Invited paper

Solar radiation in the Baltic Sea*

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Abstract

The influx of solar radiation to the Baltic Sea and its penetration into its waters is described on the basis of selected results of optical and bio-optical studies in the Baltic published by various authors during the past ca 50 years. The variability in the natural irradiance of this sea is illustrated on time scales from short-term fluctuations occurring during a single day to differences in mean monthly values over a period of many years. Data on variability of the proportions between UV,

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VIS and IR energy in the light reaching the sea surface are also discussed. Long-term monthly mean values of the incident solar radiation flux at the surface of the Baltic Proper are given; they were obtained from meteorological and solar radiation measurements and model approximations. The transmittances of these mean monthly radiation fluxes across the surface of the Baltic are given, as are the typical energy and spectral characteristics of the underwater irradiance, its attenuation with depth in the sea and the associated euphotic zone depths, as well as typical ranges of variability of these characteristics in different Baltic basins. Some of these characteristics are illustrated by typical empirical data. These mean values are not fully representative, however, because with the sole use of classical in situ measurement methods from on board research vessels in the Baltic, it has not been possible to gather a sufficiently representative set of empirical data that would adequately reflect the variability of the optical characteristics of all the basins of this sea. The article goes on to introduce the statistical model of vertical distributions of chlorophyll *a* concentration in the Baltic and the bio-optical model of Baltic Case 2 waters, the use of which contribute very significantly to this description of the optical characteristics and will enable this data set to be hugely expanded to include all the Baltic basins. This opportunity is presented by the optical parameterization of Baltic Case 2 waters, i.e. by the mathematical formulas of the model linking the coefficient of attenuation of downward irradiance with the surface chlorophyll *a* concentration, as well as the method developed for the efficient and systematic satellite remote sensing of the chlorophyll *a* concentration over the entire Baltic Sea area.

1. Introduction

Solar radiation plays a fundamental role in the marine environment. It supplies the energy that drives thermodynamic and photochemical processes in the sea, including the heating of waters, their stratification and movement, and their evaporation, also the warming of the atmosphere, the photosynthesis of the organic matter essential for the maintenance of life and the functioning of marine ecosystems, and the photo-oxidation of marine pollutants. The characteristics of light in the sea and of the light leaving the sea are also used as indicators of the concentrations of certain substances contained in sea water, e.g. chlorophyll, and are of practical significance in the satellite monitoring of the sea. For these and other reasons, much research effort, covering many facets of the problem, has gone into the investigation of the inflow of light energy to the sea and its interaction with the marine environment; with respect to the Baltic Sea, this research effort has been very great.

The inflow of solar radiation to the Baltic and its variability have been studied, modelled and described by Czyszek et al. (1979), Krężel (1982, 1985, 1997), Dera et al. (1984), Rozwadowska (1991, 1992, 1994, 1996), Dera (1995), Rozwadowska & Isemer (1998), Isemer & Rozwadowska

(1999), Berger (2002), Woźniak et al. (2003) and Krężel et al. (2008). The components of the radiation budgt have also been investigated (e.g. Pomeranec 1966, Kaczmarek & Dera 1998, Berger 2002, Lindau 2002, Omstedt & Nohr 2004), including the long-wave radiation of the Baltic surface as well as the formulas and models describing this long-wave radiation (Zapadka & Woźniak 2000, Woźniak S. B. et al. 2001, Zapadka et al. 2001, 2007, 2008).

Numerous papers have been written on the absorption, scattering and attenuation of light in Baltic Sea water, that is, on its inherent optical properties (e.g. Dera 1963, Dera et al. 1974a, 1978, Woźniak et al. 1977, Gohs et al. 1978, Dera & Sagan 1990, Sagan 1991, 2008, Olszewski et al. 1992, Sagan at al. 1992, Kowalczuk et al. 1999, Darecki et al. 2003, 2005). These studies have covered such aspects as absorption by coloured dissolved organic matter (CDOM, yellow substances) (Kowalczuk & Kaczmarek 1996, Kowalczuk 1999, Kowalczuk & Olszewski 2002, Kowalczuk et al. 2005, 2006) and the investigation and modelling of light absorption by organic matter suspended in the water, in particular by phytoplankton pigments (Woźniak & Ostrowska 1991, Woźniak et al. 1999, 2000a,b, 2005a,b, 2006, Majchrowski et al. 2000, Majchrowski 2001, Ficek et al. 2004, Woźniak & Dera 2007).

Much work has gone into the elucidation of underwater irradiance and its penetration into Baltic waters, the spectral characteristics of the underwater light field, their spatial and temporal variability, as well as the variability in range of the Baltic's euphotic zone (Dera 1970, Hapter et al. 1973, Woźniak & Hapter 1985, Woźniak et al. 1989, Sagan 1991, Sagan & Dera 1994, Dera 1995). The short-term fluctuations in underwater irradiance, including the concentrations of light energy under the wave-roughened surface of the Baltic and other seas known as underwater flashes, have also been examined (Dera & Olszewski 1967, 1978, Dera & Gordon 1968, Snyder & Dera 1970, Dera et al. 1974b, Dera & Stramski 1986, Stramski & Dera 1988, Dera et al. 1993).

The reflectance of the Baltic Sea has been studied in the context of satellite remote sensing (Olszewski & Darecki 1999, Darecki et al. 2003, 2004, 2005, Darecki & Stramski 2004), developed in the BIOCOLOR[1] (Ediger et al. 2001, Darecki et al. 2003), DESAMBEM[2] (Darecki et al. 2008,

[1]**BIOCOLOR** – The MAST III and the INCO project No. MAS3-CT97-0085: Ocean Colour for the Determination of Water Column Biological Processes.

[2]**DESAMBEM** – A large national research project No. PBZ-KBN 56/P04/2001: DEvelopment of a SAtellite Method for Baltic Ecosystem Monitoring – for deriving mathematical models and a complex algorithm for the remote sensing of the Baltic ecosystem and its primary production (Darecki et al. 2008, Woźniak et al. 2008).

Woźniak et al. 2008) and most recently in the SatBałtyk[3] programmes. Underwater visibility in the Baltic (Olszewski 1973, Sagan 2008) and the interaction between light and oil slicks on the sea surface have also been the subject of study (Otremba & Piskozub 2001, Otremba & Król 2002, Otremba 2004).

Water together with dissolved sea salt is a very strong absorber of infrared (IR) radiation (e.g. Woźniak & Dera 2007). Visible light (VIS) and ultraviolet (UV) radiation, by contrast, are absorbed by CDOM, and are also strongly absorbed and scattered by suspended particulate organic matter (Woźniak et al. 2005a,b, 2006), including phytoplankton pigments (Woźniak et al. 2000a,b, Woźniak & Dera 2007). The concentrations of these groups of substances in the Baltic are highly variable, chiefly as a result of the changeable inflows of numerous substances carried into the Baltic by the many rivers draining into it and seasonal phytoplankton blooms. This is why we observe substantial temporal changes in the optical properties of Baltic waters and their spatial differentiation. An in-depth study of these properties thus requires a copious set of data to be gathered from systematic measurements, equivalent to the continuous monitoring of this environment over many years. To date, however, such a broad range of data is unavailable owing to the prohibitive costs of deploying and maintaining a dense network of measuring stations operating 24/7 over the entire area of the Baltic. Comprehensive, remote optical measurements of the Baltic, making use of satellite technology, are unfortunately still in their infancy (Dera 2010); nonetheless, we do have at our disposal numerous data from measurements performed during a great many research and monitoring cruises, although these were not distributed evenly in time and across the whole Baltic. On this basis we can define typical and extreme values of the Baltic's optical properties, including the transmittance of radiant energy down into the sea.

The aim of this article is to select from among the numerous results of uncoordinated measurements and model calculations the numerical data, graphs and formulas that are most typical of the Baltic and of practical significance, characterizing the inflow and attenuation of the solar radiation entering the waters of this sea. Typical values are given of the solar energy influx to the Baltic and its variations in time and space. The numerous literature citations will lead the reader, according to need, to the sources

[3]**SatBałtyk** – A large-scale research project No. POIG.01.01.02-22-011/09-00: Satellite Monitoring of the Baltic Sea Environment, organized within the framework of the Innovative Economy Operational Programme (priority 1: Research and development of modern technologies, subaction 1.1.2: Strategic scientific research and development programmes (Dera 2010)).

of the characteristics, formulas and models presented here and to a more in-depth understanding of the phenomena described.

The strict, often complex, definitions of many optical properties of the sea will be found in Jerlov (1976), Højerslev (1986) and Dera (1992, 2003). Here we define only those properties that we will be focusing on in this paper (see Annex 1).

2. Solar irradiance at the top of the Earth's atmosphere

At present we can assume that the mean annual solar radiation flux reaching the top of the Earth's atmosphere, over the entire spectral range of this radiation, is $S_o \approx 1366$ W m^{-2} (e.g. Gueymard 2004, Darula et al. 2005). This is the mean annual irradiance of a surface normal to the rays at the upper boundary of the atmosphere (the top of the atmosphere (TOA) or outside the atmosphere); it is known as the solar constant. This 'constant', however, 'drifts' slowly up and down with varying periods (e.g. ca 11 years) of increasing and decreasing solar activity and also fluctuates for various other reasons; during the last 100 years it has varied by ca ±0.2%. At the same time the instantaneous values of this flux $S \equiv S(t)$ vary in time t during the year (e.g. Spencer 1971) by ±3.4% as a consequence of the Sun-Earth distance R[4] changing as our planet moves along its elliptical orbit. Assuming that from such a great distance the Sun can be treated as a point source of light, the irradiance at TOA of a plane perpendicular to the light rays emitted by that source is inversely proportional to the square of the distance R from it, that is to say, $S \sim 1/R^2$. Hence the instantaneous values of the flux S vary from ca $S_{\min} \approx 1320$ W m^{-2} (in July) to ca $S_{\max} \approx 1412$ W m^{-2} (in January).

At present the solar constant S_o is determined from solar radiation measured by accurate satellite radiometers. But if we consider the variability of the flux S and the need to take into account the energy of the entire width of the solar radiation spectrum[5] (Figure 1), (Thuillier et al. 2003, Gueymard 2004, Woods et al. 2009, Harder et al. 2010 and the papers cited therein), not to mention the geometrical complexity of the movement of our planet around the Sun, we can assume that the somewhat different values of the solar constant S_o, given in different papers, may be encumbered with an error of as much as 0.5% (see e.g. Schatten & Orosz

[4]The mean distance between the Earth and the Sun is $\bar{R} \approx 1.495 \times 10^8$ km. The Earth is closest to the Sun (at perihelion; $R = R_{\min} \approx 1.470 \times 10^8$ km) in early January, and farthest from it (at aphelion; $R = R_{\max} \approx 1.521 \times 10^8$ km) in early July.

[5]Estimates of the spectra of the solar radiation flux reaching the Earth (Woods et al. 2009, Harder et al. 2010) indicate that ca 99.9% of the energy of this flux lies within the wavelength interval 0.2–4 μm.

1990, Wilson 1993, Krommelynck & Dewitte 1997, Mecherikunnel 1998, Darula et al. 2005, ACRIM 2010).

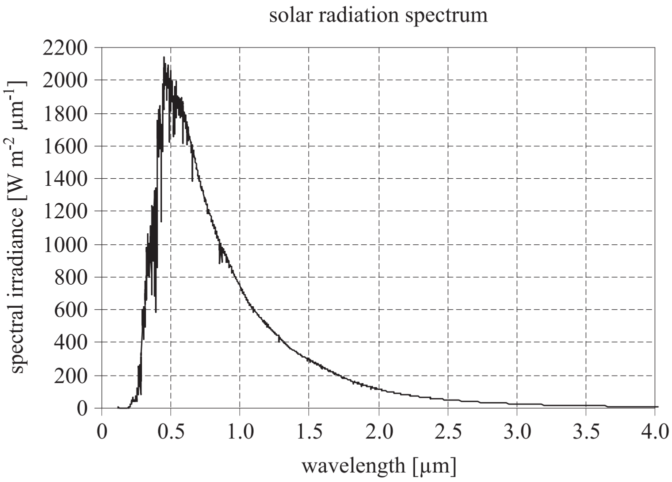

Figure 1. Spectrum of solar irradiance at the top of the Earth's atmosphere $E_S(\lambda)$ (according to Gueymard 2004)

The direction of the Sun's rays[6] reaching TOA and the direction of the local zenith form an angle θ, i.e. the angle of the Sun's inclination from the zenith, which is the angle of incidence of this radiation on a horizontal surface, i.e. perpendicular to the local zenith. The downward irradiance at TOA, over the entire spectral range of solar radiation, can thus be written as

$$E_S = S \cos\theta = S_o f \cos\theta, \tag{1}$$

where S_o – Solar constant, f – a factor describing the seasonal changes in the flux S due to changes in the Sun-Earth distance (see eq. (9)), θ – local zenith angle of the Sun as a function of the latitude Φ, local solar time t and number of the day in the year dn (e.g. Spencer 1971; see also Woźniak et al. 2008).

3. The inflow of solar radiation through the atmosphere to the surface of the Baltic Sea

As solar irradiance passes through the atmosphere to the sea surface, it is subject to complex processes – absorption, scattering, reflection and

[6]Here we assume that they are practically parallel; in fact, the discrepancy is of the order of 10^{-5} rad.

refraction by the numerous constituents of the atmosphere – which reduce part of its energy and alter its spectrum (e.g. Liu 1980, Timofeyev 1983, Warneck 1988, Thomas & Stamnes 1999, Tanaka & Wang 2004, Jin et al. 2006, Platt et al. 2007, Ota et al. 2010 and the papers cited therein). The light reaching a measuring station at sea level thus consists of a flux of rays arriving directly from the Sun, albeit attenuated in the atmosphere, and a flux of light scattered in the atmosphere coming from all directions of the sky. These two fluxes make up the irradiance of the sea surface and must be accounted for in calculations of this value (see e.g. Dera 1992, 2003). Formally, however, the downward irradiance at the sea surface in the entire spectral range can be written as

$$E_{\mathrm{d,\,surf}} = T_{\mathrm{a}}\, E_{\mathrm{S}} = T_{\mathrm{a}}\, S \cos\,\theta, \tag{2}$$

where T_{a} – the total atmospheric transmittance for the irradiance in its entire spectral range.

The instantaneous values of the solar irradiance travelling through the atmosphere to the sea surface vary dynamically with time in accordance with changes in the weather and the Sun's position in the sky. Apart from the 24 h cycle, clouds have the greatest effect on these changes in the Baltic (Figure 2).

Simplifying, the total atmospheric transmittance of the irradiance T_{a} can be written, e.g. as the product

$$T_{\mathrm{a}} = T_{\mathrm{o}}\, T_{\mathrm{clouds}}, \tag{3}$$

where T_{o} – irradiance transmittance of a clear cloudless atmosphere, T_{clouds} – irradiance transmittance of clouds, which can be empirically roughly established for different extents of cloudiness and cloud types.

Dera & Rozwadowska (1991) found that the irradiance transmittance of a clear cloudless atmosphere over the Baltic T_{o} was described to a good approximation by a formula derived earlier for the atmosphere over the North Atlantic with an atmospheric transparency coefficient $\mathrm{t}_2 = 0.75$ (after Jegorov & Kirillova 1973):

$$T_{\mathrm{o}} = 0.7650(\sin\, h)^{0.13} = 0.7650(\cos\,\theta)^{0.13}, \tag{4}$$

where $h = 90 - \theta$ is the solar altitude above the horizon in degrees.

The examples in Figure 2 show how strongly clouds attenuate solar irradiance over the Baltic (Dera & Rozwadowska 1991). Clouds cause the irradiance to fluctuate on time scales from seconds, minutes and hours (due to single clouds moving across the sky and changes in the optical thickness of clouds) to days (due to the changing synoptic systems of clouds). Sudden and profound changes in irradiance during the daytime are caused especially by convective clouds – Cumulus and Cumulonimbus. When such clouds shade the sun, the instantaneous irradiance can drop by more than one order of magnitude in comparison with its values for a cloudless sky (Figure 2b). At the same time, reflection of some of the solar radiation from single clouds towards the measurement point can increase the instantaneous irradiance to values more than 20% greater than the irradiance that would be prevailing under a cloudless sky at the same time and place. Frontal stratiform clouds bringing precipitation (e.g. Altostratus, Figure 2d) reduce the daily irradiance dose to a few percent of that obtaining under a cloudless sky. Over the Baltic the optical thickness of stratiform clouds before or after precipitation can be in excess of 40 (Rozwadowska 2004).

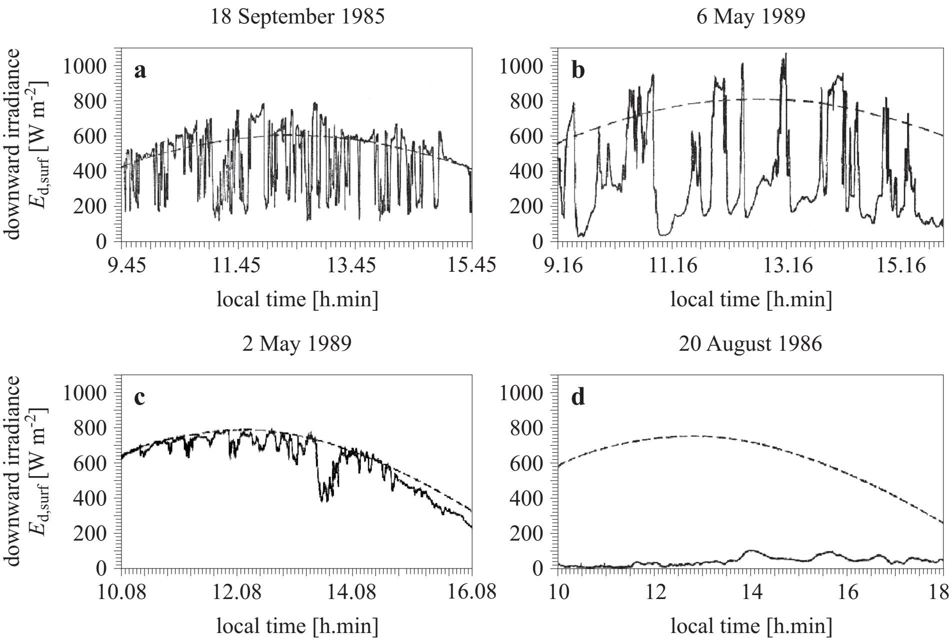

Figure 2. Examples of global solar irradiance fluctuation at the sea surface due to clouds. Cloudiness and predominant cloud types: a) 0.5 Cu med., b) 0.7–0.9 Cb, Cu con., c) 0.5–0.9 Ci, Cs, d) 1 As, Sc with precipitation. Coordinates of the measuring station: $\Phi = 54°27'$N, $\Lambda = 18°34'$E. The dashed line shows the clear sky irradiance on the day, calculated from formulas (2)–(4) and (6) assuming $T_{\text{clouds}} = 1$ (Reproduced from Dera & Rozwadowska 1991, Oceanologia 30, p. 14–16, Figures 6, 8, 9 and 10)

In her model, Rozwadowska (1981) used the following simple empirical formula to approximate the cloud transmittance T_{clouds} in the Baltic Sea region:

$$T_{\text{clouds}}(c, cc, \theta) = \begin{cases} 1 + A_L c + B_L c^2 + D_L \left[c^3 (1-c) \right] \cos\theta & \text{for} \quad cc = L \\ 1 + A_H c + B_H c^2 + C_H c^3 + D_H c \cos\theta & \text{for} \quad cc = H \\ A_M & \text{for} \quad cc = M \end{cases} \tag{5}$$

where A_L, B_L, D_L, A_H, B_H, C_H, D_H and A_M – the empirical coefficients given in Table 1, c – total cloud amount (from 0 to 1), and $cc = L$, M, H stands for a type of cloud situation defined as follows:

- cloud class L – predominantly convection clouds (Cu and Cb), low-level (St, Ns, Sc) or middle level stratiform clouds (As, Ac) covering a considerable part of the sky (c(As, Ac) ≥ 0.7); when $c > 0.95$ this class includes situations when clouds are opaque;
- cloud class M – predominantly middle-level or low-level clouds and $c > 0.95$; clouds are at least partly semi-transparent;
- cloud class H – predominantly high-level clouds (Ci, Cc, Cs) or middle-level clouds with c(As, Ac) < 0.7.

Table 1. Calculated parameters of the relations (5)

A_L	B_L	D_L	A_M	A_H	B_H	C_H	D_H
−0.660	−0.174	1.21	0.396	−0.399	0.321	−0.349	0.241

It is clear from Figure 2 and the above description that instantaneous irradiances of the sea surface are accidental, randomly variable values, which for practical purposes require the light energy Q reaching the sea surface to be averaged or summed over a particular time interval t, for example, 24 h, a month or a year. So the mean downward irradiance just above the sea surface is

$$< E_{\text{d, surf}} >_{\Delta t} = \frac{1}{\Delta t} \int_{\Delta t} E_{\text{d, surf}} dt \tag{6}$$

and the total solar energy reaching the surface in a time interval Δt is:

$$Q_{\text{surf}, \Delta t} = \int_{\Delta t} E_{\text{d, surf}} dt. \tag{7}$$

In some applications, e.g. for calculating the energy balance of a marine basin (e.g. Kaczmarek & Dera 1998, Zapadka et al. 2008) or for estimating the amount of radiant energy entering an ecosystem during photosynthesis (e.g. Woźniak et al. 2008), the crucial values are the daily or monthly totals or the mean amounts of solar radiation energy arriving at the sea surface. To calculate them we often use semi-empirical models that take account of the characteristics of the local atmosphere (e.g. for the Baltic: Krężel 1985, Kaczmarek & Dera 1998, Rozwadowska & Isemer 1998, Zapadka & Woźniak S. B. 2000, Zapadka et al. 2001, 2007, 2008, Woźniak et al. 2003, Krężel et al. 2008). The input data for such semi-empirical models are data on the state of the atmosphere recorded at hydrometeorological shore stations, on ships performing such observations, and increasingly, using remote sensing techniques (e.g. Krężel & Kozłowski 2004, Krężel et al. 2008, Woźniak et al. 2008, Zapadka et al. 2008).

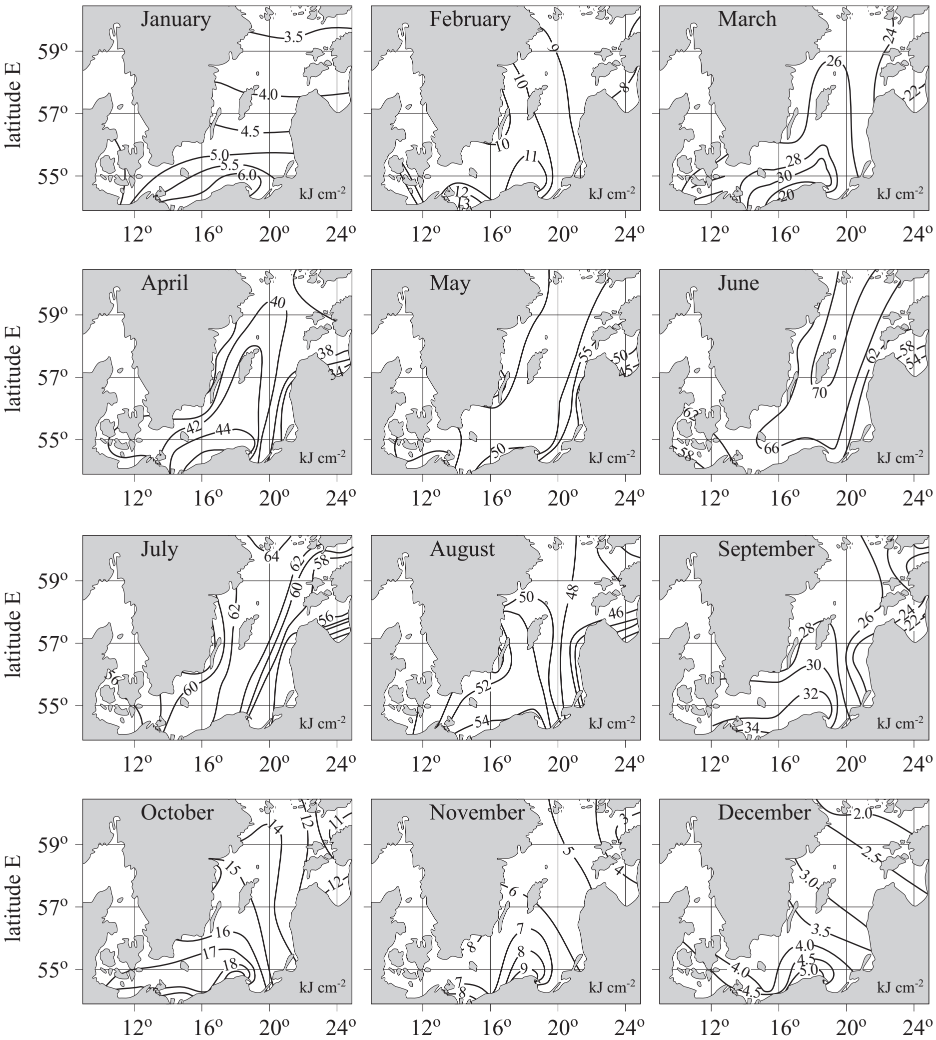

Figure 3. Long-term (1965–1974) mean monthly totals of solar radiation energy $< Q_{\mathrm{surf}} >_{\mathrm{month}}$ [kJ cm^{-2}] at the surface of the Baltic Proper (Reproduced from Krężel 1985, Oceanologia 21, p. 20–21, Figure 6)

Krężel (1982, 1985) was the first to carry out wide-ranging calculations of the solar radiation influx to the Baltic. For this he utilized meteorological data like atmospheric pressure, water vapour pressure and cloudiness from 23 weather stations located on the shores and islands of the Baltic, as well as actinometric data from six stations from the period 1965–1974. He carried out a detailed analysis of the attenuation of irradiance by an 'ideal' dry atmosphere, absorption by the water vapour in the air, and attenuation by aerosols and clouds. On this basis he derived a semi-empirical model and calculated the long-term mean monthly totals of solar radiation energy incident at the sea surface of the Baltic Proper (see Krężel 1985). The main result of these model calculations is shown in Figure 3, in which isolines of the values of $< Q_{\mathrm{surf}} >_{\mathrm{month}}$ [kJ cm^{-2}] have been drawn on maps of the Baltic.

In his later research, Krężel (1997) improved this model by introducing remotely sensed input data on the state of the atmosphere.

Another well-documented, semi-empirical model for the Baltic was developed by Rozwadowska (1991). The total downward irradiance in this model is given by the equation

$$E_{\mathrm{d,\,surf}} = \frac{S_{\mathrm{o}} f (T_{\mathrm{i}} - A_{\mathrm{wa}}) T_{\mathrm{aer}} T_{\mathrm{clouds}} \cos\theta}{1 - A_{\mathrm{sk}} A_{\mathrm{surf}}}, \tag{8}$$

where

S_{o} – the assumed solar constant (in this model 1368 W m^{-2}, according to Wilson 1993);

$f \equiv f(dn)$ – a factor describing seasonal changes in S due to changes in the Sun-Earth distance. It is given by equation (9) as a function of the day number of the year dn;

θ – solar zenith angle;

$T_{\mathrm{i}} \equiv T_{\mathrm{i}}(\theta,\ p)$ – irradiance transmittance for an ideal (dry) atmosphere (Kastrov 1956, Atwater & Brown 1974), as a function of θ and air pressure p;

$A_{\mathrm{wa}} \equiv A_{\mathrm{wa}}(e_{\mathrm{o}}, \theta)$ – absorbance of water vapour (McDonald 1960) as a function of water vapour pressure at the sea surface e_{o} and solar zenith angle θ, where $e_{\mathrm{o}} \equiv e_{\mathrm{o}}(Temp_{\mathrm{d}})$ – water vapour pressure at the sea surface as a function of the dew point temperature $Temp_{\mathrm{d}}$ (Goff 1965);

$T_{\mathrm{aer}} \equiv T_{\mathrm{aer}}(\theta, month, \Phi, \Lambda)$ – aerosol irradiance transmittance as a function of θ, month number $month$ and geographical coordinates Φ, Λ (Krężel 1985, Rozwadowska 1991);

$T_{\mathrm{clouds}} \equiv T_{\mathrm{clouds}}(c,\, cc,\, \theta)$ – cloud irradiance transmittance (Rozwadowska 1991; also eq. (5) in this paper), as a function of c – total cloud amount, cc – cloud class (L, M, H) and θ;

$A_{\mathrm{sk}} \equiv A_{\mathrm{sk}}(c,\, cc)$ – sky albedo as a function of c and cc (Rozwadowska 1991, 1998 and the papers cited therein);

$A_{\mathrm{surf}} \equiv A_{\mathrm{surf}}(T_{\mathrm{a}},\, \theta,\, ice)$ – sea surface albedo as a function of T_{a}, θ and ice (Payne 1972, Isemer 1998), where ice – fraction of the sea surface covered by ice and T_{a} – total atmospheric transmittance for the irradiance.

The factor f, according to Spencer (1971), is described as follows:

$$f = 1.00011 + 0.034221 \cos\alpha + 0.00128 \sin\alpha + 0.000719 \cos 2\alpha + \\ + 0.000077 \sin 2\alpha, \quad (9)$$

where $\alpha = 2\pi(dn - 1)/365$.

This model, in combination with voluntary observing ship meteorological data from COADS (Comprehensive Ocean-Atmosphere Data Set) (1980–1992) used as input data (Figure 4), was employed by Rozwadowska & Isemer (1998) to calculate long-term mean monthly values of the irradiance $< E_{\mathrm{d,\,surf}} >_{\mathrm{month}}$ (that is, the energy flux incident on 1 m^2

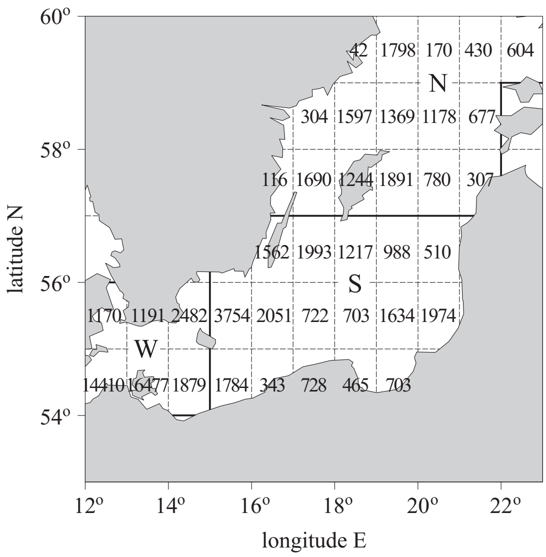

Figure 4. Division of the Baltic Proper as used in the study by Rozwadowska & Isemer (1998) (Reproduced from Rozwadowska & Isemer 1998, Oceanologia 40 (4), p. 310, Figure 1)

N – northern Baltic Proper – north of 57°N; **S** – southern Baltic Proper – south of 57°N and east of 15°E; **W** – western Baltic Proper – west of 15°E

The numbers of COADS daytime ship observations from 1980 to 1992 containing meteorological data used are given for each 1° latitude/longitude gridbox in model computations

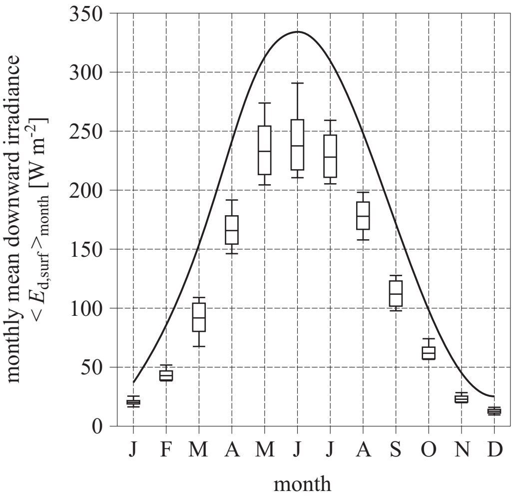

Figure 5. Long-term (1980–1992) monthly mean incident solar radiation flux (i.e. irradiance in the entire spectral range) $< E_{\mathrm{d,\,surf}} >_{\mathrm{month}}$ [W m^{-2}] at the surface of the southern Baltic Proper (box plot drawn on the basis of data from Rozwadowska & Isemer (1998))

The ends of the whiskers indicate the maximum and minimum values in different years, the boxes denote the standard deviations, and the horizontal lines in the boxes show the mean flux for a given month

The continuous line represents the approximate, modelled maximum values that would be recorded in the atmosphere over the Baltic for a cloudless sky (according to equations (2), (3), (4) and (6)) assuming $T_{\mathrm{clouds}} = 1$; see also Czyszek et al. (1979)

of sea surface) of the Baltic Proper (Figure 5). Subsequent analysis of the accuracy of these values and comparison with the results of calculations based on other empirical parameterizations (Isemer & Rozwadowska 1999) indicated that their accuracy was satisfactory and were practically useful as data on the energy reaching the Baltic Proper. The statistical error varied from ca 5–10% in the summer months to 20–25% in the winter months for very low irradiance values.

Figure 5 shows great differences in the radiant energy influx in particular months of different years: e.g. in June 1992 the flux was ca 292 W m^{-2} but in June 1987 it was only 211 W m $^{-2}$; the year-on-year differences are due mainly to changes in atmospheric conditions. The continuous line on the plot in Figure 5 shows the annual changes in the approximate model values of this flux that would be recorded at the surface of the southern Baltic, if the sky over the Baltic were endlessly cloudless (according to equations (2), (3), (4), (6) and $T_{\mathrm{a}} = 1$). For these calculations in those years the solar constant was taken to be $S = 1368$ W m^{-2} (Wilson 1993).

Table 2. Long-term monthly means of the incident solar radiation flux $< E_{\mathrm{d,\,surf}} >_{\mathrm{month}}$ in the entire spectral range at the surface of the Baltic Proper, standard deviations of the individual monthly means, absolute extreme values $< E_{\mathrm{d,\,surf}} >_{\mathrm{month,\,min}}$, $< E_{\mathrm{d,\,surf}} >_{\mathrm{month,\,max}}$ for the period 1980–1992, and mean monthly totals $< Q_{\mathrm{surf}} >_{\mathrm{month,\,max}}$ [kJ cm^{-2}] given for comparison with Krężel's data in Figure 3 (data from Rozwadowska & Isemer 1998)

- **The northern Baltic Proper – north of 57°N**

Month	Mean irradiance $<E_{\mathrm{d,\,surf}}>_{\mathrm{month}}$ [W m^{-2}]	Standard deviation [W m^{-2}]	Minimum $<E_{\mathrm{d,\,surf}}>_{\mathrm{month,\,min}}$ [W m^{-2}]	Maximum $<E_{\mathrm{d,\,surf}}>_{\mathrm{month,\,max}}$ [W m^{-2}]	Mean monthly totals $<Q_{\mathrm{surf}}>_{\mathrm{month}}$ [MJ m^{-2}]
1	2	3	4	5	6
Jan.	14.7	2.2	10.5	18.8	39.3
Feb.	42.9	5.6	32.1	53.5	103.8
March	86.4	9.6	71.3	101.4	232.8
April	163.2	15.9	124.6	181.3	423.0
May	231.7	21.3	189.9	256.9	620.5
June	248.9	23.0	217.0	282.7	645.1
July	237.2	20.9	194.6	265.2	635.4
Aug.	178.0	12.3	155.0	191.8	476.7
Sept.	108.8	12.1	89.2	127.1	282.0
Oct.	55.08	6.5	41.3	66.6	147.4
Nov.	18.90	2.6	15.2	22.9	49.0
Dec.	9.5	1.6	7.4	13.0	25.4
year	116	4	111	122	3670

- **The southern Baltic Proper – south of 57°N and east of 15°E**

1	2	3	4	5	6
Jan.	19.3	2.8	16.0	25.8	51.8
Feb.	43.3	3.8	38.9	52.5	104.8
March	91.8	11.3	66.7	108.6	245.9
April	166.4	11.5	146.5	191.0	431.4
May	234.3	20.5	206.3	274.1	627.5
June	239.1	20.8	211.0	292.3	619.7
July	229.0	17.5	205.8	259.5	613.4
Aug.	178.6	11.1	158.3	198.0	478.4
Sept.	112.5	10.4	97.3	127.6	291.7
Oct.	62.4	4.6	57.1	74.9	167.2
Nov.	23.6	2.3	20.2	27.7	61.1
Dec.	13.1	1.8	10.1	16.6	35.2
year	118	4	111	127	3730

Table 2. (*continued*)

- **The western Baltic Proper – west of 15°E**

1	2	3	4	5	6
Jan.	20.1	2.4	16.4	23.8	53.9
Feb.	45.6	5.1	36.6	53.1	110.3
March	89.6	8.2	74.5	102.2	240.0
April	163.5	18.6	132.4	194.6	423.7
May	222.0	24.8	167.0	257.5	594.7
June	223.8	31.7	179.1	287.8	590.0
July	213.4	19.3	189.9	251.3	571.5
Aug.	167.9	11.5	147.9	187.1	449.8
Sept.	111.9	10.1	96.8	127.9	290.2
Oct.	63.9	5.2	54.8	71.3	171.1
Nov.	25.4	3.8	18.3	31.2	65.8
Dec.	14.7	1.5	11.6	17.1	39.3
year	114	5	106	123	3600

- **The entire Baltic Proper**

1	2	3	4	5	6
Jan.	17.5	2.2	14.5	22.3	46.9
Feb.	43.5	3.6	36.8	48.1	105.1
March	89.3	8.8	73.2	99.9	239.1
April	164.7	12.4	135.6	184.4	426.9
May	231.6	19.9	194.9	261.4	620.2
June	241.1	21.1	213.9	287.8	624.9
July	230.3	16.7	199.3	254.7	616.9
Aug.	176.9	10.5	159.9	192.6	473.9
Sept.	110.9	9.7	93.9	126.6	287.5
Oct.	59.6	4.0	52.9	66.2	159.5
Nov.	21.9	2.0	18.0	25.0	56.7
Dec.	11.8	1.4	9.9	14.0	31.7
year	117	4	111	124	3680

Detailed numerical data on the mean monthly solar radiation flux reaching the surface of the Baltic Proper, for the three sub-areas distinguished in Figure 4, are given in Table 2.

A semi-empirical spectral model of the transmission of irradiance through the atmosphere (Woźniak et al. 2003) using hydrometeorological data after Augustyn (1985) and the spectral optical models of transmission across the sea surface (Woźniak S. B. 1996b, 1997) and through the water column (Woźniak et al. 1992a,b, Kaczmarek & Woźniak 1995), as well as the formula for the effective longwave radiation of the sea surface (Bignami et al. 1995), enabled the monthly means of the fourteen fundamental radiant energy fluxes usually distinguished in the sea-atmosphere system to be estimated for the southern Baltic. Them include the sum of the fluxes absorbed and scattered upwards (reflected) in the atmosphere, the flux reaching the sea surface, the flux entering the water column and the flux absorbed in the water column. The mean monthly values of all these fluxes for the period 1970–1990 were calculated for 20 subareas of the southern Baltic (see Kaczmarek & Dera 1998).

Table 3 lists the averaged energy fluxes reaching the sea surface of the southern Baltic Proper emerging from the three above-mentioned independent model calculations, based on different empirical data from different long-term periods. The values in column 3 of this table, distinctly greater in the summer months, result from calculations for the southern part only of the southern Baltic Proper (south of 55°30′N, excluding the coastal zone) using the set of averaged meteorological data from ships published by the Institute of Meteorology and Water Management (IMGW) (Augustyn 1985).

Table 3. Comparison of the results of three independent model calculations of the long-term monthly mean incident solar radiation fluxes $<E_{\mathrm{d,\,surf}}>_{\mathrm{month}}$ at the surface of the southern Baltic Proper

Month	$<E_{\mathrm{d,\,surf}}>_{\mathrm{month}}$ [W m^{-2}]		
	Krężel (1985) for 1965–1974*	Rozwadowska & Isemer (1998) for 1980–1992	Kaczmarek & Dera (1998) for 1970–1990**
Jan.	18.7	19.3	18.4
Feb.	43.4	43.3	52
March	100.8	91.8	120.0
April	165.9	166.4	198.4
May	224.0	234.3	261.7
June	254.6	239.1	299.9
July	224.0	229.0	275.5
Aug.	186.7	178.6	235.4
Sept.	123.5	112.5	147.8
Oct.	59.7	62.4	70.1
Nov.	27.0	23.6	24.6
Dec.	14.9	13.1	13.6
year	120	118	143

*Approximate values for the southern Baltic Proper, averaged on the basis of the data in Figure 3.
**Area of the southern Baltic Proper limited to south of 55°30′N.

At present, mean 24 h irradiance values in the Baltic are calculated by our research group using a modified version of the model by Krężel et al. (2008), which uses current satellite data on the state of the atmosphere over the Baltic (Darecki et al. 2008, Woźniak et al. 2008).

4. Spectral distribution and transfer of irradiance to below the sea surface

Before entering the atmosphere, solar radiation has a spectral distribution (Figures 1 and 6a) approximating that of the radiation emitted by an ideal black body with a surface temperature of ca 5500 K (Gueymard 2004, Woods et al. 2009, Harder et al. 2010). The real measured spectrum of this radiation is given in tabular format by Gueymard (2004). This shows that ca 7% of the energy is carried by the UV waveband, 47% by VIS light and 46% by IR waves. As this radiation travels through

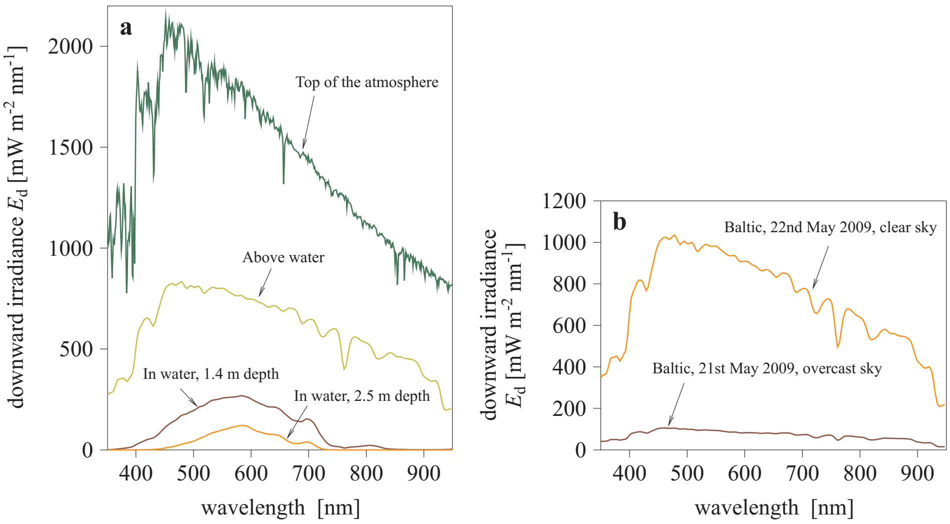

Figure 6. Solar radiation spectra

a) – at the top of the Earth's atmosphere (upper curve, according to Thuiler et al. 2003),
– just above the sea surface and below the sea surface at depths 1.4 and 2.5 m measured under cloudless conditions at the Baltic station 54°24.20′N, 19°0.00′E at 13:26 local time on 16.04.2009 (according to Mirosław Darecki, IO PAS data bank)

b) – just above the sea surface measured at the Baltic station 54°24.20′N, 19°0.00′E under a clear sky on 22 May (upper curve) and under an overcast sky on 21 May (lower curve) (according to Mirosław Darecki, IO PAS data bank) the atmosphere to the sea surface, the spectrum is modified (Figure 6) because the radiation is attenuated as a result of absorption by atmospheric gases, particularly ozone and carbon dioxide, and also water vapour, among other things (see e.g. Warneck 1988).

But these proportions of energy do not change much in the atmosphere. They depend on the state of the atmosphere – mainly cloud cover and the solar zenith angle. The empirical studies in central Poland by Podstawczyńska (2010) show that the mean ratio of UV to global solar irradiance for periods of cloudless skies varies from 3.3 to 4%, but for an overcast sky it is at least 6%. The state of the atmosphere for a given solar zenith angle can be expressed by the atmospheric transmittance T_{a}. According to empirical data from the surface of the Baltic, the ratios vary within the following limits: 5–9% UV, 35–53% VIS and 38–56% IR. The percentages of VIS and UV are largest when the atmospheric transmittance and solar zenith angle are low, i.e. $T_{\mathrm{a}} < 0.10$ and $\theta < 45°$, whereas the percentage of IR is highest when the atmospheric transmittance and solar zenith angle are both high, i.e. $T_{\mathrm{a}} > 0.40$ and $\theta > 75°$ (see Rozwadowska 1996 and the data collected by Rozwadowska in Dera 1995).

The spectrum of the daytime irradiance is further modified as this passes under the sea surface. Figure 6a shows that 2.5 m beneath the surface in Baltic water, there is practically no IR or UV light left, because of the very strong absorption of IR by water molecules and of UV by yellow organic substances, i.e. coloured dissolved organic matter (CDOM) (Woźniak & Dera 2007). Just below the surface the averaged proportions of energy in the three divisions of the spectrum are 5% UV, 50% VIS and 45% IR, as calculated, for example, in the paper by Czyszek et al. (1979).

A precise mathematical description of the transfer of the daily irradiance dose across a wave-roughened sea surface over the entire spectral range of this light is exceedingly complex. It requires the following factors to be taken into consideration: the distribution of slopes of the wave-roughened sea surface, the degree of foam coverage of the surface, and the directional distribution of the flux irradiating the surface by direct solar radiation and by light scattered in the atmosphere. In addition, the laws of reflection and refraction of light, and strictly speaking, also the spectral characteristics of the refractive index, have to be applied to the constituent parts of this flux coming from every part of the sky. The fundamentals of this process are described, for example, in Dera (1992, 2003), and detailed models and calculations, for example, in Woźniak S. B. (1996a,b, 1997), Woźniak et al. (2003, 2008). With the far simpler semi-empirical formulae, one can calculate the averaged transmittance across the sea surface from monthly mean fluxes or monthly radiant energy totals (e.g. for the southern Baltic, see Czyszek et al. 1979). By definition, this transmittance is equal to

$$< T_{\mathrm{surf}} >_{\mathrm{month}} = < E_{\mathrm{d}}(z = 0) >_{\mathrm{month}} / < E_{\mathrm{d,\,surf}} >_{\mathrm{month}}, \tag{10}$$

where $< E_{\mathrm{d}}(z = 0) >_{\mathrm{month}}$ – the mean monthly downward irradiance just below the sea surface, $< E_{\mathrm{d,\,surf}} >_{\mathrm{month}}$ – the mean monthly downward irradiance just above the sea surface (given in Tables 2 and 3).

According to the results of our numerous investigations, the transmittance of the averaged downward irradiance (or the monthly radiation total) by the real surface of the Baltic over the entire spectral range (in practice 300–4000 nm), $< T_{\mathrm{surf}} >_{\mathrm{month}}$ oscillated between 0.90 and 0.96, whereby the transmittance is least in the winter months and greatest in the summer months (see Table 4).

Table 4. Transmittance across the southern Baltic Sea surface of the long-term mean monthly solar irradiance (or mean monthly totals) over the entire spectral range $<T_{\mathrm{surf}}>_{\mathrm{month}}$

Month	Data source	
	Kaczmarek & Dera 1998	Czyszek et al. 1979
Jan.	0.91	0.92
Feb.	0.92	0.93
March	0.94	0.93
April	0.95	0.94
May	0.96	0.94
June	0.96	0.94
July	0.96	0.94
Aug.	0.95	0.94
Sept.	0.94	0.94
Oct.	0.93	0.93
Nov.	0.91	0.93
Dec.	0.90	0.92

The formulas and detailed calculations of the daily PAR dose transmitted across a wind-roughened sea surface for different wind speeds, taking into consideration separately the energy flux carried by direct solar radiation and that borne by light scattered in the atmosphere, are described in Woźniak et al. (2008) and Darecki et al. (2008).

5. Penetration of irradiance into Baltic Sea water; the euphotic zone

A considerable proportion of the total radiant energy that enters Baltic Sea water is absorbed in the thin surface layer of this water. The principal reason for this is that long-wave light is strongly absorbed by water and the short-wave radiation by CDOM, present here in large concentrations (Kowalczuk et al. 2005, 2006) as a consequence of the high trophicity of the waters. Chlorophyll a is an indicator of this trophicity, and at high concentrations of this chlorophyll ($C_{\mathrm{a}}(0) \approx 35$ mg m^{-3}), 80% of the irradiance is absorbed in a water layer 0.7 m thick, and at low concentrations ($C_{\mathrm{a}}(0) \approx 0.35$ mg m^{-3}) in a layer ca 3 m thick (Figure 7). That is why beneath the thin surface layer of the Baltic practically only visible light is present, i.e. the wavelengths lying within the PAR spectral interval (ca 400 nm to 700 nm).

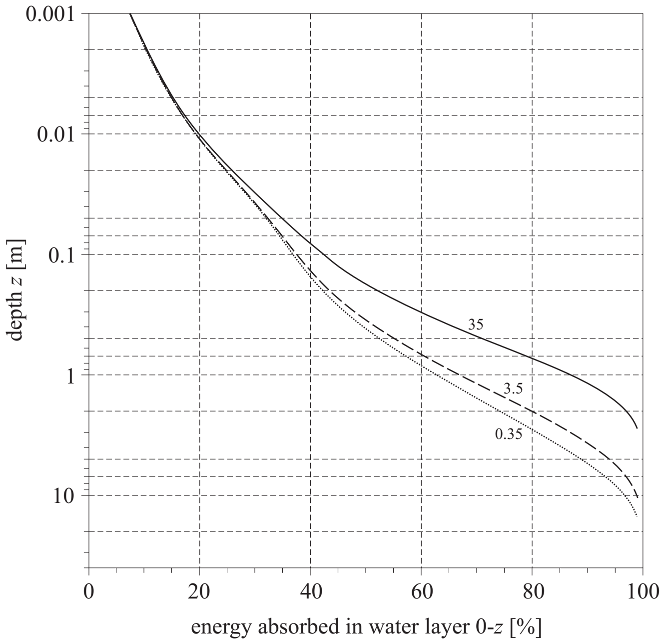

Figure 7. Radiation absorbed in the water column for three different surface chlorophyll concentrations $C_{\mathrm{a}}(0)$: 35, 3.5 and 0.35 mg m^{-3} (Reproduced from Kaczmarek & Dera 1998, Oceanologia 40 (4), p. 297, Figure 5)

Light penetrates water more selectively with respect to wavelength than in the atmosphere. This means that its spectrum changes rapidly with depth. Moreover, the absolute values of the radiant energy (well characterized by the downward irradiance $E_{\mathrm{d}}(\lambda, z)$) are much more strongly attenuated with depth than they are attenuated with height in the atmosphere. At the same time, the spectral distributions and the spatial distributions of the absolute values of irradiance at different depths in various regions of the Baltic at different times are both strongly

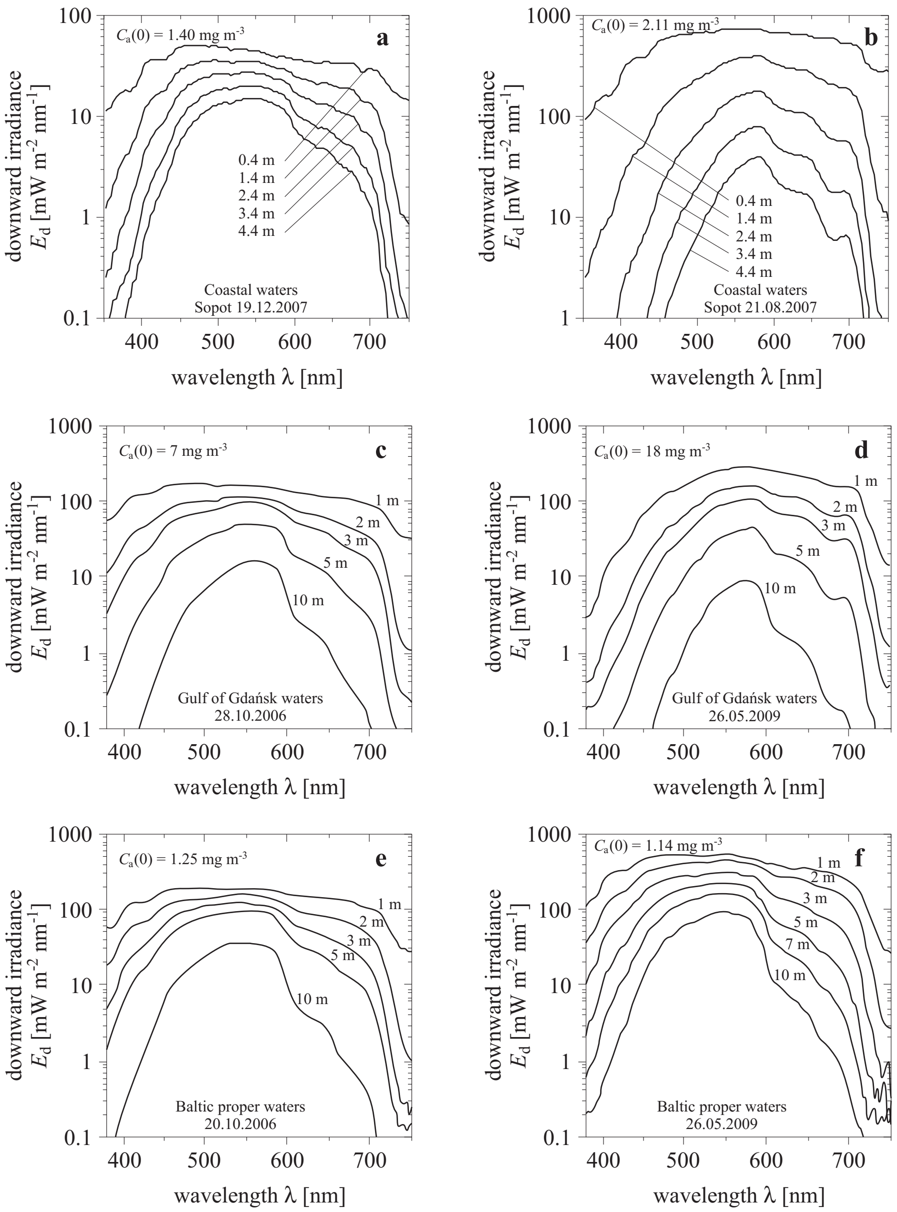

Figure 8. Spectral distributions of the downward irradiance $E_d(\lambda, z)$ at different depths in the Baltic Sea, typical of:

a) and b) shallow coastal basins (measurements by D. Ficek near Sopot Pier; IO PAS data bank)

c) and d) Gulf basins (measurements by M. Darecki at stations in the Gulf of Gdańsk, IO PAS data bank)

e) and f) Baltic open-sea waters (measurements by M. Darecki at stations in the open Baltic proper; IO PAS data bank) (*continued on next page*)

Figure 8. (*continued*) The curves in Figures a, c and e were plotted on the basis of optical measurements made in the cool seasons of the year, those in Figures b, d and f on the basis of such measurements made in spring and summer. The surface concentration of chlorophyll $C_a(0)$ [mg tot. chl a m^{-3}] on the day of measurement is given on each plot

differentiated. This differentiation in Baltic waters is far greater than the differentiation of light fields in the atmosphere, not only over the Baltic, but also over the oceans. Evidence for this is provided by the numerous measurements of underwater irradiance made in the Baltic by the authors' research teams and co-workers during the past 50 years (IO PAS oceanographic data bank). To exemplify these measurements, Figure 8 shows typical spectra of the downward irradiance $E_d(\lambda)$, recorded at different depths in the Baltic in different regions and seasons. (Note that the effects described are somewhat masked on the graphs owing to the logarithmic scales of the irradiance values).

The differentiation in irradiance in different marine basins turns out to be even stronger if we compare Baltic basins with other seas and regions of the World Ocean. An extreme example of this differentiation, illustrating how far the characteristics of underwater irradiance in the eutrophic basins of the Baltic can diverge from those in optically clear, oligotrophic oceanic waters, is shown in Figure 9. This exemplifies the spectral distributions of the downward irradiance at different depths in the sea $E_d(\lambda, z)$ in an ultra-oligotrophic basin of the Indian Ocean with a surface chlorophyll concentration $C_a(0) \approx 0.035$ mg tot. chl a m^{-3} (Figure 9a) and in the super-eutrophic[7] Puck Bay (Baltic), where the surface chlorophyll concentration $C_a(0) \approx 70$ mg tot. chl a m^{-3}.

It is clear from the curves in Figure 9 that the overall trends characterizing the differentiation in irradiance $E_d(\lambda, z)$ spectra in different types of marine basin are qualitatively similar: in both ultra-oligotrophic waters (Figure 9a) and in super-eutrophic waters (Figure 9b) the absolute irradiance falls with increasing depth z and that with increasing depth, irradiance from the middle interval of the visible light spectrum is dominant. This is due to the absorption and/or scattering, i.e. attenuation, of all the wavelengths of this light as it penetrates deeper into the water. It is always the case, however, that the attenuation of short waves (ultraviolet, violet) and of long waves (red, infrared) is naturally stronger than that of waves from the middle part of the visible light spectrum (green-blue) (see

[7]The chlorophyll a concentrations in waters of different trophicity are listed in the table in Annex 2.

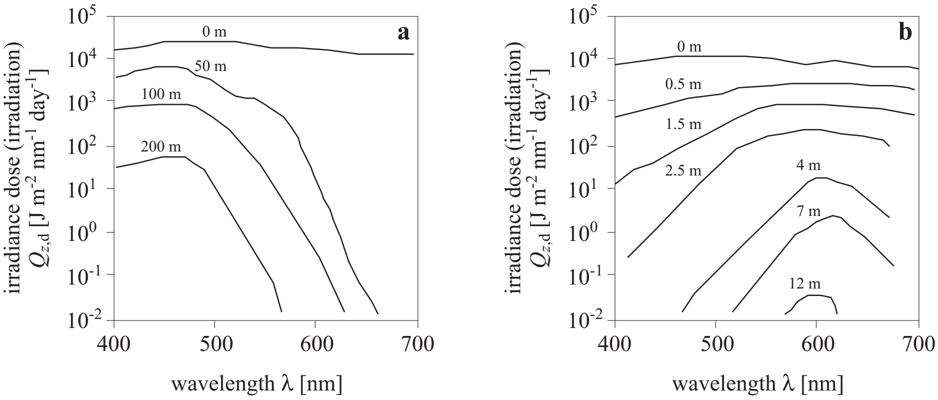

Figure 9. Comparison of the spectral distributions of downward irradiance at different depths in the sea, measured in:

a) an ultra-oligotrophic oceanic basin (Indian Ocean, near Mauritius), in which the surface concentration of chlorophyll a was $C_{\mathrm{a}}(0) \approx 0.035$ mg tot. chl a m^{-3}

b) a super-eutrophic Baltic basin (Puck Bay) with a surface concentration of chlorophyll $C_{\mathrm{a}}(0) \approx 70$ mg tot. chl a m^{-3}

(Reproduced from Ostrowska 2001)

e.g. Jerlov 1976, Dera 1992, 2003, Woźniak & Dera 2007). This narrows the irradiance spectrum at great depths down to an almost monochromatic blue-violet in the clearest oligotrophic waters and to green or even greenish-brown in eutrophic waters. But these similarities are purely qualitative; the quantitative properties of these waters are very different. To begin with, the fall in the absolute irradiance with depth z in eutrophic basins is much stronger than in oligotrophic waters. For example, in Puck Bay, in the spectral band of the light most readily penetrating the waters of this bay, 1% of all the irradiance energy entering the water reaches no deeper than ca 3–5 m. By contrast, in the Indian Ocean basin referred to in Figure 9a, 1% of the surface irradiance energy penetrates very many times deeper, as far down as 150–200 m, with the spectral maximum of light best penetrating these waters (Figure 9).

Secondly, even though the relative spectral distributions of natural irradiance just below the sea surface and at shallow depths in a basin are similar everywhere and in shape approximate the spectrum of sunlight arriving directly at the sea surface, these same distributions at greater depths are diametrically different. It is clear from Figure 9 that the position of the band of radiation penetrating Baltic waters most easily is quite different from the position of this band in oceanic waters. In the low-productivity, optically clearest oceanic basins (such as the example from the Indian Ocean in Figure 9a), the energy maximum reaching great depths in the sea comes from blue-green light and even blue light in the 440–460 nm region. In such basins, no significant changes in the position of the maximum take place with increasing depth: only the dominance of light in a narrow spectral interval over light of other wavelengths grows. In basins of greater trophicity, however, these spectral changes take a different course. With increasing depth in mesotrophic and eutrophic waters, the positions of the maxima of the spectral distributions of irradiance are increasingly shifted towards the longer wavelengths; these shifts may be greater or smaller, and in most cases are strongly correlated with the chlorophyll *a* concentration in the water. For example, Puck Bay, with a surface chlorophyll concentration $C_{\mathrm{a}}(0) \approx 70$ mg tot. chl *a* m^{-3}, is an example of a super-eutrophic basin, where the waveband most readily penetrating the water is shifted far towards the red end of the spectrum in the 600 nm region (Figure 9b), and the waters of this bay appear brown. Nevertheless, it should be mentioned that in most Baltic waters, the best penetration (i.e. maximum irradiance transmittance in the water) is achieved by light from the spectral interval 520–550 nm (see Figure 8). In daylight such waters take on a bluish-green, and sometimes a green or greenish-yellow hue. Even though the great majority of Baltic basins are eutrophic, normally containing more than 1 mg chlorophyll *a* per cubic metre, these extremely high chlorophyll concentrations, reaching several tens of mg per cubic metre of water, are sporadic and usually occur during the phytoplankton bloom, mostly in bays, coastal waters and estuarine areas.

These characteristic features, distinguishing the irradiance fields in different waters of the Baltic and other natural basins, are due to the combined attenuation of light by various kinds of suspended particulate matter and CDOM (yellow substances), the levels of which in the Baltic are considerable (e.g. Håkanson & Eckhéll 2005, Kowalczuk et al. 2005, 2006). In general, both groups of substances strongly attenuate short-wave visible light, and their concentrations in all marine basins usually rise with increasing chlorophyll concentrations. The upshot of this is the shift of the maximum penetration of irradiance towards the long-wave end of the spectrum as we go from the optically clear, oligotrophic oceanic basins through mesotrophic basins to the biologically rich eutrophic seas.

We have used the measurements of downward irradiance spectra $E_{\mathrm{d}}(\lambda, z)$, made over many years in different regions and seasons at different depths in the Baltic, not only to characterize the above-mentioned qualitative features of natural light fields in this sea, but also in attempts to capture certain statistical quantitative trends, even though they are not sufficient for a fully representative statistical analysis (Woźniak & Hapter 1985, Woźniak

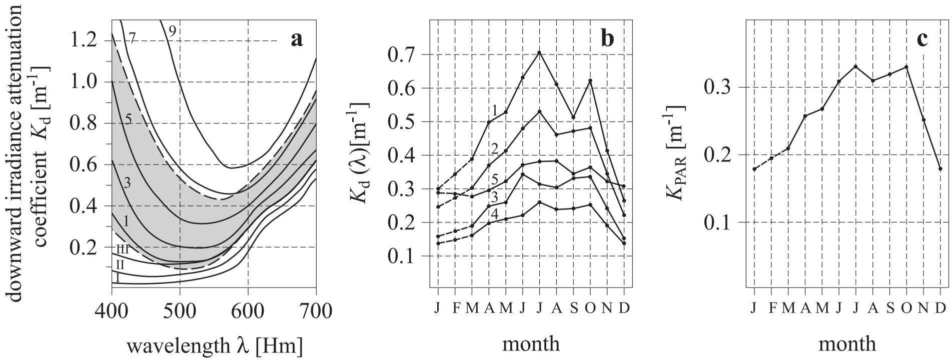

Figure 10. Typical downward irradiance attenuation coefficients K_{d} in the southern Baltic, calculated from statistically averaged empirical data for the upper layer of this sea (0–10 m), gathered between 1972 and 1984 (ca 3000 measurements) (after Woźniak & Hapter 1985)

a) Range of the most commonly measured values of the spectral downward irradiance attenuation coefficient $K_{\mathrm{d}}(\lambda)$ in the southern Baltic (the shaded area on the plot) compared with model spectra of this coefficient in different seas and oceans as classified by Jerlov (1978) (oceanic types I, II, III and coastal types 1–9)

b) Seasonal changes of typical (averaged) values of the spectral downward irradiance attenuation coefficient K_{d} for selected wavelengths (1–425 nm, 2–450 nm, 3–500 nm, 4–550 nm, 5–600 nm)

c) Seasonal changes of typical (averaged) values of the downward PAR irradiance attenuation coefficient K_{PAR}

et al. 1989). The most important results of this latter work are presented here in Figures 10 and 12, and in Tables 5 and 6. Taken together, they refer to various basins in the open southern Baltic; they were determined by averaging the 3000 or so spectral measurements of the downward irradiance $E_{\mathrm{d}}(\lambda, z)$ carried out in different Baltic basins and at different depths from 1972 to 1984 (Woźniak & Hapter 1985).

The penetration of downward irradiance into the sea is often described by the downward irradiance attenuation coefficient K_{d}. Being a function of depth z in the sea and of light wavelength λ, the spectral downward irradiance attenuation coefficient $K_{\mathrm{d}}(z, \lambda)$ is universally accepted as one of the basic optical characteristics of the sea. Its empirical values are determined from measured irradiance distributions $E_{\mathrm{d}}(\lambda, z)$, in accordance with the following definition (Jerlov 1976, Dera 1992, 2003):

$$K_{\mathrm{d}}(z, \lambda) = -\frac{1}{E_{\mathrm{d}}(z, \lambda)} \frac{dE_{\mathrm{d}}(z, \lambda)}{dz}. \tag{11}$$

Integration of equation (11) over depth z in the sea leads directly to the definition of the irradiance transmittance $T_{\mathrm{d}}(z, \lambda)$ through a layer of water from 0 to depth z:

$$T_{\mathrm{d}}(z, \lambda) = \frac{E_{\mathrm{d}}(z, \lambda)}{E_{\mathrm{d}}(z = 0, \lambda)} = e^{-K_{\mathrm{d}}(z, \lambda) z}. \tag{12}$$

This transmittance describes the relative decrease in irradiance with depth z in the sea; on the basis of the spectral downward irradiance attenuation coefficient $K_{\mathrm{d}}(z, \lambda)$ for a wavelength λ or K_{PAR}, i.e. for the entire PAR spectral interval, this decrease can be calculated, and so too the depth to which a given percentage of the surface irradiance penetrates. The transmittance thus enables the depth to be defined to which 1% of the surface irradiance penetrates, and hence the depth of the euphotic zone in a basin in accordance with the optical criterion, of which more below.

The plots in Figure 10 and the numerical data in Table 5 illustrate typical characteristics of the downward irradiance attenuation coefficients K_{d} in the upper layer of the southern Baltic. The range of the most frequently measured spectral downward irradiance attenuation coefficients $K_{\mathrm{d}}(\lambda)$ in this part of the Baltic has been defined as more or less the standard deviation of the mean of the above-mentioned large set of its empirical values. Compared with model spectra $K_{\mathrm{d}}(\lambda)$ in different seas and oceans as classified by Jerlov (1978), this range is indicated by the shaded area on Figure 10a. This shows that the waters of the southern Baltic can in most cases be allocated to coastal water types 1 to 7. Attenuation coefficients K_{d} are seasonally variable, being the lowest in winter and exhibiting two more or less distinct maxima in spring-summer and in autumn. This is illustrated by plots b and c in Figure 10 and the data in Table 5. This table lists the mean spectral downward irradiance attenuation coefficients $K_{\mathrm{d}}(\lambda)$ for particular months and for five different wavelengths, as well as the similar monthly values of K_{PAR}, i.e. the coefficient of attenuation of the total downward PAR irradiance in the spectral interval 400–700 nm. These values are typical of eutrophic basins.

Of course, it may happen, particularly in the winter months, that we come across water masses with optical properties approximating those of mesotrophic or even oligotrophic oceanic waters of type III in Jerlov's (1978) optical classification, and also super-eutrophic waters (e.g. during some spring phytoplankton blooms) resembling the coastal waters designated 7 and above in this classification. But these are rare occurrences in the southern Baltic.

Table 5. Mean values and standard deviations of the downward irradiance attenuation coefficient in the southern Baltic for each month in the year (columns 2–13) and for the whole year (column 14) (according to data collected in 1972–1984; after Woźniak & Hapter 1985)

No.	1	2	3	4	5	6	7	8	9	10	11	12	13	14
	Downward irradiance attenuation coefficient K_d (lines 1 to 10) and K_{PAR} (line 11) [m^{-1}]													
	λ [nm]	J	F	M	A	M	J	J	A	S	O	N	D	Year
1	425	0.300	0.350	0.392	0.502	0.534	0.626	0.711	0.613	0.509	0.618	0.411	0.264	0.468
				0.054		*0.069*	*0.103*	*0.153*	*0.056*	*0.076*	*0.056*	*0.036*		*0.137*
2	450	0.250	0.275	0.301	0.370	0.406	0.475	0.528	0.464	0.470	0.480	0.336	0.220	0.382
				0.054		*0.058*	*0.096*	*0.123*	*0.055*	*0.054*	*0.055*	*0.022*		*0.099*
3	475	0.195	0.210	0.235	0.293	0.322	0.379	0.398	0.370	0.414	0.398	0.289	0.181	0.306
				0.047		*0.056*	*0.079*	*0.089*	*0.040*	*0.044*	*0.051*	*0.007*		*0.081*
4	500	0.155	0.170	0.157	0.245	0.258	0.338	0.310	0.302	0.329	0.329	0.237	0.146	0.251
				0.038		*0.053*	*0.105*	*0.071*	*0.034*	*0.018*	*0.046*	*0.005*		*0.069*
5	525	0.135	0.150	0.162	0.202	0.221	0.266	0.245	0.250	0.263	0.278	0.198	0.122	0.208
				0.029		*0.033*	*0.087*	*0.053*	*0.034*	*0.020*	*0.037*	*0.004*		*0.052*
6	550	0.140	0.155	0.162	0.199	0.213	0.222	0.258	0.236	0.238	0.247	0.185	0.130	0.199
				0.029		*0.030*	*0.061*	*0.053*	*0.027*	*0.022*	*0.038*	*0.021*		*0.042*
7	580	0.220	0.229	0.219	0.248	0.264	0.286	0.323	0.306	0.280	0.299	0.244	0.225	0.261
				0.029		*0.036*	*0.075*	*0.083*	*0.027*	*0.030*	*0.043*	*0.008*		*0.036*
8	600	0.290	0.285	0.278	0.301	0.320	90.374	0.384	0.384	0.338	0.361	0.320	0.304	0.328
				0.039		*0.047*	*0.082*	*0.080*	*0.034*	*0.036*	*0.045*	*0.004*		*0.037*
9	620	0.360	0.350	0.340	0.370	0.396	0.465	0.433	0.477	0.400	0.422	0.396	0.370	0.403
				0.050		*0.062*	*0.093*	*0.084*	*0.046*	*0.044*	*0.049*	*0.026*		*0.039*
10	675	0.585	0.565	0.543	0.612	0.629	0.751	0.624	0.772	0.582	0.626	0.667	0.605	0.638
				0.080		*0.091*	*0.131*	*0.113*	*0.072*	*0.088*	*0.111*	*0.105*		*0.064*
11	400–700	0.183	0.197	0.207	0.259	0.270	0.307	0.327	0.313	0.317	0.327	0.248	0.177	

Table 6 presents the annual variations in the depth of the euphotic zone z_e in the southern Baltic (according to the optical criterion[8]), that is, the depth z to which 1% of the downward irradiance entering the sea penetrates – in this case irradiance with a wavelength corresponding to the position of the maximum irradiance penetration in the basin concerned. These variations in the euphotic zone depth during the year are linked to the growing season: this depth is far greater in winter (down to 35 m) than in the summer (ca 18 m).

Table 6. Mean depth of the euphotic zone z_e according to the optical criterion for the southern Baltic for each month of the year (according to data between 1972 and 1984; after Woźniak & Hapter 1985)

Month	J	F	M	A	M	J	J	A	S	O	N	D
z_e [m]	32	30	28	23	22	21	18	20	19	19	25	35

At this point it should be recalled that irradiance under water is subject to even more complex fluctuations than that in the atmosphere. The main reason for this is the direct transfer to the water of irradiance fluctuations due to variations in cloudiness, which were discussed in section 3 (Figure 2) and addition fluctuations caused by the wave motion of the sea surface, these last fluctuations can be described as short-term (lasting for fractions of a second) (Dera & Olszewski 1967, Dera & Gordon 1968, Snyder & Dera 1970). The complex interaction of waves with the Sun's rays crossing the water surface, and in particular the focusing of these rays by wave crests, gives rise to very strong fluctuations of the underwater irradiance in the upper water layer (Figure 11). Measured at a given point in the water, these fluctuations manifest themselves as light pulses, sometimes exceeding by two and occasionally five or more times the mean solar irradiance measured over a few minutes, depending on the optical conditions prevailing in the atmosphere and water. Such fluctuations are referred to as light flashes (Dera & Stramski 1986, Stramski 1986, Stramski & Dera 1988,

[8]The depth of the euphotic zone, i.e. the thickness of the upper layer of the sea in which the majority of the primary production of organic matter in the sea occurs, is taken to be the compensation depth of photosynthesis. This is the depth at which the rates of metabolic and catabolic processes in phytoplankton are the same. Above this depth, therefore, the rate of release of oxygen during photosynthesis is greater than its consumption during respiration, whereas below this depth the opposite situation obtains. Steemann Nielsen (1975) and Koblentz-Mishke et al. (1985) showed that the compensation depth of photosynthesis is usually very close to the depth to which penetrates 1% of the surface irradiance in the spectral band corresponding to the maximum transmittance of this irradiance in a given sea.

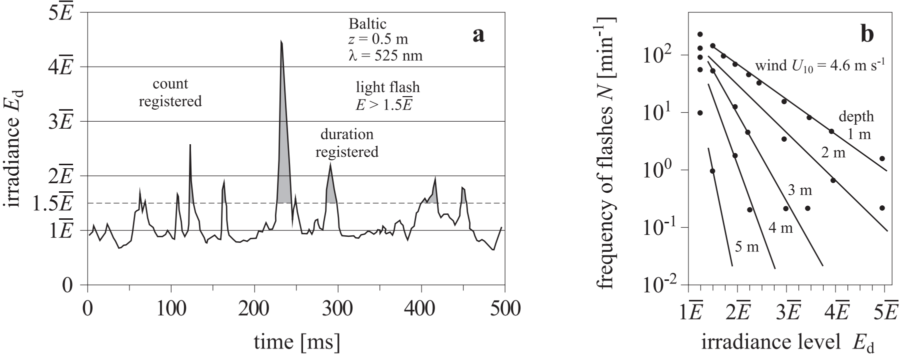

Figure 11. Short-term downward irradiance fluctuation under a wind-roughened sea surface:

a – typical time record of short-term fluctuations. Light flashes are marked as shaded pulses; z – depth in the water, λ – light wavelength

b – typical frequency distribution of flash intensity in the Baltic at wind speed $U_{10} = 4.6$ m s^{-1}, with depth as parameter (Reproduced from Dera & Stramski 1986)

Dera et al. 1993, You 2010). For lack of space in this article, we refer the reader to the detailed descriptions of these fluctuations in the above publications.

Because irradiance in the sea fluctuates strongly, its spectral and energy characteristics are usually expressed in time-averaged form, e.g. as daily

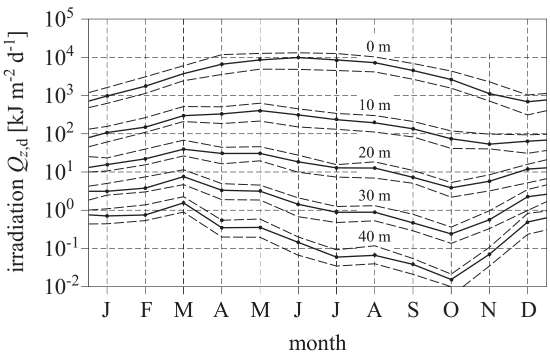

Figure 12. Seasonal variations in the typical (averaged) daily irradiance doses (i.e. irradiation), at various depths in the southern Baltic in the PAR spectral region (400–700 nm). The broken lines indicate the values for cloudless skies or minimal cloud cover, i.e. 0–0.2 on a 0–1 scale (above the solid lines) and for overcast skies, i.e. 0.8–1 (below the solid lines) (after Woźniak et al. 1989, Oceanologia 27, p. 69, Figure 5)

Table 7. Mean daily doses of downward irradiance energy Q_{PAR} [kJ m^{-2} day^{-1}] in the PAR spectral interval for each month of the year in the southern Baltic (according to data collected in 1972–1984) (based on data in Woźniak & Hapter 1985, Woźniak et al. 1989)

a, line 1 – at the sea surface, averaged over all the days in the period studied, regardless of cloudiness, **a**, line 2 – at the sea surface for cloudless days or days with minimal cloudiness up to n = 0.2 on a 0–1 scale, **a**, line 3 – at the sea surface for overcast days (cloudiness n $\geq$ 0.8);

b – at selected depths in the sea, values averaged over all the days in the period studied, regardless of cloudiness

a

	Month											
	J	F	M	A	M	J	J	A	S	O	N	D
1	862	1653	3858	6092	8260	9573	8335	7395	4321	2497	906	598
2	1360	2783	5462	8409	11556	12121	11368	9736	5793	3474	1745	979
3	590	1172	2239	3307	4504	4679	4395	3851	2340	1218	762	318

b

Depth	Month											
[m]	J	F	M	A	M	J	J	A	S	O	N	D
0	804	1628	3851	5989	8182	9475	8270	7134	4275	2416	955	572
0.05	793	1605	3797	5887	8035	9276	8096	6984	4194	2368	939	561
0.1	782	1582	3743	5785	7896	9087	7931	6842	4113	2319	923	550
0.25	751	1514	3501	5492	7487	8528	7443	6421	3877	2179	878	518
0.5	702	1408	3331	5037	6848	7675	6699	5779	3523	1967	807	471
1	613	1218	2877	4234	5727	6216	5425	4680	2898	1602	681	388
1.5	535	1052	2484	3563	4795	5031	4391	3788	2390	1305	577	319
2	463	921	2153	2989	3944	4036	3523	3039	1949	1056	484	263
4	287	555	1271	1605	2037	1914	1646	1448	936	495	265	166
6	186	348	786	904	1113	966	817	735	475	246	152	109
8	124	226	501	529	629	517	426	392	247	127	90.1	74.4
10	84.4	151	327	317	366	285	229	215	133	67.4	54.8	52.6
12	58.9	103	216	193	217	161	127	121	73.5	36.7	33.9	37.8
14	41.4	70.5	145	120	131	93.1	70.7	69.1	41.5	20.3	21.1	26.9
16	29.3	49.0	99.0	74.9	79.6	54.4	40.1	39.9	23.6	11.3	13.7	19.4
18	21.0	34.4	67.8	47.1	49.0	32.2	23.1	23.3	13.6	6.43	8.72	14.3
20	15.1	24.1	46.6	30.1	30.3	19.2	13.2	13.3	7.95	3.67	5.64	10.4
25	6.79	10.3	18.8	9.94	9.33	5.46	3.48	3.69	2.11	0.92	1.95	5.37
30	3.16	2.46	7.74	3.35	2.96	1.60	0.93	1.03	0.58	0.24	0.69	2.87
40	0.69	0.88	1.37	0.40	0.31	0.14	0.07	0.08	0.04	0.02	0.09	0.75

daily means or monthly means, or as daily or monthly doses of energy i.e. daily totals or monthly totals, or irradiation in a given time period Δt. Such statistical values of the downward irradiance doses in the PAR spectral range in [kJ m^{-2} day^{-1}] at different depths in the southern Baltic are given in Figure 12 and Table 7.

Besides the mean *PAR* irradiance doses, this figure and the table respectively show graphically (the broken curves on the figure) or list the corresponding values calculated for extreme weather conditions, i.e. for cloudless days or days with minimal cloudiness up to $n = 0.2$ on a 0–1 scale, and for overcast skies (cloudiness $n \geq 0.8$). The data shown in Figure 12 and listed in Table 7 may be useful for estimating the absolute amounts of solar radiation energy arriving in southern Baltic ecosystems at different seasons in the year.

6. Model characteristics of irradiance in the Baltic and formulas enabling their determination using remote sensing methods

Unfortunately, the statistical generalizations presented in the previous section are only fragmentary and do not adequately reflect the variability in time and space of the optical characteristics of Baltic waters that we have been discussing. This is due to objective limitations: suffice it to mention the considerable expense that the systematic measurement of the essential empirical data using classical in situ measurement techniques would incur. Fortunately, however, it is becoming easier to obtain such empirical data by means of the remote sensing techniques now routinely used in oceanography. Among other things, these enable the chlorophyll *a* concentration in the surface layer of the sea to be measured with satisfactory accuracy. The correlation of certain properties of waters with the concentration of that chlorophyll, including optical properties, makes it possible to employ these satellite techniques for the efficient and systematic determination of the optical properties of waters over large areas of the sea. To this end, however, we need a rather complicated bio-optical/mathematical model of these properties along with an appropriate algorithm, derived specifically for a particular type of sea water. As a result of our long-term bio-optical studies in the Baltic Sea (Darecki et al. 2008, Woźniak et al. 2008), our research team derived such a model and its algorithm (DESAMBEM[9])

[9]This algorithm was developed jointly by scientists from the Institute of Oceanology (Polish Academy of Sciences), the Institute of Oceanography (University of Gdańsk) and the Institute of Physics (Pomeranian Academy, Słupsk) within the framework of the project, carried out in 2001–05, commissioned by the Committee for Scientific Research (Project No. PBZ-KBN 056/P04/2001) entitled 'DEvelopment of a SAtellite Method for Baltic Ecosystem Monitoring – DESAMBEM'. It is currently undergoing further development within the framework of a new project (Project POIG.01010222011/0900) entitled 'The satellite-based monitoring of the Baltic Sea environment (SatBałtyk)',

for Baltic waters. With this algorithm the structural and functional characteristics of the Baltic ecosystem can be defined on the basis of available remote sensing data. Some of these characteristics can be defined directly using the standard formulas already applied in teledetection and also simple optical models of the atmosphere (see e.g. Krężel 1997, Ruddick et al. 2000, Darecki & Stramski 2004, Krężel et al. 2005). They include the surface water temperature in the sea, the natural irradiance of the sea surface and the concentration of chlorophyll a in the surface layer of the water. On the basis of these directly estimated characteristics, the DESAMBEM algorithm can be invoked to determine indirectly other characteristics of a whole range of processes taking place in the Baltic ecosystem, including the attenuation of irradiance and the photometric magnitudes characterizing solar radiation in this sea. This is made possible by two statistical models developed by our team for the Baltic Sea – they constitute the foundations of two integral parts of the DESAMBEM algorithm, not to mention others. They are:

- the statistical model of the vertical distributions of the chlorophyll a concentration $C_{\mathrm{a}}(z)$ in the Baltic Sea;
- the bio-optical model of Baltic Case 2 waters.

The first of these models, the statistical model of the vertical distributions of the chlorophyll a concentration $C_{\mathrm{a}}(z)$ in the Baltic Sea, was described by Ostrowska et al. (2007). The basic formula of this model links chlorophyll a concentrations at different depths $C_{\mathrm{a}}(z)$ with the surface concentration of this chlorophyll $C_{\mathrm{a}}(0)$:

$$C_{\mathrm{a}}(z) = C_{\mathrm{a}}(0) \frac{A + B \exp\left[-(z - z_m)^2 \sigma \right]}{A + B \exp\left[-(z_m)^2 \sigma \right]}, \tag{13}$$

where $C_{\mathrm{a}}(z)$ and $C_{\mathrm{a}}(0)$ are expressed in [mg tot. chl a m^{-3}], z and z_m are expressed in [m], σ is expressed in [m^{-2}], and A and B are dimensionless; in addition,

$$A = 10^{(1.38 \log(C_{\mathrm{a}}(0)) + 0.0883)},$$

$$B = 10^{(0.714 \log(C_{\mathrm{a}}(0)) + 0.0233)},$$

$$z_m = -4.61 \log(C_{\mathrm{a}}(0)) + 8.86,$$

$$\sigma = 0.0052.$$

financed from European Union funds. Besides the three institutes already mentioned, a research team from the Institute of Marine Science (University of Szczecin) is also participating in this new project.

The second model, the bio-optical model of Baltic Case 2 waters, came into being following appropriate modification and extension (see Kaczmarek & Woźniak 1995) of the earlier bio-optical model for Oceanic Case 1 waters (see Woźniak et al. 1992a,b). This model links various optical properties and the underwater irradiance field $E_{\mathrm{d}}(\lambda, z)$, e.g. with the chlorophyll a concentration at different depths $C_{\mathrm{a}}(z)$. The fundamental formula of this model describes the spectral downward irradiance attenuation coefficient $K_{\mathrm{d}}(\lambda)$ as follows:

$$K_{\mathrm{d}}(\lambda) = K_{\mathrm{w}}(\lambda) + K_{\mathrm{pl}}(\lambda) + K_{\Delta}(\lambda) + \Delta K(\lambda), \tag{14}$$

where the terms on the right-hand side of the equation stand for the partial coefficients of the downward irradiance attenuation: K_{w} – by pure water, K_{pl} – by phytoplankton in the water, K_{Δ} – by other autochthonous optically active substances (detritus, yellow substances) in the water, and ΔK – by allochthonous admixtures (mineral suspensions, river-borne material etc.).

According to Morel & Prieur (1977), to a good approximation these two cases of marine waters are distinguishable. Case 1 waters (making up ca 98% oceanic waters), in which practically all the optically active admixtures are autochthonous, come into being as a result of the functioning of the local marine ecosystem. In such waters we can assume that $\Delta K \approx 0$ and that there is a good correlation between K_{d} and the concentration C_{a} of chlorophyll a in the water, since it is the latter that governs the production of organic matter in the ecosystem. In contrast, Case 2 waters additionally contain substantial quantities of allogenous substances entering these waters from rivers, coasts and the atmosphere, which are specific to a given region and season (e.g. Darecki et al. 2003). In Case 2 waters ΔK is conspicuously larger than 0 and must be allowed for in equation (14). Analysis of the correlations between the overall coefficient $K_{\mathrm{d}}(\lambda)$ and its constituent coefficients $K_{\mathrm{pl}}(\lambda)$ and $K_{\Delta}(\lambda)$, and also the chlorophyll a concentration, followed by nonlinear approximation, yielded the final form of the function describing the dependence of $K_{\mathrm{d}}(\lambda)$ on $C_{\mathrm{a}}(z)$:

$$K_{\mathrm{d}}(\lambda, z) = K_{\mathrm{w}}(\lambda) + C_{\mathrm{a}}(z)\{C_1(\lambda)\exp[-a_1(\lambda)C_a(z)] + K_{\mathrm{d,\,i}}(\lambda)\} + \Delta K(\lambda), \tag{15}$$

where

$$\Delta K(\lambda) = \begin{cases} 0 & \text{for oceanic Case 1 waters} \\ 0.068 \exp[-0.014(\lambda - 550)] & \text{for Baltic Case 2 waters} \end{cases}$$

where ΔK is expressed in [m^{-1}] and λ in [nm].

Table 8. Values of the parameters in the bio-optical classification of seas used in model equation (15) (after Woźniak et al. 1992a,b)

λ [nm]	a_1 [m^3(mg tot. chl a)$^{-1}$]	C_1 [m^2(mg tot. chl a)$^{-1}$]	$K_{\mathrm{d,i}}$ [m^2(mg tot. chl a)$^{-1}$]	K_{w} [m^{-1}]
400	0.441	0.141	0.0675	0.0209
410	0.495	0.137	0.0643	0.0197
420	0.531	0.131	0.0626	0.0187
430	0.580	0.119	0.0610	0.0177
440	0.619	0.111	0.0609	0.0176
450	0.550	0.107	0.0569	0.0181
460	0.487	0.0950	0.0536	0.0189
470	0.500	0.0970	0.0479	0.0198
480	0.500	0.0780	0.0462	0.0205
490	0.509	0.0774	0.0427	0.0230
500	0.610	0.0672	0.0389	0.0276
510	0.594	0.0598	0.0363	0.0371
520	0.590	0.0610	0.0319	0.0473
530	0.693	0.0573	0.0288	0.0513
540	0.606	0.0506	0.0285	0.0567
550	0.514	0.0432	0.0274	0.0640
560	0.465	0.0425	0.0248	0.0720
570	0.384	0.0288	0.0240	0.0810
580	0.399	0.0230	0.0231	0.107
590	0.365	0.0180	0.0231	0.143
600	0.333	0.0171	0.0225	0.212
610	0.304	0.0159	0.0216	0.236
620	0.316	0.0150	0.0225	0.264
630	0.421	0.0183	0.0225	0.295
640	0.420	0.0216	0.0226	0.325
650	0.346	0.0164	0.0236	0.343
660	0.348	0.0141	0.0260	0.393
670	0.173	0.00939	0.0267	0.437
675	0.173	0.00436	0.0270	0.455
680	0.173	0	0.0258	0.478
690		0	0.0190	0.535
700		0	0.0125	0.626
710		0	0.0045	1.000
720		0	0.0014	1.360
730		0	0.00041	1.810
740		0	7.1×10^{-5}	2.393
750		0	1.3×10^{-5}	2.990

The constants $C_1(\lambda)$, $a_1(\lambda)$ and $K_{\mathrm{d,i}}(\lambda)$, and the attenuation coefficient for clear water $K_{\mathrm{w}}(\lambda)$, are given in Table 8.

A graphic representation of some results of calculations done using the bio-optical model of Baltic Case 2 waters, applicable to various optical characteristics of Baltic basins, is shown in Figure 13. This shows, among other things, spectra of the downward irradiance attenuation coefficients K_{d}, characteristic of the surface layers of waters in the various trophic types of Baltic basin (Figure 13a). These spectra can be calculated for any concentration of chlorophyll using the model parameterization of Baltic Case 2 waters (eq. (15)). The systematic use of remotely sensed surface concentrations of chlorophyll a $C_{\mathrm{a}}(0)$ may provide particularly significant opportunities for applying such calculations; this will lead to a fuller understanding of the variability of the energy inflow and optical properties of this sea. Some maps of the distribution of this chlorophyll in the Baltic will be found, for example, in the already cited articles describing the DESAMBEM algorithm (Darecki et al. 2008, Woźniak et al. 2008). We must always bear in mind, however, that a remotely-sensed concentration of chlorophyll provides information only about its surface concentration, that is, its concentration in a sea surface layer of undetermined thickness; such data has to suffice for calculating the optical properties of water under discussion here. This comment also applies to many other physical, chemical and biological properties of the sea associated with chlorophyll concentrations which, estimated on the basis of remotely-sensed chlorophyll concentrations $C_{\mathrm{a}}(0)$, refer solely to the surface layer of water in the basin concerned. If we wish to extend such model estimates associated with the chlorophyll a concentration to deeper water layers in the sea, then we have to take the changes in these concentrations with depth into account. That is why the model optical characteristics of Baltic waters, shown on the various plots in Figure 13, were determined on the basis of the bio-optical parameterization described above (see eq. (15)), at the same time employing the changes in chlorophyll a concentration with depth in the sea described by the statistical model of the vertical distributions of chlorophyll a concentrations $C_{\mathrm{a}}(z)$ in the Baltic (eq. (13)). Figure 13 presents the following plots for Baltic Case 2 waters of different trophicity: model spectra of the downward irradiance attenuation coefficient $K_{\mathrm{d}}(\lambda)$ (Figure 13a); model PAR irradiance transmittance $T_{PAR} = PAR(z)/PAR(0)$ (Figure 13b); example, model spectral distributions of relative irradiances $f_E = E_{\mathrm{d}}(\lambda, z)/PAR(0)$, at different depths in the Baltic, for two widely differing surface concentrations of chlorophyll a (Figures 13c and d); model dependences of the depth of PAR irradiance penetration in Baltic basins on the surface chlorophyll a concentration (Figure 13e); model dependence of the euphotic zone depth z_{e} on the surface concentration of chlorophyll a $C_{\mathrm{a}}(0)$ (Figure 13f).

The model plots in Figure 13 are a good representation of these selected optical properties of Baltic Case 2 waters. Plots c and d on this figure clearly

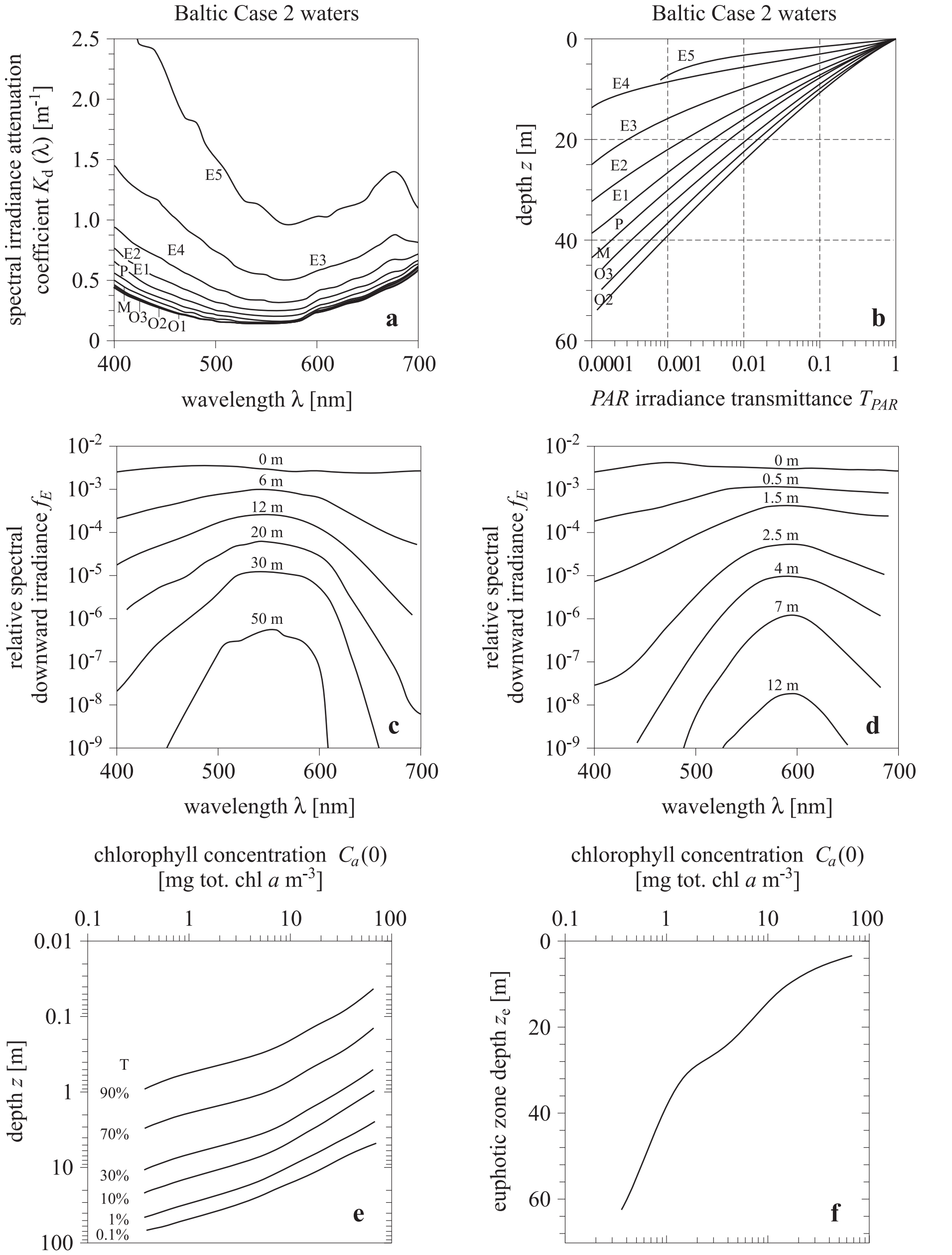

Figure 13. Graphic representation of the optical characteristics of Baltic waters calculated using the bio-optical model of the Baltic:

a – Model spectral distributions of the downward irradiance attenuation coefficient K_d in Baltic Case 2 waters of various trophic types, with surface chlorophyll a concentrations C_a [mg tot. chl a m^{-3}] as follows: 0.035 (O1); 0.07 (O2); 0.15 (O3); 0.35 (M); 0.7 (I); 1.5 (E1); 3.5 (E2); 7 (E3); 15 (E4); 35 (E5)

Figure 13. (*continued*)

b – Model *PAR* irradiance transmittance $T_{PAR} = PAR(z)/PAR(0)$ in Baltic Case 2 waters with different surface chlorophyll *a* concentrations (as in Fig. a)

c and d – Model spectral distributions of relative irradiances $f_E = E_{\rm d}(\lambda, z)/PAR(0)$, at different depths in the Baltic, for two widely differing surface concentrations of chlorophyll *a* $C_{\rm a}(z = 0)$: 3.5 mg tot. chl *a* m^{-3} (c) and 70 mg tot. chl *a* m^{-3} (d)

e – Model dependences of the *PAR* irradiance penetration in Baltic basins $z_{T=\rm const}$ for various *PAR* irradiance transmittances T_{PAR} [%] on surface chlorophyll *a* concentration $C_{\rm a}(z = 0)$

f – Model dependence of the euphotic zone depth $z_{\rm e}$ on the surface chlorophyll *a* concentration $C_{\rm a}(0)$ in southern Baltic Case 2 waters

illustrate the conspicuous narrowing of the irradiance spectrum with depth in the sea and the shift of its maximum transmittance towards the long-wave region of the spectrum with increasing chlorophyll *a* concentration. This endorses the conclusions drawn earlier on the basis of the empirical data shown in Figure 9.

7. Summary

The optical properties of Baltic waters and the influx of solar radiation energy to this sea have been the subject of a great many studies and have been described in a plethora of publications, dating largely from 1960 to the present. But because these properties are extremely variable in time and space, and because to date they have not been monitored on a systematic basis, we do not possess sufficiently reliable statistical distributions of their values. We do, however, know the long-term mean monthly values of the solar radiation influx incident on the surface of the Baltic (with an accuracy defined by standard deviations of ca 10%), and we also know, for most of the Baltic basins, the typical values, ranges of variability and spectral characteristics of the penetration of this energy into their waters: this is what we describe in the present article.

The long-term monthly mean totals of the incident solar radiation fluxes at the surface of the entire Baltic Proper vary in the annual cycle from a minimum of 31.7 MJ m^{-2} in December to a maximum of 624.9 MJ m^{-2} in June; the yearly mean total is 3680 MJ m^{-2} (see Table 2). In different years, the real monthly totals may vary widely, however (see the maximum, minimum and standard deviation in Figure 5 and Table 2). There are also conspicuous differences in this radiation influx between various regions of the Baltic (see Figure 3 and Table 2a,b,c).

The mean transmittance of the monthly mean total flux across the Baltic Sea surface is ca 93% (from 90–93% in the winter months to 94–96% in the summer months – see Table 4). Having entered Baltic waters, the broad spectrum of solar radiation wavelengths (Figure 1) already at a depth of ca 2 m narrows down to the visible light region (identical with the PAR region) (see Figure 6), and the farther this light penetrates down into the water, the narrower its spectrum becomes (see Figures 8, 9b, 13). The reason for this is that, rather than the wavelengths from the middle region of the visible spectrum becoming attenuated, it is the long waves that are attenuated by water molecules and the short waves by the ubiquitous yellow organic substances dissolved and suspended in Baltic waters. Hence, the measured and often published values of the PAR irradiance and the coefficients of its attenuation with depth in the Baltic are, with the exception of the roughly 2 m deep surface layer, equal to the irradiance and the attenuation coefficients of the entire solar radiation energy that penetrates these waters (Figures 10c, 13b; Table 5, line 11). At the same time, the spectral maximum of this penetration (the minimum of the downward irradiance attenuation coefficient $K_{\rm d}(\lambda)$) shifts from the green (ca 500 nm) to the yellow (ca 550 nm) and even brown wave interval (ca 600 nm) with increasing trophicity of the waters, as measured by the concentration of chlorophyll a in the water (see Figures 8, 9, 13a). The greatest depth to which this light can practically penetrate determines the depth of the euphotic zone, which according to the optical criterion ranges on average from ca 18 m in July to ca 35 m in December in the southern Baltic (see Table 6, Figure 13f).

Some 80% of the total solar radiation energy entering Baltic waters is absorbed in this surface layer down to a mean depth of 2 m, and in its supereutrophic basins the top half metre or so absorbs the same percentage of this energy (see Figure 7).

A crucial result of the numerous bio-optical investigations in the Baltic is the bio-optical model of Baltic Case 2 waters, outlined in section 6. This model allows us to systematize our knowledge of the quantitative characteristics of the solar radiation energy influx at different depths in the various Baltic basins by linking mathematically the coefficients of irradiance attenuation and the chlorophyll concentration in these waters, i.e. their trophicity (see formula (15)). Most importantly, however, this model is incorporated in the DESAMBEM algorithm. This provides the theoretical foundations for comprehensive, remote sensing methods (especially from on board satellites) for the monitoring and further investigation of the state and functioning of the Baltic ecosystem, including the monitoring of the characteristics of the radiation energy influx and its penetration into the sea. The accuracy of this algorithm is as good as that of similar algorithms derived for other marine areas (eg. Antoine & Morel 1996, Morel et al. 1996), indeed, in some respects it is superior, in that the errors of estimation of the relevant characteristics are distinctly smaller (see Darecki et al. 2008, Woźniak et al. 2008). The DESAMBEM algorithm also has a broader range of application. The SatBałtyk project (see Dera 2010), currently in progress, bodes well: in the near future it should be possible to effectively implement this algorithm in practice. It will facilitate the efficient remote monitoring and diagnosis of the state of the Baltic and some important processes taking place there, including the influx of solar radiation energy and its utilization by the sea's ecosystems. This will make a very significant contribution to our knowledge of the influx of light and the part it plays in the Baltic Sea.

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Annex 1

External mixing approach

Symbol	Explanation	Unit
1	2	3
CDOM	Coloured dissolved organic matter	
E_d	Downward irradiance (in the horizontal plane)	W m^{-2}
$E_{d,\,surf}$	Downward irradiance just above the sea surface	W m^{-2}
$E_d(z,\,\lambda)$	Downward irradiance of wavelength λ in the water column at depth z	W m^{-2}
$E_d(z=0,\,\lambda)$	Downward irradiance of wavelength λ just below the water surface (at depth $z=0$)	W m^{-2}
E_S	Instantaneous value of the downward irradiance at TOA at the zenith over the entire spectral range of solar radiation	W m^{-2}
$K_d(z,\,\lambda)$	The attenuation coefficient of the downward irradiance of wavelength λ at depth z in a basin: $K_d(z,\,\lambda) = -\frac{1}{E_d(z,\,\lambda)}\frac{dE_d(z,\,\lambda)}{dz}$	m^{-1}
K_{pl}	The attenuation coefficient of the downward irradiance of phytoplankton	m^{-1}
K_w	The attenuation coefficient of the downward irradiance of pure water	m^{-1}
K_Δ	The attenuation coefficient of the downward irradiance of autochthonic optically active substances other than pure water and phytoplankton (detritus, yellow substances)	m^{-1}
ΔK	The attenuation coefficient of the downward irradiance of allochthonic admixtures (mineral suspensions, river-bornematerial etc.)	m^{-1}
PAR	Spectral range of Photosynthetically Available Radiation, i.e. radiation in the spectral range from c. 400 nm to 700 nm	
PAR	Downward irradiance in the Photosynthetically Available Radiation spectral range (from c. 400 nm to 700 nm)	W m^{-2}
$Q_{surf,\,\Delta t}$	Sum of radiant energy reaching unit surface area during time interval Δt	J m^{-2}
$Q_{z,\,d}$	Sum of radiant energy reaching unit horizontal surface area at depth z in a basin during the day	J m^{-2}
S	Instantaneous solar radiation flux reaching the top of the Earth's atmosphere over the entire spectral range of this radiation (i.e. the solar irradiance of a surface normal to the rays at the upper boundary	W m^{-2}

External mixing approach *(continued)*

1	2	3
	of the atmosphere) $S = S_o f(t)$, where t – time (number of the day in the year)	
S_o	Solar constant, i.e. the mean annual solar irradiance of a surface normal to the rays at the upper boundary of the atmosphere (the top of the atmosphere (TOA) or outside the atmosphere)	W m^{-2}
T_a	Transmittance of the total downward irradiance through the atmosphere ($E_{d,\,surf}/E_S$)	dimensionless
T_{clouds}	Transmittance of the total downward irradiance through a layer of clouds	dimensionless
T_d	Irradiance transmittance through a layer of water from 0 to depth z	dimensionless
T_o	Transmittance of the total downward irradiance through a clear cloudless atmosphere ($E_{d,\,surf,\,with\,cloudless\,atmosphere}/E_S$)	dimensionless
T_{surf}	Transmittance of the total downward irradiance across the sea surface	dimensionless
TOA	Top of the atmosphere	
dn	Number of day in the year	dimensionless
h	Solar altitude above the horizon	deg
t	Time	S
z	Depth in the sea	m
Λ	Longitude	deg
Φ	Latitude	deg
θ	Angle of the Sun's inclination from the zenith i.e. solar zenith angle	deg

Annex 2

Chlorophyll *a* concentration C_a as the water trophicity index

Symbol of trophicity type	Chlorophyll *a* concentration C_a range in the water [mg m^{-3}]	Name of trophicity type
O-1	0.02–0.05	oligotrophic waters type 1 (ultra-oligotrophic)
O-2	0.05–0.10	oligotrophic waters type 2
O-3	0.10–0.20	oligotrophic waters type 3
M	0.20–0.50	mesotrophic waters
I	0.50–1.00	intermediate waters
E-1	1.00–2.00	eutrophic waters type 1
E-2	2.00–5.00	eutrophic waters type 2
E-3	5.00–10.0	eutrophic waters type 3
E-4 and more	10.0–20.0	eutrophic waters type 4 (highly euphotic)

From Woźniak et al. 1992.