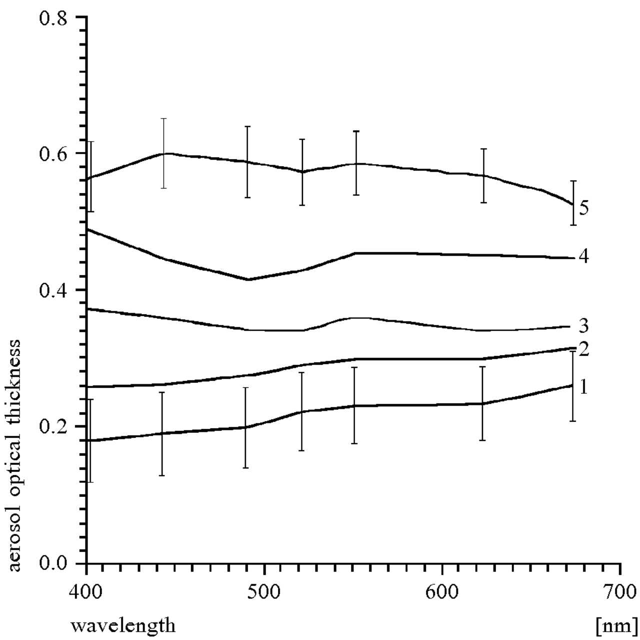

Fig. 2. Spectra of the aerosol optical thickness of the atmosphere over the Baltic Sea, recorded on 23.08.1994. The numbers correspond to the consecutive hours of measurement (GMT): 1 – 10:44, 2 – 11:44, 3 – 13:15, 4 – 13:52, 5 – 14:47. The maximum measurement uncertainty: $\Delta\tau = \pm 0.06$ (for clarity, $\Delta\tau$ is marked only for two curves). During measurements the air temperature rose from 16.9°C in the morning to 18.1°C in the afternoon

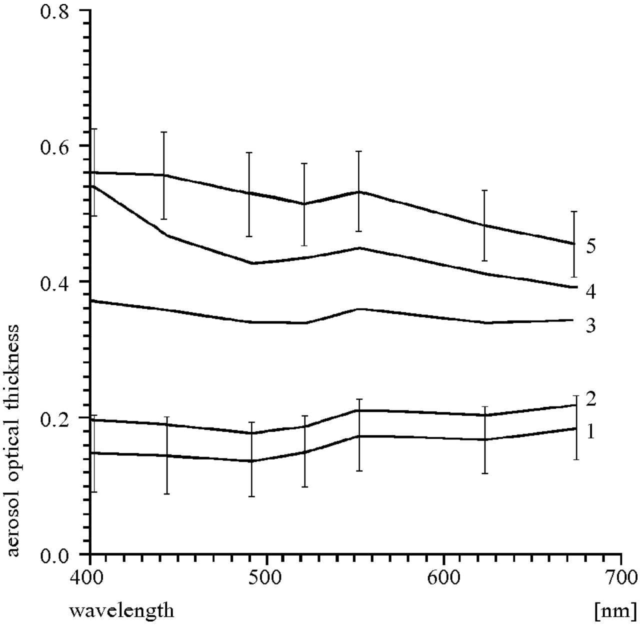

Fig. 3. Examples of different spectral types of the aerosol optical thickness $\tau_A(\lambda)$ for the atmosphere over the Baltic Sea, recorded in 1994. The numbers correspond to the measurement days (day, month) and hours (GMT) respectively: 1 – 11.05, 13:32; 2 – 11.05, 14:31; 3 – 23.08, 13:15; 4 – 08.08, 13:52; 5 – 07.08, 16:42. For clarity $\Delta\tau$ is marked only for the top and the bottom spectrum

In order to analyse more precisely the data set, which consists of 66 measurements yielding 462 empirical points at seven wavelengths, the method of empirical orthogonal functions (EOF) was applied.

4. Results

The results of the EOF method show that all spectra $\tau_A(\lambda)$ can be defined by means of one general equation, taking into account only the first mode $h_1\,(l = 1)$, because the contribution of the first eigenvalue $\wp_1$ to the total variance is 97.9% : $\aleph(1) = \imath_1 = 0.979$. Such a value allows the remaining modes $(l = 2, ..., 7)$ to be neglected (Nielsen, 1979; Wróblewski, 1986; Jankowski, 1994) and we can write $\tau_A(\lambda)$ as

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

Fig. 4 shows the amplitudes β_{il} for the first three modes $(l = 1, 2, 3)$ for each consecutive measurement station (geographical points where measurements were made). It is clear that the amplitude of the first mode reflects nearly all the differences between the aerosol optical thicknesses at the individual measurement stations. The small amplitudes of the second and third

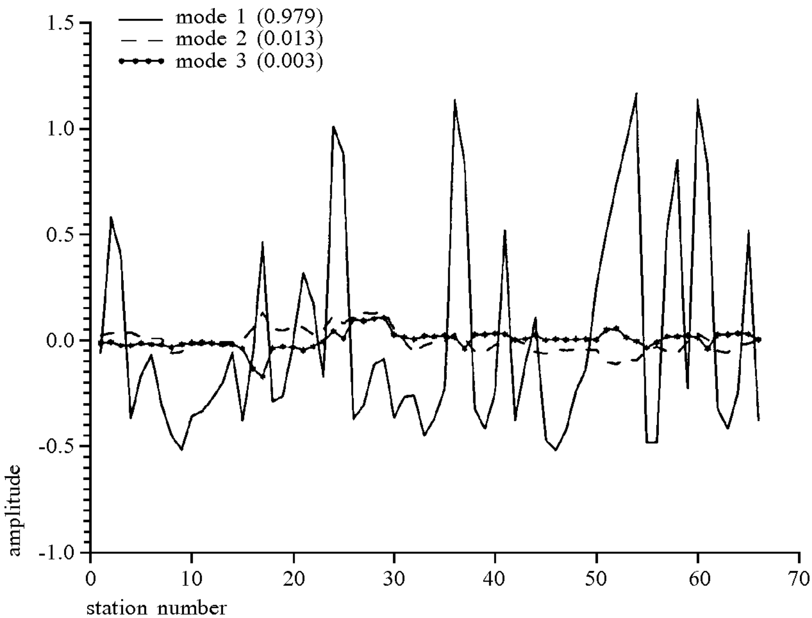

Fig. 4. Amplitudes β_{il} for the first three modes ($l = 1, 2, 3$) for each consecutive measurement station

modes, carrying in all about 2% of the total information, can be neglected and regarded as disturbances. Then, 'clear', 'filtered out' information about amplitude changes of the aerosol optical thickness is included in β_{i1}, where i is the station number (measurement number).

Fig. 5 shows the spectra of eigenvectors (modes) $h_1(\lambda_j)$, $h_2(\lambda_j)$ and $h_3(\lambda_j)$. Clearly, it is enough to take the first mode to describe the spectral structure (shape) of the aerosol optical thickness. The remaining modes, which have an infinitesimal variance, contribute little and can be omitted from the practical calculations.

Obviously, the number of data taken for the purpose of EOF affects the numerical values of the individual elements in eq. (6). However, the results of the calculations indicate that the set containing measurements from 40 profiles is sufficient to obtain the same variance contribution as for the input data (98% information included in mode 1). The mean value may also change slightly, but the changes do not exceed 2–3% at wavelength $\in$ (400 nm, 520 nm) or 0.5–1% for $\lambda \in$ (550 nm, 670 nm). Moreover, this will always be an increasing function regardless of the number of input data.

Eq. (6) describes the result of applying EOF mathematically. In order to impart a physical sense to this equation, its several terms are interpreted as follows.

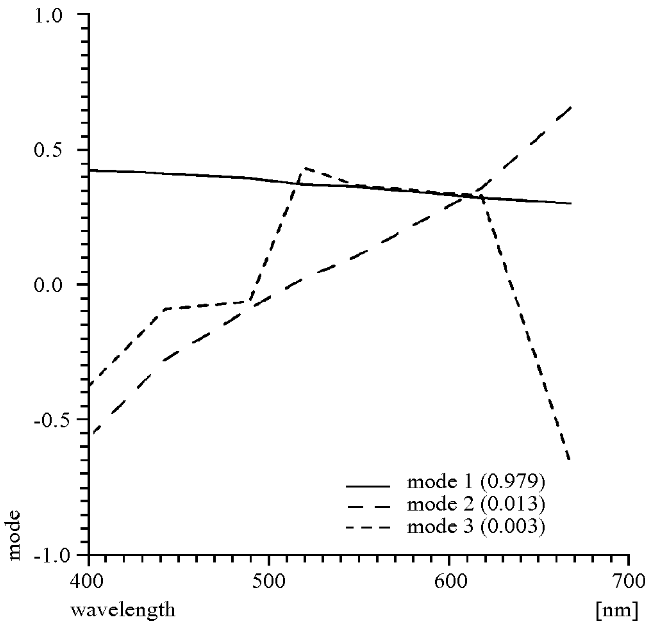

Fig. 5. Spectra of eigenvectors (modes) $h_1(\lambda)$, $h_2(\lambda)$ and $h_3(\lambda)$. The first mode reflects the shape of the aerosol optical thickness. Because of the infinitesimal variance the remaining modes can be omitted from the practical calculations

Table 1. The mean values of the aerosol optical thickness $< \tau_A(\lambda) >$

No.	λ [nm]	$< \tau_A >$	$< \tau_A >_{\max}$	$< \tau_A >_{\min}$
1	400	0.227	0.394	0.060
2	443	0.227	0.389	0.066
3	490	0.228	0.383	0.073
4	520	0.235	0.381	0.089
5	550	0.251	0.395	0.107
6	620	0.258	0.388	0.128
7	670	0.265	0.391	0.138

Fig. 6a presents two functions: $< \tau_A(\lambda) >$ (Tab. 1) and $c(\lambda)$; the latter is plotted on the basis of given values of the extinction coefficient (International Association ..., 1984) for oceanic particles. Both functions are normalised to their values for the wavelength $\lambda = 550$ nm. These two functions were compared in order to find similarities and to express the mean value $< \tau_A >$ in physical terms. Despite certain discrepancies between their numerical values (see Fig. 6a), $< \tau_A(\lambda) >$ may be interpreted as the *basic spectrum* of the aerosol optical thickness describing mainly particles

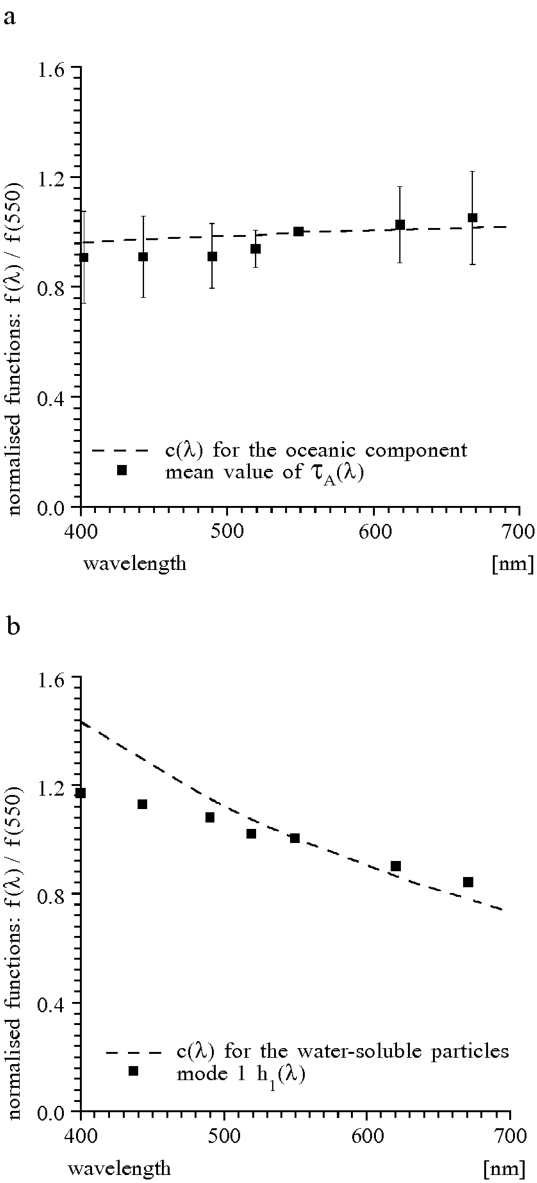

Fig. 6. Comparison between two functions, normalised to their values for $\lambda = 550$ nm. The values of the extinction coefficient $c(\lambda)$ (dashed line) are taken from International Association ... (1984); $< \tau_A(\lambda) >$ and $c(\lambda)$ for oceanic particles (a), $h_1(\lambda)$ and $c(\lambda)$ for water-soluble particles (b)

consisting of sea salt and liquid water. This *basic spectrum* changes as a consequence of various hydrometeorological factors (*i.e.* a change in wind velocity, air humidity or temperature). These changes are determined by the second part of eq. (6), which is the product of function h_1 and amplitudes β_{i1}. It is the so-called *dynamic component* of the aerosol optical thickness.

The first element of the above-mentioned product – mode 1 – can be approximated to the linear function

$$h_1(\lambda_j) = h_1(\lambda) = -0.00043\lambda + 0.603, \tag{7}$$

where λ is in [nm].

The function describing the extinction coefficient for water-soluble particles (International Association ..., 1984) is similar in shape to mode 1. Fig. 6b shows both functions ($c(\lambda)$ and h_1) normalised to their values at wavelength $\lambda = 550$ nm. Certain discrepancies at wavelengths $\in$ (400 nm, 520 nm) may have been caused by the fact that $c(\lambda)$ is also determined on the basis of the aerosol model (Rosen, 1971; Toon and Pollack, 1973) without the specific properties of Baltic aerosols having been taken into account. Obviously, this is only a suggested explanation of the above discrepancies and interpretation of h_1, which may then be regarded as a function principally describing attenuation by water-soluble aerosols.

The second element of the *dynamic component* is made up of the amplitudes β_{i1}. The direct relationship between these and the optical parameters of the atmosphere, especially the relative humidity, was analysed in depth. But only the air temperature turned out to have a significant influence on the amplitudes of the aerosol optical thickness (the detector is insufficiently temperature-sensitive to have such an influence on the results). The dependence of amplitudes on air temperature may then be defined by the linear function (see Fig. 7a–d)

$$\beta_{i1} = \beta(T_p) = \tilde{a}T_p + \tilde{b}, \tag{8}$$

with the parameters listed in Tab. 2.

Table 2. Parameter values: $\tilde{a}$ and $\tilde{b}$

Minimal temperature by day T_p [°C]	Maximum temperature by day T_p [°C]	$\tilde{a}$ [°C]$^{-1}$	$\tilde{b}$	Correlation coefficient r
7.5	10	–0.52	5.06	0.94
7.5	15	0.056	–0.82	0.96
16.5	18.5	0.65	–11.05	0.96
19.5	22	0.66	–13.43	0.96

Figs. 7a–d show the linear dependence of amplitude β_{i1} on the diurnal distribution of air temperature T_p. As a general conclusion, it can be stated that for warm days with daytime temperatures of 18.5–22°C, the amplitude

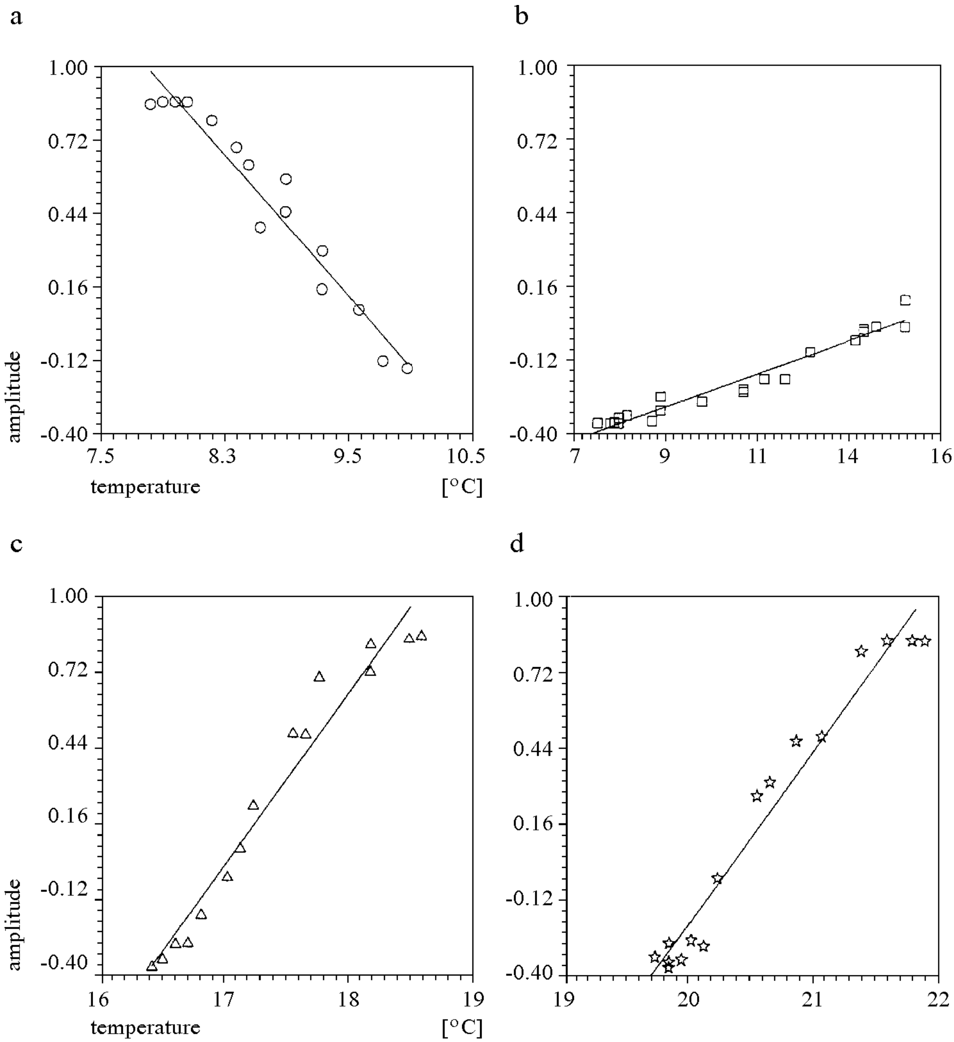

Fig. 7. Linear dependence of amplitudes β_{i1} on the diurnal distribution of air temperature T_p (a–d), (Tab. 2). As the measurements were carried out on board a ship in the open Baltic Sea, the temperature range may differ slightly in comparison to that on land

of the diurnal aerosol optical thickness will be higher (Fig. 7c, and 7d) than for cooler days when the air temperature does not exceed 10°C.

Clearly, it is very difficult to explain on the basis of one parameter only why $\tau_A(\lambda)$, or more precisely its amplitude, increases or decreases during the day. One of the reasons for this is the occurrence of numerous accompanying processes. Such additional information would be a knowledge of the vertical air temperature profile, thus allowing one to establish whether conditions existed for generating convection. However, for technical reasons such measurements were not carried out very frequently.

According to the definition of Pasquill's stability classes (Pasquill, 1961), the measurement conditions (maximum wind velocity – 4.5 m s^{-1}, daytime insolation) indicate that the atmospheric stability varied from class A – extremely unstable to class C – slightly unstable. This fact may also explain the changes in aerosol optical thickness during the day.

5. Conclusions

The results of this research are of a preliminary nature. The application of a mathematical method (empirical orthogonal functions, EOF) to aerosol optical thickness spectra has enabled us to reach certain conclusions about the spatial and temporal variability pattern. The most important conclusions are:

- All aerosol optical thickness spectra can be defined by means of one general equation which takes into account only the first mode h_1, because the contribution of the first eigenvalue to the total variance is 97.9%.
- The mean value $< \tau_A(\lambda) >$ can be interpreted as the *basic spectrum* of the aerosol optical thickness, principally describing particles consisting of sea salt and liquid water. The first mode h_1 can be regarded as a function describing mainly attenuation by water soluble aerosols.
- The temporal variability of $\tau_A(\lambda)$ is determined by the amplitudes of β_{i1}. The direct relationship between these and the optical parameters of the atmosphere, especially the relative humidity, was analysed in depth: only the air temperature turned out to have a significant influence on β_{i1}.
- The results of the EOF method are very useful for atmospheric correction models of remotely sensed data in marine applications, even if some processes are still unknown. The main advantage of this model is that it remains open to all kinds modifications.

References

Atmospheric ozone and solar radiation 1994, 1996, Publ. Inst. Geophys. PAS, D45 (279), 1–98.

Bird R. E., 1984, *A simple, solar spectral model for direct-normal and diffuse horizontal irradiance*, Solar Energy, 32 (4), 461–471.

Gordon H. R., 1978, *Removal of atmospheric effects from satellite imagery of the oceans*, Appl. Opt., 17, 1631–1636.

Gordon H. R., 1981, *A preliminary assessment of the Nimbus-7 CZCS atmospheric correction algorithm in a horizontally inhomogeneous atmosphere*, [in:] *Oceanography from space*, J. F. R. Gower (ed.), Plenum Press, New York, 257–265.

Gordon H. R., Clark D. K., Brown J. W., Brown O. B., Evans R. H., Broenkov V. V., 1983, *Phytoplankton pigment concentrations in the Middle Atlantic Bight: Comparison of ship determinations and CZCS estimates*, Appl. Opt., 22, 20–36.

International Association for Meterology and Atmospheric Physics. Radiation Commission, 1984, *A preliminary cloudless standard atmosphere for radiation computation*, Boulder, Colorado, 9–10.

Jankowski A., 1994, *The application of EOF in the analysis of the variability of water temperature, salinity and density in selected regions of the Norwegian Sea*, Oceanologia, 35, 27–60.

Kasten F., 1966, *A new table and approximate formula for relative optical mass*, Arch. Meteor. Geophys. Bioclim., B14, 206–223.

Lorenz E. N., 1956, *Empirical orthogonal function and statistical weather*, Statistical forecasting project, Sci. Rep. No. 1, Mass. Inst. Tech., Dept. Meteor., Cambridge, 49 pp.

Neckel H., Labs D., 1984, *The solar radiation between 3300 and 12500 Å*, Solar Phys., 90, 205–258.

Nielsen P. B., 1979, *On empirical orthogonal functions (EOF) and their use for analysis of the Baltic sea level*, Københavns Univ., Inst. Fys. Oceanogr., Rep. No. 40, 37 pp.

Olszewski J., Darecki M., Sokólski M., 1995a, *An instrument for measuring the spectral distribution of upward radiance above the sea*, Stud. i Mater. Oceanol., 68, 15–25.

Olszewski J., Kuśmierczyk-Michulec J., Sokólski M., 1995b, *A method for the continuous measurement of the diffusivity of the natural light field over the sea*, Oceanologia, 37 (2), 299–310.

Pasquill F., 1961, *The estimation of the dispersion of windborne material*, Meteor. Mag., 90, 33–49.

Preisendorfer R. W., 1988, *Principal component analysis in meteorology and oceanography*, Elsevier, Amsterdam–Oxford–New York–Tokyo, 425 pp.

Robinson J. S., 1985, *Satellite oceanography*, Ellis Horwood Ltd, Chichester, 455 pp.

Rosen J. M., 1971, *The boiling point of stratospheric aerosols*, J. Appl. Meteor., 10, 1044–1046.

Sturm B., 1981, *The atmospheric correction of remotely sensed data and the quantitative determination of suspended matter in marine surface layers*, [in:] *Remote sensing in meteorology, oceanography and hydrology*, A. P. Cracknell (ed.), Ellis Horwood Ltd, Chichester, 163–197.

Szydłowski H., 1981, *Theory of measurements*, PWN, Warszawa, 441 pp., (in Polish).

Toon O. B., Pollack J. B., 1973, *Physical properties of the stratospheric aerosol*, J. Appl. Meteor., 10, 725–731.

Weller M., Leiterer U., 1988, *Experimental data on spectral aerosol optical thickness and its global distribution*, Beitr. Phys. Atmosph., 61 (1), 1–9.

Wróblewski A., 1986, *Application of EOF to computations of the storm surges on the Polish Baltic coast in January 1983*, Acta Geophys. Pol., 34, 1, 63–76.