RESULTS OF INVESTIGATIONS OF THE AEROSOL OPTICAL THICKNESS AND THE WATER VAPOR COLUMN DENSITY IN THE ATMOSPHERE OF CENTRAL ATLANTIC

D.M. Kabanov and S.M. Sakerin

Institute of Atmospheric Optics,
Siberian Branch of the Russian Academy of Sciences, Tomsk
Received July 27, 1997

Spatiotemporal variability of spectral aerosol optical thicknesses and water vapor column densities in the atmosphere over the Atlantic Ocean was studied during the 39th voyage of the research vessel Akademik Mstislav Keldysh. The relative day-to-day variability of aerosol thicknesses is shown to be about 40% *and the diurnal variation has the noon maximum with the* 15–30% *amplitude relative to the morning and evening data. In spatial variability, the approach to continents is accompanied by the transformation of the spectral behavior of aerosol optical thicknesses: its selectivity increases due to the increase of small-size fraction content of the continental aerosol. As a consequence, its spectral dependence changes, namely, in the visible the differences between near-continental and central oceanic regions are manifested and in the* 1–4-μm *range the spectral behavior is close to neutral and changes less significantly.*

INTRODUCTION

In August–September 1996 as part of the program of the 39th voyage of the research vessel *Akademik Mstislav Keldysh* we made routine series of investigations into the spectral aerosol optical thickness (AOT) τ_λ^A and the water vapor column density (WVCD) w of the atmosphere over the ocean. We sailed from the Canada coast in southeastern direction to 29°N, where an experimental site for oceanographic investigations was organized and the greater part of the data was obtained. The atmospheric-climatic conditions at the experimental site were determined by its position on the periphery of the tropical zone and the northeastern trade wind (4000 km apart from the continent; the temperature was 25.5°–27.5°C; the humidity was 15.4–19.6 g/m^3; the wind velocity was 0.3–8.5 m/s). Ten days later the voyage was continued in the northeastern direction across the English Channel to Kaliningrad. The measurements began during a stop in Halifax and ended in the North Sea on September 18, 1996. The absence of continuous cloudiness enabled us to make a series of continuous daytime observations during 25 days.

In the investigations of the AOT and WVCD of the atmosphere, a multiwave solar photometer[1] with thirteen light filters tuned on atmospheric transparency windows was used. Unlike the previous works, the AOT was analyzed in a wider spectral range from 0.37 to 4.0 μm. To this end, a new procedure was developed to consider transmission functions of gas components.[2] Note that the AOT studies over the ocean in the spectral range more than 1 μm were practically not performed. The data with sufficient statistics are presented for this spectral range only in Refs. 3–5.

The atmospheric WVCD was determined on the basis of measurements of solar radiation transmission near the 0.94-μm absorption band by the procedure described in Refs. 6 and 7. More than three thousand spectra of vertical atmospheric transmission were measured over the expedition period. When analyzing the spectral AOT and WVCD of the atmosphere, the hourly and daytime average data were used. The general patterns of variations of the main characteristics studied over the voyage period are shown in Fig. 1.

ATMOSPHERIC OPTICAL THICKNESS

Preliminary analysis of the AOT measurement demonstrated the advisability of data separation into two groups. The atmospheres of central-oceanic (CO) region spaced at more than 500–1000 km from the coast and near the continents (NC) of the North America and Europe were analyzed separately. Note that the above-mentioned classification by the distance from dry land is tentative, because the aerosol composition and content in the air depend substantially on the type of air mass, trajectory of its motion, and air mass transformation. The atmospheric coastal zone could be classified more definitely (up to ~100 km), where additional periodical mechanism of mixing of continental and marine aerosols acts caused by the breeze circulation. However, the number of measurements in this zone was insufficient.

0235-6880/97/12 913-06 $02.00

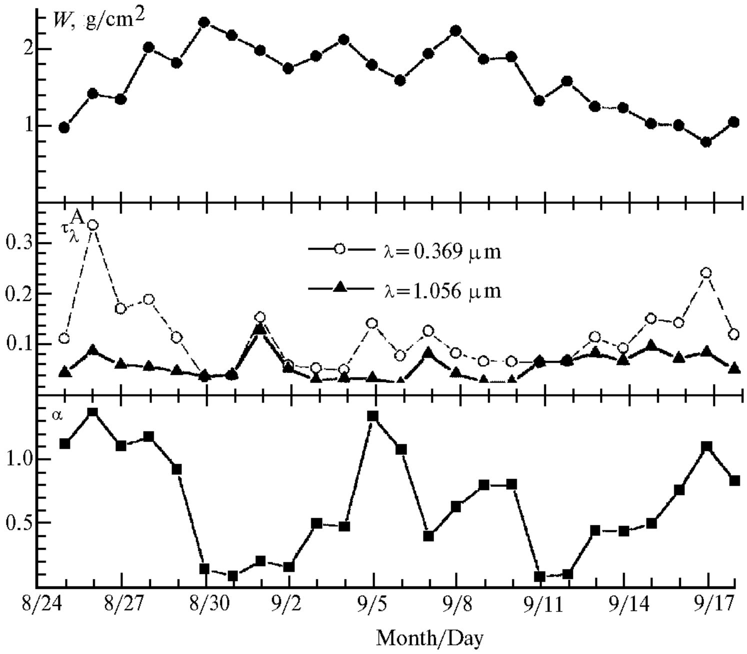

FIG. 1. Variations of WVCD and AOT of the atmosphere and the parameter α over a period of the 39th voyage.

TABLE I. Statistical characteristics of the daytime (d) and hourly average (h) values of τ_λ^A and α for different regions.

λ, nm	Combined data (d)		Regions near the continents (d)		Central oceanic region (d)		Central oceanic region (h)	
	$\bar{\tau}_\lambda^A$	σ_τ	$\bar{\tau}_\lambda^A$	σ_τ	$\bar{\tau}_\lambda^A$	σ_τ	$\bar{\tau}_\lambda^A$	σ_τ
369	0.118	0.068	0.186	0.077	0.084	0.036	0.083	0.037
408	0.116	0.062	0.178	0.062	0.085	0.036	0.083	0.038
423	0.111	0.062	0.176	0.067	0.080	0.034	0.079	0.036
438	0.110	0.059	0.173	0.062	0.081	0.033	0.079	0.035
484	0.100	0.050	0.153	0.054	0.077	0.031	0.076	0.032
513	0.089	0.045	0.136	0.051	0.071	0.030	0.070	0.031
558	0.083	0.041	0.125	0.044	0.069	0.030	0.069	0.032
637	0.075	0.034	0.109	0.032	0.062	0.029	0.060	0.029
671	0.069	0.031	0.099	0.027	0.058	0.027	0.056	0.028
871	0.065	0.026	0.080	0.019	0.060	0.028	0.057	0.030
1056	0.060	0.026	0.072	0.020	0.055	0.028	0.052	0.031
α	0.670	0.410	1.000	0.260	0.490	0.370	0.530	0.430

Statistical characteristics of spectral AOT for day-to-day variability are given in Table I. Presented here are the Angström parameters α characterizing the selectivity of spectral behavior of AOT in the range 0.37–1.06 μm

$$\tau_\lambda^A = \beta\lambda^{-\alpha}, \qquad (1)$$

where $\beta = \tau_1^A$ is the turbidity factor. The main difference between the two regions consists in higher atmospheric turbidity and more selective AOT near the continents. In the short-wavelength range of spectrum (near 0.37 μm) the average values of AOT differ more than twice and for 1.06 μm they differ by 25%. The standard deviation σ_τ of day-to-day variations of AOT duplicated the spectral behavior of τ_λ, and its relative variability was 40–50% for CO regions and 30–40% for NC regions. Our data are in good agreement with our previous results[8,9] obtained in 1989–1995 and, in most cases, with the results of investigation of other authors generalized in Refs. 10–11 (Table II). Exceptions are the data reported in Refs. 12 and 13, which for the central oceanic region are, in our opinion, overestimated. In the first paper, the measurements were carried out to the northwest of the Stream of Gibraltar; therefore, the effect of the continental aerosol removed from the Pyrenees peninsula and Africa could be observed. The results of Ref. 13 are doubtful, because the AOT values obtained at the centre of the Atlantic Ocean (to the north of 30°N) are in the rahge 0.11–0.61, actually, as in the zone of intense dust aerosol removal. Note that the summarized data given in Table II do not include the results of investigations made in 1991–1993 and 1982–1984 after the powerful Pinatubo and El Chichon eruptions.

TABLE II. Comparison of our data with the results of other investigations.

Observation period	Number of days (N)	$\bar{\tau}^{A}_{0.55}$	σ_τ	$\bar{\alpha}$	σ_α	Reference
		Central oceanic regions				
08–09.96	16	0.069	0.030	0.49	0.37	39th voyage
09–12.89	27	0.070	0.040	0.75	0.76	8
1979, 1986–1990	> 40	0.07–0.12	0.02–0.10	0.1–1	0.2–0.45	10, 11
1985, 1988	17	0.16–0.18	0.09	0.56–1	0.2–0.32	12, 13
		Regions near the continens				
08–09.96	9	0.125	0.044	1.0	0.26	39th voyage
1989–1995	45	0.06–0.18	0.03–0.16	0.72–0.84	–	8, 9
1979, 1982	–	0.097–0.172	0.028–0.052	0.2–1.4	0.02–0.2	11
1986, 1988–1990	56	0.04–0.21	0.02–0.09	0.56–1.17	0.17–0.38	10

The number of AOT measurements in the 2–4-μm wavelength range was smaller (13 + 6 measurement days); therefore, their statistical characteristics were calculated separately (Fig. 2).

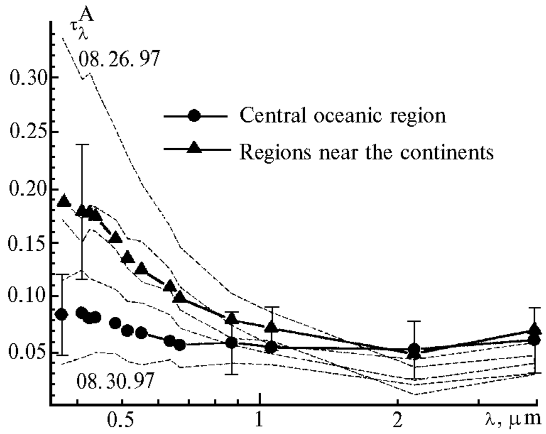

FIG. 2. Spectral variation of AOT and σ_τ for the two regions and transformation of τ^A_λ when we sailed from the continent from August 26 *to* 30.

In the short-wavelength region of spectrum (0.37–1.06 μm), the average AOT and σ_τ practically coincide with those for the combined data arrays (see Table I). As to the Angström parameter α (see Table I), for the central oceanic region it was 0.49 and near the continents its values were intermediate between the purely marine and continental ($\alpha = 1.3$) atmospheres. In this case, the approximation error of AOT spectral variation by Eq. (1) was, on average, 10%.

In the range more than 1 μm, the AOT power-law decay described by Eq. (1) is violated and, on average, a small increase in τ^A_λ is observed when its values are within the range from 0.05 to 0.07. Estimates of errors in using Eq. (1) for a wide spectral range indicated that they doubled. As for the AOT average values, it should be noted for comparison that Villevalde et al.[3] and Wolgin et al.[5] for the North Atlantic and wavelengths of 1.24 and 1.64 μm reported the results close to our data, that is, varying within 0.04–0.08.

Thus, from the results obtained it follows that the AOT values of the atmosphere of NC and CO regions in the range 1–4 μm are close and the main differences between them are manifested in the visible range. Transformation of the spectral behavior of τ^A_λ during several days of measurements, when we sailed from the continent deep into the ocean, is illustrated by dashed curves in Fig. 2.

The daytime variability of AOT over the ocean, as judged from the literature data available, was not analyzed previously. It is probable that this is due to insufficiently complete observation results and the assumption that less pronounced (as compared with the continent) diurnal variations of the meteorological parameters should not lead to noticeable variability of AOT. Really, consideration of diurnal variations of AOT (see the last column of Table I) practically does not increase the variance. This testifies that the daytime variations are insignificant. For individual realizations of $\tau^A_\lambda(t)$, no regularity is observed as well. Nevertheless, for the averaged data we have already noted[9] the diurnal variation of AOT with maximum at noon. The estimates of $\tau^A_\lambda(t)$ for CO region supported this fact (Fig. 3). Along with spectral peculiarities, the common regularity is the increase of τ^A_λ from morning till noon followed by its slow decrease in the evening. The daytime amplitude of AOT variation is, on average, 15–30%. The above-considered regularity of $\tau^A_\lambda(t)$ can be explained by the combined effect of the wind velocity and relative himidity, whose daytime variations are correlated, on the aerosol. An interesting consequence of spectral peculiarities of daytime behavior after noon is a well-defined change of the selectivity of τ_λ^A, namely, the normalized value of α increases in the evening by a factor of 2 to 2.5.

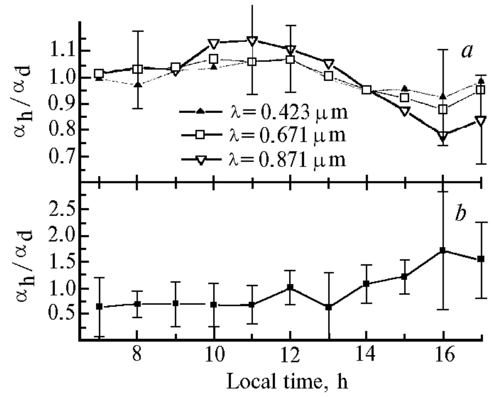

FIG. 3. Daytime variability of spectral components of AOT (a) and the Angström parameter (b).

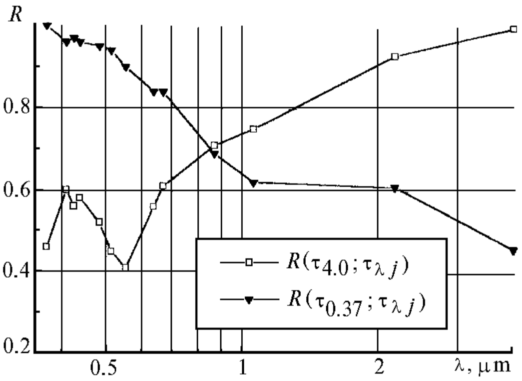

FIG. 4. Spectral variation of the correlation coefficients $R(\tau_{0.37}; \tau_{\lambda j})$ and $R(\tau_{4.0}; \tau_{\lambda j})$. *The level of significant correlation with a* 0.95 *confidence level is* 0.2.

Cross correlation of AOT in different spectral ranges is illustrated by Fig. 4. Among the peculiarities of spectral variation of the correlation coefficients $R(\tau_{\lambda i}; \tau_{\lambda j})$, it should be noted that the maxima at ~0.4 and 2 μm are superimposed on the monotonic decrease of $R(\tau_{\lambda i}; \tau_{\lambda j})$ as the wavelength difference $(\lambda_i - \lambda_j)$ increases. Such behavior of the cross-correlation coefficients may be attributed to a common physical nature of the two maxima in the aerosol particle size distribution over the ocean[14,15] centered at radii of about 0.4 and 1.5 μm, respectively. The availability of mutual response of two fractions to the change of environmental conditions (wind velocity, humidity, and so on) is manifested through the enhanced AOT correlation in the spectral ranges more sensitive to the above-indicated radii of aerosol particles.

WATER VAPOR COLUMN DENSITY (WVCD) OF THE ATMOSPHERE

Total change of the water vapor column density of the atmosphere during the voyage was 0.8–2.4 g/cm^2 (see Fig. 1) and was primarily connected with spatial variability when the vessel sailed between 29 and 54°N. The latitudinal dependence of WVCD was clearly defined (the correlation coefficient was 0.86) with a mean gradient of the W decrease of about 0.5 g/cm^2 per 10° of latitude (Fig. 5*a*). To reduce the effect of spatial component, the data obtained in the 29–40°N latitude belt were analyzed separately. Statistical characteristics of day-to-day variability of WVCD (Table III) indicated that the water vapor column density in the examined oceanic region over a period of observations varied insignificantly (V_W = 14%) and its value was less compared with a mean value for these latitudes[16] of about 2.6 g/cm^2. Higher amplitudes of variations of W observed for the combined data array were connected with the spatial inhomogeneities of the humidity field. The estimates of daily variability of WVCD pointed to the absence of regular component in the diurnal variation of W.

To study the effect of the meteorological parameters on the WVCD formation we analyzed the correlation with the water vapor pressure e and the wind velocity V in the atmospheric layer adjacent to the water surface. The WVCD dependence on the humidity for the combined data array was linear with a correlation coefficient of 0.92 (Fig. 5*b*). If only the data for the CO region are considered, the correlation $R(W; e)$ decreases down to 0.5.

The situation is just the opposite when the correlation between WVCD and the wind velocity is analyzed. Due to the strong effect of latitudinal dependence of WVCD, the correlation with the wind velocity for the combined data array was not observed. When we took into account the effect of the zonal component (that is, analyzed only the data for the CO region), the correlation $R(W; V)$ became significant and equal to 0.56 (Fig. 5*b*), that is, the increase of evaporation from the water surface with the increase of the wind velocity affects not only the humidity in the lower atmospheric layer, but also the column density of water vapor.

Such opposite tendencies of $R(W; e)$ and $R(W; V)$ when passing from the combined data array to the individual region seem strange at first glance. The reason is that WVCD and the water vapor pressure are of the same physical nature. They characterize the humidity in a local volume and in the atmospheric column, respectively. Therefore, for vertical profile of humidity analogous for all atmospheric conditions (on average, its exponential decrease), the correlation $R(W; e)$ increases for wider range of variations of the parameters that are compared. Widening spatiotemporal range of observations intensifies this effect. For a finite data sample, the correlation $R(W; e)$ deteriorates due to the local inhomogeneities and short-period humidity variations, which are less pronounced for the water vapor column density W.

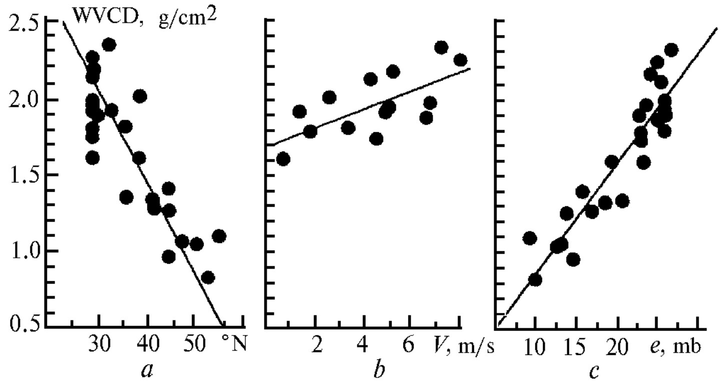

FIG. 5. Dependence of WVCD on the latitude, wind velocity, and water vapor pressure in the atmospheric layer adjacent to the water surface.

TABLE III. Statistical characteristics of day-to-day variability of WVCD.

Statistics of WVCD	$\bar{W}$	σ_W	V_W	min	max	Number of days (N)
Central oceanic region (29–40°N)	1.91	0.26	0.14	1.36	2.35	16
Combined data (29–54°N)	1.64	0.44	0.27	0.84	2.35	25

The WVCD dependence on the wind velocity is more complicated and masked because the variability of W is determined by many other factors (circulation processes, temperature regime, and so on). Therefore, a condition for the manifestation of the correlation $R(W; V)$ is the exception or reduction of the effect of other factors. In our case, this occurs when considering the data for a bounded region under approximately constant atmospheric conditions affecting the water vapor column density.

CONCLUSION

Our investigations of the spectral atmospheric transmission above the ocean enabled us to obtain the new data on the regularities of variability of two main (except for cloudiness) components – AOT and WVCD determining the radiation transfers – with sufficient statistics. Main conclusions can be summarized in the following form:

1. The oceanic atmosphere near the continents, even beyond the coastal zone, differs from that of the central oceanic regions by higher aerosol turbidity and selectivity of the AOT spectral variation ($\overline{\tau_{0.56}}$ equals to 0.125 and 0.069 and α equals to 1.0 and 0.49, respectively).

2. The main contribution to the variance of AOT variability is made by the synoptic processes (coefficient of day-to-day variations is about 40%) and in the daytime behavior the noon maximum with total amplitude of 15–30% is observed. The AOT decrease after noon is accompanied by the increase, on average, of its spectral selectivity.

3. In the spectral range more than 1 μm AOT of the investigated regions has quasineutral spectral behavior with average values within the range 0.05–0.07.

4. The WVCD variability of the atmosphere over the period of investigations was determined by the latitudinal dependence (about 0.5 g/cm^2 per 10° of latitude) and the amplitude of synoptic variations was small ($V_W \approx 14\%$) as compared with AOT.

REFERENCES

1. D.M. Kabanov, S.M. Sakerin, and S.A. Turchinovich, in: *Multiwave Solar Photometers for Investigating Directly Transmitted Radiation and Aerosol-Gaseous Composition of the Atmosphere. II. Regional Monitoring of the Atmosphere*, M.V. Kabanov, ed. (Publishing House Spektr, Tomsk, 1997), pp. 131–145.

2. D.M. Kabanov and S.M. Sakerin, Atmos. Oceanic Opt. **10**, No. 7, 540–545 (1997).

3. Yu.V. Villevalde, A.V. Smirnov, et al., J. Geophys. Res. **99**, 20, 983–20, 988 (1994).

4. F. Volz, J. Atmos. Sci. **27**, 1041–1046 (1970).

5. V.M. Wolgin, V.F. Radionov, and V. Leiterer, Z. Meteorol. **41**, 267–272 (1991).

6. D.M. Kabanov and S.M. Sakerin, Atmos. Oceanic Opt. **8**, No. 6, 442–446 (1995).

7. D.M. Kabanov and S.M. Sakerin, in: *Abstracts of Reports at the Fourth Symposium on Atmospheric and Oceanic Optics*, Tomsk (1997), pp. 140–141.

8. S.M. Sakerin, S.V. Afonin, T.A. Eremina, et al., Atmos. Oceanic Opt. **4**, No. 7, 504–510 (1991).
9. V.E. Zuev, D.M. Kabanov, and S.M. Sakerin, Proc. SPIE, Ocean Optics XIII (1997) (to be published).
10. A. Smirnov, O. Yershov, and Y. Villevalde, Proc. SPIE, Atmospheric Sensing and Modeling II **2582**, 203–214 (1995).
11. O.D. Barteneva, N.I. Nikitinskaya, et al., *Transparency of the Atmospheric Column in the Visible and Near-IR Spectral Ranges* (Gidrometeoizdat, Leningrad, 1991), 224 pp.
12. A. V. Smirnov, Yu. V. Villevalde, et al., J. Geophys. Res. **100**, No. D8, 16,639–16, 650 (1995).
13. P.J. Reddy and F.W. Kreiner, Global Biogeochem. Cycles, No. 4, 225–240 (1990).
14. S.M. Sakerin, D.M. Kabanov, and V.V. Pol'kin, Atmos. Oceanic Opt. **8**, No. 12, 981–988 (1995).
15. D.M. Kabanov, M.V. Panchenko, V.V. Pol'kin, and S.M. Sakerin, in: *Abstract of Reports at the Fourth Symposium on Atmospheric and Oceanic Optics*, Tomsk (1997), pp. 116–117.
16. S.E. Tuller, Monthly Weather Review **96**, No. 11, 785–797 (1968).