The aerosol optical thickness of the atmosphere over the Norwegian Sea obtained from different experimental data

OCEANOLOGIA, No. 34
pp. 27–37, 1993.
PL ISSN 0078-3234

Aerosol optical thickness
Atmosphere
Arctic Sea
Norwegian Sea

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Manuscript received July 9, 1993, in final form December 3, 1993.

Abstract

The aerosol optical thickness at visible wavelengths for the marine aerosol is analysed. Several methods prior to atmospheric correction are compared for the case when the only known optical parameter is one of the following: irradiance, visibility, relative humidity or aerosol size distribution. Experimental data measured in 1987–1989 and 1991–1992 in the Norwegian and Arctic Seas are used to verify the theoretical models.

1. Introduction

A knowledge of the aerosol optical thickness is of importance both in environmental research and in attempts at correcting the influence of the atmosphere on satellite data. It is beyond doubt that absorption and scattering by aerosol particles in the atmosphere exert an important effect on radiative transfer and the radiation budget of the Earth. The aerosol optical thickness – the vertically integrated extinction due to aerosols – is thus an important parameter affecting climate.

In the absence of direct spectral measurements of the aerosol optical thickness, *e.g.* transmission measurements, the only method is to estimate it theoretically. Therefore, it is very important to find the method giving the most accurate results.

In this paper the aerosol optical thickness is determined simultaneously for the same geographic point by four different methods; these take into account visibility (Sturm, 1981), relative humidity (Fitzgerald, 1989), size distribution for aerosols (Guzzi *et al.*, 1987) and irradiance. The results are then applied to the solar spectral model for irradiance (Bird, 1984) and compared with irradiance measurements. All calculations were carried out for cloudless days.

The results are based on experimental data obtained by the Institute of Oceanology of the Polish Academy of Sciences in 1987–1989 and 1991–1992 in the Norwegian and Arctic Seas.

2. Methods

2.1. Method dependent on visibility V

One of several methods describing the aerosol optical thickness τ_a is the analytical approximation formula proposed by Sturm (1981), according to whom τ_a can be calculated from the relation

$$\begin{aligned}\tau_a(\lambda) &= (\frac{3.912}{V} - 0.0116)(\frac{0.55}{\lambda})^{\alpha-3}\left[H_1(1-\exp(\frac{-5.5}{H_1}))+\right.\\ &+ \left. 12.5\exp(\frac{-5.5}{H_1}) + H_2\exp(\frac{-5.5}{H_1})\right], \qquad (1)\end{aligned}$$

where $H_1 = 0.886 + 0.0222\,V$ (km), $H_2 = 3.77$ (km) and α is some constant. The knowledge of the number of particles (cm^3)μm^{-1} radius interval and the assumption that the aerosol in the whole depth of the atmosphere is described by a Junge distribution of spherical particles enable the constant α to be calculated. For the Norwegian and Arctic Seas α is of the order of 3.67 and describes the decrease in particle density with radius in the size distribution.

Since H_1 is known in terms of V and H_2 is a known constant, it is possible to calculate the aerosol optical thickness from a single parameter V (km), the visibility range or meteorological range.

2.2. Method dependent on relative humidity

Another method is based on the model described by Fitzgerald (1989), where the aerosol optical thickness in a well-mixed marine boundary layer is estimated, taking into account relative humidity, temperature and extinction coefficient at the surface. $\tau_a(\lambda)$ can be obtained from

$$\tau_a(\lambda) = \beta_e(h_0)\int_{h_0}^{H}\left[\frac{\beta_e(h)}{\beta_e(h_0)}\right]dh, \qquad (2)$$

where h is the height of the boundary layer, $h_0 = 10$ m; it is the height of the bottom of the mixed layer, *i.e.* the height of shipboard observations. The extinction coefficient $\beta_e(h_0)$ is expressed by

$$\beta_e(h_0) = \int_0^\infty \pi r^2 Q_e n(r)dr, \qquad (3)$$

where $n(r)$ is the number of particles $(\mathrm{cm}^3)\mu\,\mathrm{m}^{-1}$ radius interval and Q_e is the extinction efficiency factor computed from Mie theory for spherical homogeneous particles. Fitzgerald (1989) recommends that the extinction coefficient at height h should be calculated from the formula

$$\beta_e(h) = \beta_e(h_0)\left[\frac{1.017 - S_0}{1.017 - S(h)}\right]^{0.84}. \tag{4}$$

The saturation ratio at height h, $S(h)$, is given by

$$S(h) = \exp\left[\frac{19.5(T_{d0} - T_0) + 0.156(h - h_0)}{269.9 + T_0 + T_{d0} - 0.0115(h - h_0)}\right]. \tag{5}$$

Eq. (5) gives the relative humidity at height h as a function of temperature T_0 and dew point T_{d0} at the surface. The saturation ratio at the surface S_0 can be obtained from

$$S_0 = \exp\left[\frac{19.5(T_{d0} - T_0)}{269.9 + T_0 + T_{d0}}\right]. \tag{6}$$

2.3. Method dependent on size distribution

One standard description of the aerosol optical thickness is the classic formula proposed by Angström. This relation has been modified in different ways. For example, Guzzi *et al.* (1987) assumed that the aerosol in the whole depth of the atmosphere is described by a Junge distribution of spherical particles, whose concentration varies as a function of height, and obtained the following formula for τ_a:

$$\tau_a(\lambda) = H_a(h_0)A\lambda^{-B}, \tag{7}$$

with $H_a(h_0) = 1 - \exp(\frac{-h}{H_p})$, where H_p is the scale height given by Penndorf (Guzzi *et al.*, 1987) ($H_p = 0.97 - 1.4$ for the first 5 km above the Earth's surface).

Taking into account experimental data obtained in 1987–1989 and 1991–1992 in the Norwegian and Arctic Seas, the exponent B (the Angström exponent) was found to be $B = 0.67$. The parameter A was assumed to be 0.125.

2.4. Method dependent on total irradiance E_{tot}

This paper presents the analytical approximation formula describing the dependence of the aerosol optical thickness on the total irradiance E_{tot} at the sea surface.

The first assumption, justified for visible wavelengths, is that the total optical thickness is the sum of the molecular optical thickness τ_r and the aerosol optical thickness τ_a. Then

$$\tau(\lambda) = \tau_a(\lambda) + \tau_r(\lambda). \tag{8}$$

According to the solar spectral model proposed by Bird (1984), the total irradiance at the sea surface is the sum of the direct and the diffuse irradiances

$$E_{tot}(\lambda) = E_s(\lambda)\cos(\theta) + E_d(\lambda), \tag{9}$$

where θ is the solar zenith angle (Krężel, 1985).

The direct clear sky irradiance can be expressed by

$$E_s(\lambda) = F_s(\lambda)\beta^{-1}T_r(\lambda)T_a(\lambda)T_w(\lambda)T_0(\lambda)T_u(\lambda), \tag{10}$$

where λ is the wavelength, $F_s(\lambda)$ is the extraterrestial spectral irradiance, and T_r, T_a, T_w, T_0 and T_u are the respective transmittance functions for Rayleigh scattering, aerosol extinction, water vapour absorption, ozone absorption and uniformly mixed gas absorption.

Taking into account Eq. (8), Eq. (10) reduces to

$$E_s(\lambda) = F_s(\lambda)\beta^{-1}T_r(\lambda)T_a(\lambda), \tag{11}$$

where β is the correction factor for the Earth-Sun distance. The equation for the scattered component on a horizontal surface at wavelength λ is (Bird, 1984)

$$E_d(\lambda) = [E_r(\lambda) + E_a(\lambda)]\,C_\lambda + E_g(\lambda), \tag{12}$$

where $E_r(\lambda)$ is the Rayleigh scattered irradiance on a horizontal surface at wavelength λ, $E_a(\lambda)$ is the aerosol scattered component on a horizontal surface at wavelength λ, $E_g(\lambda)$ is the ground/air reflected irradiance on a horizontal surface at wavelength λ, and C_λ is a correction factor that is wavelength- and zenith-angle-dependent. In order to estimate the aerosol optical thickness, we can assume that $E_d(\lambda)$ is described only by $E_r(\lambda)$. This assumption enables the 'zero approximation' of τ_a to be obtained. It can be treated as the 'first step' in determining the aerosol optical thickness as a function of total irradiance.

Taking into account the expression for $E_r(\lambda)$ (Krężel, 1985) and Eq. (8), we can write

$$E_r(\lambda) = \cos(\theta)F_s(\lambda)T_{aa}(\lambda)\beta^{-1}\left[1 - T_r^{0.95}(\lambda)\right]0.5, \tag{13}$$

where $T_{aa}(\lambda) = \exp[-(1-\omega_0(\lambda))\tau_a(\lambda)m]$, ω_0 is the single scattering albedo of the aerosol and m, according to Kasten (1966), is expressed by

$$m = \left[\cos(\theta) + 0.15(93.885 - \theta)^{-1.259}\right]^{-1}. \tag{14}$$

Combining Eqs. (9), (11) and (13) yields

$$\begin{aligned} \exp(\tau_a m)\frac{E_{tot}(\lambda)}{F_s(\lambda)\beta^{-1}\cos(\theta)} &- \exp(\omega_0\tau_a m)\left[1 - T_r^{0.95}(\lambda)\right]0.5 - \\ &+ T_r(\lambda) = 0. \end{aligned} \tag{15}$$

The approximate analytical solution for Eq. (15) can be written as

$$\tau_a(\lambda) = 0.145 \left[\frac{E_{tot}(\lambda)}{F_s(\lambda)\beta^{-1}\cos(\theta)T_r(\lambda)} - \frac{(1 - T_r^{0.95}(\lambda))}{2T_r(\lambda)} \right]^{\frac{0.945}{m}} \tag{16}$$

Eq. (16) allows the aerosol optical thickness to be estimated for the case when the only known optical parameter is the total sea surface irradiance. This relationship (16) is very useful, especially when the aerosol size distribution is unknown.

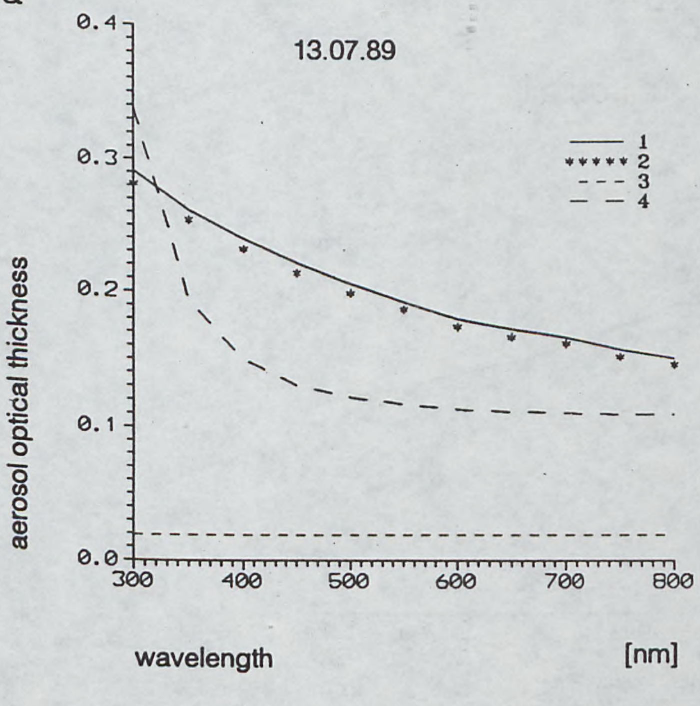
3. Results

The results are shown on the Figures.

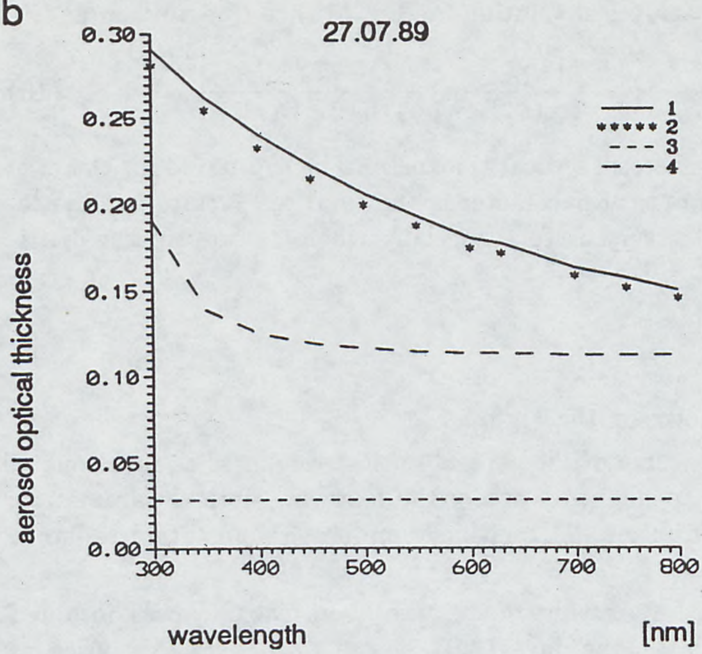
Figs. 1a–d show spectra of the aerosol optical thickness as functions of visible wavelengths, dependent on one of the following optical parameters: visibility (1), size distribution (2), relative humidity (3) and total irradiance at the sea surface (4).

In order to illustrate the results of the aerosol optical thickness obtained by four different methods, the days 13.07.89 and 27.07.89 were chosen as examples for the period 1987–1989 and the days 24.06.91 and 13.07.91 as examples for the period 1991–1992.

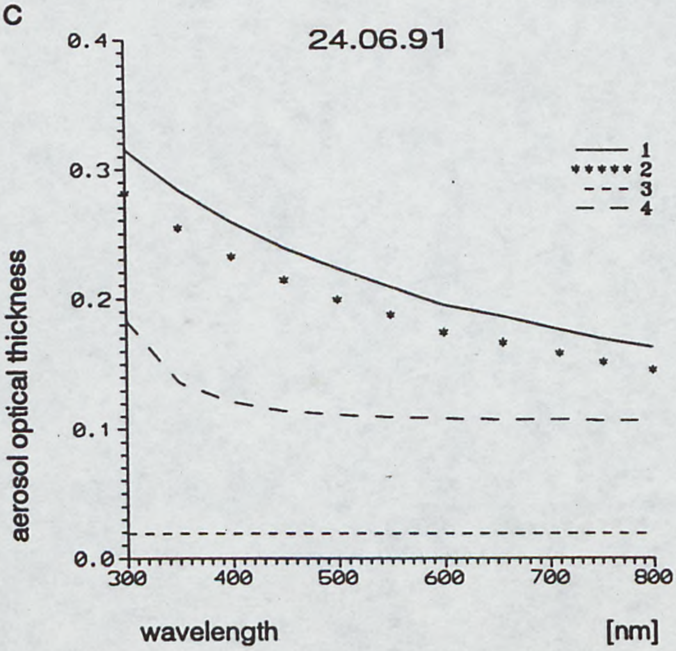
a



b


27.07.89
aerosol optical thickness
0.30
0.25
0.20
0.15
0.10
0.05
0.00
300
400
500
600
700
800
wavelength
[nm]
1
2
3
4

c


24.06.91
aerosol optical thickness
0.4
0.3
0.2
0.1
0.0
300
400
500
600
700
800
wavelength
[nm]
1
2
3
4

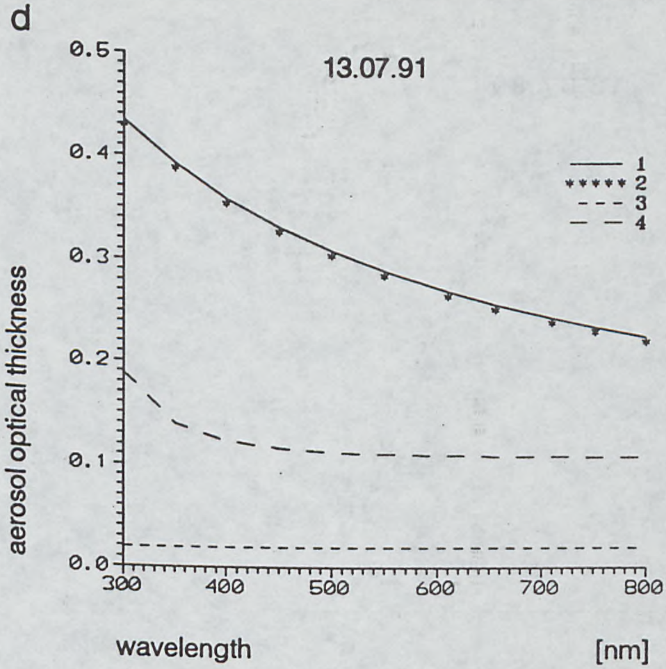
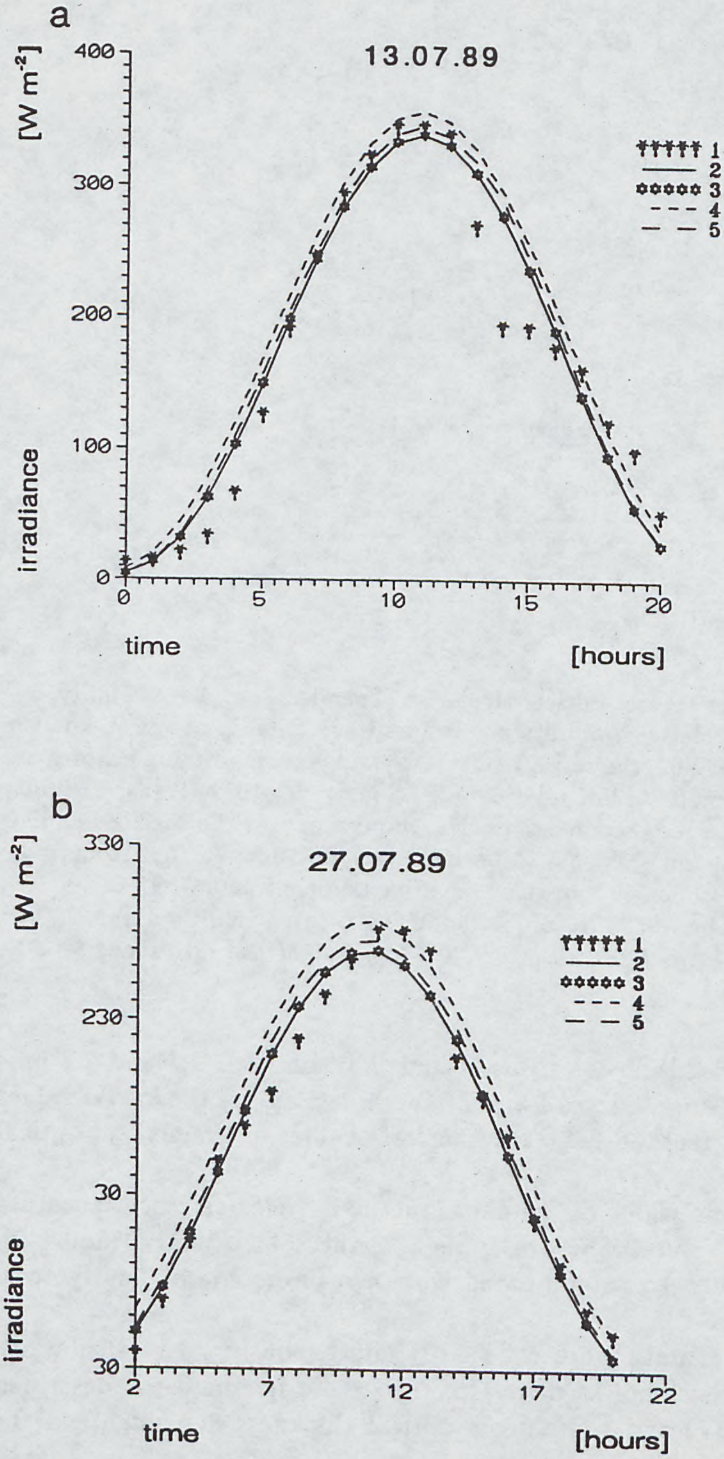


Fig. 1. Spectra of the aerosol optical thickness dependent on: 1 – visibility, 2 – size distribution, 3 – relative humidity, 4 – irradiance. Visibility about 50 km, the Angström exponent = 0.67, the respective surface and dew point temperatures are $T_0 = 8.4(°C)$, $T_{d0} = 6.0(°C)$, latitude: 69°12'N, longitude: 10°24'E (a); Visibility about 50 km, the Angström exponent = 0.67, temperatures at the surface and dew point are $T_0 = 5.1(°C)$ and $T_{d0} = 3.3(°C)$ respectively, latitude: 75°07'N, longitude: 14°5'E (b); Visibility about 40 km, the Angström exponent = 0.67, $T_0 = 6.9(°C)$, $T_{d0} = 5.0(°C)$, latitude: 73°03'N, longitude: 15°04'E (c); Visibility about 20 km, the Angström exponent = 0.67, $T_0 = 9.9(°C)$, $T_{d0} = 9.3(°C)$, latitude: 68°03'N, longitude: 00°04'E (d)

As a general conclusion it can be stated that methods 2.1 and 2.3 produce very similar results (curves 1 and 2), method 2.2 gives the lowest values of τ_a (curve 3) and method 2.4 – intermediate values of the aerosol optical thickness (curve 4).

In order to decide which of the above methods yields the most accurate results, calculated values of τ_a are applied to the solar spectral model for irradiance (Bird, 1984) and compared with measurements of hourly total irradiance (Figs. 2a–d).

All experimental data were collected simultaneously. All calculations were carried out for cloudless days. In Figs. 2a–d irradiance is described by measurements (1) and the aerosol optical thickness dependent on the


a
13.07.89
[W m^{-2}]
irradiance
400
300
200
100
0
0
5
10
15
20
time
[hours]
1
2
3
4
5
b
27.07.89
[W m^{-2}]
irradiance
330
230
30
30
2
7
12
17
22
time
[hours]
1
2
3
4
5

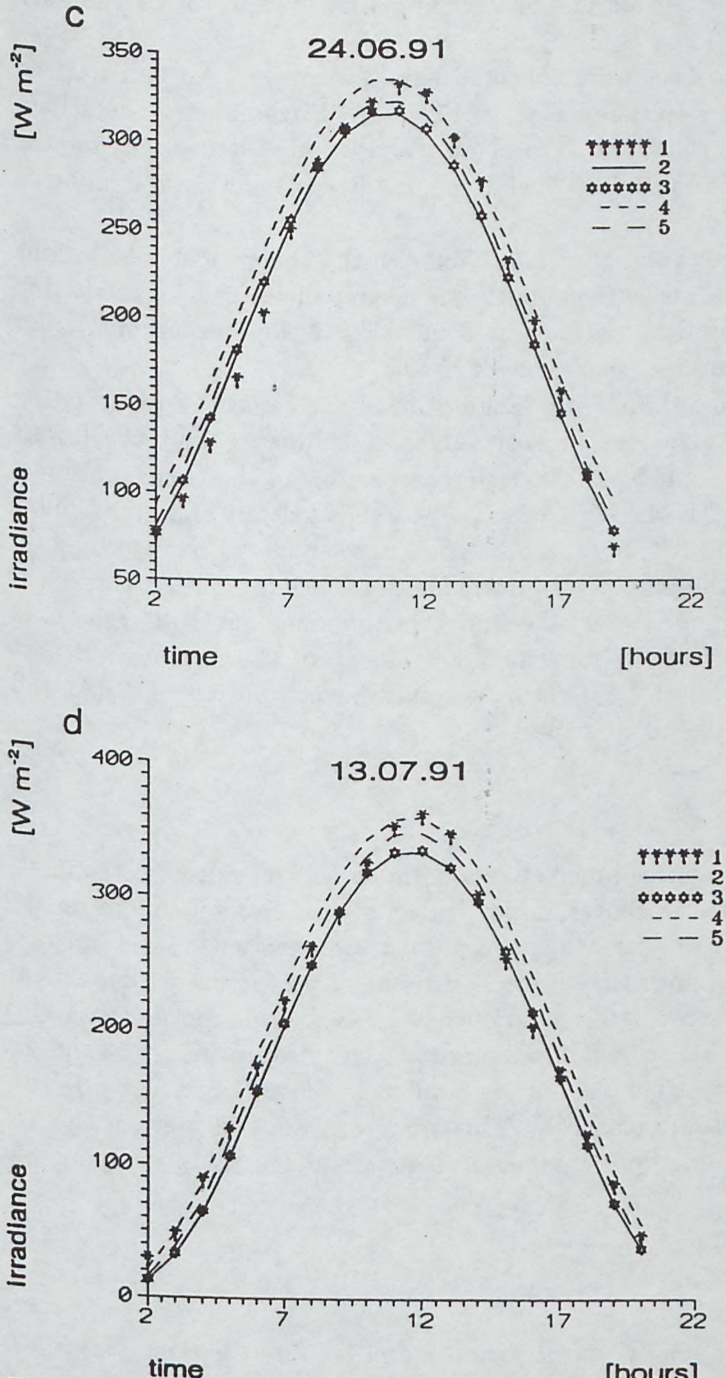


Fig. 2. Hourly total irradiance described by measurements (1) and the aerosol optical thickness dependent on: visibility (2), size distribution (3), relative humidity (4), irradiance at the sea surface (5)

visibility (2), size distribution (3), relative humidity (4) and total irradiance at the sea surface (5).

All experimental data were collected simultaneously. All calculations were carried out for cloudless days. In Figs. 2a–d irradiance is described by measurements (1) and the aerosol optical thickness dependent on the visibility (2), size distribution (3), relative humidity (4) and total irradiance at the sea surface (5).

In general, there is good agreement between the theoretical calculations and the experimental measurements. Some discrepancies can be attributed not only to the theoretical methods used but also to the measurements, for example, the momentary appearance of clouds.

Nevertheless, some simplifying assumptions have caused errors. For example, method 2.2 gives the highest values of irradiance and the lowest values of the aerosol optical thickness in comparison to the other methods. Therefore this model is the least precise. Methods 2.1 and 2.3 give similar results (curves 2 and 3). Both are in good agreement with experimental data (curve 1). Method 2.4 gives intermediate results (curve 5). Curve (5) is a mean between models (2.1, 2.3) and 2.2. Some negligible discrepancies between curves (1) and (5) may have arisen out of the assumption that $E_d(\lambda)$ is described only by $E_r(\lambda)$. A more precise estimation of $E_d(\lambda)$ will be presented in a subsequent paper.

4. Conclusions

The methods presented in this paper seem to be well suited for estimating the aerosol optical thickness, in particular, the models dependent on the visibility, aerosol size distribution and total irradiance at the sea surface. The error associated with the determination of τ_a by the model dependent on relative humidity is relatively large; nevertheless, this method can also be used when the other optical parameters (visibility or irradiance at the sea surface) are unknown. The results are similar to those obtained by Shifrin (1992) for the Mediterranean Sea. This work was done in view of the extensive use of these methods by meteorologists and scientists studying the atmosphere.

References

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

Fitzgerald J. W., 1989, *Model of the aerosol extinction profile in a well-mixed marine boundary layer*, Appl. Opt., 28, 15, 3534–3538.

Guzzi R., Rizzi R., Zibordi G., 1987, *Atmospheric correction of data measured by a flying platform over the sea: elements of a model and its experimental validation*, Appl. Opt., 26, 15, 3043–3051.

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

Krężel A., 1985, *Solar radiation at the Baltic Sea surface*, Oceanologia, 21, 5–32.

Shifrin K. S., 1992, *Optical properties of the atmosphere over the ocean*, SPIE, Optics of the Air-Sea Interface, 1749, 151–161.

Sturm B., 1981, *The atmospheric correction of remotely sensed data and the quantitative determination of suspended matter in marine water surface layers*, [in:] *Remote Sensing in Meteorology, Oceanography and Hydrology*, A. P. Cracknell (ed.), Ellis Horwood Limited, Chichester, 163–197.