STATISTICAL DEPENDENCE OF OUTGOING SHORT–WAVE RADIATION OVER THE OCEAN ON THE CLOUD AMOUNT

S.I. Avdyushin, A.G. Laktionov, N.A. Lutseva, V.A. Smerkalov, V.S. Suetin, and S.P. Shutikov

E.K. Fedorov Institute of Applied Geophysics, Moscow,
Marine Hydrophysical Institute, Sevastopol'
Received August 18, 1991

The effect of cloud field on the spectral brightness of outgoing short–wave radiation is studied. The statistical processing of the specrtral measurements of B_λ in the "ocean–atmosphere" system and of the cloud amount n over the Atlantic in the 0–30° N *latitude belt is carried out. The analytic dependence of $\bar{B}_\lambda$ on n, wavelength λ, and atmospheric optical mass $m_\odot$ is proposed based on sufficient statistics.*

The majority of theoretical and empirical models intended to calculate the outgoing radiation are based, as a rule, on the study of the processes in the cloudless atmosphere.[1,2,3] In reality the cloudless weather (with the cloud amount 0–2) is observed over the ocean, on the average, not more than in 15% cases.[4]

For this reason it seems important to create the statistical experimental model of the outgoinging radiation taking into account the variations in the cloud amount.

The existing models of the albedo of the "ocean–atmosphere" system were created using the results of the airborne radiation measurements.[5] At present the satellite data on the total cloud amount have most sufficient statistics.

The model proposed is based on the comparison of two satellite data arrays of cloudiness and the outgoing short–wave radiation (0.4–0.8 μm). The measurements were timed and related to one region of the Atlantic (in the 0–30° N latitude belt).

The measurements of the outgoing radiation were carried out for one and a half year (1981–1982) on board the satellite "Intercosmos–21" with the help of a multichannel spectrophotometer. The readings were taken during continuous sounding along the flight path with sighting on nadir. The field–of–view angle of the spectrometer was ~ 1°, the total error in the measurements of the spectral brightness of the "ocean–atmosphere" system was ~ 3%. The value B_λ was recorded in 13 spectral channels of the visible spectral range (0.415, 0.449, 0.483, 0.534, 0.569, 0.621, 0.676, 0.758, 0.761, 0.763, 0.767, 0.794, and 0.823 μm).

In all about 47,000 data points were statistically processed. They were grooped by five solar zenith angle ranges $Z_\odot$: 0–60, 60–75, 75–80, 80–85, and 85–90°. Insignificant seasonal trends in the brightness spectra (not more than 30%) were observed over the Atlantic in the 0–30° N altitude belt, so subsequent data treatment disregarded time of year.

Data of daily satellite observations of clouds given by The All–Union Scientific Research Institute of Hydromachinery Engineering and by the World Data Center (WDC) were referenced to the nodes of the regular grid 5°× 10°. The results of calculations of statistical characteristics of cloudiness obtained during 1981–1982 and related to the Atlantic zone in the 0–30° N altitude belt were used for the creation of the model. The measurement array (~ 25,000 data points) and calculated statistical parameters (the mathematical expectation, rms error, etc.) were in a good agreement with analogous values obtained during 1971–1980.[4]

The albedo of clouds was much greater than the albedo of the ocean, so there was rather a close relation between the upwelling radiation fluxes and the cloud amount n over the oceans. Starting from this fact, each reading of the spectral brightness of the "ocean–atmosphere" system B_λ was characterized by a certain value of n. The model of the relation between the outgoing radiation and the total cloud amount was based on the hypothesis of correspondence between the available arrays of B_λ and n. When creating the model, the dependence $B_\lambda(n)$ was approximated by the following relation:

$$B_\lambda(n) = B_{0\lambda} + n^\alpha (B_{1\lambda} - B_{0\lambda}) , \tag{1}$$

where $B_{0\lambda}$ is the spectral brightness of the outgoing radiation of the "ocean–atmosphere" system for $n = 0$, $B_{1\lambda}$ is the same brightness for the continuous cloudiness ($n = 1$), and n is the cloud amount (the cloud cover index devided by 10).

The value of α was determined from Eq. (1) assuming the correspondence between the distribution functions of the quantities n and B_λ in addition, the values of n and B_λ, corresponding to the 10 and 90% probability of their recording, were taken as the reference data points. For the data arrays under consideration the 10% probability of recording of the spectral brightness of the outgoing radiation corresponded to the cloud amount being equal to 2, and the 90% probability corresponded to the cloud amount being equal to 9.

The analysis of variation in the exponent α entering into Eq. (1) showed that it depended on the optical mass of the atmosphere $m_\odot$ (which is a function of the solar zenith angle $Z_\odot$) and for all wavelengths under consideration can be represented in the form

$$\alpha = 2.5\, m_\odot^{-0.1} . \tag{2}$$

Further $B_{0\lambda}$ and $B_{1\lambda}$ were determined for all five ranges of the solar zenith angles from Eq. (1) by virtue of Eq. (2). It was found that the dependences of $B_{0\lambda}$ and $B_{1\lambda}$ on the wavelength λ and the atmospheric optical mass $m_\odot$ can be fitted with satisfactory accuracy by the relations

0235-6880/92/01 59-02 $02.00

$$B_{0\lambda} = (0.008/\lambda^4 + 0.025)(m_\odot + 1)^{-1} S_{\odot\lambda} \,, \quad (3)$$

$$B_{1\lambda} = 0.47 m_\odot^{-0.1} S_{\odot\lambda} \,, \quad (4)$$

where $E = \pi S_{\odot\lambda}$ is the solar spectral constant.

The model was verified by way of calculating the spectral brightness of the "ocean–atmosphere" system with the mean value $\bar{n} = 0.6$ for the region under consideration and with $n = \bar{n} \pm \sigma/2$ (σ is the rms error). Calculated values of $B_\lambda(n)$ were compared with statistical characteristics of measurements $\bar{B}_\lambda$ and $\bar{B}_\lambda \pm \sigma/2$. Mean error in the approximation did not exceed 10%.

It is should be noted that the wavelength 0.761, 0.763, and 0.767 μm, where the absorption by molecular oxygen was presented, were not studied. In addition, the measured values of B_λ at the wavelength $\lambda = 0.823$ μm for small cloud amount appeared to be close in values to the noise level of the instrumentation. This leads to larger approximation errors.

In the range of the solar zenith angles 85–90°, where the variation in the optical mass of the atmosphere was significant and the variance of measured values of B_λ was large, the proposed model provided only rough estimates of spectral brightness of the outgoing radiation. Mean approximation error in this range was 30%.

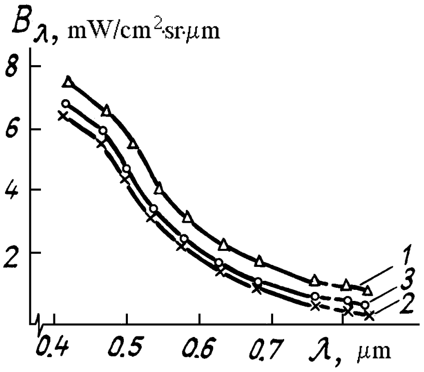

FIG. 1. Average spectral distribution in the brightness of the "ocean–atmosphere" system ($m_\odot = 1.36$): 1) calculated result over the Atlantic taken from Ref. 2, 2) calculated result over the Pacific taken from Ref. 3, and 3) calculated result over the Atlantic according to Eq. (3).

Empirical relations between the estimate of spectral brightness B_λ in the cloudless atmosphere in the 0–30° N altitude belt over the Atlantic and in the 0–30° N altitude belt over the Pacific were obtained in Refs. 2 and 3. Calculated data on B_λ (curve *1* refers to the 0–30° N altitude belt over the Atlantic[2] and curve *2* refers to the 0–30° N latitude belt over the Pacific[3]) for $Z_\odot = 43.8°$ are shown in Fig. 1. Results of calculations from Eq. (3) are shown in Fig. 1 (curve *3* for $Z_\odot = 43.8°$). As can be seen from Fig. 1, the difference between the values of $B_{0\lambda}$ (curve *3*) and B_λ (curve *1*) does not exceed 30% when the wavelength $\lambda \le 0.6$ μm and increases up to 100% at $\lambda = 0.823$ μm. This can be considered as quite a satisfactory agreement taking into account a large variance of B_λ in the red spectral range and substantial difference in the techniques of calculation of $B_{0\lambda}$ and B_λ. The values of B_λ over the Pacific appeared to be quite close to $B_{0\lambda}$ calculated for the equatorial region of the Atlantic.

Proceeding to the analysis of the constructed model (relations (1)–(4)) we must compare it with the existing models of the albedo of the "ocean–cloudy atmosphere" system. The correction for the albedo ΔA due to the presence of clouds was found in Ref. 5

$$\Delta A = 0.1n/(0.826 - 0.6\, n) \,. \quad (5)$$

Relation (5) can be approximated for $n > 0.4$ by the function

$$\Delta A = 0.43\, n^{2.5} \,, \quad (6)$$

where n is the cloud–cover index devided by 10. The recurrence of the albedo of the upper cloud boundary A_n for $n = 0.8$–1.0 was also analyzed in Ref. 5. The most probable value of A_n was 0.41–0.49. Dependences of albedo (6) and of spectral brightness of the "ocean–atmosphere" system (1)–(2) on the cloud amount were close to each other.

Thus, the model can be used for calculation of the spectral brightness of the outgoing short–wave radiation over the equatorial regions of the Atlantic for an arbitrary cloud cover index.

REFERENCES

1. G.I. Marchuk, ed., *Monte Carlo Method in the Atmospheric Optics* (Nauka, Novosibirsk, 1976), 233 pp.
2. A.G. Laktionov, N.A. Lutseva, V.A. Mironenko, and V.S. Suetin, Atm. Opt. **3**, No. 1, 84–88 (1990).
3. A.G. Laktionov, N.A. Lutseva, V.A. Mironenko, and V.S. Suetin, Atm. Opt. **4**, No. 1, 86–88 (1991)
4. L.T. Matveev, ed., *Global Cloud Field* (Gidrometeoizdat, Leningrad, 1986), 278 pp.
5. E.M. Feigel'son, ed., *Radiation in the Cloudy Atmosphere* (Gidrometeoizdat, Leningrad, 1981), 280 pp.