Regularities of the Spatial-Temporal Variability of the Aerosol Optical Thickness Over Atlantic Ocean

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Introduction

Measurements of the atmospheric transparency over the ocean were carried out episodically in separate regions due to the difficulty and high cost of marine experiments. Judging by the review (Smirnov et al. 1995), about 50 expeditions were performed since the end of 60-th, but a significant portion of the results was obtained only for several days of measurements or in a narrow spectral range (at one or two wavelengths). The total bulk of the data obtained in the most investigated part of the World Ocean-Atlantic is comparable with the two-year cycle of observation of one continental station, and the regularity of measurements (the ratio of the days of measurements to the total expedition duration) is $0.3 \div 0.6$. The most detailed investigations were carried out in 60-th to 80-th by two groups, Barteneva et al. (1991) and Shifrin et al. (1989). In 1989-1996, we performed the new cycle of investigations that allowed us to fill up a gap in the knowledge of the spectral transparency of the marine atmosphere. In particular, we have succeeded in dividing the aerosol turbidity of the atmosphere over the Atlantic into zones, determining the characteristics of synoptic and diurnal variations, and estimating the relations with meteorological parameters.

Characteristic of Investigations

The multiwave sun photometer, which was designed for ship-borne measurements, was used for measuring the spectral transparency of the atmosphere. Automation of the experiment and the technique developed for determining aerosol optical thickness (AOT) of the atmosphere (τ^A) provided the possibility of continuous measurements and obtaining reliable data (Kabanov and Sakerin 1997). The principal data was obtained in the wavelength range $0.37 \div 1.06$ μm; the error in determining τ^A was estimated as $0.005 \div 0.01$. The hourly mean and daily mean values were used for physical analysis. The bulk of the data obtained was sufficiently representative in the degree of regularity of measurements (~0.9), geographical area (from 10°S to 60°N), and amount (264 days of measurements, more than one thousand hourly mean values AOT of the atmosphere). The regions of marine expeditions are shown in Figure 1.

It is well known that the spatial distribution of τ^A over the ocean is very inhomogeneous, and the total range of variations reaches two orders of magnitude. There are differences not only in the τ^A values but also in the character of variations and the spectral behavior. It is unlikely to expect equal characteristics of atmospheric turbidity over the large area of ocean bordering the clear polar regions and the continents, which are the generators of different aerosols. The results shown in Figure 2 illustrate

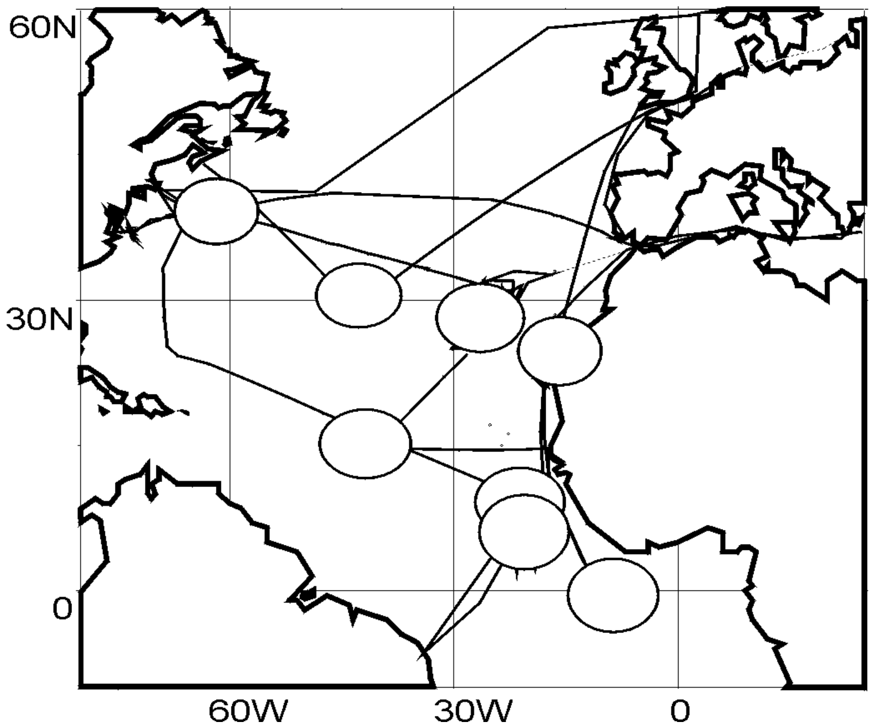

Figure 1. Routes of five scientific expeditions (sites of the most intensive observations of the transparency are marked by circles).

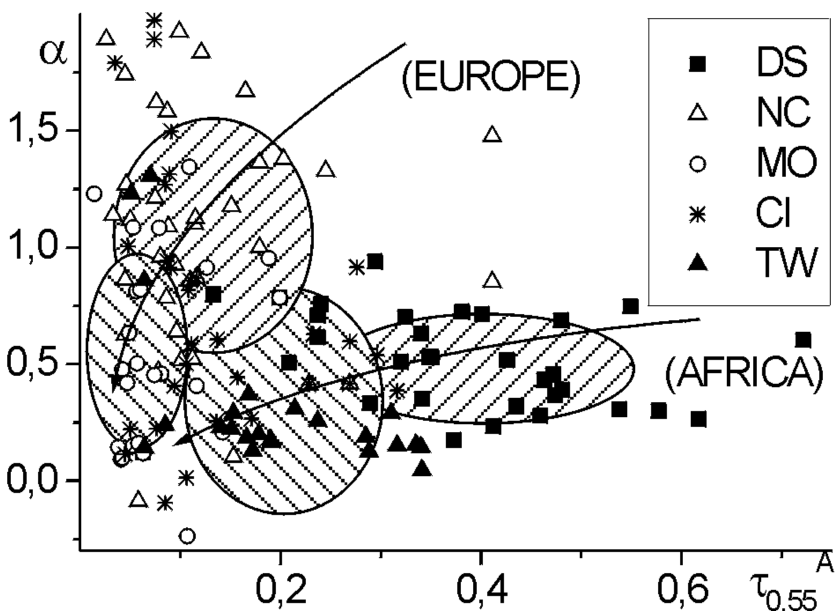

Figure 2. Illustration of correlation between $\tau^A_{0.55}$ and α over the ocean (the areas of location of the characteristics of the principal regions are marked by circles).

the range of variability of the τ^A and the Angstrom parameter α that characterizes the spectral dependence $\tau^A(\lambda) = \beta \cdot \lambda^{-\alpha}$. At such a variety of values $\tau^A(\lambda)$, the necessity of dividing into zones is obvious.

One can select two groups of classification in the available methods of dividing into zones, the "sign" and "genetic" (e. g., Samoilenko 1983). An analogous approach should be undertaken for aerosol as the element of the climatic system. The model of the AOT distribution over the ocean is presented below on the basis of two classifications (see Figure 3).

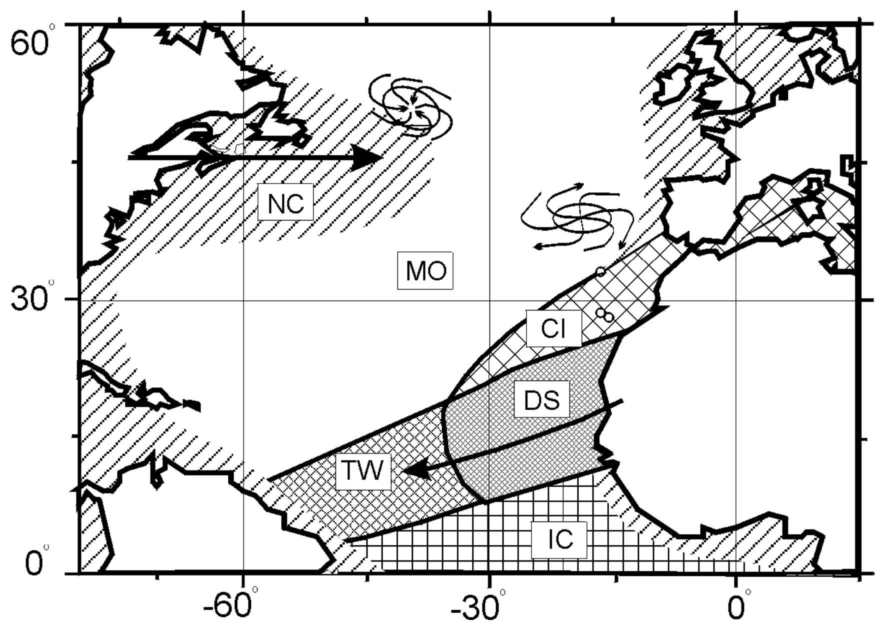

Figure 3. Location of the "aerosol areas" as result of the genetic classification.

Genetic Dividing Into Zones

The advantage of the genetic classification is that it is based on revealing the principal processes or factors of formation of the physical fields of the atmosphere, and, in conditions of deficiency of data, allows the expansion of properties or features to the less studied regions. Atmospheric circulation and peculiarities of the underlying surface play an important role in many classifications. The majority of the aerosol sources is also determined by the type of the underlying surface, and the spatial distribution depends on the motion of air masses. The results of optical and microphysical investigations of aerosol over the ocean show that, in spite of the homogeneity of the underlying surface (the source of marine aerosol), variations of the characteristics are significant and are related to the intrusion of different types of aerosol from the continent. That means, the spatial inhomogeneities of AOT over the ocean are caused by the peculiarities of the continental aerosol transfer. Thus, classification of the aerosol fields should take into account the combined effect of two factors: 1) the sources (types) of continental aerosol prevalent in each latitude zone; and 2) the prevalent circulation (transfer) of air masses that determine the direction and distance of aerosol intrusion into the ocean atmosphere.

The more fine aerosol relative to the marine one is prevalent in the mid-latitude atmosphere (e.g., Kondrat'ev et al. 1983). Breeze circulation occurring near the continent boundary favors regular mixing of continental aerosol with the marine one and enriches it with fine fraction. Then the littoral zone (LT) up to 10 km can be selected. Further transport of continental aerosol into the ocean is provided by the cyclonic activity developed in these latitudes and by westward transfer. Then one can select the near-to-continent zone (NC) and the mid-ocean (MO) zone. The random character of the transfer does not allow one to exactly determine the length of the NC zone deep into the ocean. So let us use the semi-qualitative estimate based on the cyclonic formation scale. Taking into account the different (between latitudes) intensity of the aerosol sources and transfer, one can estimate the length as 100 ÷ 200 km at the boundaries of the midlatitude zone up to ~1000 km in the middle part.

The next zone is located in the tropics. The stable trade-wind with emission of the coarse dust aerosol from the Sahara desert prevails here. It is known from the results of continental and ship-borne measurements (e.g., Barteneva et al. 1991; Kondrat'ev et al. 1983) that the content of Sahara aerosol in the atmosphere gradually decreases but is observed right up to the America coast. It is favored by the inversion and relatively high velocity of aerosol transfer in the layer 2 km to 4 km characteristic of the trade wind zone (TW). The part of the TW most filled by aerosol (to the east from 35°W) is well known as the "Dark Sea" (DS). The intertropical convergence zone (IC) is the natural southern boundary of DS and TW. The mean position of the northern boundary of IC is near 7°N. The IC axis, as the trade-wind axis, is deflected to the NE-SW direction.

The mixed zones, Mediterranean Sea (MS) and Canary Islands (CI), can be selected at the north of the DS areas. Depending on the season, they are alternatively affected by emission of coarse aerosol from Africa and fine aerosol from Europe. The change of circulation is caused here by the neighborhood of the trade wind with the Azor anticyclone, one of the main action centers.

At least, the equatorial, or IC zone is located in the south part. The following conditions are characteristics of this zone: weak wind, developed convection, filtration of continental aerosol by cloudiness, and frequent precipitation. The complex of meteorological conditions forms here the character of the aerosol turbidity different from the neighbor TW.

"Sign" Dividing Into Zones

Selection of zones in the "sign" classification is performed based on the numerical values of some measured parameters of the atmosphere. As for AOT, it was shown earlier (Volgin et al. 1988), that it is enough to use two "signs," $\tau^A_{0.55}$ and the parameter α, for determining the typical zones. The procedure of dividing into zones included two stages. First, the maps with isolines of the characteristics under study were constructed (Figure 4). The values τ^A and α were averaged with the step of 5° over longitude and latitude, and the absent points were reconstructed by inter- and extrapolation of the data. The distributions obtained are in good agreements with the aforementioned features. For example, when approaching the continent in the midlatitudes, the value of turbidity increases, as well as the selectivity of the spectral dependence. The parameter α near the coast tends to be the values characteristic of the continent (1.3). It does not occur in the trade wind zone. Here, when approaching Africa, τ^A increases, the parameter α remains small (coarse aerosol is prevalent).

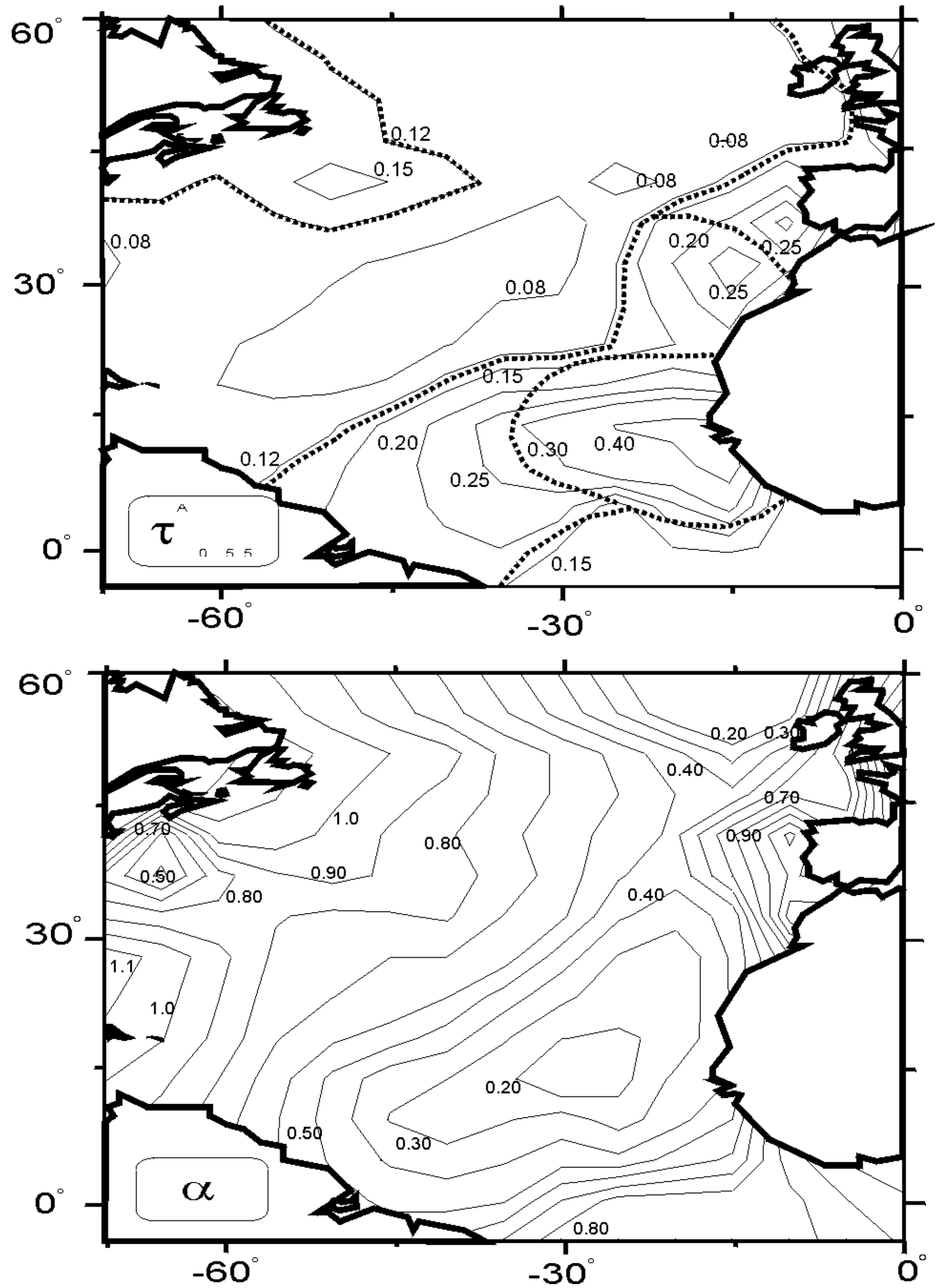

Figure 4. Isolines of the spatial distribution of $\tau^A_{0.55}$, α and the results of "sign" dividing into zones (dotted line).

The second stage is determination of the zone boundaries. The criterion for selection of the MO zone is finding the values τ^A and α obtained in the central midlatitude part of the ocean within the interval "mean value $\pm\,\sigma$." Actually, the zones MO and NC are separated by the values τ^A. Criteria for determination of the boundaries in tropical and equatorial zones are the qualitative differences in the character of the spatial distribution of τ^A and α. The boundaries are drawn in the areas of the minimum gradient or the minimum values of the parameters. As result, the "aerosol provinces" are determined, sufficiently close to the data of genetic dividing into zones. Let us also note that the samples obtained in different zones are statistically divisible by at least one parameter with confidence probability not less than 0.9.

Statistics of the daily mean values τ^A and α for the elected areas are presented in Table 1. The data of Volgin et al. (1988) are presented for the LT zone near coasts due to the poor bulk of our data. As a whole, the total range of variability of the AOT characteristics over ocean is comparable with the results obtained in continental conditions ($\tau = 0{,}01 \div 0{,}7$; $\alpha = -0{,}24 \div 2{,}42$).

Table 1. Statistical characteristics of $\tau^A_{0.55}$, α.

Regions (N - number of days)	$\bar{\tau}$	σ_τ	V_τ	$\bar{\alpha}$	σ_α	V_α
MO (26)	0,076	0,047	0,62	0,59	0,47	0,80
NC (25)	0,149	0,097	0,65	1,11	0,50	0,45
LT *	0,20	0,10	0,49	0,90	0,43	0,48
CI (31)	0,135	0,081	0,60	0,73	0,54	0,74
MS (6)	0,072	0,048	0,66	1,0	0,60	0.60
DS (32)	0,381	0,143	0,38	0,50	0,22	0,45
TW (23)	0,196	0,095	0,49	0,31	0,34	1,11
IC (25)	0,141	0,073	0,52	0,85	0,55	0,65

Temporal Variability

The classification makes it possible to separate the spatial inhomogeneities of AOT and temporal variability. The statistical characteristics presented in Table 1 give an indication of the most strong component, inter-day variations caused by the change of air mass. Spatial distribution of σ_τ approximately repeats the distribution of the mean values τ^A. The values σ_τ are comparable to the continental ones (Kabanov and Sakerin 1996), and the relative variations $\left(V_\tau = \sigma_\tau / \bar{\tau}\right)$ are greater due to less turbidity of the marine atmosphere. The maximum coefficient of variations V_τ = 60-80% are characteristic of the areas MO, NC, and CI, which are situated in the active cyclonic zone. Variations of parameter α (especially relative ones V_α) are also more significant than in continental conditions $0{,}45 \div 1{,}11$ compared to $0{,}3 \div 0{,}4$ in Tomsk.

Clarification of the distribution law is necessary for the more full description of the AOT variations. For this purpose, we constructed the histograms (Figure 5) of the hourly mean values $\tau^A_{0,48}$ and compared them with a number of theoretical laws of the probability density distribution. The histograms are divided into two types: single-mode for the MO, NC, and DS zones and spread for the IC, CI, and TW zones. The second mode is weakly seen in the histograms of the second group due to the episodic intrusions of dusted air from the DS side. Only three first zones were taken for the subsequent analysis.

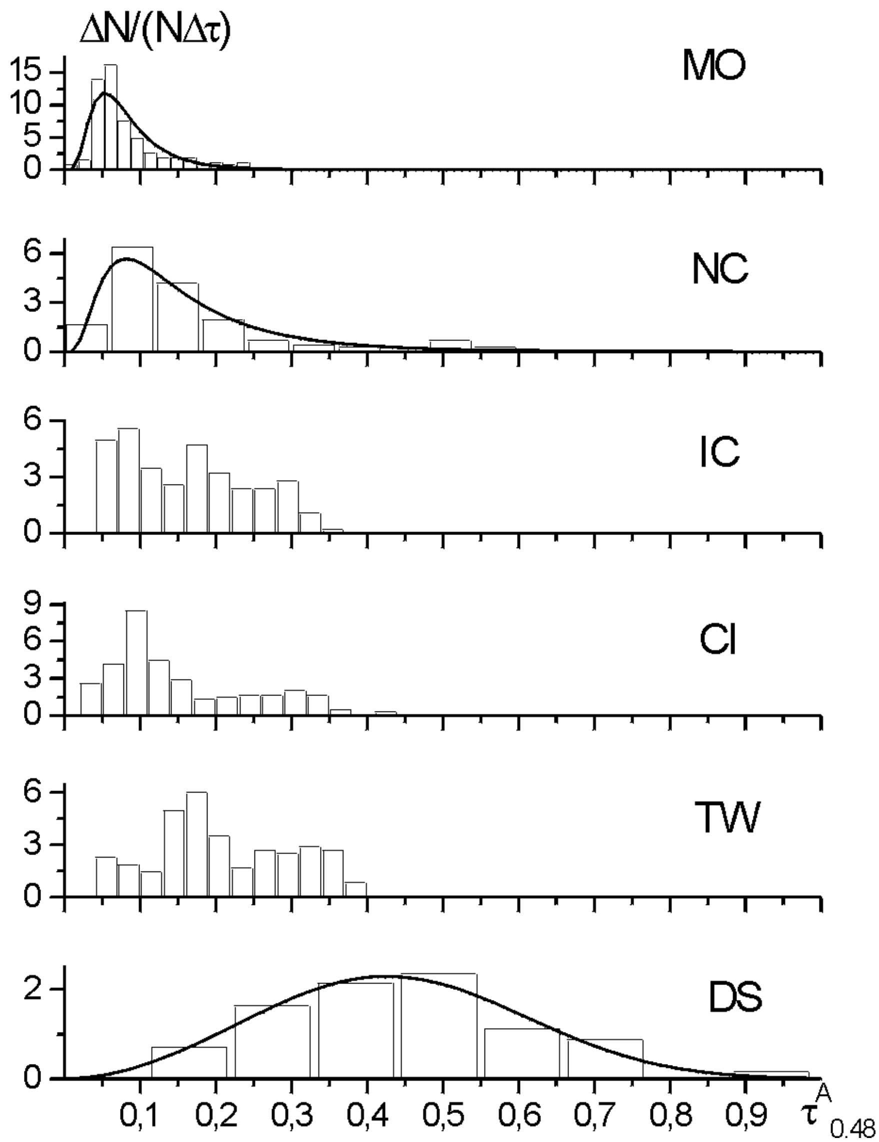

Figure 5. Histograms of τ^A and the theoretical distribution laws.

It is seen in their statistics (Table 2) that the distributions are extended to the side of greater turbidities ($\gamma_1 > 0$) and have a sharper top than the normal distribution ($\gamma_2 > 0$). The absolute values of the coefficients γ_1 and γ_2 significantly exceed their root mean square (rms) errors.

Table 2. Statistic of $\tau^A{}_{0,48}$ (γ_1, γ_2 - coefficients of skewness and kurtosis).

	$\bar{\tau}$	σ_τ	V_τ	Min	Max	Median	γ_1	γ_2	N
MO	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
DS	0,45	0,18	0,40	0,07	1,26	0,44	0,81	1,83	179

The parameters of theoretical distributions were selected by the momentum method using the sample initial and central moments of the respective order. Estimation by the χ^2 criterion shows that the lognormal law is the best approximation of empirical histograms for the MO and NC zones. Lognormal distribution is the only acceptable law for the MO zone with confidence probability 0.95. Weibull distribution is better for the DS zone (probably due to the insufficient duration of measurements for characteristic variation of AOT in this zone).

We have succeeded in revealing the weaker component (diurnal change) of AOT in some areas. Judging by the literature data, the diurnal change was not studied at all because of the absence of regular observations, small (in comparison with the continent) amplitude of diurnal variations of meteorological characteristics, and the relatively big influence of synoptic variations. As before (Panchenko et al. 1999), to select the diurnal change, we should normalize AOT to the daily mean values and average the data for each hour of measurements. As a result, the mean diurnal change was determined for the ocean regions with weak synoptic influence (Figure 6). Different from the continental atmosphere (Panchenko et al. 1999), there was a slow increase of AOT, which finished just before noon, and then followed with a longer and stronger decrease. The mean diurnal amplitude of the AOT changes were 15% to 30%. Besides, there were spectral differences in the diurnal change of $\tau^A(t)$, which determined almost twice the increase of the parameter α.

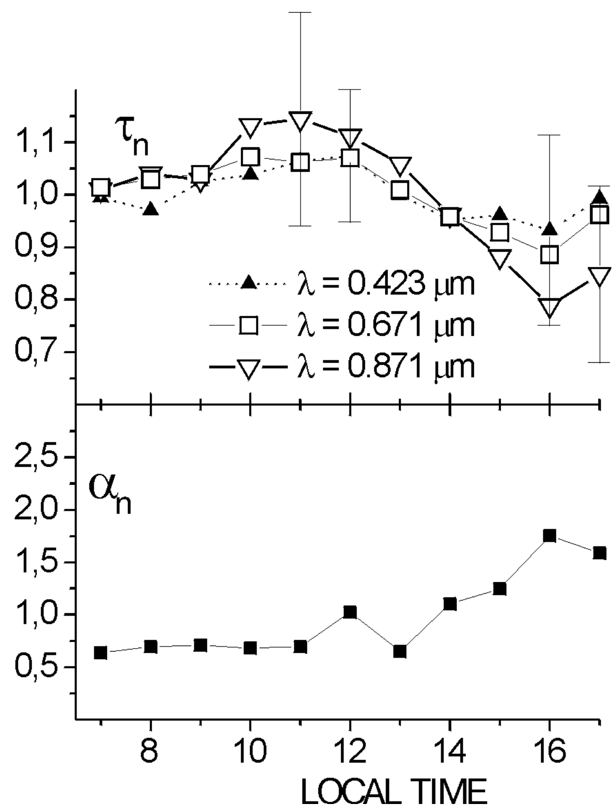

Figure 6. Illustration of the diurnal change of $\tau^A(\lambda)$ and α in far ocean area.

One can explain the considered peculiarities of $\tau^A(t)$ and $\alpha(t)$ by the combined effect of wind velocity and humidity on the marine aerosol, which gave similar diurnal variations. In particular, at the decrease of wind velocity in the evening, the generation of marine aerosol also decreased, coarse particles settled quicker, and selectivity of the spectral behavior increased.

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