Papers

Finding the aerosol optical thickness over the Baltic Sea – comparison of two methods*

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Abstract

The results of two methods used to estimate the aerosol optical thickness over the Baltic Sea are compared. The standard method is based on measurements of the direct component of the downward irradiance at the sea surface in 8 spectral bands (412, 443, 490, 510, 555, 670, 765, 865 nm – the same as SeaWiFS). In the pyranometric method, Baltic aerosols are assumed to be a mixture of model aerosol types with strictly defined optical properties, *i.e.* maritime, continental and stratospheric types. Their proportion in the Baltic aerosol is found from broadband spectral downward irradiance measurements (VIS, IR) using the radiative transfer model. Simultaneous measurements of the spectral downward irradiance and its direct component on cloudless days in the southern Baltic were used in the comparison. The pyranometric method of estimating the aerosol optical thickness proved to be a satisfactory tool. Depending on the wavelength, the statistical errors in it are not in excess of ± 0.06 to ± 0.08.

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1. Introduction

Aerosols are an important component of the atmosphere. First of all, they interact with visible and infrared solar radiation; secondly, they participate in cloud processes; and lastly, sea-salt particles are chemical carriers of species containing Cl, Br, I and S. Aerosols therefore play a significant role in the earth's energy budget and climate (Gong *et al.*, 1997b; Coakley *et al.*, 1983).

Correct parametrisation of aerosol optical properties is important in modelling the radiation fluxes in the atmosphere and at the earth's surface, and also for the atmospheric correction. The aerosol optical thickness, the single scattering albedo and the phase function are the fundamental parameters describing aerosol optical properties. Methods for determining the optical characteristics of aerosols have been given in a number of papers (*e.g.* Bokoye *et al.*, 1997; Tarasova *et al.*, 1992; Weller and Leiterer, 1988). The aerosol optical thickness is the parameter most often assessed. It routinely is determined on the basis of narrow-band spectral measurements of the direct component of the irradiance, or alternatively, the total downward irradiance and its diffuse component (Amato *et al.*, 1995; Smirnov *et al.*, 1994; Weller and Leiterer, 1988; Villevalde *et al.*, 1994). Spectral channels that lie outside any significant absorption bands are usually employed.

The EOF (Empirical Orthogonal Functions) method, as applied to aerosol optical thickness spectra determined from narrow-band spectral measurements of the direct component of the downward irradiance over the southern Baltic, showed that the first mode accounted for 95% of the total variance in the aerosol optical thickness (Kuśmierczyk-Michulec and Darecki, 1996; Kuśmierczyk-Michulec *et al.*, in press). The shapes of the mean aerosol optical thickness spectra and the first mode obtained by these authors are in good agreement with those of the optical thickness spectra for continental and maritime model aerosols given in McClatchey *et al.* (1984). This provided the incentive to study the possibility of utilising our extensive database of broadband spectral pyranometer measurements from the Baltic Sea area to retrieve the aerosol optical thickness. The 'pyranometric method' was developed as a result.

The present paper compares the pyranometric method of estimating the aerosol optical thickness over the Baltic Sea with the standard technique. In the pyranometric method, Baltic aerosols are assumed to be a mixture of model aerosol types with strictly defined optical properties, *i.e.* stratospheric, maritime and continental types. Their proportion in the Baltic aerosol and the total aerosol optical thickness for 550 nm are found from the broadband spectral downward irradiance measurements (VIS, IR)

using the radiative transfer model. The standard method is based on measurements of the direct component of solar irradiance at the sea surface in 8 spectral bands (412, 443, 490, 510, 555, 670, 765, 865 nm) (Olszewski *et al.*, 1995). Simultaneous measurements of the downward spectral irradiance and its direct component carried out during cloudless days in the southern Baltic were used in the comparison.

2. Measurements

The comparison was made on cloudless days during several cruises to the Baltic (May and August 1994, May and September 1997, May 1998) and on the beach at Sopot (March 1998). The locations of the measurement sites and the meteorological conditions obtaining during the measurements are listed in Tab. 1.

Broad-band downward irradiance was measured by means of two Eppley Precision Spectral Pyranometers fitted with hemispherical Schott filters GG395 and RG695 (the numbers denote the cut-off wavelengths in nm). The pyranometers were calibrated once a year against a Kipp and Zonen CM5 reference pyranometer calibrated by the manufacturer. The measurement error, comprising all error sources of the actinometric measurements on the ship, did not exceed ± 3–5%. The 5-minute means of irradiance were used in calculations.

A shadow-band spectrophotometer was used for the simultaneous measurements of the total and the diffuse solar spectral irradiances in eight spectral channels (each channel of ± 10 nm width): 412, 490, 510, 555, 670, 765 and 865 nm (the same range of wavelengths as the SeaWiFS radiometer uses). The instrument was calibrated against a Multiwavelength Environmental Radiometer MER2041 (Biospherical Instruments Inc.). The measurement precision was estimated at ± 2–3% for the total irradiance and ± 6–9% for the diffuse irradiance. Irradiances averaged over 1 minute were used in the calculations. In 1994, measurements were made with a similar instrument but using slightly different spectral channels (400, 443, 490, 520, 550, 620 and 670 nm). The spectral irradiances for the 412, 510 and 555 nm channels were interpolated. The measurement method is described in detail in Olszewski *et al.*, 1995.

Only measurements with the solar zenith angle $\vartheta < 70^\circ$ were used in the comparison. At the same time, the standard meteorological observations were made. The atmospheric ozone amount, necessary in both methods, was not measured during the cruises. Instead, the long-term monthly means for Belsk were used in this comparison (Dziewulska-Łosiowa, 1991).

Table 1. Coordinates of the measurement sites and meteorological conditions during the measurements

Date	Latitude [°N]	Longitude [°E]	Air temperature [°C]	Relative humidity	Air pressure [hPa]	Wind speed [m s^{-1}]	Wind direction [°]
11 May 1994	54.6	14.4–14.5	10.6–12.6	0.82–0.88	1015–1017	4–5	90
22 August 1994	54.8–55.1	16.1–16.2	17.0–17.1	0.75–0.85	1017–1018	4–5	260–280
23 August 1994	54.6	16.3–16.4	17.0–17.5	0.81–0.82	1018	1–2	270–340
24 August 1994	54.7	16.9	14.2	0.96	1015	1	110
24 May 1997	54.8–54.9	19.3	7.7–8.6	0.61–0.71	1023–1024	6–7	10
25 May 1997	55.0–55.2	17.1–17.5	7.0–7.9	0.86–0.92	1027	3–5	230–280
1 September 1997	54.1–54.3	13.9–14.1	21.5	0.66–0.73	1019	3–5	90–120
5 September 1997	54.1–54.5	14.6–15.8	18.7–20.9	0.71–0.85	1017–1018	2–4	180–250
25 March 1998	54.4	18.6	1.7–3.1	0.78–0.84	1040–1042	–5	40–60
10 May 1998	55.3	18.0	13.4–14.6	0.82–0.88	1020–1022	0.5–1	140–160

3. Methods

In the standard method the aerosol optical thickness $\tau_{aer}(\lambda)$ is obtained from the following expression:

$$\tau_{aer}(\lambda) = m^{-1} \ln \left(T_{aer,\,dir}(\lambda)\right)^{-1}, \tag{1}$$

where

m – atmospheric optical mass (by Kasten, 1966),

$T_{aer,\,dir}(\lambda)$ – transmittance function for aerosols:

$$T_{aer,\,dir}(\lambda) = \frac{E^{emp}_{d,\,tot}(\lambda) - E^{emp}_{d,\,dif}(\lambda)}{S_{\oplus,\,\lambda}\ f_S\ T_R(\lambda,p)\ T_{o3}(\lambda,O_3)\ T_w(\lambda,w(e),p)\ T_g(\lambda,p)\cos\vartheta}, \tag{2}$$

$S_{\oplus,\,\lambda}$ – spectral distribution of the mean extraterrestrial irradiance (Neckel and Labs, 1981; from Bird and Riordan, 1986),

f_S – extraterrestrial irradiance correction for the Earth-Sun distance (Gordon *at al.*, 1983; Paltridge and Platt, 1976),

ϑ – solar zenith angle,

$E^{emp}_{d,\,tot}(\lambda)$, $E^{emp}_{d,\,dif}(\lambda)$ – the respective total and diffuse narrow-band downward irradiances measured at the sea surface,

$T_R(\lambda,\, p)$, $T_{O_3}(\lambda,\, O_3)$, $T_w(\lambda,\, w(e),\, p)$, $T_g(\lambda,\, p)$ – the respective transmittance functions for Rayleigh scattering (Van Stokkom and Guzzi, 1984), ozone, water vapour and uniformly mixed gas absorptions (Leckner, 1978),

O_3 – atmospheric ozone amount [atm–cm],

$w(e)$ – precipitable water vapour in a vertical path [cm],

e – surface water vapour pressure [hPa],

p – atmospheric pressure [hPa].

In the pyranometric method the real southern Baltic aerosol is assumed to consist of the model aerosols by McClatchey *et al.* (1984), which have strictly defined optical properties (absorption c and scattering b coefficients, asymmetry parameter of the scattering function g). The extinction coefficients given in McClatchey *et al.* (1984) are normalised to 1 km^{-1} at $\lambda = 550$ nm (Fig. 1).

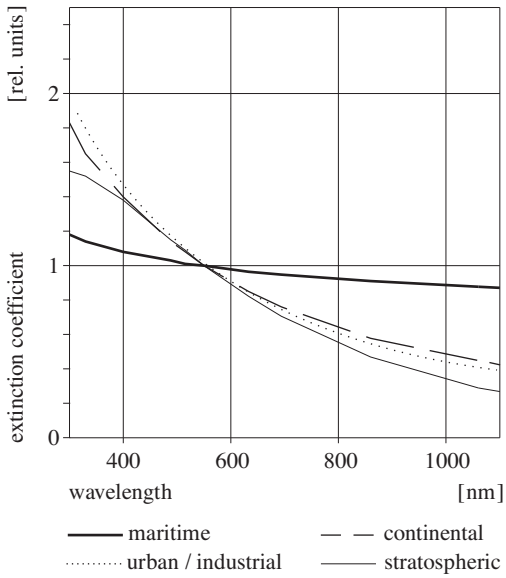


Fig. 1. Extinction coefficients of model aerosols by McClatchey *et al.* (1984) vs wavelength, normalised to $c(550)$

Then, the total aerosol optical thickness is defined as

$$\tau_{aer}(\lambda) = \sum_{i=1}^{3} \tau_{aer,\,i}(550) \frac{c_{aer,\,i}(\lambda)}{c_{aer,\,i}(550)}, \tag{3}$$

where

i – model aerosol type by McClatchey *et al.* (1984): 1 – maritime, 2 – continental, 3 – stratospheric.

The optical thickness of the stratospheric aerosol is assumed constant $\tau_{aer,\,3}(550) = 0.0047$.

The optical thickness of the continental and maritime components can be found from the numerical solution of the following set of equations:

$$\begin{aligned} E^{emp}_{d,\,tot}(VIS) &= E^{emp}_{d,\,tot}(\mathrm{GG395}) - E^{emp}_{d,\,tot}(\mathrm{RG695}) = \\ &= \int_{295}^{2800} (f_{\mathrm{GG395}}(550) - f_{\mathrm{RG695}}(550)) \times \\ &\times\ E^{\mathrm{mod}}_{d,\,tot}(\lambda, \tau_{aer,\,1}(550), \tau_{aer,\,2}(550)) d\lambda, \\ E^{emp}_{d,\,tot}(IR) &= E^{emp}_{d,\,tot}(\mathrm{RG695}) = \int_{295}^{2800} f_{\mathrm{RG695}}(550) \times \\ &\times\ E^{\mathrm{mod}}_{d,\,tot}(\lambda, \tau_{aer,\,1}(550),\ \tau_{aer,\,2}(550)) d\lambda, \end{aligned} \tag{4}$$

where

$E^{emp}_{d,\,tot}$ – measured broad-band total downward irradiance at the sea surface,

f_{GG395}, f_{RG695} – Schott filter transmittance functions,

$E^{\mathrm{mod}}_{d,\,tot}$ – modelled total downward irradiance:

$$E^{\mathrm{mod}}_{d,\,tot}(\lambda) = S_{\oplus,\,\lambda}\; f_S\; \cos\vartheta\; T_{atm}(\lambda, \vartheta, p, w(e), T, O_3, \tau_{aer,\,1}(550), \tau_{aer,\,2}(550)), \tag{5}$$

T_{atm} – irradiance transmittance for a homogenous atmosphere computed using the δ approximation of the scattering function and single scattering radiative transfer model (quasi-single scattering model) (Woźniak *et al.*, in press),

T – air temperature.

When the point $(E^{emp}_{d,\,tot}(VIS),\, E^{emp}_{d,\,tot}(IR),\, w(e))$ lies outside the model space limits, the aerosols are assumed to be continental if $\frac{\tau_{aer,\,1}}{\tau_{aer,\,1}+\tau_{aer,\,2}} < 0$ and maritime if $\frac{\tau_{aer,\,1}}{\tau_{aer,\,1}+\tau_{aer,\,2}} > 1$. Then, $\tau_{aer,\,1+2}$ is found from the relationship

$$E^{emp}_{d,\,tot}(VIS+IR) = E^{emp}_{d,\,tot}(\mathrm{GG395}) = \int_{295}^{2800} f_{\mathrm{GG395}}(550) \times \times\, E^{\mathrm{mod}}_{d,\,tot}(\lambda, \tau_{aer,\,1+2}(550))\, d\lambda. \tag{6}$$

4. Results and discussion

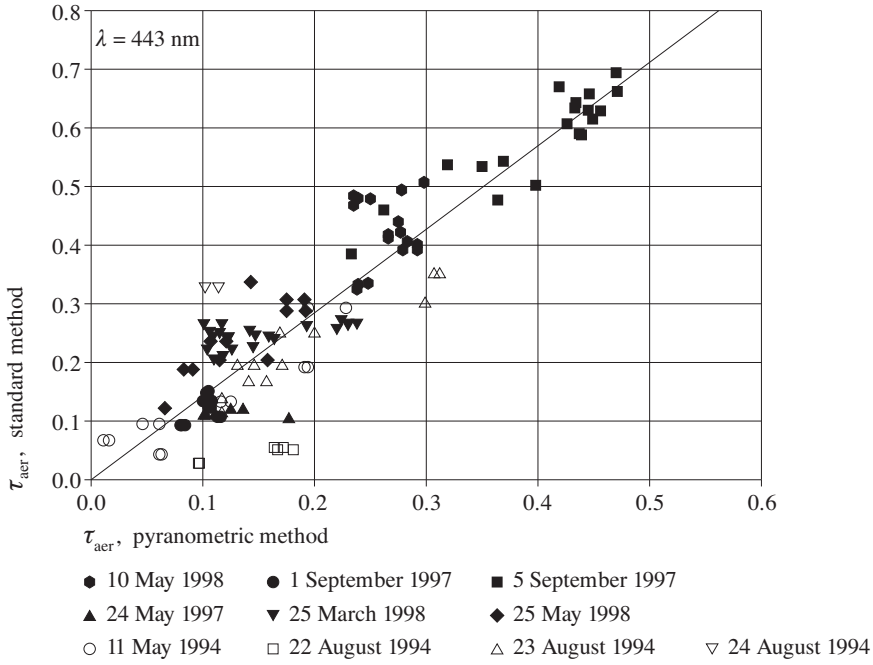
Comparison of zenithal aerosol optical thicknesses obtained by the standard and pyranometric methods yielded a linear relationship

$$\tau^{dir}_{aer}(\lambda) = a_1(\lambda)\, \tau^{pyr}_{aer}(\lambda) + a_0(\lambda). \tag{7}$$

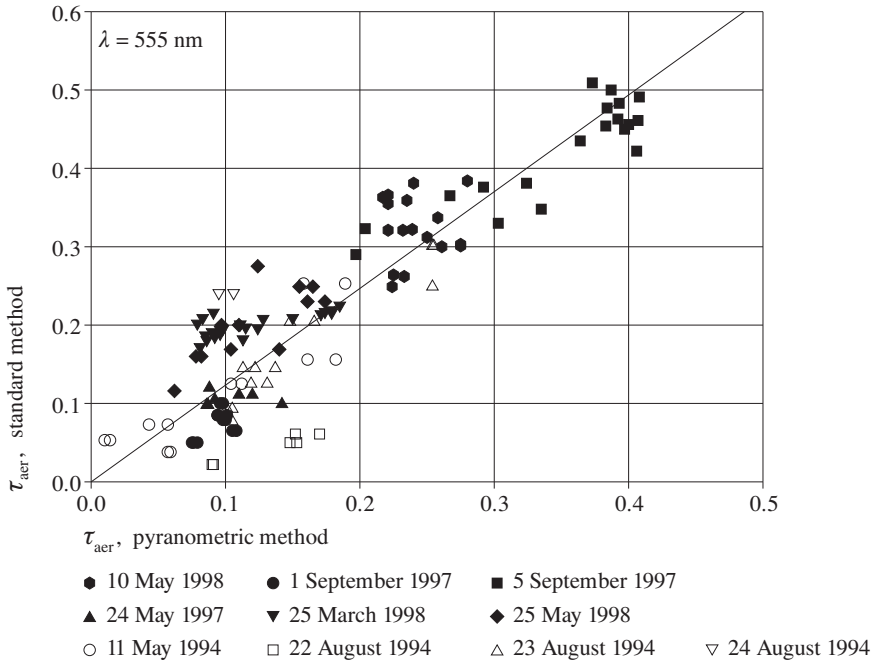
Some examples for selected wavelengths are shown in Figs. 2a, 2b and 2c. In the linear approximation, $a_0(\lambda) = 0$ was assumed. The regression coefficients obtained for the spectrophotometer spectral bands and the linear correlation coefficients are given in Tab. 2. The dependence of a_1 on the wavelength in the 412–865 nm range can be approximated by the polynomial (Fig. 3)

$$a_1(\lambda) = 2.339 - 2.713 \times 10^{-3}\, \lambda + 1.358 \times 10^{-6}\, \lambda^2. \tag{8}$$

a


λ = 443 nm
τaer, standard method
τaer, pyranometric method
0.0
0.1
0.2
0.3
0.4
0.5
0.6
0.7
0.8
10 May 1998
1 September 1997
5 September 1997
24 May 1997
25 March 1998
25 May 1998
11 May 1994
22 August 1994
23 August 1994
24 August 1994

b


λ = 555 nm
τaer, standard method
τaer, pyranometric method
0.0
0.1
0.2
0.3
0.4
0.5
0.6
10 May 1998
1 September 1997
5 September 1997
24 May 1997
25 March 1998
25 May 1998
11 May 1994
22 August 1994
23 August 1994
24 August 1994

c

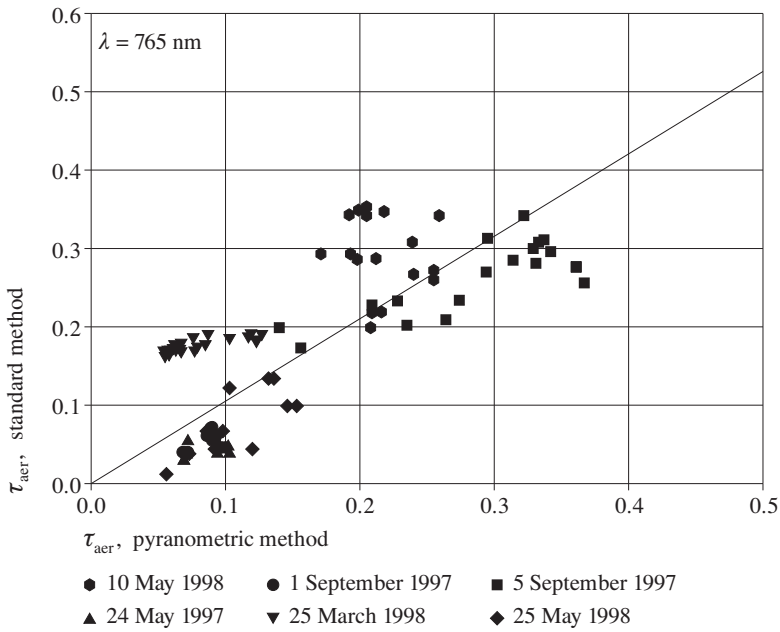


Fig. 2. Comparison of the zenithal aerosol optical thicknesses at $\lambda = 443$ nm (a), $\lambda = 555$ nm (b), $\lambda = 765$ nm (c), obtained by the standard and pyranometric methods. Measurement points from one day are denoted by the same symbol

Table 2. Coefficient $a_1(\lambda)$ of the relation between the zenithal aerosol optical thicknesses obtained by the standard and pyranometric methods (eq. (7)) for various spectrophotometer channels, and the respective linear correlation coefficients

Wavelength [nm]	a_1	Correlation coefficient
412	1.446	0.824
443	1.424	0.828
490	1.336	0.829
510	1.296	0.816
555	1.234	0.805
670	1.154	0.688
765	1.052	0.566
865	1.007	0.522

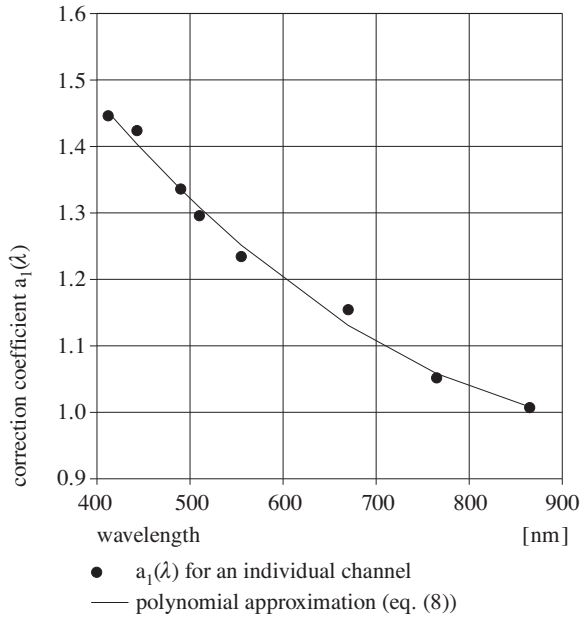


Fig. 3. Linear regression coefficient $a_1(\lambda)$ of the eq. (7) (correction coefficient for the pyranometric method) vs wavelength and a polynomial approximation of the relation

The values of $a_1(\lambda) \neq 1$ indicate systematic discrepancies between the methods being compared. Given that the standard method is precise, that is to say systematic errors in the standard method are negligible or at least much lower than those in the pyranometric technique, the latter method systematically underestimates aerosol optical thickness, especially in the short-wave part of the spectrum. Model assumptions and parametrisations appear to be the main sources of the discrepancies, which are discussed further in detail. Bearing the above in mind, eqs. (7) and (8) (or eq. (7) with coefficient a_1 from Tab. 2) may serve as corrections for the aerosol optical thickness obtained by the pyranometric method.

Having applied the corrections, the uncertainties in the pyranometric method with respect to the direct method were calculated. The discrepancies are expressed as systematic (bias) and statistical (random) errors defined as follows:

- systematic error:

$$e_{s,\,pyr/dir} = \frac{1}{\mathrm{N}} \sum_{i=1}^{N} \varepsilon_{i,\,pyr/dir}, \tag{9}$$

where

$$\varepsilon_{pyr/dir} = \tau_{aer}^{pyr}(\lambda) - \tau_{aer}^{dir}(\lambda), \tag{10}$$

$\tau_{aer}^{dir}(\lambda)$ – zenithal aerosol optical thickness obtained by the standard method,

$\tau_{aer}^{pyr}(\lambda)$ – zenithal aerosol optical thickness obtained by the pyranometric method,

- statistical error:

$$e_{st,\,pyr/dir} = \frac{1}{\mathrm{N}} \sum_{i=1}^{N} (\varepsilon_{i,\,pyr/dir} - e_{s,\,pyr/dir})^2. \tag{11}$$

The errors are set out in Tab. 3. As a result of applying the correction relationship, the systematic errors are very low, one order of magnitude lower than the corresponding statistical errors, for most of the spectral bands being analysed. Application of the correction relationship (7) without the assumption that $a_0(\lambda) = 0$ would considerably reduce the systematic errors calculated for the given data sets (down to –0.00001), but the values of $a_0(\lambda)$ thereby obtained would then increase with λ, which would distort the spectral shape of the corrected aerosol optical thickness. Moreover, for the last 3 channels $\tau_{aer}(\lambda)$ values lower than $a_0(\lambda)$ thus obtained (rejecting the assumption that $a_0(\lambda) = 0$ and for the given data set) are quite probable in the case of northerly winds. Therefore the assumption that $a_0(\lambda) = 0$ seems justified, even though it results in slightly higher systematic errors. The statistical errors vary from 0.056 for $\lambda = 555$ nm to 0.077 for $\lambda = 412$ nm.

Table 3. Systematic and statistical errors in the pyranometric method of finding the aerosol optical thickness calculated against the standard method

Wavelength [nm]	Systematic error	Statistical error
412	–0.002	0.077
443	–0.002	0.072
490	–0.002	0.063
510	–0.004	0.061
555	–0.005	0.057
670	–0.011	0.061
765	–0.013	0.070
865	–0.013	0.069

In the standard method, however, there is also an inherent random error. It can be proven that the real statistical error of the pyranometric method, *i.e.* calculated with respect to the 'real' aerosol optical thickness values instead of the measured ones, can be expressed by the following relation:

$$e^2_{st,\,pyr} = e^2_{st,\,pyr/dir} - e^2_{st,\,dir}, \tag{12}$$

where

$$e_{st,\,pyr} = \frac{1}{\mathrm{N}} \sum_{i=1}^{N} (\varepsilon_{i,\,pyr} - e_{s,\,pyr})^2, \tag{13}$$

$$e_{s,\,pyr} = \frac{1}{\mathrm{N}} \sum_{i=1}^{N} \varepsilon_{i,\,pyr}, \tag{14}$$

$$\varepsilon_{pyr} = \tau^{pyr}_{aer}\,(\lambda) - \tau^{true}_{aer}\,(\lambda), \tag{15}$$

τ^{true}_{aer} – true (ideal) zenithal aerosol optical thickness

$$e_{st,\,dir} = \frac{1}{\mathrm{N}} \sum_{i=1}^{N} (\varepsilon_{i,\,dir} - e_{s,\,dir})^2, \tag{16}$$

$$e_{s,\,dir} = \frac{1}{\mathrm{N}} \sum_{i=1}^{N} \varepsilon_{i,\,dir}, \tag{17}$$

$$\varepsilon_{dir} = \tau^{dir}_{aer}\,(\lambda) - \tau^{true}_{aer}\,(\lambda). \tag{18}$$

Because the random uncertainty in the aerosol optical thickness obtained by the standard method is never equal to zero, the following relation is true:

$$(e_{st,\,pyr})^2 < (e_{st,\,pyr/dir})^2. \tag{19}$$

Hence, the values of the error given in Tab. 3 may serve as the 'upper limits' of the statistical errors inherent in the pyranometric method.

Each of the methods under discussion is subject to uncertainties specific to them.

In the case of the standard method, the uncertainties in the optical thickness stem mainly from errors in the total and diffuse irradiance measurements. These, in turn, result from the calibration uncertainty, estimated at $\pm 2\%$. Moreover, the shadow-band method of screening the Sun also contributes to the uncertainty in the diffuse irradiance measurements. A detailed discussion of the errors inherent in the diffuse and total irradiance measurements will be found in Olszewski *et al.* (1995). The total statistical error in the zenithal aerosol optical thickness obtained by the standard method is estimated at about $\pm 0.02 - \pm 0.04$, whereas the systematic errors

are estimated at about 0.03 for the 'blue' channels and <0.01 for the red and infrared channels.

In the pyranometric method there are two sources of error, *i.e.* derived from the accuracy of the model, and associated with the measurement precision. The single scattering assumption used in the model may result in the underestimation of the VIS radiation by the model. A rough estimate shows that respective direct and single scattered radiation amounts to about 92, 97 and 99% of the total for the wavelengths 400, 500 and 600 nm reaching the sea surface after having travelled through an atmosphere similar to that over the Baltic (Woźniak, personal communication). Another assumption that may bias the aerosol optical thickness found is that the atmosphere is uniformly mixed. The real atmosphere is non-homogenous, with the tropospheric aerosols generally confined to the boundary layer. In a non-absorbing atmosphere, the flux transmissivities for diffuse radiation are independent of the vertical structure of a non-homogenous plane parallel atmosphere. However, this is not true when the atmosphere absorbs radiation (Coakley *et al.*, 1983). This assumption may then lead to errors in calculating the aerosol optical thickness. The errors discussed in this paragraph are of a systematic nature rather than statistical.

Further sources of systematic and statistical errors in the aerosol optical thickness estimation are the uncertainties in the parametrisations of the solar radiation attenuation by atmospheric constituents, including the uncertainties in the parametrisation of atmospheric absorber and scatterer amounts, mainly that of water vapour. Estimation of the precipitable water vapour content in the vertical path in the atmosphere w is based on surface measurements. In the model, the formula from Timofeyev (1983) formula designed for the marine atmosphere was used. The comparison of this formula with radiosonde measurements taken in summer 1997 over the Baltic suggested that the random error of the formula was about $\pm 25\%$. The influence of the uncertainty in estimating the precipitable water vapour on the retrieval error was estimated by means of the model for $\vartheta = 45^\circ$ (Fig. 4). The aerosol optical thickness is sensitive to the error in precipitable water vapour estimation mainly in the range of low w. For instance, the change in the value of w from 1 to 1.6 results in a 30–40% decrease in $\tau_{aer,\,1+2}(550)$, whereas a similar change from 2.8 to 3.4 results in a 5–30% decrease, depending on the value of $\tau_{aer,\,1+2}(550)$. Another important source of the retrieval error is the assumption concerning the model properties of the aerosol. Model aerosols have fixed composition and size distributions. In nature, composition and size distribution depend on aerosol sources and history, changing as a result of transport processes (turbulent diffusion, advection, vertical convection of the atmosphere),

dry and wet deposition, chemical and physical transformation, and coagulation. Moreover, the particle sizes of hygroscopic components such as sea salt aerosols increase with the ambient relative humidity (Gong *et al.*, 1997a).

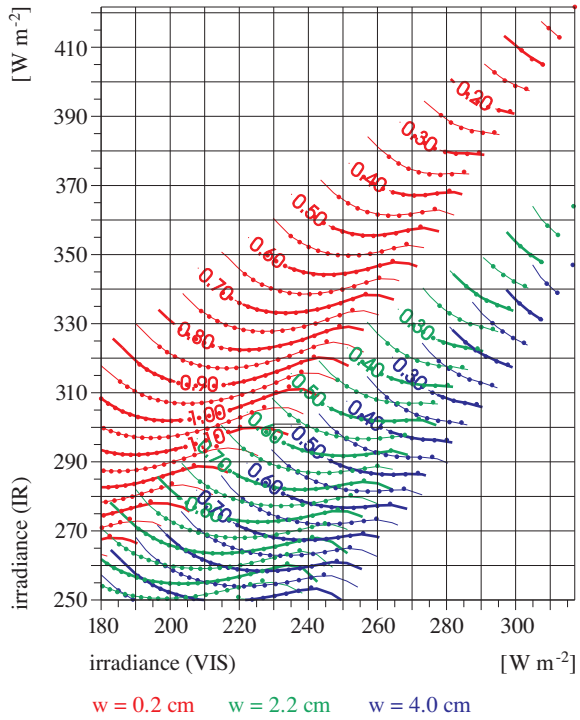


Fig. 4. The dependence of the zenithal aerosol thickness $\tau_{aer,\,1+2}(550)$ (isolines) found by the pyranometric method on the input parameters: the VIS and IR irradiance, and the precipitable water vapour in the zenithal path w. Computations for $\vartheta = 45°$

The sensitivity of the values of $\tau_{aer}(550)$ obtained to the precision of the irradiance measurements was tested for $\vartheta = 45°$ (Fig. 4). In the 0.1–0.2 range of $\tau_{aer,\,1+2}(550)$ and $w = 1$ cm, a decrease in IR of about 3% results in an increase in $\tau_{aer,\,1+2}(550)$ of about 36%, whereas a similar increase in IR causes changes in $\tau_{aer,\,1+2}(550)$ of –21%. Analogous changes in VIS result in the respective shifts in $\tau_{aer,\,1+2}(550)$ of +14% and –17%. In the case of a large aerosol optical thickness (about 0.5), the $\tau_{aer,\,1+2}(550)$ found is less sensitive to the uncertainties in the irradiance measurements. A change in VIS irradiance of about $\pm 4\%$ results in $\mp 1 - \mp 6\%$ changes in the aerosol optical thickness. The influence of the IR is higher and amounts to $\mp 15 - \mp 20\%$ for an irradiance change of about $\pm 3\%$. As far as the

parameter $\frac{\tau_{aer,1}}{\tau_{aer,1}+\tau_{aer,2}}$ that controls the spectral dependence of the found aerosol optical thickness is concerned, changes in VIS or IR irradiance of $\pm 3 - \pm 4\%$ result in a parameter change of from ± 50 to over $\pm 100\%$, depending on the values of $\tau_{aer,1+2}(550)$ and w (Fig. 5). An increase in VIS raises the value of the parameter, whereas a rise in IR acts in the opposite direction. The uncertainty in the irradiance measurements consists of the calibration error and a number of random components. In the case of long-term measurements taken with different instruments (or the same set of pyranometers but with different calibrations), the calibration uncertainty can be treated as random rather than systematic.

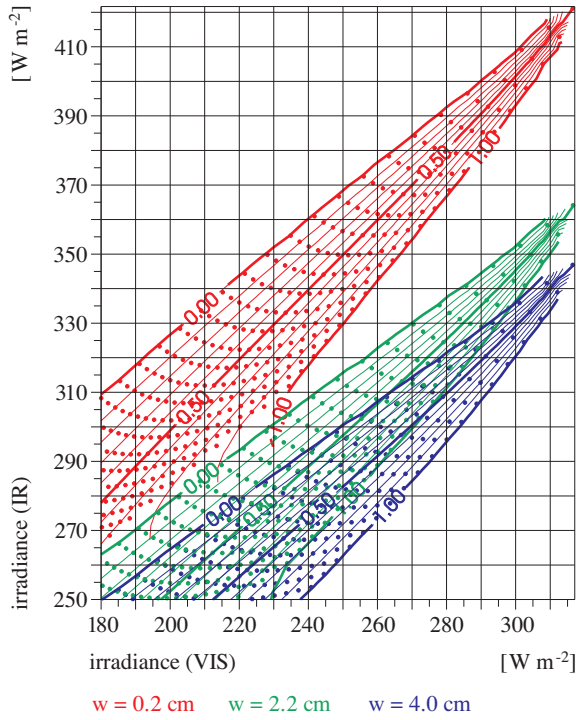


Fig. 5. The dependence of the parameter $\frac{\tau_{aer,1}}{\tau_{aer,1}+\tau_{aer,2}}$ (isolines) found by the pyranometric method on the input parameters: the VIS and IR irradiance, and the precipitable water vapour in the zenithal path w. Computations for $\vartheta = 45°$

Assuming that uncertainties in the estimation of precipitable water vapour and irradiance measurements constitute the main sources of the random uncertainty in the estimation of optical thickness by the pyranometric method, the statistical (random) errors for $\tau_{aer,1+2}(550)$ vary from ± 0.06 to ± 0.12 which, expressed as relative errors, are from

$\pm 18 - \pm 37\%$ for $\lambda = 555$ nm to $\pm 30 - \pm 53\%$ for $\lambda = 865$ nm and $\pm 22 - \pm 44\%$ for $\lambda = 412$ nm, depending on the aerosol optical thickness. The random errors estimated above are comparable to those obtained by the comparison of methods (Tab. 3). The accuracy of the pyranometric technique decreases for low values of the aerosol thickness.

5. Conclusions

The pyranometric method of estimating the aerosol optical thickness proved to be a satisfactory tool. Although the test of the pyranometric method of retrieving the aerosol optical thickness against the standard one reveals considerable systematic discrepancies, they can be removed by the application of correction coefficients. The discrepancies stem mainly from the model's assumptions, *e.g.* concerning single scattering and a homogenous atmosphere, and also from parametrisations of the model, *e.g.* solar radiation attenuation by atmospheric constituents, including the parametrisation of atmospheric absorber and scatterer amounts.

The statistical errors are no greater than ± 0.06 to ± 0.08, depending on the wavelength (Tab. 3). The errors are twice as high as those expected in the standard method but are nevertheless quite acceptable. In the case of low values of the aerosol thickness the uncertainties in the estimation increase. The pyranometric method of finding the optical aerosol thickness is very sensitive to the precision with which irradiances are measured and the water vapour content in the atmosphere estimated; these seem to be the main sources of the random errors.

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