Emissivity, thermal albedo and effective emissivity of the sea at different wind speeds

Emissivity of the sea
Radiation exchange
Radiation balance
Effective emissivity of the sea
Émissivité de la mer
Échange radiatif
Bilan radiatif
Émissivité efficace de la mer

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ABSTRACT

Sea surface emissivity ε, sea albedo for thermal atmospheric radiation x and effective emissivity δ have been calculated at different wind speeds v from current data on sea water optical constants with regard to the real sea surface roughness. The values obtained are markedly different from those used at present.

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RÉSUMÉ

Émissivité, albédo pour le rayonnement thermique et émissivité efficace de la mer

L'émissivité ε de la surface de la mer, l'albédo x de la mer pour le rayonnement thermique de l'atmosphère et l'émissivité efficace δ de la mer ont été calculés pour différentes vitesses du vent v à partir des caractéristiques optiques modernes de l'eau marine et en tenant compte de la rugosité réelle de la surface de la mer. Les résultats obtenus diffèrent notablement de ceux qui sont admis actuellement.

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INTRODUCTION

In calculations of the radiant heat exchange between sea and atmosphere, the magnitude of the sea surface emissivity ε and the sea albedo for thermal atmospheric radiation x are of significant importance. When calculating the radiation balance components, the sea albedo for the atmospheric thermal radiation x is often taken to be equal to $(1-\varepsilon)$. In practice, since the spectral distribution and the angular structure of atmospheric radiation and of the black body (BB) are markedly different, the magnitudes x and $x'=1-\varepsilon$ are also different. If this is taken into account, the effective sea surface emissivity, *i.e.* radiation heat losses of the sea, is determined from the following formula

$$E_{eff}=(1-x')\,\sigma T_{up}^4-(1-x)\,E_a. \qquad (1)$$

Here T_{up} is the sea upper layer temperature and E_a is the atmospheric radiation. The first addend on the right in equation (1) is the sea thermal radiation; $\varepsilon=1-x'$ is the sea surface emissivity. The second addend is the atmospheric radiation absorbed by the sea. It reduces radiation heat losses.

In radiation climatology, calculations of sea radiation losses involve effective emissivity of the sea δ. This magnitude is defined from the following formula

$$\delta=\frac{E_{eff}}{\sigma T_{up}^4-E_a}. \qquad (2)$$

From equations (1) and (2) for the magnitude δ we find

$$\delta=(1-x')+(x-x')\frac{\eta}{1-\eta},$$

$$\eta=\frac{E_a}{\sigma T_{up}^4}. \qquad (3)$$

This paper presents calculation results for x', x and δ under average standard conditions.

In radiation calculations it is usually assumed that $x'=x=0.09$ and, correspondingly, $\delta=0.91$. However, this assumption is somewhat approximate. Its error will be estimated below.

The magnitudes x', x and δ depend on the wind speed v over the sea. The objective of the present paper is to study this dependence. Our new ascertained values are listed in Tables 3 and 4. They may be used to define E_{eff} from data on T_{up} and E_a.

The E_a value, unless it is a known quantity, may be estimated from semi-empirical formulas, from radiation nomograms (Perrin de Brichambaut, 1963, Ch. 13; Ivanoff, 1974), or from Tables by Girdiuk *et al.* (1982; these tables include air temperature and cloudiness as input data).

CALCULATIONS OF MAGNITUDES x', x AND δ

The angular spectral emissivity $\varepsilon(\lambda, \vartheta)$, total spectral emissivity $\varepsilon(\lambda)$ and total emissivity ε of the sea are defined by the relations

$$\varepsilon(\lambda, \vartheta) = 1 - \int_0^{\pi/2} \rho(\lambda, \vartheta, \vartheta') \sin 2\vartheta' \, d\vartheta',$$

$$\varepsilon(\lambda) = \pi \int_0^{\pi/2} \varepsilon(\lambda, \vartheta) \sin 2\vartheta \, d\vartheta,$$

$$\varepsilon = \frac{\pi \int_0^{\pi/2} \varepsilon(\lambda) B(T_{up}, \lambda) \, d\lambda}{\sigma T_{up}^4}. \tag{4}$$

Here $\rho(\lambda, \vartheta, \vartheta')$ is the sea surface radiance factor for different angles of incidence ϑ' and viewing ϑ. To avoid overburdening the formulas, we consider the sea reflection and radiation, as well as the atmospheric radiation, to be azimuth-independent quantities. The instantaneous picture is, certainly, anisotropic. It manifests itself, for example, in the distribution of facets according to inclinations. However, in climatologic calculations our concern is usually with magnitudes averaged over the azimuth.

Equation (4) follows directly from the Kirchhoff law. This law is quite applicable to the thermal sea radiation despite of the fact that significant temperature gradients are observed in the vicinity of the surface since the thickness of layer l that generates thermal radiation of the sea is significantly less than the value L for which the water temperature changes significantly:

$$l \ll L. \tag{5}$$

The value of l equals $3\alpha^{-1}$, where α is the sea water absorption coefficient for IR radiation. Using the values of α in Shifrin (1983) we find that $l \simeq 50\text{-}100\,\mu m$, whereas $L \simeq 1\text{-}10$ cm. The estimate (5) also follows from direct measurement data by McAlister and McLeish (1969; for details, *see* Dubov, 1974).

The values of $\rho(\lambda, \vartheta, \vartheta')$ have been calculated in the spectral range of $2.5-250\,\mu m$ from current data on the optical constants of sea water with standard salinity S(3.5%). These data may be found, for example, in a monograph by Shifrin (1983). When we pass from pure to sea water, the changes are due to the absorption band of sulphates. They are not large, but lie in a spectral range from 9 through $15\,\mu m$, *i.e.* precisely where the radiation transfer in the atmosphere takes place. The calculation has been performed in an approximation to the stochastically distributed facets method (Mullamaa, 1964). Distribution of facets according to inclinations at different wind speeds was taken according to Cox and Munk (1954). In contrast to Mullamaa (1964), our calculations account for the facet mutual shadowing function $Q(\vartheta', \vartheta)$ for different angles of incidence ϑ' and viewing ϑ.

Table 1
Shadowing functions $Q(\vartheta', \vartheta)$ for $\vartheta' = 75°$.

ϑ	30	40	50	60	70	80	85	88	89.5
Q	0.88	0.88	0.88	0.88	0.75	0.66	0.44	0.21	0.06

The function was chosen according to Bass and Fuks (1972). It indicates the portion of the illuminated area that is not shaded for given directions ϑ' and ϑ and takes part in generating the reflected signal.

To evaluate the shadowing effect, Table 1 gives the values of $Q(\vartheta', \vartheta)$ at $\vartheta' = 75°$ for different ϑ and a typical sea roughness (the sea wave height/sea wave length ratio is taken to be equal to 1/7).

We add that the method of stochastically distributed facets taking into account the Cox and Munk (1954) distribution originally developed by Mullamaa (1964) was applied in a paper by Takashima and Takayama (1981) to determine the reflective properties of the sea surface for 3 spectral ranges that correspond to channels of the AVHRR radiometer aboard the NOAA-6 satellite. Our calculations differ from those in that paper in three respects: 1) by allowance for shadowing; 2) by using optical constants of the sea rather than fresh water; 3) by studying the whole range of thermal radiation, from 2.5 through 250 μm.

The sea albedo x for thermal atmospheric radiation is given by

$$x = \frac{\int_0^\infty d\lambda \int_0^{\pi/2} A(\lambda, \vartheta') B(\lambda, \vartheta') \sin 2\vartheta' \, d\vartheta'}{\int_0^\infty d\lambda \int_0^{\pi/2} B(\lambda, \vartheta') \sin 2\vartheta' \, d\vartheta'}. \tag{6}$$

Here $B(\lambda, \vartheta')$, $A(\lambda, \vartheta')$ are the sky thermal radiance and the sea surface albedo for a parallel radiation beam. The albedo $A(\lambda, \vartheta')$ has been calculated using the radiance factor $\rho(\lambda, \vartheta, \vartheta')$ according to

$$A(\lambda, \vartheta') = \int_0^{\pi/2} \rho(\lambda, \vartheta, \vartheta') \sin 2\vartheta \, d\vartheta,$$

and data on the angular structure of atmospheric thermal radiance $B(\lambda, \vartheta')$ have been taken from Zolotova (1971) who treated measurements of $B(\lambda, \vartheta')$ taken by Ashcheulov *et al.* (1968) at Kaunas, USSR. Experimental data show that $B(\lambda, \vartheta')$ may be conveniently given as

$$B(\lambda, \vartheta') = B_0(T_a, \lambda) - f(\lambda, \vartheta'). \tag{7}$$

Here $B_0(T_a, \lambda)$ is the Planck function for the near-to-ground atmospheric surface layer temperature T_a and $f(\lambda, \vartheta')$ is a certain correction. In cloudless conditions the correction $f(\lambda, \vartheta')$ is of significant magnitude in the range from 7 through $15\,\mu m$. This is the range of a "window" where the atmospheric radiation is formed

Table 2
Values of $f(\lambda, \vartheta')$ in W/m^2 μm ster.

$\lambda \setminus \vartheta'$	0	60	70	80	85	90
6	0	0	0	0	0	0
7	0.60	0	0	0	0	0
8	5.0	3.4	2.2	1	0.40	0
9	7.4	6.4	4.8	3.3	1.4	0
9.5	5.9	4.8	3.8	2.8	1.3	0
10	7.5	6.5	5.7	3.8	1.8	0
12.5	3.3	1.8	0.8	0.5	0	0
15.0	0.65	0.55	0	0	0	0

by thick air layers and is therefore markedly lower than $B_0(T_a, \lambda)$. Outside the "window" $f(\lambda, \vartheta')=0$, *i. e.* $B(\lambda, \vartheta')=B_0(T_a, \lambda)$ there. In conditions of complete cloudiness the magnitude $B(\lambda, \vartheta')$ within the "window" approximately corresponds to radiation of the black body whose temperature is 5° lower than T_a; outside the "window" $B(\lambda, \vartheta')$ coincides with $B_0(T_a, \lambda)$.

The magnitude $f(\lambda, \vartheta)$ certainly depends on the temperature profiles, humidity, aerosol and the amount of ozone, and may vary occasionally. However, small variations in it are not of significant importance as the magnitude $f(\lambda, \vartheta')$ enters the formula for x under the integration sign. Values of $f(\lambda, \vartheta')$ for $\lambda=6.0$ (0.5) 15 μm have been obtained by smoothing experimental data. Some of them referring to $T_a=290°$ K are given in Table 2.

The minimum at $\lambda=9.5$ μm is associated with radiation of a strong ozone absorption band at 9.6 μm. In order to control the values of $f(\lambda, \vartheta')$ obtained, the magnitude $2\pi \int_6^{15} d\lambda \int_0^{\pi/2} f(\lambda, \vartheta') \sin\vartheta' \, d\sin\vartheta'$, *i. e.* the effective radiation of the earth surface, was calculated (Zolotova, 1971). It proved to be equal to 0.136 cal/cm² min, which is close to the value 0.142 cal/cm² min obtained for standard cloudless atmosphere, the ARDC model 1959 (*see* Kondratiyev, 1969, Tab. 8.4).

We applied the model of thermal atmospheric radiation (Zolotova, 1971) to the marine atmosphere. Assuming that T_a coincides with T_{up} (temperature of the upper sea layer), we performed test calculations of the atmospheric thermal radiation flux E_a for cases of a cloudless atmosphere ($p=0$) and a continuous cloud cover ($p=1$). For $T=290°$ K we obtained 320 and 383 W/m², respectively (Gardashov, Shifrin, 1983; Gardashov *et al.*, 1985). These values, in practice, coincide with those recommended in tables by Girdiuk *et al.* (1982; 320 and 380 W/m², respectively) for the cases in question.

Results of calculations of the atmospheric thermal radiation E_a are available for four atmospheric models (Schlüssel, Grassl, 1987, priv. comm.), namely: 1) midlatitude summer; 2) midtropical; 3) subarctic winter; and 4) extreme tropical (with water vapour column content of 7-8 g/cm²!). Table 3 gives a comparison of these calculations with data taken from tables by Girdiuk *et al.* (1982). These tables summarize the many thousands of ocean measurements performed on weather ships.

It follows from Table 3 that, on the average, the atmosphere over the ocean is closest to the first model, *i. e.* to the midlatitude summer atmosphere and farthest from the third and fourth (extreme) models. We note that the flux E_a calculated for $T_{up}=294$°K by our

Table 3
Atmospheric thermal radiation fluxes $E_a (W/m^2)$.

Model N°	T_{up} K	Schlüssel and Grassl	Girdiuk *et al.*
1	294	351	343
2	300	397	382
3	271	182	212
4	300	435	382

simple scheme [equation (7) and Tab. 2] is equal to 345 W/m², *i. e.* coincides, in practice, both with tables by Girdiuk *et al.* (1982), and with calculations of Schlüssel and Grassl (1987) for the first model. Radiative sea properties x_0 and δ_0 vs wind speed calculated below are thus related to a certain mean sea atmosphere which appears to be close to the midlatitude summer atmospheric model.

We note that in radiation calculations we do not usually have aerological data. Therefore, when plotting radiative balance maps we are compelled to use data related to the mean sea atmosphere. However, we do have data on the wind over the sea. This is why the clarifications of κ_0 and δ_0 proposed below are significant. Certainly, calculations with other functions $B(\lambda, \vartheta')$ may also be repeated provided the data are available.

Knowing the value $B(\lambda, \vartheta')$ we use equation (6) to calculate the thermal sea albedo under cloudless conditions ($p=0$) and for a continuous cloud cover ($p=1$). The magnitudes obtained are designated x_0 and x_1, respectively, and are given in Table 4 for different wind speeds v (m/s). The values of the magnitude $x'=1-\varepsilon$ are also given there. It has the meaning of albedo for the BB radiation.

Table 4
Sea albedo for thermal atmospheric radiation x and BB radiation x' (%).

v, m/s	0	5	10	15
x_0	10.5	9.4	8.6	8.1
x_1	9.4	8.4	7.6	7.2
x'	9.2	8.2	7.5	7.1

So far we have neglected the effect of a cold "film" on the upper sea surface. It has been known that thermal sea radiation is determined by the temperature T_{rad} of the upper radiative layer of thickness l ($l \simeq 50-100$ μm) rather than by the upper sea layer temperature T_{up} usually measured by contact sensors at depths ≈20-50 cm. Analysis of the magnitude $\Delta=T_{rad}-T_{up}$ (Roll, 1965; Dubov, 1974) suggests that the maximum recurrence of the magnitude Δ amounts to $-0.4°$ and $-0.2°$ at $v=0$ and 20 m/s, respectively. The cold "film" on the surface may be accounted for, if we are to formally decrease the emissivity of water by $\delta\varepsilon$. It is obvious that the surface temperature decrease by ΔT is equivalent to a decrease in ε by $\delta\varepsilon=\varepsilon(4\Delta T/T)$. So, according to data on ΔT from Dubov (1974), we find the values of $\delta\varepsilon$ indicated in Table 5.

Table 5
Allowance for the "film" effect.

v, m/s	0	5	10	15	20
$10^3 . \delta\varepsilon$	5	5	4	3	2
$10^2 . \bar{x}'$	9.7	8.7	7.9	7.3	6.9
$\bar{\varepsilon}$	0.903	0.913	0.921	0.927	0.931
δ_0	0.935	0.942	0.948	0.950	0.951

Table 5 shows the sea water emissivity and the magnitude $1-\varepsilon=x'$ at $T_{up}=290°K$ with allowance for the "film" effect (we indicate them with an overbar). It may be seen that the magnitude ε increases by 3% as v increases from 0 to 20 m/s. It changes but slightly with water temperature and so the values of ε in Table 5 may be used at any temperature. According to Dubov (1974), the "film" correction is not necessary under continuous cloud cover conditions.

When calculating the net radiation of the sea, we are usually interested in the total effect, the sea surface effective radiation. From equation (3), using data on the magnitude E_a and T_{up} from Girdiuk *et al.* (1982), and the values x', x', x^0 and x_1 from Tables 4 and 5, we have calculated δ_0 for a cloudless atmosphere and δ_1 for a continuous cloud cover. The values of δ_0 at $T_{up}=290°K$ for different values of v are also shown in Table 5. With a complete cloud cover the values of δ_1 proved to be independent of the wind speed v and equal to approximately 0.95.

DISCUSSION AND CONCLUSIONS

The value of δ is usually taken to be equal to 0.91 in radiation calculations (Girdiuk *et al.*, 1982). We can see that if the data on the wind speed and cloudiness are absent, the value 0.95 should be used for δ, *i. e.* the usual value is underestimated by 4.4%.

Let us estimate the value of our correction for calculations of the radiation losses E_{eff}. We rewrite equation (2) as

$$E_{eff}=\delta\sigma T_{up}^4(1-\eta).$$

If T_{up} and E_a are exactly known to us, then $\Delta\eta=0$ and

$$\Delta E_{eff}/E_{eff}=\Delta\delta/\delta+4\Delta T_{up}/T_{up}.$$

So, the 4% error in δ corresponds to 1% uncertainty in T_{up}, *i. e.* 3°. This is not very small, if one takes into account the fact that in the tropical zone of the ocean the sea surface temperature annual amplitude is around 2°.

In a general case it may be seen from equation (3) that the relative error $\Delta=\Delta E_{eff}/E_{eff}$ is

$$\Delta=[(x-x')/\delta][\eta/(1-\eta)].$$

For a cloudless atmosphere, Δ amounts to $\approx$ 6% without allowance for the film and to 3% including the film; for a continuous cloud cover $\Delta=4\%$.

If the wind speed v m/s is known, we recommend that values for x_0, x_1, ε and δ_0 be used from Tables 4 and 5 and that $\delta_1=0.95$. Under conditions of broken clouds it is necessary to interpolate between magnitudes with the index 0 (cloudless sky with cloud number $p=0$) and the index 1 (continuous cloud cover, $p=1$).

In conclusion, we note that our data may be used only for $v \ll 15$ m/s since the foam on the sea surface is not taken into account.

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