Transmission of the atmosphere over the Atlantic Ocean. Part 2. Variations of the aerosol optical thickness and moisture content

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Analysis is being continued in this paper of the time variation of the atmospheric transmission characteristics and their interconnection with the meteorological quantities within the near-water layer over the ocean. Statistical characteristics of the most pronounced diurnal variations of the aerosol optical depth and moisture content in the oceanic regions, identified earlier (see Part 1. Spatial inhomogeneities of the transmission), are determined. A weaker component of the diurnal behavior of the atmospheric aerosol optical depth is revealed that has maximum in the forenoon hours. The influence of wind manifests itself in the non-monotonic increase in the aerosol turbidity (coarse aerosol fraction) with the increasing wind speed. Using data collected in the equatorial zone, as an example, the interconnection with relative humidity was examined and empiric dependence proposed allowing one to calculate the component of atmospheric optical thickness due to extinction by small-size aerosol. Estimates of the contributions coming from different factors in variations of the atmospheric transmission characteristics are presented.

Introduction

The first part of the paper[1] dealt with the spatial variability of the characteristics of the spectral and integral atmospheric transmission over the Atlantic Ocean caused by the differences in the aerosol optical thickness (AOT) and column density of the atmospheric water vapor (CDWV). The main result was in isolating the oceanic regions having same characteristics of AOT of the atmosphere $\tau(\lambda)$, *viz.,* open ocean (OO), near-to-continent (NC), Canary Islands (CI), "Sea of Darkness" (SD), trade wind (TW), and equatorial (E) zones. Separating out of these zones of the ocean made it possible to analyze the characteristics of time variation and their relations with the meteorological quantities of the near-water layer of the atmosphere separately. Coastal regions and Mediterranean Sea were excluded from the subsequent analysis because of a smaller bulk of data compiled.

1. Variations of AOT of the atmosphere in different regions

Statistical characteristics of the interdiurnal variability caused by the change of air masses were considered in our previous papers (Refs. 2–4 and others) using the daily mean values. The property common for all regions is the spectral behavior of some characteristics, i.e., τ_{min}, τ_{max}, and σ_τ decrease with the increasing wavelength as the AOT does. The coefficients of variation ($V_\tau = \sigma_\tau/\bar{\tau}$) have no any pronounced spectral dependence, and their values are higher than that in the continental atmosphere. For example, in the OO, CI, and NC regions the value V_τ is 0.6–0.8 as compared with 0.3–0.5 in Tomsk region.[5] Analogous excess is observed in the variability of the Angström parameter α that characterizes the spectral behavior $\tau(\lambda)$:

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

The coefficients of variation V_α for regions over the ocean are in the range 0.45 to 1.11 in contrast to 0.3–0.4 over the continent. The absolute variation (σ_τ) has maximum value in the "Sea of Darkness" and minimum in the OO zone. The values σ_τ over the ocean are comparable with the continental ones, but mean $\bar{\tau}$ in the majority of regions is lower, what is just the cause of the increased relative variations of AOT in the atmosphere over the ocean.

To comprehensively describe the random variations of τ_λ, it is necessary to find the distribution law. To do this, the frequency histograms were constructed using hourly mean values τ(0.48 μm) and the comparison was made with some theoretical laws of the probability density distribution (lognormal, Rayleigh, Weibull, etc.) Histograms of the distributions $\tau_{0.48}$ were divided into two types: single-mode, in the regions OO, NC, and SDand diffuse in the other regions.[2] Then only single-mode distributions were analyzed. The results of analogous investigations conducted in summer on the continent (Tomsk[5]) were considered for a comparison. It is seen from the histograms (Fig. 1) and statistical data given in Table 1 that the distributions of τ are elongated toward stronger turbidity ($\gamma_1 > 0$) and have sharper peaks than the probability density of the normal distribution ($\gamma_2 > 0$). The absolute values of the coefficients γ_1 and γ_2 significantly exceed their rms errors in all samples.

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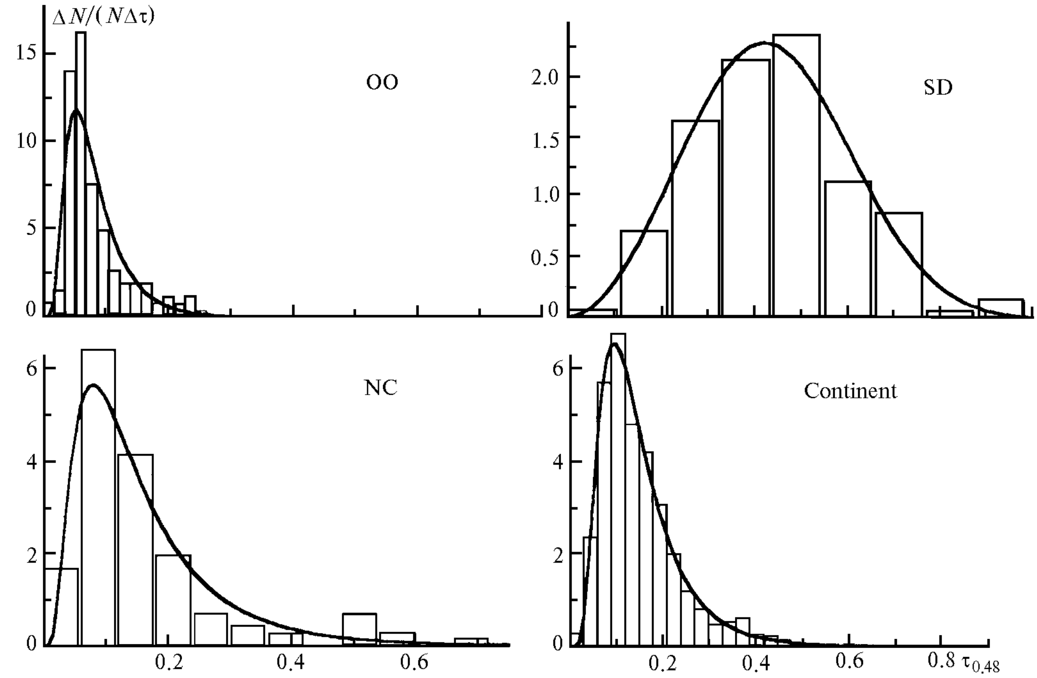

Fig. 1. Histograms of distribution of $\tau_{0.48}$ and theoretical probability densities for three regions of the Atlantics and continent (Tomsk).

Table 1. Statistical characteristics $\tau_{0.48}$ (γ_1 and γ_2 are the asymmetry and excess coefficients)

Region	$\bar{\tau}$	σ_τ	V_τ	Min	Max	Median	γ_1	γ_2	N
OO	0.08	0.05	0.59	0.01	0.26	0.07	1.16	2.03	152
NC	0.16	0.13	0.78	0.02	0.72	0.12	2.08	4.32	120
SD	0.45	0.18	0.40	0.07	1.26	0.44	0.81	1.83	179
Tomsk	0.15	0.09	0.58	0.015	0.65	0.13	1.57	3.26	983

The parameters of theoretical distributions were selected from the known relationships[6] using the sample initial and central moments of the corresponding orders. Estimation by the chi-square criterion has shown that the best approximation of empiric histograms for the regions of Tomsk, OO, and NC is the lognormal law. Lognormal distribution is the only acceptable for OO with the confidence probability of 0.95. The Weibull distribution better suits for describing the histograms in the SD zone (possibly, because of insufficient duration of measurements for revealing the characteristic variations of the AOT in this region).

One can explain the coincidence of the distribution laws for three geographical regions of different type (Tomsk, NC, and OO) by the existence of a general factor that determines the AOT variability. Estimates of the dependence of normalized variance (σ_i^2/σ_τ^2) on different intervals of averaging have shown that the principal characteristic decrease is observed in the range of 1–10-day averaging (Fig. 2), i.e., synoptic variations with the period of 3–5 days make the main contribution to the AOT time variation. Coincidence of the statistics of daily mean and hourly mean values of $\tau_{0.48}$ is also an evidence of this fact. Transition to the hourly mean values of τ that bear information about the diurnal variability, practically does not change the variance.

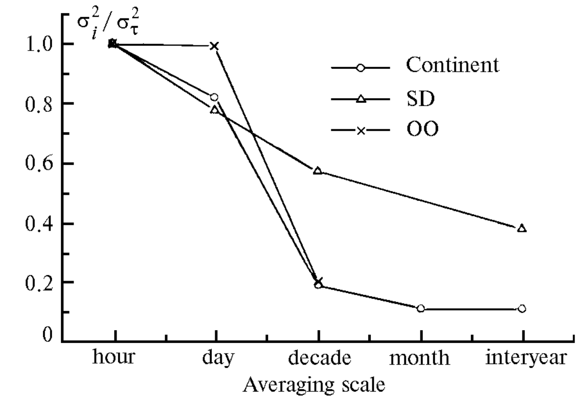

Fig. 2. Normalized values of the variance of AOT as a function of the duration of averaging intervals.

Indeed, the regular component of the diurnal behavior of τ is essentially lower than the synoptic variability and was not analyzed before our investigations.[3,4] We have succeeded in revealing the diurnal variability only for separate regions with weak synoptic influence (OO, E, CI) using averaging and normalizing to the daily mean values. Summarizing the results from Refs. 3 and 4, let us present the mean dependence $\tau(t)$ for three regions (Fig. 3). The principal peculiarity is the increase of τ by noon and then its more rapid fall off by the evening. One can estimate the relative diurnal change of τ_λ as 15–30%. Another peculiarity is related to the change of selectivity in the spectral behavior of τ. The parameter α decreases a little bit by 11 a.m., and then it increases by 27% relative to the mean level. The relative increase and subsequent decrease (sink) of the content of large particles in the air correspond to that dependence. The significance of the diurnal behavior of τ and α is confirmed by high confidence probability p calculated using the Student criterion.

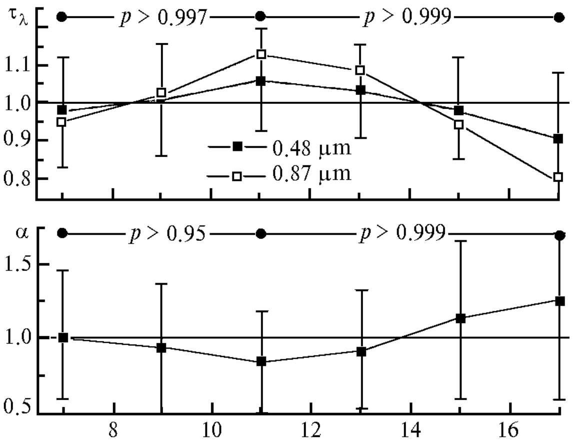

Fig. 3. Illustration to the diurnal behavior of the normalized values τ_λ and α for three region: OO, CI, and E (p is the confidence probability of the significance of the extremes).

Diurnal and synoptic variations of $\tau(\lambda)$ during transfer and transformation of air mass occur under the effect (or taking into account) the change of meteorological conditions. Therefore the traditional problem is revealing the relations with meteorological parameters which directly affect the content and disperse composition of aerosol.

2. Relations of AOT of the atmosphere with meteorological parameters

It is known that revealing the relations is difficult even for the scattering by aerosols of the near-water layer[7] because of the variety of factors and complex dependence of the aerosol characteristics on meteorological data. Moreover, one cannot expect simple highly significant correlation with the integral parameter τ. One should also take into account that mutual diurnal variations of meteorological and aerosol characteristics decrease when moving from the underlying surface, while the role of synoptic processes increases.

Preliminary analysis has shown that the relations of τ with the characteristics of humidity, temperature, and wind velocity are not stable and can be revealed only in some individual samples (regions, expeditions). The significant though weak correlation remains, in the total bulk of data, only with two parameters, wind speed S_w that determines the generation of marine aerosol, and relative humidity (RH) that affects condensation-caused change of the particle size. To decrease the distorting effect of the synoptic factor, subsequent analysis was carried out for two regions with minimum effect of aerosol emissions from continents – OO and E.

Effect of wind speed

When investigating the dependence of $\tau(\lambda)$ on wind speed, additional selection of data was performed. The readouts were selected, in which the humidity did not exceed 75%, i.e., out of zone of the effect of condensation mechanism. The tendency toward increase in the aerosol turbidity as S_w increases is seen in the obtained dependencies (Figs. 4*a–c*, thin lines). The increase of τ_λ in relative units is better seen in the long-wave region of the spectrum. The consequence of the spectral peculiarities is the inverse dependence of the selectivity parameter α on the wind speed (Fig. 4*d*), which is evidence of the primary increase of the concentration of coarse particles. Correlation coefficients of the considered relations are 0.29 for $\tau_{1.06}$ and −0.46 for α at the level of significant correlation of 0.106 with the confidence probability of 0.95. As for results of other investigations of $\tau(\lambda)$ over the ocean, one can refer only to the data in Ref. 8, where analogous correlations were revealed: positive with τ and negative with α.

Mean values and rms errors of τ were calculated in separate intervals of S_w for more detailed analysis of the effect of wind factor (filled circles and solid lines in Figs. 4*a–c*). The obtained dependences are not monotonic with maximum near 3 m/sec, minimum in the range 6–8 m/sec, and then again increase of τ. The extreme points are distinguishable with confidence probability not less than 0.9. Similar results can be obtained for the dependence of the parameter β in the Angström formula (1), which well correlates with the coarse component of AOT (Fig. 4*e*). Minimum β values obtained in a wide range of S_w correspond to situations of the weakened effect of all other factors: the change of air mass (emission of aerosol from continents), condensation growth of particles, etc. Therefore, the envelope of the scatter plot from below is the consequence of the nonlinear dependence of β on the wind.

The non-monotonic dependence similar to the considered one (at the first interval 0–6 m/sec) was revealed earlier for the aerosol scattering coefficient over sea surface in coastal regions.[7] Hence, new confirmation is obtained of the effect in the oceanic regions and for the whole atmospheric column. In physical interpretation one should agree with the authors of Refs. 7 and 21 that the cause of the nonlinear dependence is the difference and redistribution of the role of the mechanisms of aerosol generation in different ranges of wind velocity. According to some data,[9–11] the bubble mechanism is the main at small S_w: burst of the gas bubbles floating to the surface. Generation of big particles ($r > 0.5$ μm) is more effective at the wind velocity up to 2–3 m/sec due to the break-up of the water jets thrown out of the bubble at its burst. The process is accompanied by generation of smaller particles (r < 0.1–0.2 μm) resulting from the break of the bubble film. However the role of this fraction becomes prevalent at S_w = 3–7 m/s when injection of big particles is hindered. At least, one mechanism more is "switched on" at the wind velocity above 7 m/s: the generation of the largest particles at blowing splashes from crests of waves and subsequent drying. Thus, the obtained optical results $\tau_\lambda(S_w)$ are in a good agreement and confirm the current idea on the peculiarities in generation of marine aerosol under conditions of variable wind field.

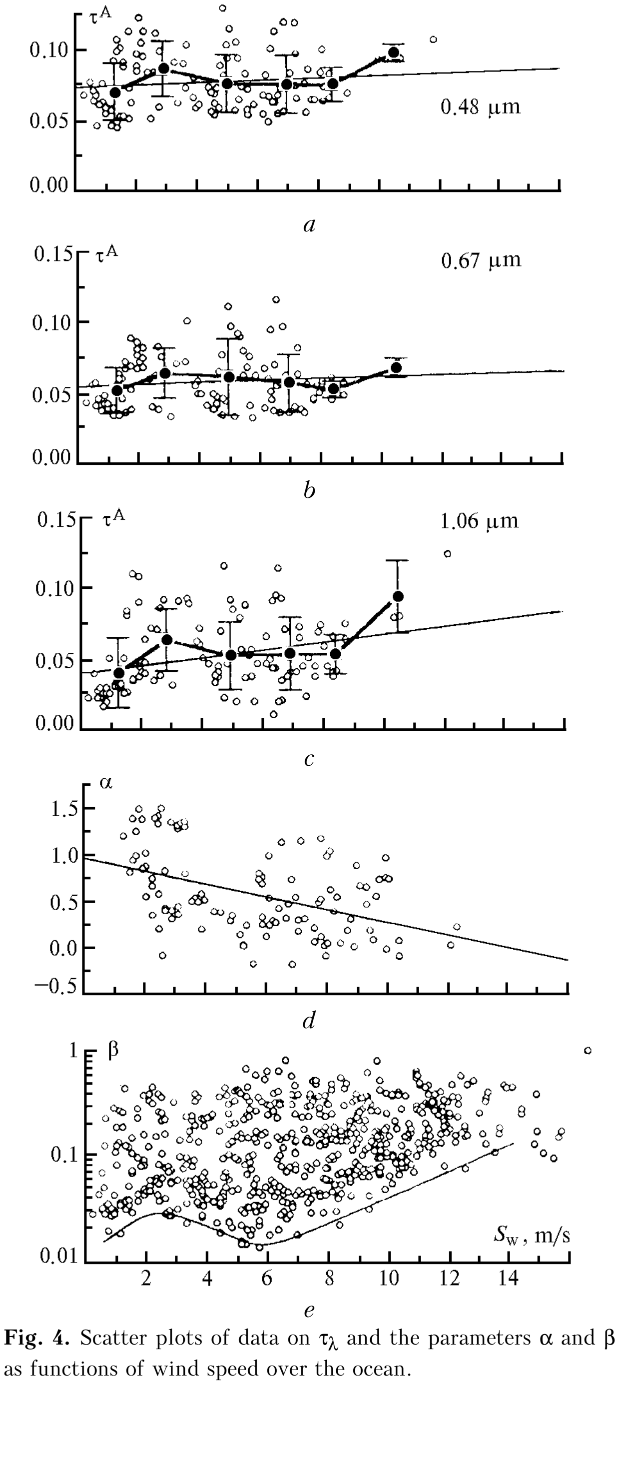

Fig. 4. Scatter plots of data on τ_λ and the parameters α and β as functions of wind speed over the ocean.

Effect of relative humidity

In analyzing the relations of τ_λ with humidity RH, the data obtained at S_w < 7 m/s were selected, that made it possible to decrease the effect of the wind factor. The dependences obtained (Fig. 5) are agree with the known facts and ideas[7,10–12,etc.] on the change of the aerosol scattering coefficient under variable humidity field: increase of extinction of radiation in the range RH > 75% and neutral behavior at RH less than 75%. Spectral peculiarities of τ_λ(RH) due to primary condensation growth of small particles are also well seen. The steepness of the ($\Delta\tau_\lambda$/RH) dependence decreases in the range of high humidity from 0.022 (for $\tau_{0.37}$) to the statistically insignificant value ~0.006 in the range of 1–4 μm. Transformation of the aerosol disperse composition at RH > 75% leads to change in the selectivity of the spectral behavior $\tau(\lambda)$: the parameter α increases, on the average, almost twice from 0.7 to 1.3.

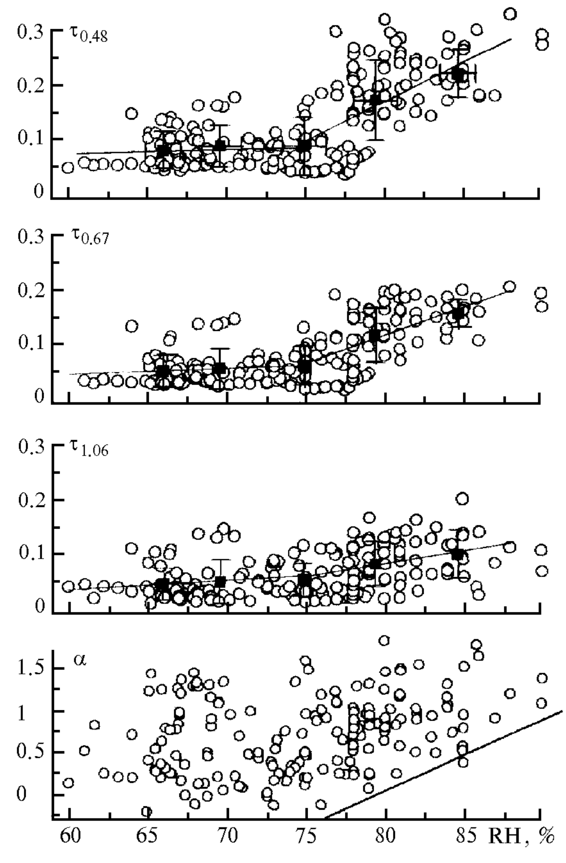

Fig. 5. Relation of τ_λ and α to relative humidity of air in the near-water layer.

One can trace the change in the selectivity of the spectral behavior at high humidity by the "method of minimum points" (Fig. 5). The role of small particles (accumulative fraction) at low humidity is insignificant. Therefore, the minimum AOT have neutral spectral dependence (α is near zero). Optical contribution of accumulative particles becomes more significant in the range RH > 75% under the effect of condensation growth, and the minimum possible values of AOT gradually increase.

Let us consider the results obtained in the equatorial zone in more details. The mean values of the wind speed here are, on the average, lower than in other regions (wind factor is weakened) and the variations of humidity occur in the range of high RH values.

Dominating effect of the condensation mechanism on the variations of τ_λ gives us the grounds for selecting empirical relationship, which makes it possible to calculate τ_λ from the known RH values. The sought model dependence was set to have the form

$$\tau_\lambda(\mathrm{RH}) = a_\lambda + b_\lambda\ \mathrm{RH}\ . \quad (2)$$

It was revealed in calculating the linear regression that the spectral behavior of the parameters a_λ and b_λ is well approximated by the relationship of the type:

$$a_\lambda(b_\lambda) = c + d\ \exp(\eta\lambda). \quad (3)$$

The values of the parameters c, d, and η are presented in Table 2, and the spectral dependence of a_λ and b_λ is illustrated in Fig. 6. Comparison of the measured values τ_λ with that obtained from calculations has shown that the absolute error of the model representation (2) and (3) is 0.026–0.055, and the relative error is about 30%.

Table 2. The parameters of the approximation dependence

Parameter	c	d	η
a_λ	−0.521	−4.68	−3.89
b_λ	0.00747	0.0633	−3.72

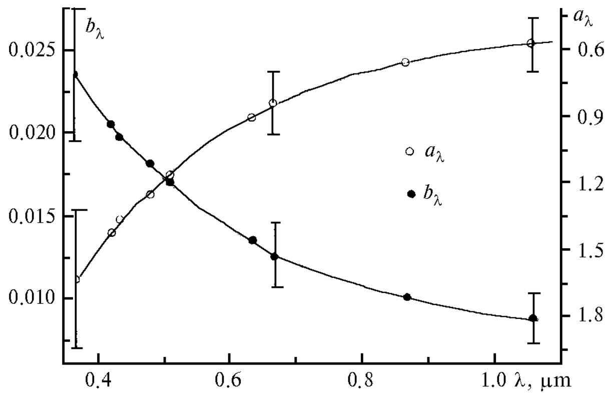

Fig. 6. Spectral dependence of the parameters of linear regression a_λ and b_λ.

Redistribution of the role of two factors, S_w and RH, determining the variability of AOT of the atmosphere in the considered regions can be illustrated by the example of correlation coefficients of τ_λ with differential number density of particles in the near-water layer N_r (the data were obtained by V.V. Pol'kin[13]). It is seen from the plots of the coefficients $R_{\tau,N}$ (Fig. 7) that significant correlation with the number density of large particles generated under the effect of wind factor is observed in the mid-latitudes (OO). The contrary situation is observed in the equatorial zone. Mutual variations of τ and N_r (and significant correlation) occur in the range of accumulative fraction under the action of relative humidity and condensation mechanism.

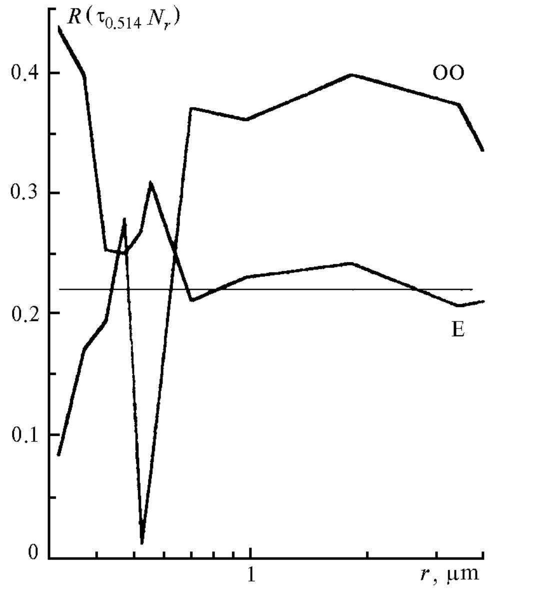

Fig. 7. Correlation coefficient $R_{\tau,N}$ as a function of the particle radius (thin line shows the level of significant correlation)

3. Variations of column density of water vapor and the height H_0

Taking into account the peculiarities in the spatial distribution of the column density of water vapor W and the height of the mixing layer $H_0 = W/a_0$ (a_0 is absolute humidity in the near-water layer),[1] we have divided the entire array of data into three latitudinal zones: mid-latitude, tropical, and equatorial or the inter-tropic convergence zone (ICZ). The mean values obtained from measurements in tropics are different, because they were obtained in different seasons and at different distance from continent: near Canary Islands in 1994, and in the central part of the ocean in 1995. So, the two sub-arrays are denoted, by analogy to the aerosol zones, as CI and TW. The characteristics of temporal variability of W and H_0 were compared with the data obtained in Tomsk in summer[14] for comparison with the continental conditions (see Table 3).

It follows from the statistical characteristics that the relative day-to-day variations (V_W) over the ocean are in the range 12–34%, and the mean values W during the period of measurements are close to the long-term data.[15] The equatorial zone is characterized by the minimum values V_W at maximum mean value of W, that is in a good agreement with the peculiarities of meteorological conditions in this region.[1,16] The height

H_0 is characterized by even higher stability: $V_H = 13–19\%$. That means, that the character of the profile $a(h)$ varies insignificantly at the change of air mass. Geographical variations of H_0 are more essential. The maximum values $H_0 = 2–3$ km are observed in the equatorial zone. These are almost two times lower in other regions. The minimum values $H_0 \approx 1$ km are characteristic of the tropical atmosphere near African coast. The height H_0 over continent takes an intermediate value between the trade wind and equatorial zones but is characterized by a stronger synoptic variability.

Thus, one can conclude based on the statistical data that the characteristics of the column density of water vapor are more stable than that of the AOT of the atmosphere on the scale of synoptic variations (variations of W are significantly weaker than those of τ). Taking into account the results from Refs.15–17 one can suppose that the contrary situation occurs in the range of seasonal variability: annual behavior of W is well pronounced (except for the equatorial zone), comparable with synoptic variations and is more significant than seasonal variation of AOT of the atmosphere.

For a conclusion, let us present the approximate estimates of the ratio of different types of variation of the AOT and CDWV of the atmosphere. One can take as a basis for making such an estimation the variation coefficients V, which are used for estimating relative variability of the random variations (synoptic variation of W and τ in this case). Let us suppose that seasonal behavior of W has a sinusoidal dependence, the variance of which is presented by the amplitude A in the form $\sigma^2 = A^2/2 = (W_{max} - W_{min})^2/8$. Then one can write for the variation coefficient of the seasonal variations the following formula:

$$V_W = \frac{(W_{max} - W_{min})}{2\sqrt{2}\,W} = \frac{(W_{max} - W_{min})}{\sqrt{2}\,(W_{max} + W_{min})}. \quad (4)$$

The data for the central part of the ocean[15] (50°N, 30°W) were used in calculating V_W. Seasonal variations of AOT of the atmosphere were estimated by the formula analogous to Eq. (4). Mean values τ for the ocean of the northern hemisphere in the period 1985 until 1989 reconstructed from satellite measurements (NOAA–9)[18] were used as the initial data. The comparison of the values of diurnal variation was also made using Eq. (4) based on the data on the mean diurnal behavior of the normalized values τ (see Fig. 3). At least, the mean values W and τ in the regions selected before (see the initial data in the zones of the total transmission[1]) were used for estimating the spatial variability.

The obtained estimates (Table 4) are conditional, but they provide for understanding the general idea of the role of different factors in variations of the characteristics determining the atmospheric transmission over the ocean. The most significant are synoptic and spatial variations of AOT of the atmosphere. Variations of the column density of water vapor of different types (spatial, seasonal, synoptic) are approximately two times lower and comparable with each other. Seasonal and diurnal behaviors of τ are insignificant in comparison with other variations. We have not revealed any significant regular component of the diurnal behavior of W over the ocean.

As to the relations between CDWV of the atmosphere and meteorological parameters, the maximum correlation is observed with the related characteristic – absolute humidity of air in the near-water layer. According to data of other authors[19,20,etc.] the relation is close to linear, and the correlation coefficients for different regions and scales of averaging lie in the range from 0.7 to 0.98. According to our investigations, the correlation coefficient is 0.76 for all regions outside of the ICZ. Besides, the correlation of W with wind speed is observed in the central oceanic region.[3] For example, the correlation coefficient of the daily mean values is 0.67. That means, that the increase of evaporation from surface at the increase of wind speed affects not only the humidity in the near-water layer, but also the CDWV of the atmosphere.

Table 3. Statistical characteristics of W (g/cm^2) and H_0 (km) in three latitudinal zones of the Atlantic and in Tomsk (July 1992–1997)

Latitudinal zone	Latitude	Mean		Min		Max		V	
		W	H_0	W	H_0	W	H_0	W	H_0
Equatorial	2°S – 7°N	5.36	2.47	4.18	1.95	6.25	3.11	0.12	0.15
Tropical (TW)	7° – 30°N	1.96	1.4	1.04	0.92	4.49	2.29	0.34	0.18
Tropical (CI)		1.23	1.02	0.94	0.73	1.83	1.39	0.17	0.19
Mid-latitude	30° – 65°N	1.64	1.11	0.84	0.90	2.35	1.57	0.27	0.13
Tomsk, summer	56°N	2.47	1.92	1.19	1.17	3.74	2.74	0.27	0.23

Table 4

Type of variation	Diurnal	Synoptic	Seasonal	Spatial
V_τ, %	5	70	7	53
V_W, %	–	27	20	33

Conclusion

One can generalize the new quantitative data and regularities revealed in investigating the temporal variability of the AOT and CDWV of the atmosphere as follows.

1. Statistical parameters have been determined of the day-to-day variability of $\tau(\lambda)$ and W in different regions of the ocean. Correspondence of the variations of τ in mid-latitudes to the lognormal distribution of probability density is shown.

2. Weaker component of the temporal variability – diurnal behavior of $\tau(\lambda)$ with the maximum before noon and amplitude 15–30% was revealed in some regions of the ocean. Another peculiarity of the diurnal variability is the increase of the selectivity of the spectral behavior $\tau(\lambda)$ in the evening.

3. Quantitative data and the character of relations of $\tau(\lambda)$ with the wind speed and relative humidity were determined. The dependence on wind is observed as non-monotonic increase of AOT mainly in the IR wavelength range resulting from the increase of number density of large particles. The dependence of $\tau(\lambda)$ on humidity is observed in the short-wave range what agrees with the assumption on the condensation growth of small particles. Redistribution of the role of meteorological parameters in different latitudinal zones can be presented as follows:

– variations of τ_λ in mid-latitudes are mainly caused by cyclonic transfer of fine continental aerosol and generation of coarse particles of marine aerosol under conditions of variable wind field;

– cyclic emissions of dust aerosol from Africa and generation of large marine particles play the main role in variations of τ in trade wind zone;

– variations of τ_λ in the equatorial zone are mainly caused by the condensation mechanism of transformation of the accumulative aerosol fraction in the variable humidity field.

4. Variability of the total transmission of the atmosphere over the ocean is determined, primarily, by spatial inhomogeneities and synoptic variations of the aerosol component (effect of continents, difference and change of air masses). The next, by significance, factors are spatial, seasonal, and synoptic variations of the column density of water vapor of the atmosphere that are comparable in strengths. The role of seasonal and diurnal variations of AOT is minimum though being statistically significant.

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