Transmission of the atmosphere over the Atlantic Ocean. Part 1. Spatial inhomogeneities of the transmission

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Spatial distribution of the atmospheric transmission over the Atlantic Ocean is discussed based on the results of five research missions (1989–1996). The transmission inhomogeneities are shown to be mostly due to variations of the atmospheric aerosol optical thickness affected by the carry-over of continental aerosol. The carried over aerosol is small-sized in midlatitudes and coarse disperse one in the trade-wind zone. The characteristics of moisture content variability and the height of homogeneous layers in the atmosphere are presented and the necessity to distinguish between two latitude zones in tropics is underlined. A model of the transmission spatial distribution is proposed based on the consistent results of genetic and descriptor-based zoning of the atmospheric aerosol turbidity. Peculiarities of the spectral behavior of aerosol optical thickness are treated and its parameters in 0.37–1.06 and 0.37–4 μm are determined for different oceanic regions. The contribution from fine and coarse aerosols to the aerosol optical thickness and total transmission of the atmosphere are discussed.

Introduction

The influx of direct solar radiation under cloudless conditions is determined by the total transmission function of the atmosphere $T^{\Sigma} = T^{A} \cdot T^{W} \cdot T^{C}$, that depends on “aerosol” T^{A}, “moisture” T^{W}, and the components due to transmission by other gases T^{C}. Important peculiarities of the principal components, T^{A} and T^{W}, are their regional features and strong synoptic variability due to change of air mass. Other components can be taken into account as constants or calculated from the mean model data for specific seasons and latitudinal zones. So, if not considering the long-term trends, the variations of T^{Σ} in different regions are practically completely determined by the variations of T^{A} and T^{W}.

The continental network of observation stations provides for a vast experimental material that enables revealing the regularities in the spatiotemporal variability of the atmospheric transmission, water vapor column density, and aerosol optical thickness (AOT) $\tau(\lambda)$. The information on optical characteristics of the atmosphere over oceans is not complete and reliable, because it is based on the results of a few in number shipborne missions and data of coastal stations affected by the continents.

In 1989–1996 we have carried out atmospheric optics investigations in some regions of the Atlantic Ocean, which made it possible, to a certain degree, to fill the gap in our knowledge about the transmission of the marine atmosphere.[1–4] The measurements were carried out by means of the multiwave sun photometer in the transmission windows of the atmosphere, in the wavelength range 0.37–1.06 μm (some of these data were obtained in the range near 4 μm). The moisture content of the atmosphere was determined by use of a differential absorption technique within the water vapor absorption band at 0.94 μm. The instrumentation used in the study was considered in a more detail in Refs. 5, 6, and others. The results of investigation into the spatial distribution of the transmission characteristics, peculiarities of diurnal behavior of AOT and moisture content of the atmosphere over some parts of the ocean are generalized in this paper.

1. Spatial distribution of the AOT of the atmosphere

Genetic zoning of the aerosol turbidity of the atmosphere over the Atlantic Ocean into zones was proposed in our previous paper,[1] which takes into account the combined effect of two factors, the sources (types) of continental aerosol prevalent in each latitudinal zone and circulation (transport) of the air masses. Based on literature data we have isolated the following zones: open ocean (OO), coastal (C), near-to-continent (NC), Canary Islands (CI), trade wind zone (TW), “Sea of Darkness” (SD), and the equatorial zone (E). The characteristics of the transmission themselves have been used only for the indirect assessment of the classification.

Different approach was realized in testing the results of genetic zoning – the “descriptor-based” classification of the regions with the characteristics $\tau(\lambda)$ of the same type. Isolating particular regions in the “descriptor-based” classification[7] was based on the quantitative data on some directly measured characteristics of the atmosphere. As was shown in the case with AOT[8] the two "descriptors," namely, the value of optical thickness τ (0.55 μm) and the Angström parameter that characterizes the spectral dependence

$$\tau(\lambda) = \beta\, \lambda^{-\alpha}. \qquad (1)$$

are enough for distinguishing among typical zones.

The procedure isolating the zones included two stages. First, the maps were constructed with the isolines of the characteristics under study (Fig. 1). The values $\tau_{0.55}$ and α were averaged over the squares with the steps of 5° in latitude and longitude, and the missing points were calculated by means of inter- and extrapolation. The results obtained are in a good agreement with the conclusions drawn in making the genetic zoning.[1] For example, the distribution of the isolines in mid-latitudes is an evidence of the increase in $\tau_{0.55}$ and in the selectivity of the spectral behavior of α when approaching continents both in westward and eastward directions. Another situation is observed in the trade wind zone. The value of the atmospheric turbidity continuously increases from America to Africa, while the parameter α is minimum in the central part of the ocean and it has the concentric distribution of isolines (increases in all directions).

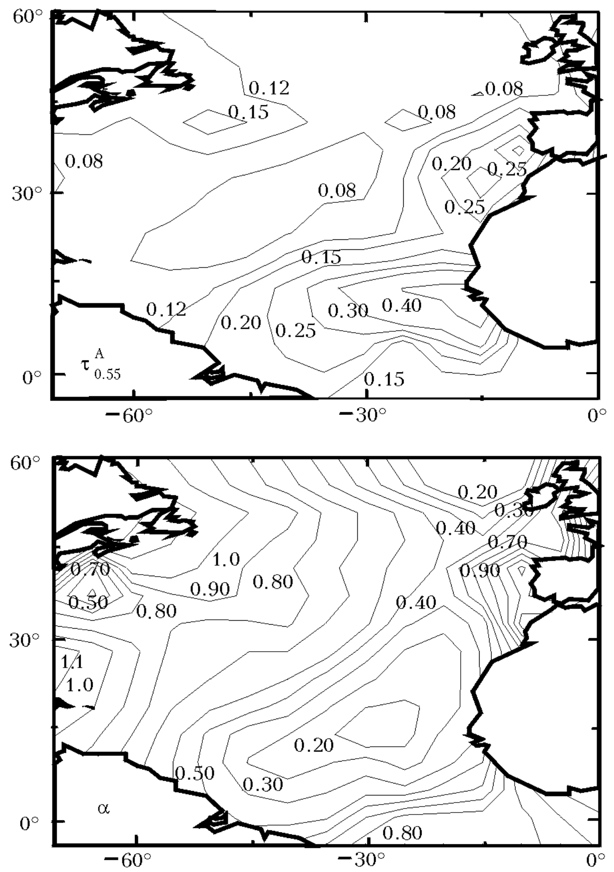

Fig. 1. Isolines of the spatial distribution of $\tau_{0.55}$ and the parameter α.

The second stage is the determination of the zone boundaries. The criterion used in selecting the OO zone was that the values $\tau_{0.55}$ and α fall within the range "mean ± standard deviation" obtained in the central part of the mid-latitude ocean. Separation between the OO and NC zones was practically determined by the $\tau_{0.55}$ isolines. Criteria for selecting the boundaries in tropical and equatorial zones were the qualitative differences in the spatial distributions of $\tau_{0.55}$ and α. The boundaries were drawn in the regions of maximum gradients or minimum values of the parameters in moving from one zone to another. As a result, the "aerosol provinces" separated out in this way were quite close to those resulting from the genetic zoning (Fig. 2). The samples obtained for separate zones are statistically distinguishable between each other by at least one parameter with the confidence probability not less than 0.9, and the mean values of $\tau_{0.55}$ and α are in a good agreement with the results of other observations carried out in similar zones.[8–10] Let us also note that the principal differences in $\tau_{0.55}$ of different zones are observed in the wavelength range up to 1 μm (see below). So the results of the zoning are valid at least up to 4 μm (i.e., in the entire wavelength range where solar radiation plays most important part).

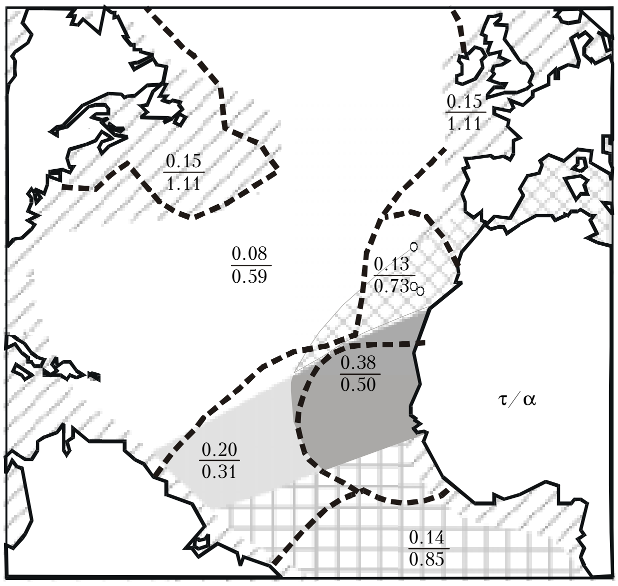

Fig. 2. The results of the "descriptor-based" (dashed lines) and genetic zoning (the mean values of $\tau_{0.55}/\alpha$ are shown as fractions).

2. Spectral behavior of the atmospheric AOT in different regions

The differences in $\tau(\lambda)$, that are observed in some wavelength ranges, are not that significant because of a low content of fine aerosol in the marine atmosphere. Nevertheless, monotonic decrease of τ with the increasing wavelength is observed in the range from 0.37 to 1 μm (details see in Ref. 1). Deviations from the Angström power-law approximation (1) usually do not exceed 3–6%, and 12% in the OO zone. Mean values of the power parameter α are within the limits from 0.3 in the trade wind zone to 1.1 near continents (see Fig. 2).

As it follows from literature data, the measurements of the atmospheric transmission in the range of wavelength longer than 1 μm are only occasional. The results of our investigations have shown that the spectral dependence $\tau(\lambda)$ in the long-wave region changes from power-law to the neutral one (Fig. 3). The results of other investigations carried out under similar conditions[11–14] are shown in Fig. 3 for a comparison. Let us pay attention to the closeness between the data obtained in the "Sea of Darkness" to the values $\tau(\lambda)$ measured during the events of dust emissions from Senegal[11] and the South of France.[12] The measurements presented in Refs. 13 and 14 were carried out in the Norwegian Sea (relatively close to the shore) so some data are intermediate between the spectra $\tau(\lambda)$ in the OO and NC zones. On the whole, one can note that the AOT of the atmosphere over the regions far from dust emissions (i.e., in the OO, NC, and E regions) measured in the IR range is approximately the same (~0.06) and is mainly caused by coarse particles of marine aerosol itself. The enhanced content of large dust particles in other regions leads to an increase of the AOT in IR range by several times.

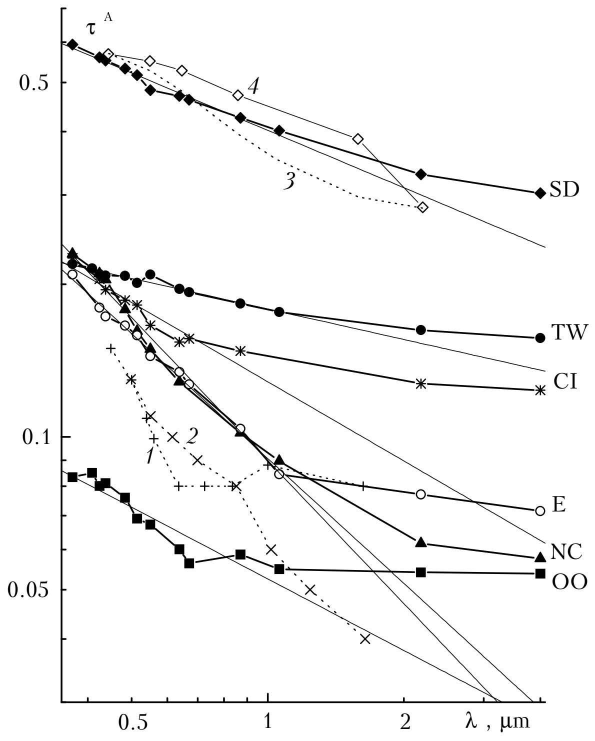

Fig. 3. Mean spectra $\tau(\lambda)$ in different regions: (thin lines are approximation by a power-law dependence; *1* and *2* are results of measurements in the Norwegian Sea[13,14]: *3* and *4* are AOT during dust emissions from Senegal[11] and from the south of France.[12]

As was noted by many authors, description of the spectral behavior $\tau(\lambda)$ by formula (1) in a widened wavelength range (0.37 to 4 μm) is not justified. The mean, over the data array considered, values of $\tau(\lambda)$ deviations from the power-law dependence undergo a many-fold increase in different geographical regions. Thus, the increase of this parameter measured in the spectral range up to 2 μm reaches 2 to 12 times, and by 4 to 18 times at 4 μm. Representation of $\tau(\lambda)$ in the form of a sum of two components caused by the effect of fine and coarse aerosol is more appropriate:

$$\tau(\lambda) = \tau_f + \tau_c = m\,\lambda^{-n} + \tau_c. \qquad (2)$$

By analogy with the Eq. (1), the parameter n is related to the refractive index and particle-size distribution function of fine aerosol, while the coefficient m characterizes the number density of particles. The values τ in the spectral range at 4 μm ($\tau_c = \tau_4$) can be used as the component of AOT due to coarse aerosol that has neutral spectral behavior.

The parameters in the representation (2) calculated by the method of least squares for mean $\tau(\lambda)$ are shown in Table 1. The data obtained make it possible to estimate the contribution of fine and coarse aerosol particles to AOT in different regions and wavelength ranges. In particular, the value $\gamma = (m\lambda^{-n}/\tau_c)$ characterizes redistribution between the effects of the two fractions in the wavelength range from 0.37 to 1.06 μm. Distribution of τ_c over regions, with the maximum in the "Sea of Darkness" seems to be quite obvious. Since the values m and γ are high in the SD zone, this region is characterized by a high content of both the coarse and fine aerosol. The zones are divided into three groups by the degree of selectivity of $\tau_f(\lambda)$: minimum values are observed in the trade wind zone, the intermediate ones are in the CI, E, and NC zones, and the maximum values are in the open ocean. Insufficient data does not allow an unambiguous interpretation of the latter fact. One can only suppose that the distribution functions of fine aerosol particles in clear air of OO is the most narrow and is displaced to the range of smaller size.

Table 1. Parameters of the $\tau(\lambda)$ model and the estimates of the contribution coming from different aerosol fractions

Regions	OO	TW	CI	SD	E	NC
$\tau_c = \tau_4$	0.054	0.157	0.124	0.303	0.071	0.058
m	0.002	0.020	0.016	0.089	0.022	0.027
n	2.83	1.30	1.85	1.30	1.91	2.06
$\gamma = \tau_f/\tau_c$	0.62– –0.03	0.46– –0.12	0.81– –0.12	1.07– –0.27	2.07– –0.28	3.61– –0.41

The spectral behavior of $\tau(\lambda)$ in OO regions is of independent interest from the standpoint of their background character. The matter is, that the composition of the atmosphere over all other regions is a mixture of the marine aerosol with different types of aerosol generated over the land. The continental air mass undergoes significant transformations during its long-range transport, and practically losses its continental features when reaching the remote areas of the ocean (of coarse, except for the zone of the north-east trade wind in the Atlantics). That means, that the atmosphere losses the regional dependence on the sources (emissions) of continental aerosol and is approximately the same everywhere over the World Ocean. The results of observations in the Pacific and Indian oceans[8–10] confirm this conclusion.

It is seen from Fig. 3 that $\tau(\lambda)$ in the OO reaches the level of neutral wavelength dependence in the spectral range from 0.6 to 1 μm much sooner. In this case one can estimate the component τ_c due to coarse aerosol from the minimum values of AOT in a range of shorter waves ($\tau_c = \tau_{min}$) and simulate the spectral behavior of $\tau(\lambda)$ in the form similar to Eq. (2) (the notation of the parameters being m_0 and n_0). Hence, there is a possibility of estimating the spectral dependence better if using the array of data over the entire spectral region (0.37 to 4 μm) and not only the limited (0.37 to 1.06 μm) one. If excluding extraordinary situations of "splashes" in the values τ in the short-wave range resulting from the long-range transfer of fine aerosol from continents, we obtain the spectral dependence under really background conditions of the ocean (Table 2).

Table 2. Statistics of the parameters of the spectral dependence $\tau(\lambda)$ for the ocean regions not affected by continents (V is the coefficient of variation)

Parameter	Mean	Standard deviation	V	Min	Max	N
$\tau_c = \tau_{min}$	0.035	0.011	0.31	0.019	0.064	48
m_0	0.0038	0.0018	0.49	0.001	0.008	48
n_0	2.18	0.66	0.30	0.76	3.15	48

One can judge about the content and stability of the two main fractions, coarse marine particles and secondary, evidently, sulfate aerosol, from the statistical characteristics. Rough estimates show that for the background spectral dependence to occur the aforementioned fractions can be separated on the size scale. Obviously, fine particles (in the range lower than the first maximum of the scattering efficiency factor) do not exceed 0.2-μm size, and coarse particles are concentrated in the size range greater than 1 μm. Otherwise, the power-law decrease of $\tau(\lambda)$ should be more slow and extended to 1 or even 4 μm wavelength.

The spectral behavior $\tau(\lambda)$ is closely related to the question on the correlation among AOT values at different wavelengths. Analysis of the correlation coefficients $R(\tau_{\lambda_i}; \tau_{\lambda_j})$ has shown high correlation (as in a narrow wavelength range[2,8]) that decreases as the spectral range widens ($\lambda_i - \lambda_j$). For example, the correlation coefficients $R(\tau_{0.37}; \tau_4) = 0.787$ and $R(\tau_{1.06}; \tau_4) = 0.855$ at the critical correlation value 0.082 (with the confidence probability of 0.95). High level correlation, (τ_{λ_i}; τ_{λ_j}), provides the grounds to write the regression equations for estimating AOT in the IR based on data obtained in the short-wave region. The parameters of regression formulas for $\tau_{0.514}$ and $\tau_{1.06}$ are given in Table 3, and Fig. 4 illustrates real scatter of measured and calculated values τ_4 and $\tau_{2.18}$. It follows from the data presented that the error in determining the AOT based on a single input parameter (τ_{λ_0}) can be assessed to be quite satisfactory.

Table 3. Parameters of the linear regression $\tau_\lambda = a + b\,\tau_{\lambda_0}$ for two typical regions (Δ and δ are the absolute and relative regression errors, respectively)

λ, μm	λ_0, μm	a	b	Δ	δ,%
		Open ocean			
2.18	0.514	0.004 ± 0.05	0.755 ± 0.065	0.02	37
	1.06	0.010 ± 0.004	0.792 ± 0.053	0.02	31
4.0	0.514	0.012 ± 0.007	0.624 ± 0.088	0.03	50
	1.06	0.011 ± 0.005	0.753 ± 0.070	0.02	41
		"Sea of Darkness"			
2.18	0.514	0.005 ± 0.018	0.602 ± 0.036	0.04	15
	1.06	0.005 ± 0.014	0.750 ± 0.035	0.04	12
4.0	0.514	0.036 ± 0.031	0.472 ± 0.061	0.07	26
	1.06	0.029 ± 0.024	0.638 ± 0.061	0.06	23

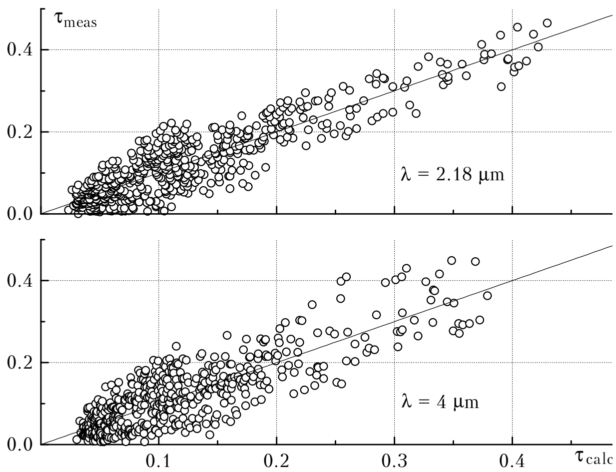

Fig. 4. Comparison of AOT values measured and calculated by regression equations (at $\lambda_0 = 0.514$ μm).

3. Moisture content and height of the homogeneous atmosphere H_0

In analyzing the variability of the moisture characteristics, we used the data on the column density of water vapor W and height of the homogeneous atmosphere $H_0 = W/a_0$ (a_0 is the absolute humidity in the near-water layer).[15] Principal regularity of the spatial distribution of CDWV over the ocean is its dependence on latitude. (We did not observe any significant dependence of W when approaching the continents in our investigations).

Isolating 3 or 4 latitudinal zones[16] is most widely used. Those are the tropical zone from 0 to 25°, midlatitudes 45–65° and the arctic zone 65–90°. The shipborne observations[3,17] have shown the necessity of a different division into zones in the tropics. The matter is, two areas of tropical latitudes, namely, the intertropical convergence zone (ICZ) and the trade wind zone have significant differences in atmospheric conditions.[7,19] The trade wind zone (~30–7°N) is characterized by the stable strong wind, small cloud amount and subsidence inversion that prevents the spread of water vapor aloft. The ICZ is a relatively narrow lane near the equator and is characterized by weak wind, developed convection, and cloudiness.

The consequence of the aforementioned differences is a large gradient of W (0.35 g/cm^2 per one degree of latitude) and the height H_0 at the northern boundary of the ICZ (Fig. 5). The division of the tropical zone into two parts becomes natural based not only on the data of a single observation. It follows from the data of long-term observations[18] that the spatiotemporal variations of the ICZ position leads to a certain smoothing of the boundary. Nevertheless, even the averaged data reveal the enhanced gradient of W at the ICZ boundary quite distinctly, especially in the eastern part of the Atlantics (~0.2 g/cm^2 per one degree of latitude). Thus, by analogy to the aerosol zoning, three zones were isolated to the south of 60°N, within which the characteristics of time variation of W and H_0 were studied.[22] The considered peculiarity was also observed in zoning the total atmospheric transmission.

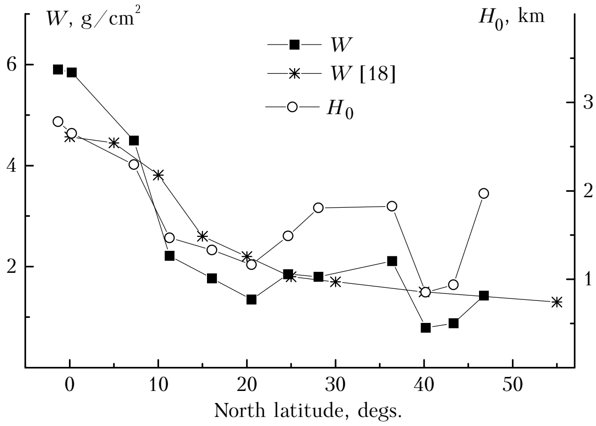

Fig. 5. The dependence of W and H_0 on latitude obtained during the 35th mission of the research vessel *Akademik Mstislav Keldysh*[3] and calculated from the data of long-term observations[18] (April; 20°W).

4. Spatial distribution of the integral atmospheric transmission

Determination of the spatial distribution of the integral (over the wavelength range 0.37 to 4 μm) atmospheric transmission and the direct solar radiation S was based on the results of AOT zoning. That means, that first we calculated the aerosol-caused components of the atmospheric transmission T^{A} in the selected regions (see Sect. 1), then the mean values W and T^{W} were determined for the same regions, and the integral transmission T^{Σ} was determined at the concluding stage taking into account constant gases. The calculations were made for the elevation angle of the Sun of 30° using the LOWTRAN-7 computer code[20] by the following formulas:

$$T^{\mathrm{A}} = \int S_{0\lambda} \exp(-\tau_\lambda M)\, \mathrm{d}\lambda \Big/ \int S_{0\lambda}\, \mathrm{d}\lambda; \tag{3}$$

$$T^{\mathrm{W}} = \int S_{0\lambda}\, T_\lambda(W M)\, \mathrm{d}\lambda \Big/ \int S_{0\lambda}\, \mathrm{d}\lambda; \tag{4}$$

$$T^{\Sigma} = \int S_{0\lambda}\, T^{\mathrm{C}}_\lambda\, T^{\mathrm{W}}_\lambda\, T^{\mathrm{A}}_\lambda\, \mathrm{d}\lambda \Big/ \int S_{0\lambda}\, \mathrm{d}\lambda; \tag{5}$$

$$S = \int S_{0\lambda}\, T^{\mathrm{C}}_\lambda\, T^{\mathrm{W}}_\lambda\, T^{\mathrm{A}}_\lambda\, \mathrm{d}\lambda, \tag{6}$$

where $S_{0\lambda}$ is the spectral solar constant,[21] $M = 2$ is the optical mass of the atmosphere. The function T^{W} was calculated based on the data of long-term observations on W obtained by Tuller.[18] The mean value τ_λ were used in calculating T^{A} in the spectral range from 0.37 to 1.06 μm, and neutral spectral behavior of AOT was assumed at λ longer than 1 μm. The latter was caused by poorer reliability (smaller bulk) of data on $\tau(\lambda)$ in the IR range. However, it is more important that in the spectral range $\lambda > 1$ μm the transmission T^{Σ} is mainly determined by the moisture component T^{W}. Thus, some annually mean distribution of the transmission and the incoming direct radiation over the Atlantic Ocean was obtained (Table 4). The annual mean distribution needs an explanation.

The results of investigations[1–4,8–10] of $\tau(\lambda)$ were obtained in different periods, but the data are not still sufficient for revealing any seasonal peculiarities. The results of the whole-year spaceborne observations of τ appeared in recent years refer to a single wavelength and are not sufficiently accurate in the range of small values. Let us also note that the seasonal variability of the AOT over the ocean should be less pronounced, because of the generation of marine aerosol that occurs whole year round, and the effect of regions with mild climate with no changes of the land cover is prevalent in the emissions of the continental aerosol. Approximate estimates of the seasonal variability of τ are presented in Ref. 22. Quite different situation is observed with the moisture content. The amount of data available is significantly greater, and the annual behavior of W is quite well pronounced. It was for this reason that the obtained values $\tau(\lambda)$ are assumed to be close to the annual mean ones, and, correspondingly, the data of long-term observations[18] on W were used in calculating T^{W} and not our data for some months.

Table 4. Results of zoning of the components of the integral atmospheric transmission and the influx of the direct solar radiation at the elevation angle of the Sun of 30° (regions within the latitudinal zones are presented consequently from west to east; *X* is the total ozone content)

Latitudinal zone	region	X, D.U.	W, g/cm^2	T^{A}	T^{W}	T^{Σ}	S, W/m^2
Mid-latitudes	NC	331.6	1.27	0.792	0.840	0.561	767
	OO		1.80	0.872	0.823	0.605	827
	NC		1.65	0.792	0.828	0.553	756
Subtropics	NC	300	3.35	0.792	0.787	0.526	719
	OO		3.2	0.872	0.790	0.581	794
	CI		2.4	0.796	0.807	0.542	741
Tropical (trade wind) zone	NC	277.3	4.2	0.792	0.771	0.516	705
	OO		4.05	0.872	0.774	0.570	779
	TW		3.65	0.683	0.781	0.451	617
	SD		3.3	0.498	0.788	0.331	452
Equatorial (ICZ)	E	277.3	4.5	0.787	0.767	0.510	697

It is seen from Table 4 that the integral transmission varied almost twice due to the spatial inhomogeneities in the aerosol and moisture content. The aerosol component plays more significant role in variations of T^{Σ}: the ranges are $T^{A} \approx 0.5 - 0.87$ and $T^{W} \approx 0.77 - 0.84$. The greater variations of T^{A} are characteristic of both the latitudinal and meridional sections. According to calculations, the influx of direct radiation reaches maximum values in the regions of open ocean of mid-latitudes (0.82 kW/m^2) and minimum in the “Sea of Darkness” (0.452 kW/m^2).

One more issue was considered based on the data obtained: the effect was estimated of fine and coarse aerosol on the integrated (over the wavelength range from 0.37 to 4 μm) AOT of the atmosphere (for $M = 1$):

$$\tau^* = -\ln T^{A} =$$
$$= -\ln\left[\int S_{0\lambda} \exp(-\tau_{\lambda})\, d\lambda \Big/ \int S_{0\lambda}\, d\lambda\right]. \quad (7)$$

As before, the coarse component τ_c^* was determined from the AOT value at 4 μm, and the fine component was calculated from the difference $\tau_f^* = (\tau^* - \tau_c^*)$. The calculated results have shown that the coarse aerosol mainly affects τ^* and the influx of direct solar radiation in the majority of regions (Table 5). The fine fraction of aerosol becomes prevalent only in the coastal areas of mid-latitudes (NC).

Table 5. Comparison of the contributions of the components due to fine (τ_f^*) and coarse ($\tau_c^* = \tau_4$) aerosol fractions to the total (τ^*) AOT of the atmosphere over different regions of Atlantics

Regions	OO	TW	CI	SD	E	NC
τ^*	0.061	0.188	0.157	0.443	0.117	0.120
τ_f^*	0.007	0.031	0.033	0.140	0.046	0.062
τ_f^*/τ^*	0.12	0.17	0.21	0.32	0.39	0.52
τ_c^*/τ_f^*	7.71	5.07	3.76	2.16	1.54	0.94

Conclusion

Taking into account the results obtained by other authors, the main peculiarities of the variability of the atmospheric transmission over the ocean have been determined, the necessity is noted of regional approach to description of its properties and the model is proposed of the transmission spatial distribution. It is shown that component of the transmission due to aerosol T^{A}, or the AOT of the atmosphere, is the most variable. Spatial inhomogeneities in $\tau(\lambda)$ are primarily determined by different content of continental aerosol: fine in mid-latitudes and coarse dust particles in the tropics. The parameters of empiric dependences $\tau(\lambda)$ provide for the quantitative characterization of the mean spectral behavior of AOT in different regions and are an evidence of the priority role of the coarse aerosol over the most part of the ocean. In particular, the contribution of fine aerosol to the integral AOT dominates only in mid-latitudes near the continents.

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