Papers

Solar radiation at the Baltic Sea surface*

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Solar radiation
Atmosferic
transmittance
Baltic Sea

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Abstract

A semi-empirical model for computing monthly totals of solar radiation at the Baltic sea surface has been carried out. Readily available meteorological data, like: atmospheric pressure, water vapour pressure and cloudiness, measured near the sea surface constituted the input parameters. In comparison to well-known, simplified empirical methods of computations, the present model specifies local atmospheric parameters effect on solar radiation incoming on the sea surface. Using mean monthly values of above-mentioned meteorological parameters for 10 years (1965—1974), monthly totals of global solar radiation were calculated for 23 meteorological stations at the Baltic area. On the basis of these calculations and the actinometric data collected at 6 existing stations, there were made the interpolated distributions of mean monthly totals of solar radiation at the Baltic surface for every month in a year.

1. Introduction

The Earth as a planet receives virtually all of its energy from solar radiation. The solar radiation incoming at a point on the earth surface within a given period is subject to changes defined basically by the spherical form of the Earth and laws describing the position with respect to the sun. Further differences in the amount of solar radiation, incident on the earth surface, result from the geographical variability of physical properties of the earth's atmosphere and its albedo.

On the other hand, the knowledge of the radiative energy input to the surface of the sea as a part of the earth surface is essential for evaluation of the heat budget of the sea and its effect on the climate, for the modeling of hydrodynamic and thermodynamic processes associated with the sea-air interaction, for studies covering such domains as the biology of the sea, photochemistry *etc.*

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In the Baltic Sea, where the exchange of waters (and thus also the energy exchange) with the ocean is slight, the solar radiation is the basic source of energy. Up to now the input of solar radiation energy to this region has not been recognized sufficiently well in spite of its evident importance to diverse environmental investigations. The actual value of this parameter – at any time and any place on the surface of the Baltic – may be evaluated only in virtue of generalized radiation atlases (such as *Atlas teplovogo balansa okeanov*, 1970) or by direct measurements.

The few existing works, of Pomeraniec (1970), Laevastu (1960), Hankimo (1964), Defant (1970), Dera and Wensierski (1979), Czyszek *et al.* (1979), either consider the solar radiation at the Baltic Sea surface as one of many components of the heat budget (which leads to a somewhat generalized evaluation) or apply to some specific location or time period.

Non-uniform distribution and small number of actinometric stations in the Baltic Sea area make necessary to utilize such indirect method of evaluating the amount of incident solar radiation; that would allow to calculate this amount based on standard meteorological data.

Thus, to characterize the problem under consideration three basic objectives have been formulated and their solution is given in this paper. These are:

(i) Development of a semi-empirical model for calculation of the energy of solar radiation at the Baltic Sea surface based on the readily available standard meteorological data;

(ii) Calculation, using the model developed, of mean monthly totals of global solar radiation in the Baltic Sea area and making interpolated distributions of this parameter for each month of the year;

(iii) Evaluation of contribution of the principal components of solar radiation attenuation in the atmosphere over the Baltic Sea to the overall attenuation of the radiation along its path through the atmosphere down to the sea surface.

2. Calculation of the radiation flux

2.1. Basic equations and assumptions

The solar radiation flux penetrating the earth's atmosphere is subject to the following processes:

– molecular scattering (Rayleigh scattering),

– absorption by water vapour,

– scattering and absorption by so called atmospheric aerosols,

– absorption by ozone,

– absorption by other gaseous components of the atmosphere, regarded in what follows, and called the constant components of the atmosphere.

These processes affect both the directly incident and scattered radiation. They are selective, that is their intensity depends considerably on the incident radiation wavelength.

Thus for determination of the total energy of the whole spectrum of solar radiation that reaches the sea surface at any time and location the processes as mentioned above should be accounted for in the radiation transfer equation, which has for the earth's atmosphere the following form:

$$\frac{dL(\vec{r}_p,\vec{\xi})}{dr} = -cL(\vec{r}_p,\vec{\xi}) + \frac{b}{4\pi}\int_{\Omega} L(\vec{r}_p,\vec{\xi}')\,p(\vec{r}_p,\vec{\xi},\vec{\xi}')\,d\omega(\vec{\xi}'), \tag{1}$$

where:

$L(\vec{r}_p,\vec{\xi})$ – the radiance impinging on the volume element at $d\vec{r}$,

$\vec{r}$ – the directional vector,

$\vec{\xi}$ – $\left(\frac{dx}{dr}, \frac{dy}{dr}, \frac{dz}{dr}\right)$,

$\frac{d}{dr} = \vec{\xi}\nabla$ – the Lagrange derivative,

c; b – the total volume coefficients of light attenuation and scattering respectively,

p – the phase function (the scattering indicatrix) defining the amount of radiative energy scattered at given point into a solid angle element $d\omega(\vec{\xi}')$,

Ω – the solid angle of 4π.

Equation (1), has been written in a time-independent form with the black body radiation sources in the atmosphere omitted, as the rate of radiation changes in the atmosphere is very low relatively to the velocity of light and the subject of interest is the solar radiation the energy of which is in 99% contained within the spectral range of 0.28 – 5 μm. It follows from the equation that for determination of the radiation field in the atmosphere knowledge of spatial distributions and dependence on time of the total attenuation coefficient c, scattering coefficient b and phase function p is necessary (all of them represented in a wavelength-dependent form), together with formulation of appropriate boundary conditions.

However, variability and complex structure of the earth's atmosphere bulk make such time-and co-ordinate-dependent representation impossible. Hence, equation (1) should be subjected to further simplifications to become analytically solvable. Various assumptions have allowed to solve this equation using methods known in the literature of the subject under the following names (reflecting the most characteristic features of the simplifying assumptions): two-stream approximation (Chandrasekhar, 1950; Goody, 1964; Shettle and Weinman, 1970; Wang and Domoto, 1974; Liou, 1974; Paltridge and Platt, 1976; Schaller, 1979; Schmetz and Raschke, 1979), discrete-ordinates method (Chandrasekhar, 1950; Goody, 1964; Liou, 1975) and numerical methods applied to simplified equation (1) and using: the Monte Carlo method (Colling and Wells, 1965), 'adding' (Van de Hulst, 1963; Lacis and Hansen, 1974), iterative solutions (Herman and Browning, 1965; Braslau and Dave, 1973) and other methods.

Solutions obtained for L by one of these methods may be used for determining the irradiance according to the formula:

$$E_{\downarrow}(\vec{r}_p)=\int_{2\pi}|\vec{\xi}\vec{n}|\,L(\vec{r}_p,\vec{\xi})\,d\omega(\vec{\xi}), \tag{2}$$

where:
$E_{\downarrow}$ – downward irradiance,
$\vec{n}$ – an upward-directed unit vector normal to the horizontal plane,
$\vec{r}_p$ – the position vector.

Integration with respect to time will provide totals of radiative energy at the surface of the sea.

Leaving out the labour consumption involved the principal shortcoming of such a method of calculation lies in the necessity of knowing the c, b and p fields in the atmosphere. This difficulty may be avoided by adopting for the calculations one of known models of the atmosphere (*eg* Mc Clatchey *et al.*, 1972; Glagolev, 1970) which, however, lead to considerable averaging of the actual conditions.

Consequently, results obtained for a given place and time have relatively low accuracy while the labour consumption and cost of calculations are disproportionately large. This is the reason for using empirical or semi-empirical methods (Dera, 1983) when calculating the input of solar radiation energy to the surface of the sea in wide spectral ranges.

The main shortcoming of simple empirical methods consists in that the amount of radiation at the sea (earth) surface is regarded as a function of only one variable describing the state of the atmosphere (mostly cloudiness or insolation). In a view of the observed strong dependence of the transmission of radiation through the atmosphere on a number of other, independent parameters defining the state of the atmosphere, adoption of the known empirical methods may lead to large errors (especially for the totals of radiation incident on the surface within short time intervals).

The method as presented below allows to take into account in the calculations a qualitative and quantitative description of the processes of radiation attenuation in the medium as a function of several parameters describing the state of the medium at the given time. Thus, the method may be regarded as a semi-empirical one.

Consider for the beginning a homogeneous non-stratified cloudless atmosphere and introduce for a single scattering process (Woźniak, 1974; Jerlov, 1976) the notion of the backward scattering attenuation coefficient b_b defined by the relation:

$$b_b L(\tau,\mu,\varphi)=bL(\tau,\mu,\varphi)-\frac{bE^F_{\downarrow}(\lambda,\mu_0,\varphi_0)}{4\pi c}\,p(\tau,\mu,\varphi,\mu_0,\varphi_0)\exp(-\tau m_{r0}^{-1}), \tag{3}$$

where:
b_b – the backward scattering attenuation coefficient,
μ – cos ϑ,
ϑ – the zenith distance,
τ – the zenith optical depth of the atmosphere,

m_{ro} – the relative optical air mass,
φ – the azimuth,
λ – the wavelength,
$E_{\downarrow}^{F}$ – the downward irradiance at the upper boundary of the atmosphere.
Then for the solar direct beam radiation equation (1) reduces to:

$$m_{ro}\frac{dL(\tau,\mu_0,\varphi_0)}{d\tau}=L(\tau,\mu_0,\varphi_0), \tag{4}$$

where

$$\tau=\int_{h=0}^{\infty}(b_b+a)\,dh. \tag{5}$$

Here ∞ denotes the upper boundary of the atmosphere and h is the height above the sea level.

Integration of this simplified transfer equation yields a solution in the known form of the Bouguer-Lambert law (index S in upper fraction means that given value refers to the solar direct beam radiation):

$$L_{\downarrow}^{S}(\lambda)=L^{F}\exp(-\tau m_{ro}^{-1}), \tag{6}$$

which takes for the irradiance the following form:

$$E_{\downarrow}^{S}(\lambda)=L^{F}\mu_0\exp(-\tau m_{ro}^{-1}) \tag{7}$$

where L^F describes the monochromatic solar radiation reaching the upper boundary of the atmosphere directly (at the direction defined by μ_0, φ_0).

The contribution of the principal processes and components, as mentioned above, that lead to attenuation of the radiation in the atmosphere may be accounted for in equation (7), if the optical thickness of the atmosphere is written as a sum (Paltridge and Platt, 1976):

$$\tau=\tau_R+\tau_a+\tau_W+\tau_{OZ}+\tau_G \tag{8}$$

each term of which is defined for monochromatic radiation. The terms correspond to the molecular scattering, attenuation by aerosols, absorption by water vapour, absorption by ozone and absorption by constant components of the atmosphere (in our case: CO_2 and O_2), respectively.

As it was already mentioned, relation (7) is satisfied for monochromatic radiation. For some spectral range of the incident radiation appropriate integration should be done. In view of relation (8) we have:

$$E_{\downarrow}^{S}(\Delta\lambda)=\int_{\lambda_1}^{\lambda_2}L^{F}\mu_0\exp[-(\tau_R+\tau_a+\tau_W+\tau_{OZ}+\tau_G)\,m_r^{-1}]\,d\lambda. \tag{9}$$

It is possible, without making an excessive error, to determine for any interval $\Delta\lambda$ mean values of the coefficients of molecular scattering and aerosol attenuation (*Dinamicheskaya meteorologiya*, 1978) and have them removed from the integral (9). Simultaneous division and multiplication of the integral on the right-hand side of equation (9) by $E_{\downarrow}^{F}(\Delta\lambda)=\int\limits_{\lambda_1}^{\lambda_2} E_{\downarrow}^{F}(\lambda)d\lambda$ yields:

$$E_{\downarrow}^{S}(\Delta\lambda)=E_{\downarrow}^{F}(\Delta\lambda)\exp[-(\bar{\tau}_R+\bar{\tau}_a)\,m_{ro}^{-1}]\,T_W^S(\Delta\lambda)\,T_{OZ}^S(\Delta\lambda)\,T_G^S(\Delta\lambda), \tag{10}$$

where $\bar{\tau}_R$ and $\bar{\tau}_a$ are the mean values of the zenith optical depth of the atmosphere for the scattering on molecules and aerosols, defined for the interval $\Delta\lambda$ where:

$$T_W^S(\Delta\lambda)\,T_{OZ}^S(\Delta\lambda)\,T_G^S(\Delta\lambda)\equiv\int\limits_{\lambda_1}^{\lambda_2}\frac{E_{\downarrow}^{F}(\lambda)}{E_{\downarrow}^{F}(\Delta\lambda)}\exp[-(\tau_W+\tau_{OZ}+\tau_G)\,m_{ro}^{-1}]\,d\lambda. \tag{11}$$

For the whole spectrum (index Q in lower fraction) of solar radiation ($\Delta\lambda\to(0,\infty)$) there is:

$$E_{Q\downarrow}^{S}=F_{SQ}\,\mu_0\exp[(-\bar{\tau}_R-\tilde{\tau}_a)\,m_{ro}^{-1}]\,T_{WQ}^S\,T_{OZQ}^S\,T_{GQ}^S, \tag{12}$$

where F_{SQ} is the solar constant*.

The product of functions, describing the respective processes of radiation attenuation in the atmosphere – on the right-hand side of equation (12) – is called the atmospheric transmission (Dera, 1983) (here this concerns the direct solar radiation):

$$T_Q^S=\frac{E_{Q\downarrow}^{S}}{F_{SQ}\,\mu_0}. \tag{13}$$

For the global radiation (directly incident and scattered):

$$T_Q=\frac{E_{Q\downarrow}}{F_{SQ}\,\mu_0}. \tag{14}$$

Relations (5) – (14) make it possible to calculate the irradiance and, after integration within some time interval, solar radiation totals $Q_{\Delta t}$ at the surface of the sea. To this end the empirical formulae relating the respective transmissions to the parameters of the atmosphere as measured in the course of routine observations at meteorological stations are used.

It follows from observation of irradiance under a clear sky that for the zenith distances of the sun $\vartheta_0<80°$ the diffuse component $E_{\downarrow}^{D}$ ($h=0$) equals to about an order of magnitude less than direct radiation. Hence, for practical purposes one can calculate the total irradiance assuming, without making an excessively large error, that the diffuse radiation L^D is isotropic, *ie*:

$$L^D(\vartheta,\varphi)\equiv L^D=\mathrm{const}, \tag{15}$$

which yields:

$$E_{\downarrow}^{D}=\int\limits_0^{2\pi}\int\limits_0^{\frac{\pi}{2}} L^D\sin\vartheta\cos\vartheta\,d\vartheta\,d\varphi=\pi L^D. \tag{16}$$

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* It was assumed after Thekaekara, 1973, that the solar constant equals 1353 $\mathrm{Wm^{-2}}$.

The relation, as shown above, for the direct solar radiation propagating in a specific direction through a cloudless sky makes it possible to include the diffuse radiation in the global radiation calculations by using a single parameter depending solely on the solar zenith distance (Paltridge and Platt, 1976).

2.2. Attenuance of an 'ideal' atmosphere

In this paper the atmosphere is assumed to be 'ideal' when the radiation propagating through it is attenuated only due to the molecular scattering and absorption by constant gaseous components (namely O_2, O_3^*, CO_2). Kastrov (1956) determined transmission for an atmosphere of such characteristic, assuming that the molecular scattering coefficient $b_R(\lambda) = 1.14 \times 10^4 \times (n-1)^2 \times \lambda^{-4}$ (n – refractive index) and using the Fowle's data on the absorption of radiation by O_2, O_3 and CO_2. His results may be expressed with the accuracy of 0.003 by the following formula:

$$T_{RQ}^S T_{OZQ}^S T_{GQ}^S = \frac{E_{i\downarrow}^S(h=0, m_{ro}, P)}{E_{Q\downarrow}(h=\infty)} = 1.041 - 0.16\sqrt{m_{ro}\left(0.949\frac{P}{P_0} + 0.051\right)}, \qquad (17)$$

where: $T_{RQ}^S = \exp(-\tau_R m_{ro}^{-1})$, $E_{i\downarrow}^S$ is the irradiance due to direct solar radiation on the sea level in the ideal atmosphere, and P is the atmospheric pressure [hP], $P_0 = 1000$ hP.

Atwater and Brown (1974) modified relation (17) so as to include also the contribution of diffuse radiation. Then, the resulting relation describing the transmission of global solar radiation (directly incident and diffuse) through the 'ideal' earth's atmosphere to the sea surface has the following form:

$$T_{RQ} T_{OZQ} T_{GQ} = \frac{E_{iQ\downarrow}(h=0, m_{ro}, P)}{E_{Q\downarrow}(h=\infty)} = 0.485 + 0.515 T_{RQ}^S T_{OZQ}^S T_{GQ}^S. \qquad (18)$$

2.3. Absorbance of water vapour

To determine the amount of energy absorbed by water vapour from solar radiation going through the atmosphere (from its upper boundary down to the sea surface) one should solve the following two problems:

(i) determine the total content of water vapour along the path of radiation in the atmosphere,

(ii) determine the amount of radiation absorbed along a unit path in the medium characterized by a given content of water vapour.

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* When the entire spectrum of solar radiation is considered, O_3 may be regarded as one of constant gaseous components and its mean content in the atmosphere may be used in calculations. Variations of O_3 content have negligible influence on the total amount of radiation reaching the surface of the sea.

There exist empirical relations expressing the dependence of the total content of water vapour in an atmospheric air column of unit base area on such parameters of the state of atmosphere (provided by routine measurements on the sea level, made at meteorological stations) as the air pressure and water vapour pressure (Reitan, 1960; Timofeev, 1965; Shatunov, 1970) or the dew point temperature (Smith, 1966; Atwater and Ball, 1976).

The respective relations differ—but slightly—from one another (Fig. 1). Therefore the convenient form, as obtained by Reitan, has been used for the calculations:

$$u=(0.123+0.152e_0)\frac{P}{1000}. \tag{19}$$

Here e_0 and P denote the water vapour pressure and the atmospheric pressure on the sea level [hP], respectively, and u is the height [cm] of the entire amount of condensed water vapour contained in an atmospheric air column of unit base area.

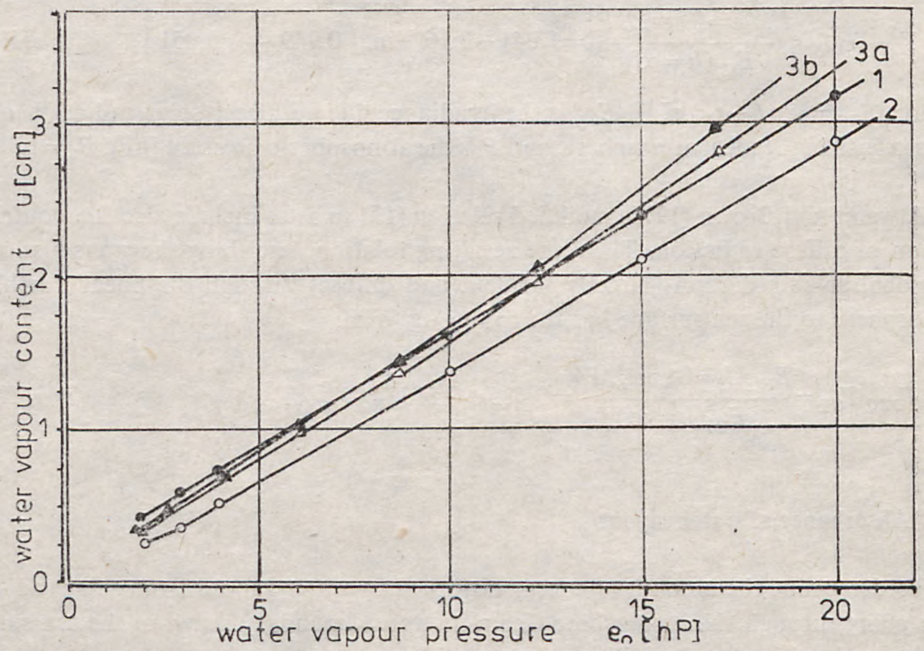


Fig. 1. Water vapour content in an atmosphere vertical column *vs* water vapour pressure on the sea level. 1—Reitan (1960); 2—Timofeev (1975); 3a—Smith (1966) (IV—VI); 3b—Smith (1966) (VII—III)

The amount of solar radiation energy absorbed (in the whole spectral range), depending on the water vapour content in the atmosphere, has been evaluated using the relation of McDonald (1960), based on Fowle's laboratory data (1912, 1915):

$$T_{WQ}=1-0.077(u\sec\vartheta_0)^{0.30}. \tag{20}$$

The author did not apply more recent and perhaps the most authoritative to date relations obtained by Yamamoto (1962), where also the presence of O_2 and CO_2 in the atmosphere had been taken into account, as the influence of these gases on attenuation of radiation in the atmosphere was accounted for in relations (17) and (18). On the other hand, the term sec ϑ_0 describing the relative optical mass of a non-refracting, horizontally plano-parallel atmosphere was substituted with a relation, obtained by Schmidt (1938), in which the curvature of the atmosphere and the way water vapour content in the atmosphere changes with the height above the sea level have been taken into account:

$$m_{rw}=[\cos\vartheta_0+0.0548(92.65-\vartheta_0)^{-1.452}]^{-1}. \tag{21}$$

Finally, substitution of formulae (19) and (21) into (20) yielded the transmittance of the earth's atmosphere for the solar radiation (within the whole spectral range), controlled by the presence of water vapour, as a function of the water vapour and atmospheric pressures on the sea level, as well as the zenith distance of the sun. The resultant formula was used in the calculations.

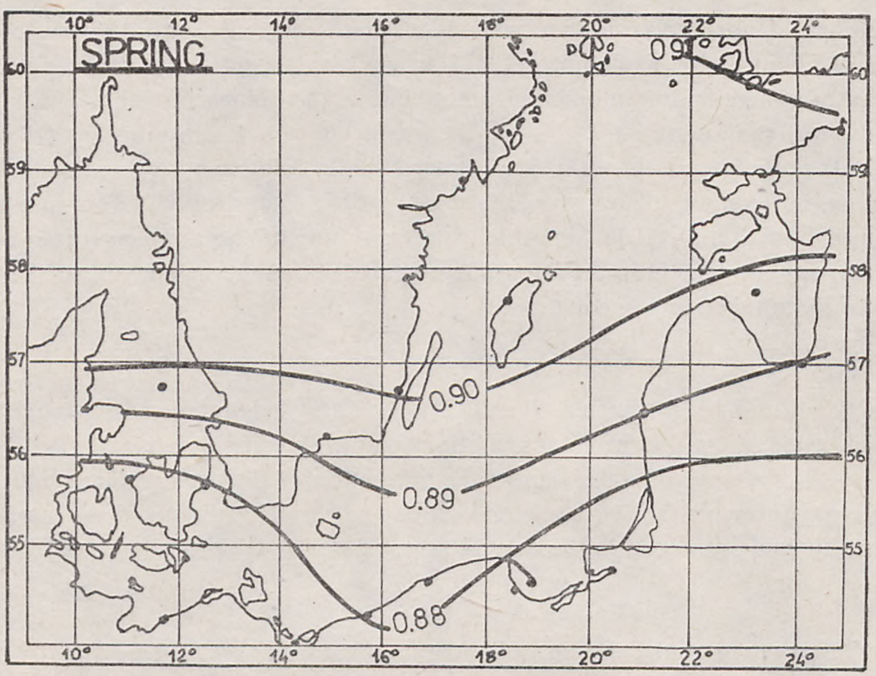
2.4. Aerosol attenuation of global solar radiation in the atmosphere over the Baltic Sea

Routine measurements (at meteorological stations) of parameters characterizing the state of the atmosphere do not include such parameters that could be used for determination of aerosol influence on the transmission of radiation through the atmosphere. For this reason the influence of aerosols on the total attenuation of light will be determined based on the actinometric data.

It can be assumed that for a cloudless atmosphere the transmission functions (as described in paragraphs 2.2 and 2.3) cover all the essential processes of radiation attenuation except the absorption and scattering by aerosols. Comparison of actinometric data for cloudless days with daily totals of solar radiation calculated for these days based on meteorological data utilized in relations (18), (19), (20) and (21) yields the values of radiation transmission determined by the presence of aerosols in the atmosphere.

For the five stations in the Baltic Sea area at which daily totals of global solar radiation have been determined together with the cloudiness, air pressure and water vapour pressure, cloudless days in the 1965–1974 decade were selected. A day was assumed to be 'cloudless' if the cloud cover (based on three recordings per day) had not exceeded 20%. The numbers of such days at Helsinki, Stockholm, Copenhagen, Gdynia and Kołobrzeg were 407, 298, 342, 378, and 218, respectively. For each of them values of the mean daily transmission of the radiation total (T_{aQ}) were determined. Averaging of these values over the respective seasons made it possible to determine the spatial variability of the mean seasonal distributions of these parameter values in the atmosphere over the Baltic Sea. For the results see Figure 2.

It can be seen that in general the transmission increases when moving north-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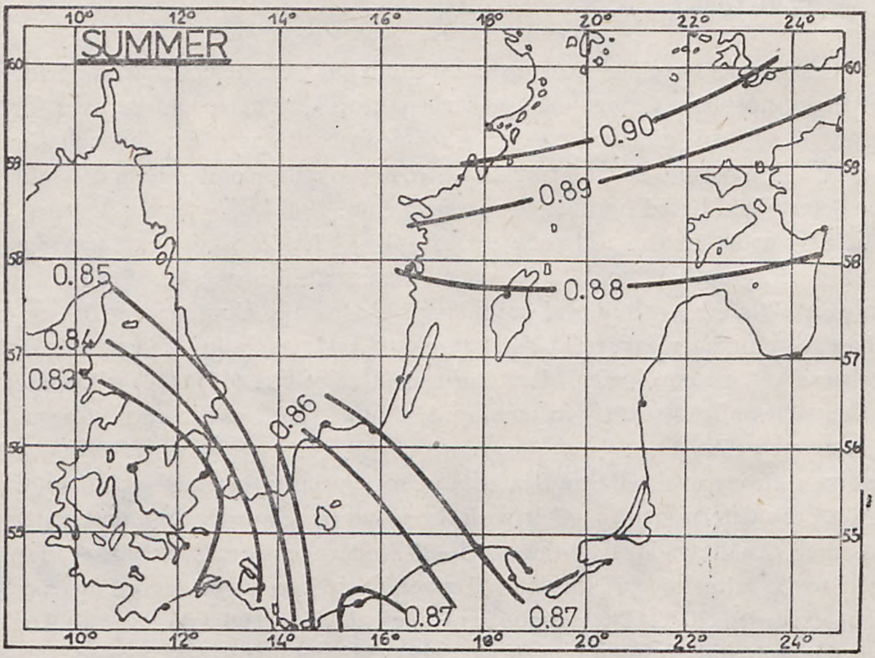
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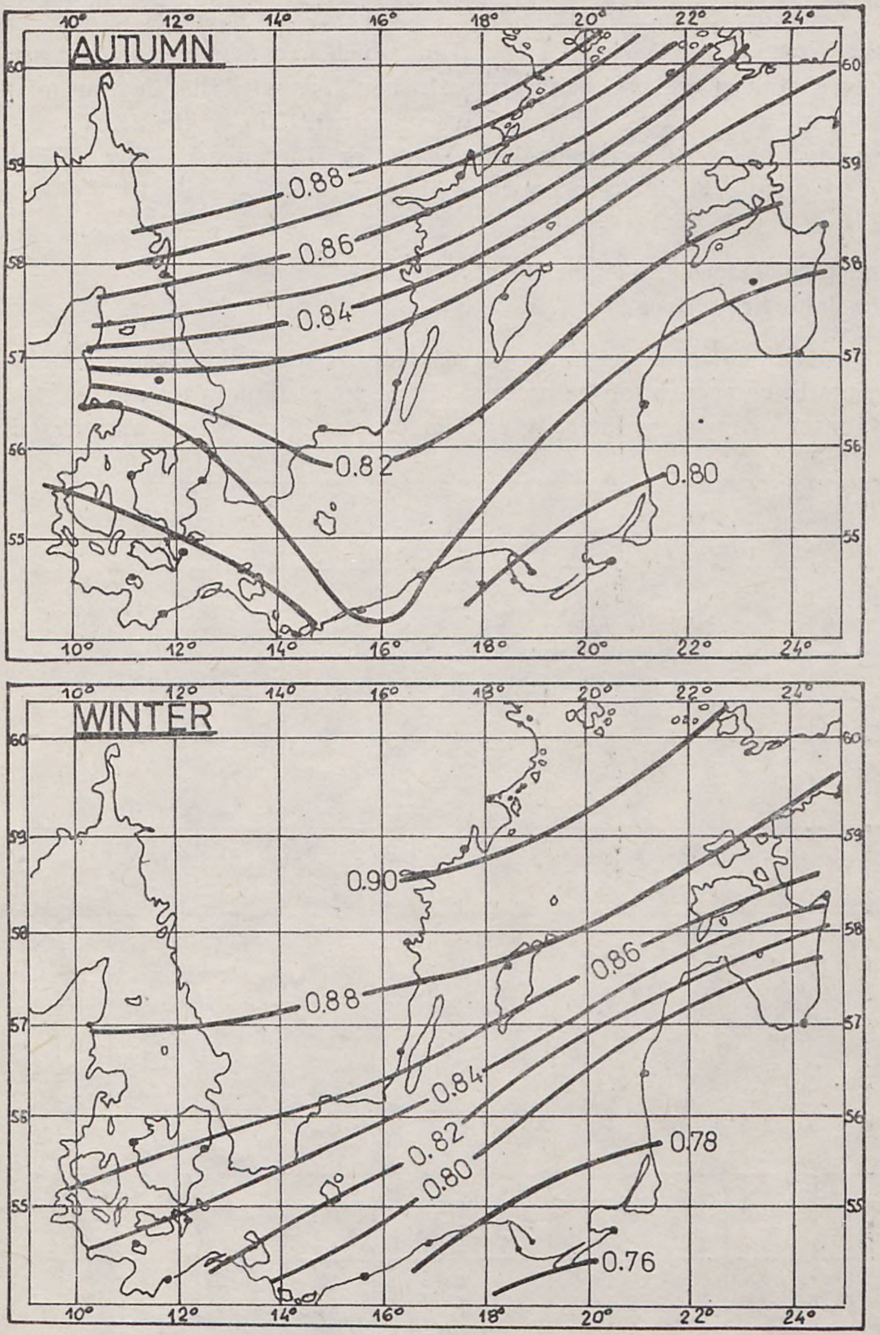


Fig. 2. Transmission of solar radiation through a cloudless atmosphere due to attenuation by aerosols. Computed by the radiation balance method on the basis of actinometric and meteorological data from 1965 – 1974

wards. The increasing distance from the industrial regions and intensive agriculture areas of the Western and Central Europe, from which vast amounts of dust are emitted into the atmosphere, may be the cause. It should be noted that the transmission determined, based on the model atmosphere rather, than on the actual conditions, will decrease with increasing latitude (cf. *eg* Braslau and Dave, 1973), contrary to the calculation results as presented above.

2.5. The influence of cloudiness on the global radiation totals at the Baltic Sea surface

The extent of cloud cover and the way it changes over the Baltic Sea area make the cloudiness the component of the process of light attenuation in the atmosphere that affects most strongly the intensity of light, its variation in time and spatial

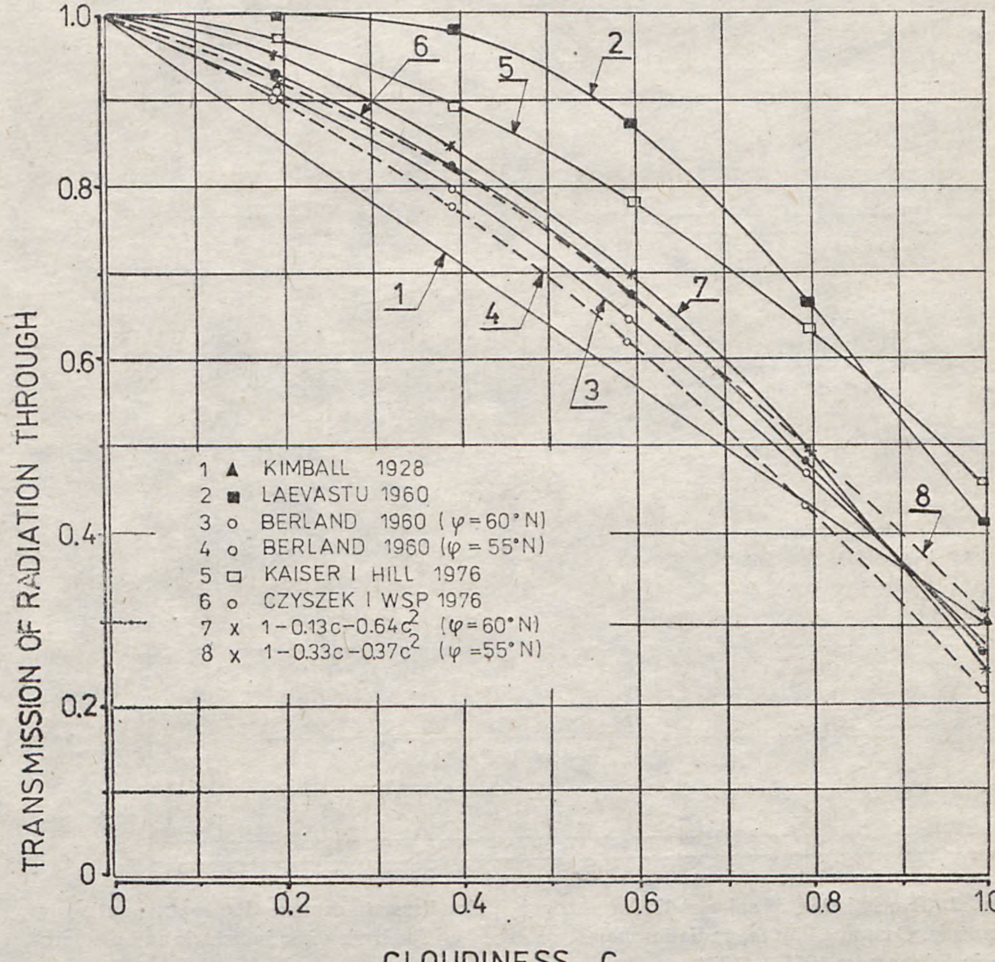


Fig. 3. Influence of cloudiness on the global radiation transmission through the earth's atmosphere

distribution over the sea surface. The extent of cloud cover (the fraction of sky covered by clouds) and the so called insolation (the time for which the sun is not covered with clouds) are the only parameters of cloudiness that are commonly and continually observed. Incidentally, insolation is observed at some meteorological stations only. This has been the reason for adopting the traditional approach to the problem of determining the influence of clouds on radiation transmission through the atmosphere. A review of available papers aimed at determining the influence between the cloudiness and the amount of solar radiation at the earth's surface showed that the results differed significantly from one another (see Fig. 3). Differences in the features of the areas under consideration and duration of the averaging periods having been the cause.

To overcome this difficulty relations defining the cloudiness-dependent transmission were determined for the Baltic Sea area, based on the available series of monthly totals of global solar radiation for a ten-year period (1965–1975) as well as appropriate meteorological data and calculations of these totals for a cloudless atmosphere. Two regression curves were determined – one based on the data for Helsinki and Stockholm (240 data pairs) and another one for Copenhagen, Gdynia and Kołobrzeg (300 data pairs). The following relations were obtained:

$$Q=Q_c(1-0.13c-0.64c^2) \quad \text{for } \varphi_G>57°N, \tag{22}$$

$$Q=Q_c(1-0.33c-0.37c^2) \quad \text{for } \varphi_G<57°N, \tag{23}$$

where Q and Q_c denote the monthly totals of solar radiation incident on a unit horizontal area under actual atmospheric conditions and for a cloudless atmosphere, respectively, c is the cloudiness in tenths.

Relations (22) and (23) are also shown in Figure 3.

3. Real totals of solar radiation incident on the Baltic Sea surface

The atmospheric transmittances for the global radiation, as presented above, and data obtained on the contribution of aerosols to light attenuation allow to calculate, by appropriate integration (see Supplement) the daily and monthly solar radiation totals at the Baltic Sea surface for cloudless atmosphere.

Addition of formulae (22) and (23) to the calculation algorithm makes it possible to determine the monthly total radiation for any meteorological station in the Baltic Sea region (for $\varphi_G \leqslant 60°N$), where systematic observations of the atmospheric pressure, water vapour pressure and cloudiness are made. Figure 4 shows the stations for which such calculations were performed.

The accuracy of the results obtained was estimated by comparing calculations carried out for Gdynia and Stockholm with the corresponding actinometric data. The comparison is presented in Figure 5. The correlation coefficient between the series of observational data and calculation results was computed. It amounted to 0.998 for Stockholm and 0.983 for Gdynia.

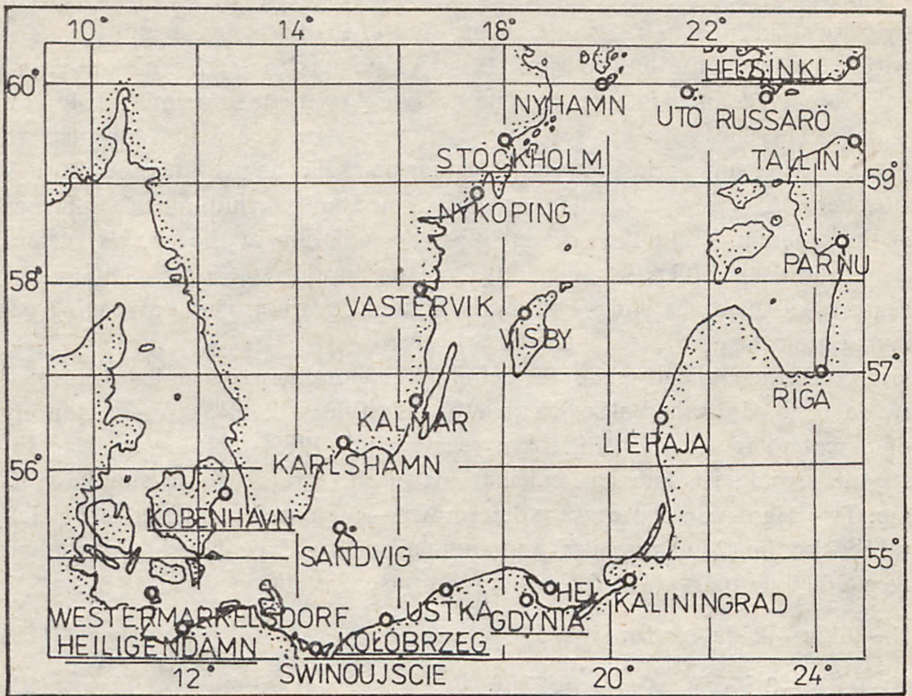


Fig. 4. Meteorological and actinometric (underlined) stations data from which were used in the study

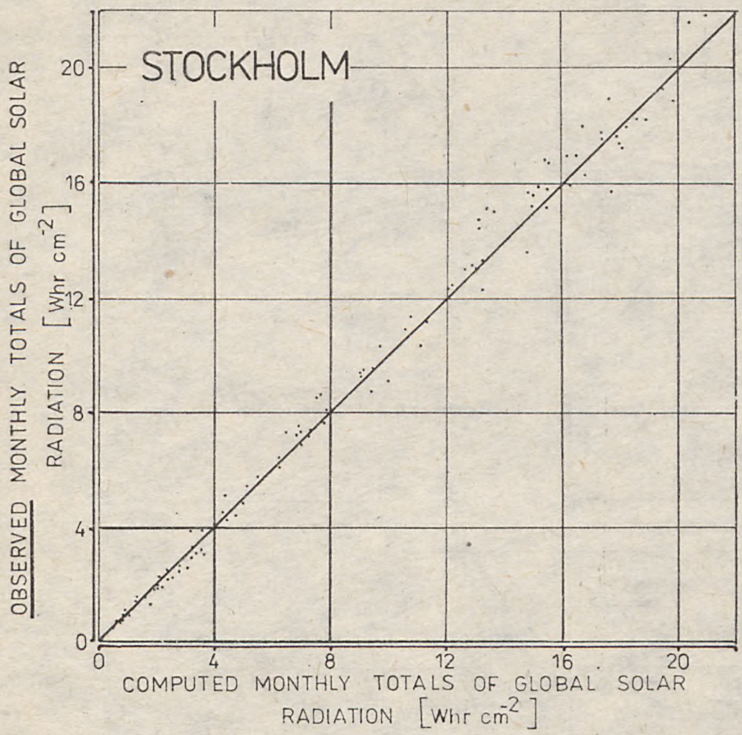
According to the author's evaluations, the model – as presented in the paper – designed for calculation of the monthly totals of global solar radiation reaching the sea surface provides the appropriate values with a mean error exceeding 5% of the actual value only for winter months, where the absolute values of the parameter calculated are relatively low.

Finally, monthly totals of solar radiation incident on the Baltic Sea surface were calculated in virtue of the mean monthly values of the above specified necessary meteorological parameters recorded in the years 1965 – 1974 for 23 stations situated in the region. The mean values (averaged over 10 years) are presented in Table 1. Based on them 12 distributions of mean monthly values of this parameter were determined (Fig. 6).

An analysis of these results revealed the following characteristic features:

– in the winter months (December, January) the radiation totals decrease (in the northern direction) in almost perfect agreement with the latitude increase. This shows that astronomical factors (sun's altitude and the duration of the day) decide in that period on the spatial distribution of the amount of radiation incident onto the sea surface, the spatial differentiation of meteorological factors being of secondary importance;

– in the spring and summer (May, June, July) the monthly radiation totals increase northward. This is an evidence of predominance of meteorological factors



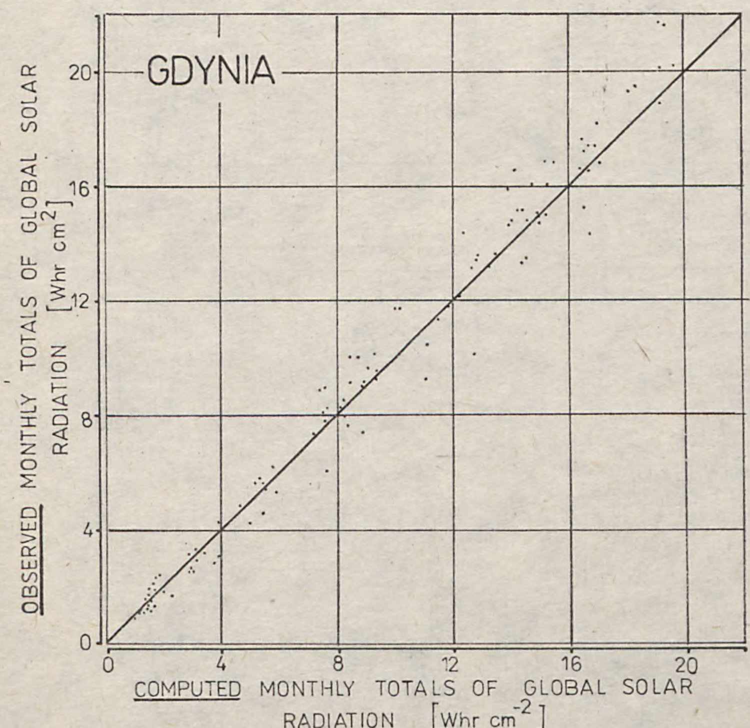
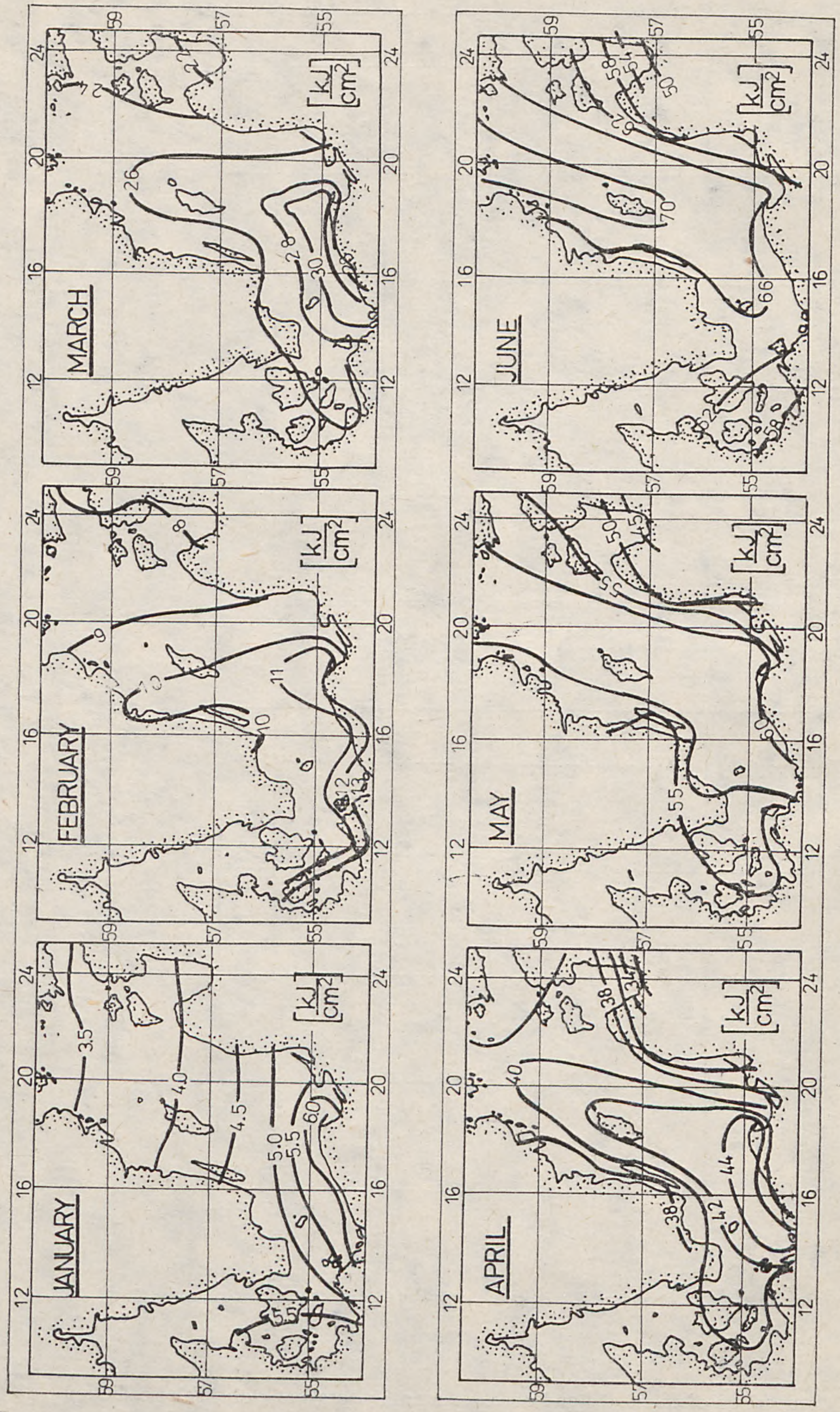


Fig. 5. Comparison of calculated and observed monthly totals of global solar radiation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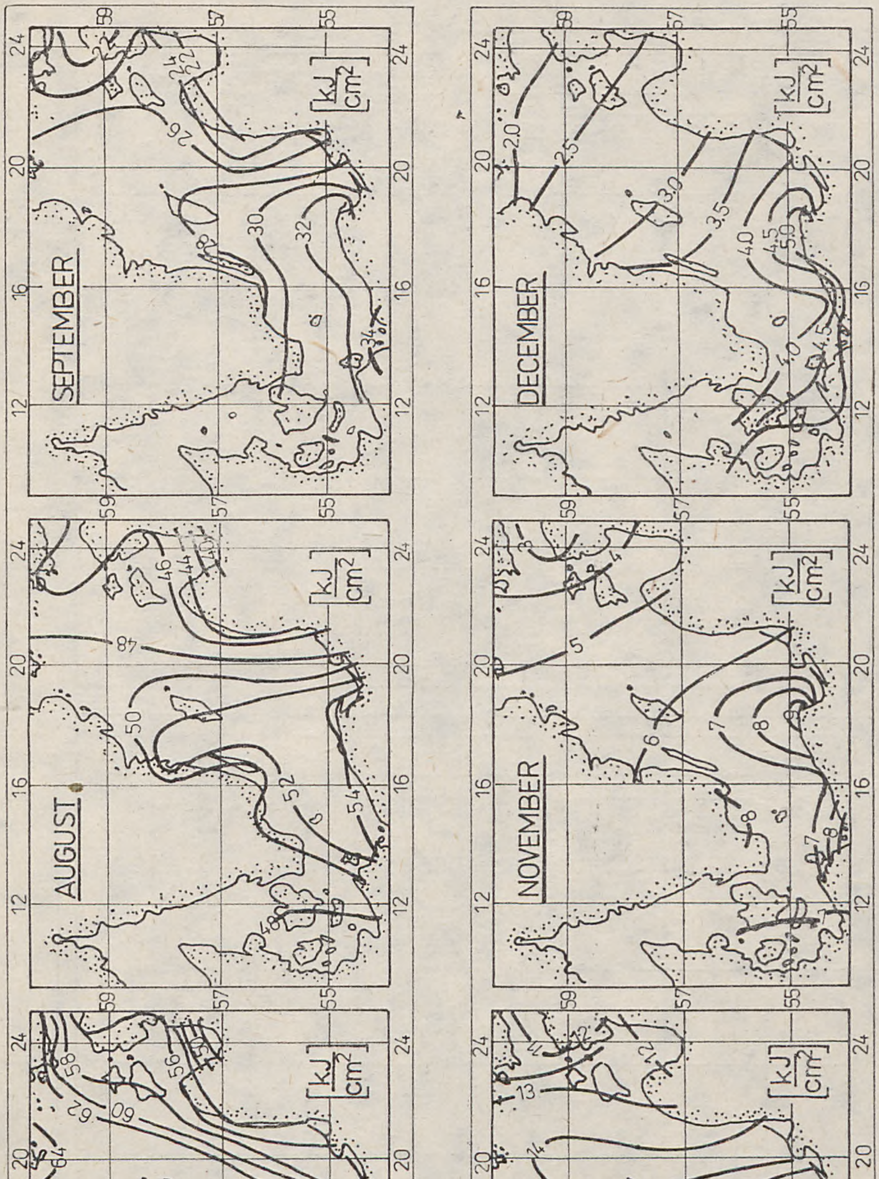
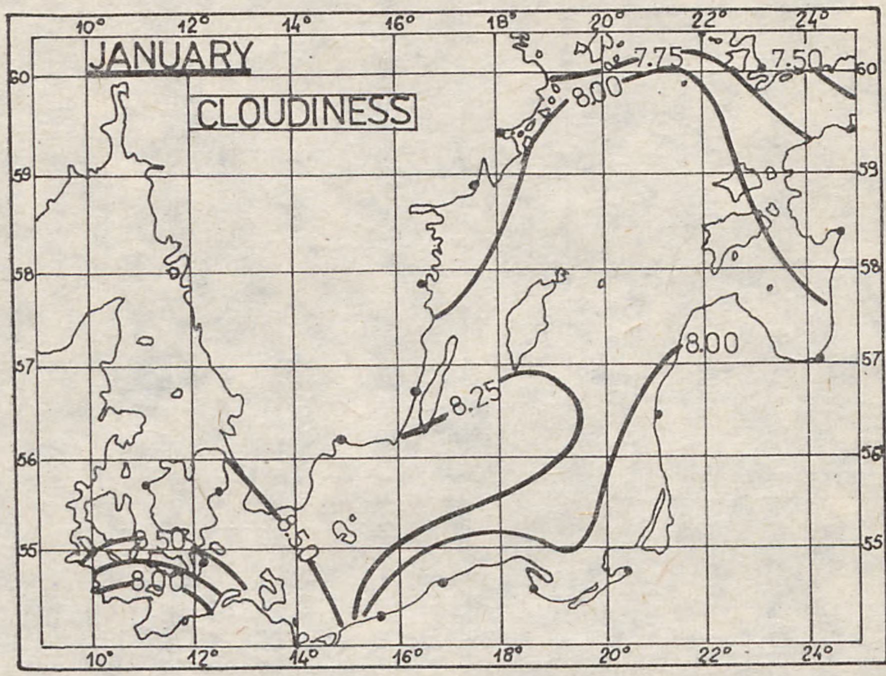


Fig. 6. Mean monthly totals of solar radiation at the Baltic Sea surface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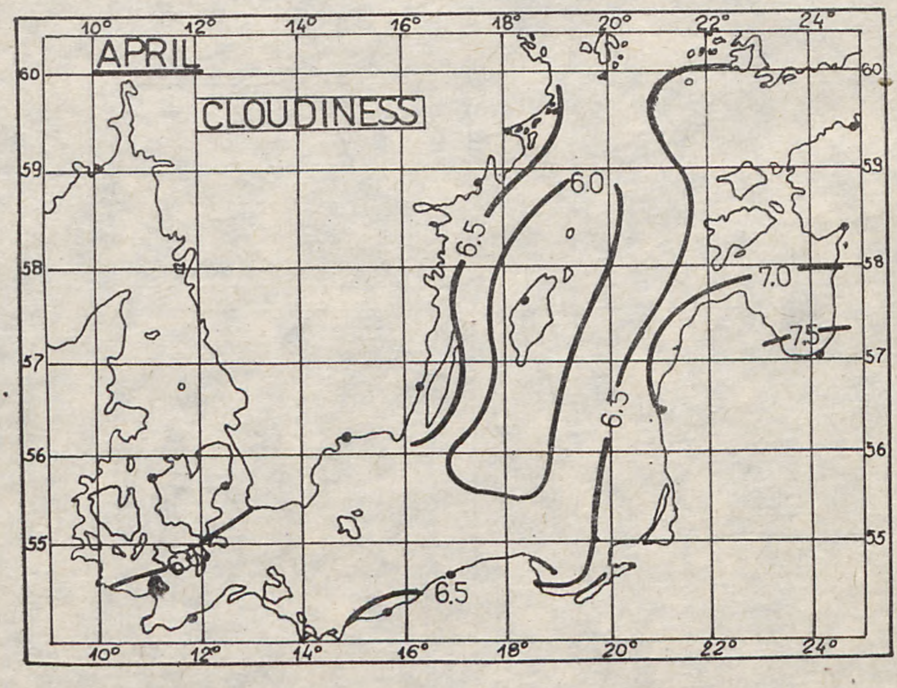
APRIL
CLOUDINESS
6.0
6.5
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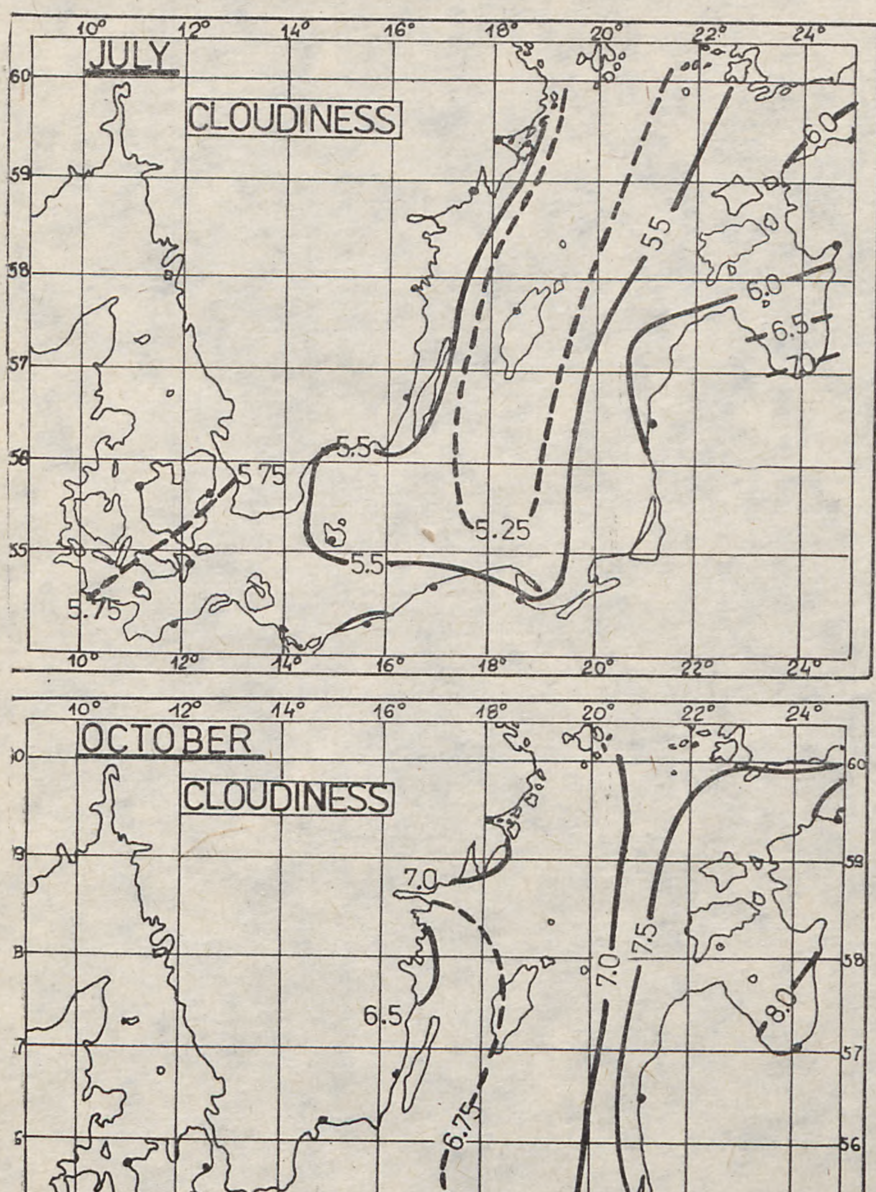

Fig. 7. Mean monthly cloudiness over the Baltic Sea (scale 0 – 10)

Table 1. Mean monthly totals of solar radiation [kJ. cm^{-2}] for the Baltic Sea area (1965 – 1974)

Station	Month												Year
	Jan	Feb	Mar	Apr	May	Jun	Jul	Aug	Sep	Oct	Nov	Dec	
Helsinki	3.08	8.13	23.05	37.62	57.22	67.20	61.98	47.74	25.19	11.79	3.58	1.56	348.14
Nyhamn	3.18	8.91	24.50	39.23	61.95	71.38	64.06	48.98	26.09	13.69	5.02	2.04	369.03
Utö	3.09	8.29	24.08	37.31	61.51	69.20	62.22	46.91	24.67	13.27	4.08	1.97	356.60
Russarö	3.29	8.39	23.03	37.14	59.14	64.06	59.59	45.03	23.30	11.89	3.92	1.91	340.68
Tallin	3.50	7.31	22.39	36.17	55.51	63.06	55.07	45.25	21.59	10.53	3.05	1.84	325.26
Stockholm	3.65	9.23	24.81	38.07	58.79	67.81	60.21	49.29	27.94	13.64	5.29	2.61	361.35
Nykoping	3.80	9.50	24.00	36.58	57.27	61.32	60.07	49.94	27.93	14.64	5.67	2.85	353.58
Pärnu	3.92	8.00	22.88	39.00	53.69	62.85	57.48	46.44	24.04	12.28	3.86	2.32	336.75
Västervik	4.26	10.31	25.47	35.60	56.14	66.05	58.42	54.26	24.80	16.10	5.84	3.53	360.78
Visby	3.99	9.42	26.95	42.12	61.01	71.13	63.08	50.50	27.86	15.08	5.92	3.00	380.01
Riga	4.10	7.84	21.75	32.00	44.01	49.63	46.35	38.06	20.43	11.39	4.65	2.68	282.89
Kalmar	4.70	9.82	24.41	37.33	53.66	61.84	57.07	47.21	27.23	15.52	6.32	3.86	348.98
Liepaja	4.64	8.71	24.26	35.67	50.67	57.41	53.51	43.77	22.01	13.60	5.32	3.00	322.58
Karlshamn	4.64	10.08	26.40	39.55	63.64	65.35	58.60	50.46	28.39	15.98	5.94	3.88	372.89
København	4.78	10.66	25.81	40.69	55.11	63.84	56.61	48.38	30.06	16.42	6.78	3.82	362.86
Kaliningrad	5.54	9.75	26.01	37.73	51.55	59.54	56.26	48.05	27.73	14.91	6.25	4.05	347.37
Ustka	6.76	12.06	29.50	44.34	60.67	65.01	62.74	53.89	33.29	19.12	8.74	5.14	401.27
Hel	6.56	11.70	30.57	44.77	61.41	66.99	57.77	55.29	33.00	18.93	9.13	4.90	400.99
Westermarkelsdorf	6.02	12.89	26.85	40.76	55.34	58.69	54.71	46.93	30.62	17.67	7.54	4.97	362.97
Gdynia	5.30	10.31	27.38	40.21	55.73	65.46	58.87	56.85	31.10	17.04	6.69	4.16	379.10
Kołobrzeg	5.97	10.31	26.87	41.22	57.23	64.53	61.06	54.18	32.79	17.49	6.73	3.87	382.26
Heiligendamm	4.93	10.13	25.18	39.27	54.40	60.94	55.31	49.37	31.85	16.11	6.87	4.19	358.54
Świnoujście	7.38	13.16	32.14	45.79	60.63	63.10	61.90	54.31	34.30	20.30	9.49	5.80	408.30

over astronomical ones in the influence on the amount of radiation reaching the sea surface. The distribution of radiation totals over the region under consideration follows fairly well the cloudiness distribution (Fig. 7);

– one can assume that from March till September the local atmospheric conditions affect strongly the input of solar radiation to the earth's surface in the Baltic Sea region. It is worth noting that the amount of incident radiation increases with the distance from the coast, to the centre of the sea following the cloudiness distribution;

– the fact that in the spring and summer the radiation totals are higher in the northern parts of the central Baltic than in the southern Baltic area may be explained by decreasing contribution of aerosols in the total attenuation of radiation (the influence of aerosols decreases when moving from S and SW towards N).

4. Participation of basic processes in total attenuation of solar radiation in the atmosphere over the Baltic Sea

The calculation method as described in this paper allows to evaluate the contribution of the respective components of attenuation process in the total attenuation of solar radiation passing through the atmosphere over the Baltic Sea. Some examples have been presented in Figures 8 and 9 based on calculations performed for Stockholm and Gdynia. Figure 8 shows seasonal changes of the attenuation processes for a cloudless atmosphere and constant solar zenith distance of 85° (which corresponds to the relative optical air mass equal to 10.16). Under such conditions the atmosphere over the northern part of the Baltic region under consideration transmits

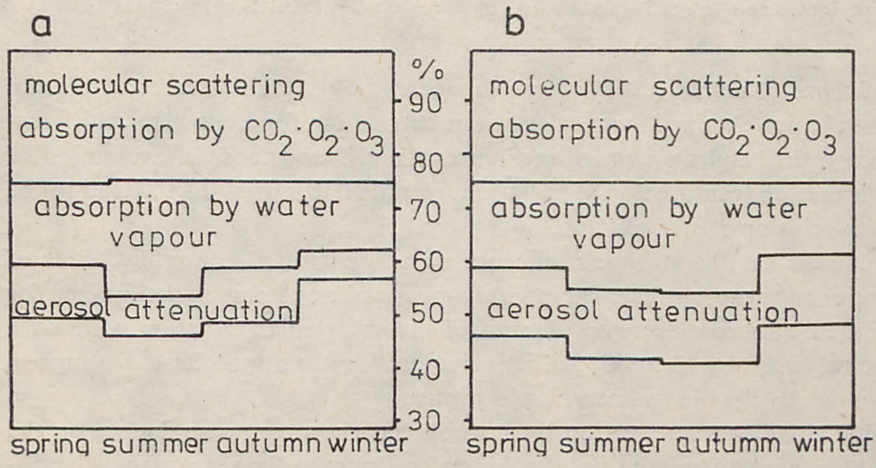

Fig. 8. Contribution of principal processes to attenuation of solar radiation for cloudless atmosphere at $\vartheta_0 = 85°$. a – based on the data from Stockholm and Helsinki, b – based on the data from Gdynia, Kołobrzeg and Copenhagen

about 5% of radiation more than over the southern part. This is caused by lower absorption by the water vapour as well as lower attenuation by aerosols. The maximum attenuation occurs in the northern part of the region under consideration – in summer, and in its southern part – in autumn, the differences being however small. For the same zenith distance of the sun and cloudless sky the amount of radiative energy reaching the sea surface is in the winter larger by nearly 10% than in the summer and autumn.

The annual variations of the actual contribution of basic attenuation processes in the global attenuation of solar radiation in the atmosphere over the Baltic are presented in Figure 9. The figure is based on calculations of daily transmittances of the atmosphere at Stockholm and Gdynia.

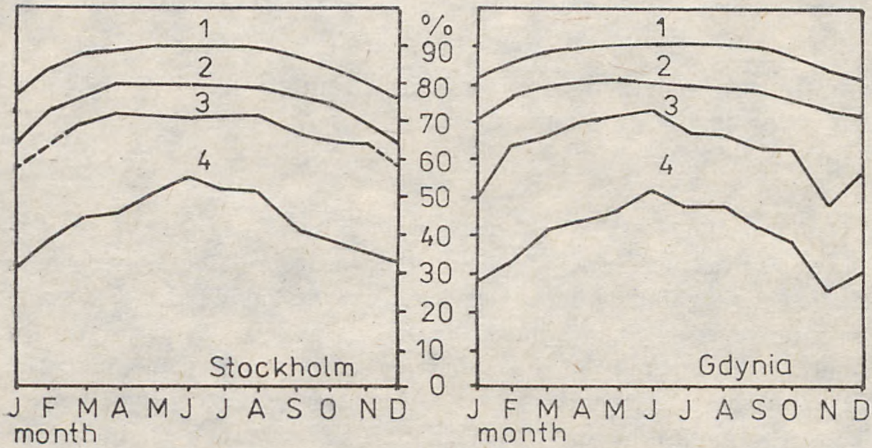

Fig. 9. Contribution of principal processes to attenuation of solar radiation by the atmosphere over the Baltic Sea during a year. 1 – molecular scattering and absorption by CO_2 and O_2; 2 – absorption by water vapour; 3 – attenuation by aerosols; 4 – attenuation by clouds

The influence of annual variations of the mean daily zenith distance of the sun reverses the picture obtained from calculations carried out for a constant zenith distance. Finally, the results presented in Figures 8 and 9 and those of evaluation of aerosol influence on transmission of radiation over the Baltic area (Fig. 2) lead to the following conclusions:

– cloudiness has the strongest influence on amounts of the monthly totals of solar radiation at the sea surface. The monthly means of cloudiness as determined for the Baltic area cause the decrease of sums of incoming solar radiation within the whole spectral range by 10 to 40% of its value outside the atmosphere, the highest attenuation due to cloudiness occurring in winter;

– beside cloudiness attenuation by aerosols has the strongest influence on the spatial and seasonal differentiation of solar energy input at the sea surface. The differnces are distinctly smaller in the northern part of the region for any month of a year. Clearly the larger distance from the industrial centers and areas of intensive agriculture of the Western and Central Europe is responsible for that effect;

— water vapour absorbs 10—15% of solar radiation in the atmosphere. In winter this effect becomes more marked in the northern part of the region under consideration, which is due mainly to the substantial increase of the optical mass of water vapour in the atmosphere for the direct solar radiation, as the mean zenith distances of the sun are large in that season. In the remaining months the absorption of radiation by water vapour increases southwards following the increase of water vapour contents in the atmosphere.

5. Concluding remarks

The results of calculations as quoted in paragraphs 2.4, 2.5, 3, and 4, and the conclusions based on them are the first ones of this type and scale of accuracy published for the Baltic Sea.

The model used for calculation of the global solar radiation based on readily available meteorological data allows to evaluate quickly the amount of radiation for the Baltic Sea area without the necessity of making actinometric measurements. Owing to the fact that the influence of the respective parameters of the state of atmosphere on all the essential processes of radiation attenuation was taken into account in the model, it can be used in the case of cloudless sky, for short averaging periods, and even for evaluation of instantaneous values. However, in the latter case improvement of the model towards more precise definition of aerosol attenuation would be necessary.

Calculations concerning relatively short averaging periods (a decade, week or day) may be carried out also for any cloudiness.

However, it would be necessary, especially when evaluating the daily totals of global solar radiation, to adopt evaluation of the cloudiness influence on the transmission of solar radiation through the atmosphere more precise than that used in our work (for monthly totals). Due to the observed characteristic distribution of the parameters of the state of atmosphere that strongly affect the propagation of radiation from the upper boundary of the atmosphere down to the surface of the sea it could be useful, in order to obtain more accurate results, to use in calculation data from observations carried out on the sea level (rather than using observational data from coastal stations) as it appears that in spite of its small size and location inside the European continent, the Baltic Sea very strongly affects the features of the state of atmosphere that are essential for the transmission of solar radiation.

Using the methodics as described by Tooming and Nijlisk (1967), and Rusin (1979) one can utilize the obtained results of calculation of global solar radiation at the Baltic surface to compute the appropriate totals in the range defined as the photosynthetically active radiation. An attempt at such calculations has been made by the author and the results will be published before long.

Acknowledgements

This study makes up a part of doctoral dissertation which has been done under the supervision of Professor Jerzy Dera. The author gratefully acknowledge Professor J. Dera for the encouragement to undertake the subject as well as for the valuable discussion and remarks during the preparation of this work.

Supplement

To obtain the daily and monthly energy doses the relations presented in the paper which describe the dependence of the irradiance on the sea level on the parameters of the state of atmosphere and astronomical factors were integrated based on assumptions as formulated by Hapter *et al.* (1974). Assuming for a clear sky a symmetry of irradiance before and after the apparent noon one can write the daily dose of the radiative energy incident onto the sea surface $Q_d(h=0)$ as:

$$Q_d = \int_{t_w}^{t_z} E_{Q\downarrow}(h=0)\,dt \cong 2\int_{t_M}^{t_z} E_{Q\downarrow}(h=0)\,dt, \tag{S.1}$$

where:
t_z – the sunset hour angle,
t_M – the hour angle of sun culmination at noon,
t_w – the sunrise hour angle.
For the variable changing (in order to obtain the dependence on the latitude φ_G and declination of the sun δ) one can use a relation expressing the zenith distance of the sun in terms of time, declination and latitude:

$$\cos\vartheta_0 = \sin\varphi_G \sin\delta + \cos\varphi_G \cos\delta \cos\tau, \tag{S.2}$$

where τ is the apparent time, measured in degrees $(-180° \leqslant \tau \leqslant 180°)$, and related to the time t measured in hours by the formula:

$$\tau = 15t - 180°. \tag{S.3}$$

After variable changing:

$$Q_d = 2\int_{90°}^{\vartheta_{0\min}} E_{Q\downarrow}(h=0, \vartheta_0)\frac{dt}{d\vartheta_0}\,d\vartheta_0. \tag{S.4}$$

To calculate the derivative of time with respect to the zenith distance of the sun, that appeared under the integral sign in expression (S.4) time t is determined from relations (S.2) and (S.3):

$$t = \frac{1}{15}\arccos\left(\frac{\sin\varphi_G \sin\delta - \cos\vartheta_0}{\cos\varphi_G \cos\delta}\right), \tag{S.5}$$

$$\frac{dt}{d\vartheta} = -\frac{\sin\vartheta_0}{15\cos\varphi_G\cos\delta}\left[\sqrt{1-\left(\frac{\sin\varphi_G\sin\delta-\cos\vartheta_0}{\cos\varphi_G\cos\delta}\right)^2}\,\right]^{-1}. \tag{S.6}$$

In order to make possible integration of (S.4) in spite of the complex form of function $E_{Q\downarrow}(h=0, \vartheta_0)$ it is assumed that this function is constant and equal to $E_{Q\downarrow}(h=0, \vartheta_0)$ in intervals between $\vartheta_{0i}+\Delta\vartheta_0$ and $\vartheta_{0i}-\Delta\vartheta_0$. Thus:

$$Q_d=2\sum_{\vartheta_0=90^\circ}^{\vartheta_{0\min}} E_{Q\downarrow}(h=0,\vartheta_{0i})\int_{\vartheta_{0i}+\Delta\vartheta_0}^{\vartheta_{0i}-\Delta\vartheta_0} -\frac{\sin\vartheta_0}{15\cos\varphi_G\cos\delta}\left[\sqrt{1-\left(\frac{\sin\varphi_G\sin\delta-\cos\vartheta_0}{\cos\varphi_G\cos\delta}\right)^2}\right]^{-1} d\vartheta_0=$$

$$=2\sum_{\vartheta_0=90^\circ}^{\vartheta_{0\min}} E_{Q\downarrow}(h=0,\vartheta_{0i})\left[-\frac{1}{15}\arcsin\frac{\sin\varphi_G\sin\delta-\cos\vartheta_0}{\cos\varphi_G\cos\delta}\right]\Bigg|_{\vartheta_{0i}-\Delta\vartheta_0}^{\vartheta_{0i}+\Delta\vartheta_0}. \tag{S.7}$$

For a given $\Delta\vartheta_0$ (here 1° was chosen) we obtain finally:

$$Q_d=2\sum_{i=1} E_{Q\downarrow}(h=0,i)\frac{12}{\pi}\left\{\arcsin\frac{\sin\varphi_G\sin\delta-\cos\left[\frac{\pi}{180}(92-2i)\right]}{\cos\varphi_G\cos\delta}-\right.$$

$$\left.+\arcsin\frac{\sin\varphi_G\sin\delta-\cos\left[\frac{\pi}{180}(90-2i)\right]}{\cos\varphi_G\cos\delta}\right\} \tag{S.8}$$

where angles are measured in radians and

$$i=1,2,3,\ldots,i_{\max}=\frac{1}{2}\left\{\frac{90}{\pi}\left(91\frac{\pi}{180}-\varphi_G-\delta\right)+E\left[\frac{90}{\pi}\left(\frac{\pi}{2}-\varphi_G+\delta\right)\right]+0.5\right\}. \tag{S.9}$$

Here $E[\ ]$ denotes the entire function.

The declination of the sun appearing in relations (S.2)–(S.9) may be given with the accuracy of 0.0006 radians by the formula (Spencer, 1971):

$$\delta=0.006918-0.399912\cos\theta_0+0.070257\sin\theta_0-$$

$$+0.006758\cos 2\theta_0+0.000907\sin 2\theta_0-0.002697\cos 3\theta_0+0.001480\sin 3\theta_0, \tag{S.10}$$

where: $\theta_0=\frac{2\pi d_n}{365}$, $d_n=0, 1, \ldots, 364$ denotes the respective days of the year, *ie* 0→1.01, 1→2.01 *etc.*

Due to the fact that the solar constant has been determined for the mean distance between the sun and the earth, a correction should be introduced into calculations of the radiation total for a given day of the year that takes into account the deviation of the actual distance from the mean distance. The flux of radiation incident onto a unit surface is inversely proportional to the square of the distance from the source. Therefore the actual value of the radiation flux (corresponding to the actual distance between the earth and the sun) given by the solar constant and the ratio of the mean distance $\bar{R}$ to the actual one R is equal to:

$$F_R=F_{SQ}\frac{\bar{R}^2}{R^2}. \tag{S.11}$$

The ratio $\beta=\frac{R^2}{\bar{R}^2}$ can be calculated for each day of the year with sufficient accuracy from the formula (Paltridge and Platt, 1976):

$$\beta^{-1}=1.0001100+0.034221\cos\theta_0+0.001280\sin\theta_0+$$

$$+0.000719\cos 2\theta_0+0.000077\sin 2\theta_0, \tag{S.12}$$

where θ_0 has the same meaning as in (S.10).

For the Baltic Sea latitudes the mean zenith distance of the sun in the winter, defined (Manabe and Möller, 1961) as:

$$\cos\bar{\vartheta}_0 = \frac{\int\limits_{t_w}^{t_z} \cos\vartheta_0 \, dt}{\int\limits_{t_w}^{t_z} dt}, \tag{S.13}$$

exceeds 80°, and for the northern part of this region even 85°. Thus it seems necessary to take into account (in calculations of the radiation flux) the curvature of the atmosphere. In this case an approximate formula for the relative optical air mass as given by Kasten (1966) for the standard atmosphere and radiation wavelength $\lambda = 0.7$ μm may be used:

$$m_{r_0} = [\cos\vartheta_0 + 0.1500(93.885 - \vartheta_0)^{-1.253}]^{-1}. \tag{S.14}$$

According to Kondratiev (1969) for the short-wave radiation (0.3 – 5 μm), $m_{r_0}(\vartheta_0\ \lambda) \cong m_{r_0}(\vartheta_0)$ may be assumed.

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