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Variability of aerosol extinction of solar radiation in Antarctica

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Abstract: Temporal variations of the aerosol optical depth and transmission coefficient of the atmosphere are considered using data from Mirny Observatory, Antarctica. Year-to-year variability of these parameters is determined mainly by stratospheric aerosol pollution due to volcanic activity. A considerable increase of atmospheric turbidity has been observed since the end of September 1991. This phenomenon seems to be associated with the Mount Pinatubo volcanic eruption.

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Introduction

Studies of atmospheric aerosol are of interest because of its considerable influence on radiative processes. Since the late 1970s a great number of observations of the aerosol extinction of solar radiation in various regions of the Earth have been made (Bodhaine 1983, Kondratyev 1978, Kondratyev & Binenko 1981, Bodhaine 1989, Sakunov *et al.* 1986). Such measurements in Antarctica, which is not directly affected by man-made aerosol, have revealed the main features of the variability of the background aerosol and atmospheric optical characteristics on a range of time scales (ranging from time scales of less than one day to interannual ones). They have, in particular, corroborated the episodic data obtained earlier (Liljequist 1957, Kuhn 1972, Sakunov 1975) showing an annual cycle of aerosol extinction with a summer (December–January) maximum in Antarctica. Information on the variation of the optical state of the atmosphere, covering a few decades, may be obtained from actinometric measurements. A detailed analysis of observed time series of direct solar radiation at the Antarctic stations Novolazarevskaya (70°46'S, 11°50'E), Molodezhnaya (67°40'S, 45°50'E), Mirny (66°33'S, 93°01'E) and Vostok (78°28'S, 106°48'E) from the beginning of observations to 1975 is given in the monograph of M.S. Marsunova (1980). Direct measurements of aerosol optical depth in the atmosphere, started in 1979 at Mirny Observatory, have provided an opportunity to compare them with the results of actinometric measurements.

Methods

Transmission coefficient

The following characteristic of the optical state of the atmosphere is used to interpret actinometric observations from the Russian network of stations:

$$P_2 = \sqrt{S_{M,2}/S_0} \qquad (1)$$

P_2 is the coefficient of the total transparency (transmission coefficient), $S_{M,2}$ is the intensity of the direct solar radiation at the mean distance between the Sun and the Earth, incident on a surface normal to the solar ray direction and attenuated by two optical air masses (m = 2), and S_0 is the solar constant. The correction of the measured radiative intensity $S(h)$ (h is the Sun elevation) to allow for the variation in the distance to the Sun is made by standard methods.

The direct solar radiation $S(h)$ was measured by the thermoelectric actinometer M-3 which is an analogue of the Eppley actinometer. All the instruments were calibrated in St. Petersburg before and after the study. Additionally, the actinometers were verified at least once a month at the stations using a standard actinometer, which was annually calibrated against the Ångström pyrheliometer.

The accuracy of the measurement of the flux of direct solar radiation is ± 10 Wm^{-2}. The values of $S(h)$ are reduced to a Sun elevation h = 30° (i.e. m = 2) by Sivkov's method (Sivkov 1968) using empirical tables. This method eliminates the Forbes effect (i.e. an apparent increase of transmittance at large air masses) over a broad range of atmospheric turbidity changes. The empirical relation between $S(h)$ and $S(30°)$ used in this method is

$$S(30^\circ) = 1367 \left[\frac{S(h)}{1367}\right]^{1.42\sin h + 0.29} \qquad (2)$$

(Evnevich & Savikovsky 1989) where $S(h)$ is in Wm^{-2}. The error in computing $S(30°)$ using (2) averages at 16 Wm^{-2}. P_2 is calculated using

$$P_2 = \left[\frac{S(h)}{1367}\right]^{(\sin h + 0.204)/1.41} \qquad (3)$$

To compute P_2 the measured data $S(h)$ for h >10° and with no clouds closer than 20° to the solar disc are used. The variability of the daily average of P_2, usually estimated from five to seven separate measurements, does not exceed 2%. The determination of the transmission coefficient P_2 is similar to a procedure for estimation of a so-called 'apparent transmission' (Bodhaine *et al.* 1989). The value of P_2 is determined from values of direct solar radiation in a wide range of wavelengths from 0.4 μm to 4 μm and it therefore depends on extinction due to Rayleigh scattering, Mie scattering and selective absorptions by various

Table I. Spectral maxima and halfwidths (in brackets) of interference filters (nm).

	Channels								
Periods	1	2	3	4	5	6	7	8	9
January 1983– March 1986	406 (13)	458 (12)	504 (9)	551 (11)	682 (10)	744 (9)	798 (10)	995 (20)	-
January 1987– March 1987	413 (8)	442 (15)	454 (12)	497 (7)	560 (7)	666 (11)	771 (10)	-	
December 1987– April 1988	355 (17)	407 (14)	460 (13)	505 (12)	552 (11)	680 (9)	750 (7)	797 (9)	996 (18)
January 1989– December 1989	384	413	455	498	560	667	797	1023	-
January 91– present	389	413	483	587	649	783	858	1037	-

Table II. Mean monthly values of $\beta_{0.5}$, $\beta_{1.0}$ and α.

Year	Parameter	Sep	Oct	Nov	Dec	Jan	Feb	Mar
				Molodezhnaya				
1977	$\beta_{0.5}$	-	-	-	0.06	0.06	0.05	0.04
1978	$\beta_{1.0}$	-	-	-	0.02	0.01	0.01	0.01
	α	-	-	-	2.1	2.0	2.0	1.7
				Mirny				
1979–	$\beta_{0.5}$	0.03	0.04	0.05	0.05	0.05	-	-
1980	$\beta_{1.0}$	0.01	0.02	0.02	0.02	0.02	-	-
	α	1.5	1.5	1.7	1.5	1.4	-	-
1981	$\beta_{0.5}$	-	-	-	-	0.04	0.04	0.04
	$\beta_{1.0}$	-	-	-	-	0.02	0.02	0.02
	α	-	-	-	-	1.2	1.4	1.2
1981–	$\beta_{0.5}$	0.04	0.05	0.06	0.07	0.08	0.07	0.05
1982	$\beta_{1.0}$	0.02	0.02	0.03	0.03	0.03	0.02	0.02
	α	1.3	1.3	1.2	1.2	1.4	1.6	1.8
1982–	$\beta_{0.5}$	0.06	0.06	0.10	0.12	0.13	0.12	0.11
1983	$\beta_{1.0}$	0.02	0.02	0.03	0.04	0.04	0.05	0.04
	α	1.6	1.8	1.8	1.6	1.6	1.5	1.4
1983–	$\beta_{0.5}$	-	0.07	0.07	0.08	0.08	0.07	0.06
1984	$\beta_{1.0}$	-	0.03	0.02	0.03	0.02	0.02	0.02
	α	-	1.4	1.6	1.6	1.7	1.8	1.3
1984–	$\beta_{0.5}$	0.05	0.06	0.06	0.06	0.07	0.06	0.04
1985	$\beta_{1.0}$	0.03	0.03	0.03	0.03	0.03	0.03	0.02
	α	1.0	0.9	0.8	0.9	1.0	0.8	0.8
1985–	$\beta_{0.5}$	0.04	0.04	0.03	0.04	0.05	0.04	0.05
1986	$\beta_{1.0}$	0.01	0.01	0.01	0.01	0.01	0.01	0.01
	α	1.3	1.6	-	1.7	2.3	2.5	2.2
1986–	$\beta_{0.5}$	0.04	0.04	0.04	0.04	0.04	-	-
1987	$\beta_{1.0}$	0.01	0.01	0.01	0.01	0.01	-	-
	α	2.0	2.2	2.5	2.5	2.6	-	-

atmospheric gases (H_2O, O_2, O_3, CO_2, NO_2). Major factors influencing P_2 variations are changes in atmospheric water vapour and aerosol content; contributions of other gases are considered to be negligible.

Turbidity parameters

Three parameters characteristic of the atmospheric aerosol turbidity were estimated:

1) aerosol optical depth at wavelength $\lambda = 0.5\ \mu m$, $\beta_{0.5}$,

Table III. Mean monthly values of the transmission coefficient at Mirny Observatory.

$(P_2 \times 10^3)$

	Month								
Year	1	2	3	4	5	9	10	11	12
1956				814		816	826	810	806
1957	811	834	860	842		852	845	846	834
1958	836	837	839	851		809	809	829	827
1959	830	837	839	850		820	820	830	820
1960	830	837	839	848		820	830	830	820
1961	830	837	839	847		833	835	832	810
1962	821	838	830	848		834	836	831	829
1963	839	838	838	851		822	826	823	729
1964	714	706	719	677		744	758	740	738
1965	749	782	778	804		822	820	811	816
1966	823	818	829	833		839	841	833	830
1967	824	826	827	836		836	838	836	828
1968	803	812	826	837		833	828	814	820
1969	823	830	842	848		842	842	838	820
1970	823	829	829	850		843	848	827	825
1971	830	839	856	843		864	871	842	843
1972	841	838	838	837	858	851	850	839	829
1973	830	840	840	849		847	838	835	834
1974	835	838	844	856		856	843	851	836
1975	823	836	837	853	853	853	843	837	837
1976	834	837	846	850		842	835	838	832
1977	821	837	818	823	845	845	841	843	827
1978	833	834	837	853	856	848	842	840	828
1979	837	840	840	847		850	847	848	841
1980	828	829	838	848		848	835	836	825
1981	832	826	837	844		850	840	840	828
1982	830	827	838	853		844	843	814	794
1983	777	791	799			822	830	819	805
1984	808	806	819			839	839	831	826
1985	819	827	838	844		850	847	848	841
1986	836	832	835	859		854	851	841	835
1987	835	850	855	852		859	850	848	831
1988	831	837	832	846		848	832	837	829
1989	830	830	835	843	857	850	840	836	827
1990	838	835	858	860	868	862	843	849	850
1991	846	844	848	861	861	825	766	730	741
1992	759	749				792	799	783	781
1993	763	761	794	814		811			

2) Ångström law exponent, α

3) Ångström turbidity factor, $\beta_{1.0}$ (i.e. aerosol optical depth at the wavelength 1.0 μm).

The Russian background network is equipped with thermoelectric actinometers with a set of glass bandpass filters with their short-wavelength cutoffs at 0.38, 0.47, 0.53, 0.63 and 0.71 μm. The measurement and data-processing techniques are described by Rusina (1977). As a rule, 7–10 measurements per day were made. The error in a single value of $\beta_{0.5}$ is within 25–30% and that of $\beta_{1.0}$ reaches 50%. The standard deviation of individual values divided by a daily mean (daily variability) is within 8–10%. The standard deviation of the daily mean values divided by the monthly mean (monthly variability) is within 25–30%.

Aerosol optical depth

The measurement of the spectral aerosol optical depth, τ_λ, were made using a sun photometer with narrow-band interference filters. The values of τ (λ) are then given by

$$\tau(\lambda) = 1/m \ln\left[\frac{N_{0\lambda}}{N_\lambda K}\right] - \tau_{RE}(\lambda) - \tau_{abs}(\lambda) \qquad (4)$$

where m is the optical air mass, $N_{0\lambda}$ is the extra-terrestrial irradiance, N_λ is the measured value of irradiance, K is the Sun to Earth distance factor, τ_{RE} is the pressure-corrected optical depth due to Rayleigh scattering and τ_{abs} is the optical depth due to selective absorption by atmospheric gases. The extraterrestrial irradiance is obtained by the Langley plot method. The value of $\tau_{RE}(\lambda)$ was calculated using

$$\tau_{RE}(\lambda) = 0.00879\, \lambda^{-4.09} \qquad (5)$$

The ozone absorption in the Chappuis band (450–750 nm) has been taken into account (Radionov & Marshunova 1992). Measurements of total ozone from Mirny were used for this purpose. The daily average of τ (λ), calculated from series of 10–12 measurements, can be retrieved with an accuracy of not worse than ± 0.01. Spectral maxima of the interference filters used at Mirny Observatory are listed in Table I.

During the period 1983–1988 the spectral aerosol optical depths were measured by a filter sun photometer of the same design as the Linke-Feussner panzeractinometer (Sonntag 1970). From January 1989 the data were collected using the sun photometer ABAS (Leiterer & Weller 1988).

Results

The values of $\beta_{0.5}$, $\beta_{1.0}$, $\tau(0.5)$, $\tau(1.0)$ and P_2 are now considered. The spectral dependence of τ (λ) was discussed earlier by Sakunov *et al.* (1986), Leiterer & Sakunov (1989) and Sakunov *et al.* (1990).

The mean monthly values of $\beta_{0.5}$, $\beta_{1.0}$ and α at Mirny Observatory are listed in Table II. The mean monthly values of P_2 are given in Table III and mean monthly values of τ (1.0) are listed in Table IV. The mean monthly values of P_2 and $\beta_{0.5}$ from 1980–1987 at Mirny (Fig. 1) demonstrate that the lower transmission coefficient in the summer is accompanied by higher aerosol extinction.

The mean monthly values of total water content, w, and their standard deviations, Δw, during 1957–1980 at Mirny Observatory are given in Table V.

Thus the minimum of total transparency in the summer months can be attributed both to the increased absorption of solar radiation by water vapour and to aerosol extinction.

Rusina (1977) has found that results obtained by wide- and narrow-band techniques coincide within experimental error for the range $\beta_{0.5}$ = 0.1–0.5 and the range $\beta_{1.0}$ = 0.05–0.22. In the Antarctic, values of these parameters are usually considerably lower than this. To test the comparability of narrow-band and wide-band measurements in the Antarctic, concurrent

Table IV. Mean monthly values of the aerosol optical depth at wavelength 1.0 μm in Antarctica (x 10^3)

		Month								
Station	Year	1	2	3	4	9	10	11	12	Source
Maudheim 71°S, 10.9°E	1950–1951	25	18	10	10	18	18	18	25	1
Plateau 79.2°S, 40°E	1967–1968	21	25	23	–	–	16	16	17	2
Mirny 66.5°S, 93°E	1966, 1968 1969	24	24	20	16	16	20	22	21	3
Molodezhnaya 67.7°S, 45.8°E	1987	–	–	–	–	–	–	21	18	5
Georg-Forster	1988	–	–	–	31	26	27	37	31	4
70.8°S, 11.8°E	1989	27	31	26	–	–	–	–	–	4
	1990	21	24	19	15	–	–	–	20	4
Mirny	1988	27	20	19	16	–	–	–	–	5
66.5°S, 93°E	1989	13	19	23	27	42	27	21	31	5
	1991*	19	21	23	23	44	74	104	104	5
	1992*	105	98	–	–	77	85	76	81	5
	1993*	97	108	101	102	70	48	–	–	5

*aerosol optical depth at wavelength λ = 1037 nm.

Sources: 1. Liljequist (1957). 2. Kuhn (1972). 3. Sakunov (1975). 4. Leiterer & Herber (1992). 5. Herber *et al.* and this paper.

observations by both techniques were conducted at Mirny Observatory between October 1983 and March 1986. The results are presented in Fig. 2. Regression analysis has shown (Radionov & Marshunova 1992) that the mean monthly values of aerosol optical depth for both wavelengths (0.5 μm and 1.0 μm) yielded by the two techniques are comparable.

For convenient comparison of broadband and spectral optical characteristics of the atmosphere it seems reasonable to introduce a parameter

$$\tau_2 = -\ln P_2 \tag{6}$$

This may be formally referred to as an optical depth of the atmosphere for broadband radiation. Long-term variations of monthly averages of τ_2 and $\beta_{1.0}$ in December at Mirny Observatory are shown in Fig. 3. Within the observational period starting in 1956, dramatic increases of τ_2 during 1964–65 and 1983–84 were recorded. The first event is attributed to aerosol pollution of the stratosphere after the volcanic eruption of Mount Agung (8.2°S) in March 1963. At Mirny Observatory the growth of τ_2 was detected in December 1963. After the El Chichon volcanic eruption (17.3°N) in April 1982, the atmospheric turbidity increase was recorded in November 1982. The increase of τ_2 in this case was, however, smaller than that after the Agung eruption. The Arctic atmosphere became more turbid in the spring of 1983 than in 1964 (Marshunova & Radionov 1988). In both the Arctic and the Antarctic, the decrease of transmission coefficient after the volcanic eruption persisted for 1.5–2 years.

In Antarctica in late 1991, a dramatic increase of τ_2 was again recorded. Fig. 4 presents mean daily values of τ_2 and τ (0.5) obtained at Mirny Observatory during 1991–1992. The growth of the aerosol optical depth was recorded on 18 September 1991.

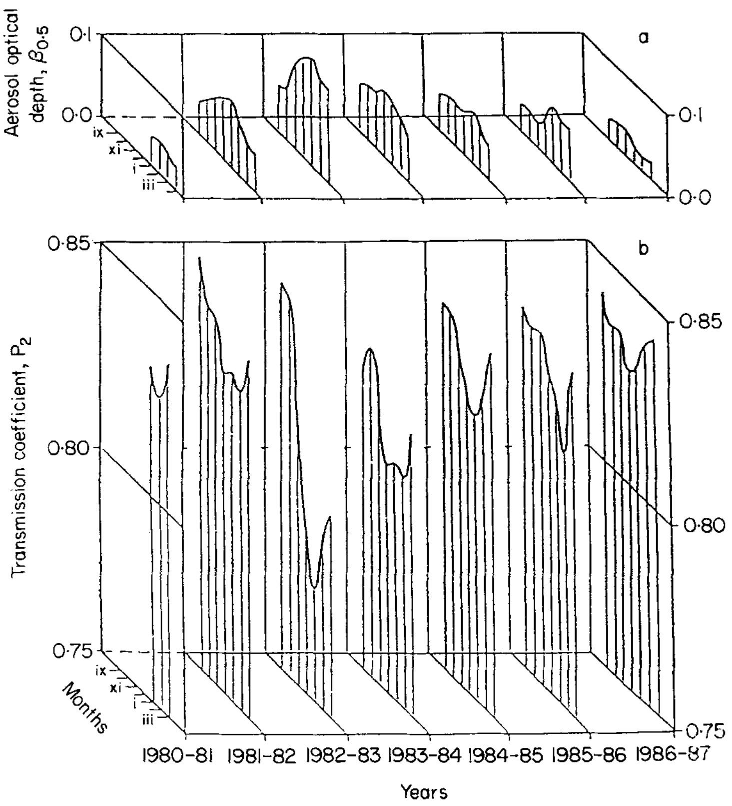

Fig. 1. Mean monthly values of **a.** aerosol optical depth $\beta_{0.5}$; **b.** transmission coefficient P_2 at Mirny Observatory.

Table V. Total water content *w* and its standard deviation Δw in mm, at Mirny Observatory.

Month	Sept	Oct	Nov	Dec	Jan	Feb	Mar
w	2.5	3.0	4.4	6.0	6.7	5.3	4.2
Δw	0.4	0.6	0.7	0.8	0.9	0.5	0.5

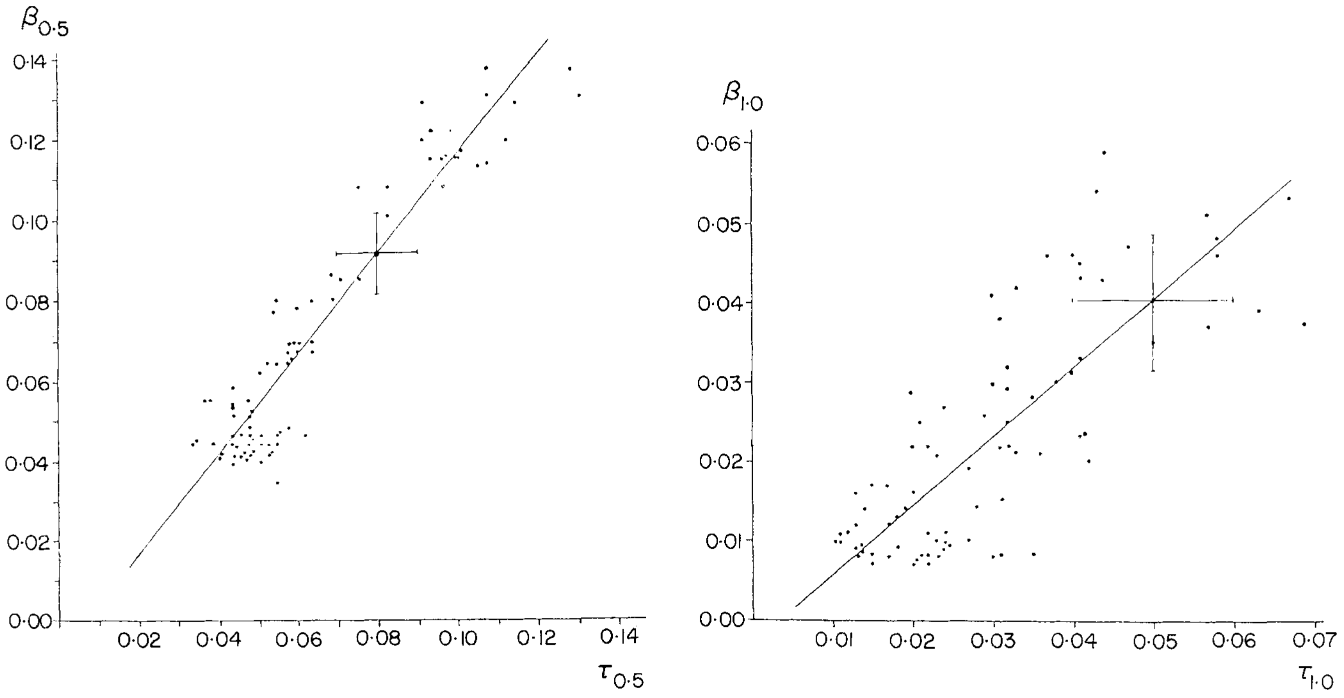

Fig. 2. Comparison of wide- and narrow-band measurements of aerosol optical depths.

A similar increase to that observed at Mirny during September–December 1991 was also seen at Georg Forster Station (Leiterer & Herber 1992). This effect is most probably caused by the stratospheric turbidity following the eruptions of Pinatubo (15°N) in June 1991 and Hudson (46°S) in August 1991. The increase in atmospheric turbidity during the period late 1991 to early 1992 at Mirny Observatory is comparable to that after the Agung eruption. The effect is expected to be of a global character. In the Arctic the highest aerosol turbidity was recorded from March 1992 onwards after the end of the polar night. The monthly means of τ_2 at some Arctic stations for the period from 1966–1980 and for 1992 are given in Table VI.

The analysis of long-term actinometric observations from the Arctic network (Radionov & Marshunova 1992) has shown that in spring (March–May) a downward trend of transparency is observed. This trend has been shown to be significant at the 99% confidence level with a mean rate of decrease of P_2 of 0.2% y^{-1} in the Western Arctic (Kara Sea) and 0.1% y^{-1} in the Eastern Arctic (Chukchi Sea) from the 1940s to 1982. The summer trend is negligible at the 95% confidence level. There is no corresponding downward trend in the Antarctic.

The possibility of calculating the parameter τ (0.5) using P_2

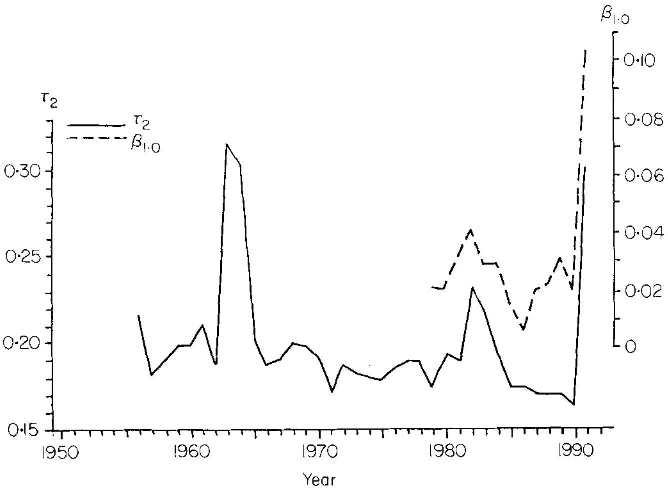

Fig. 3. Long-term variations of τ_2 and $\beta_{1.0}$ in December at Mirny Observatory.

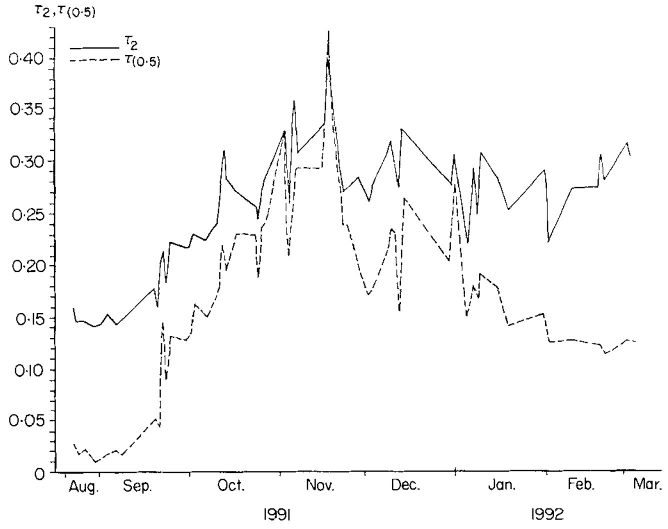

Fig. 4. Mean daily values of τ_2 and $\tau(0.5)$ during 1991–1992 at Mirny Observatory.

Table VI. Monthly mean of τ_2 and its standard deviations (x 10^2) in the Russian Arctic.

		Stations				
		Muostakh Is.	Chetyrekh-stolbovoy Is.	Uelen	Vankarem	Wrangel Is.
Month	Years	71.6°N 130°E	70.6°N 162.5°E	66.2°N 169.8°W	67.8°N 175.8°W	71°N 178.5°W
March	1966-88	26(4)	27(3)	24(4)	26(3)	25(3)
	1992	33(3)	34(4)	37(3)	37(2)	42(4)
April	1966-88	27(4)	27(3)	27(4)	26(4)	26(3)
	1992	38(3)	37(3)	37(3)	38(4)	38(3)
May	1966-88	27(4)	26(3)	27(4)	26(3)	25(4)
	1992	40(3)	37(3)	37(3)	36(3)	38(3)
June	1966-88	26(4)	26(3)	25(3)	25(3)	25(4)
	1992	37(3)	34(1)	34(3)	36(3)	38(4)
July	1966-88	25(4)	25(3)	26(3)	25(3)	26(4)
	1992	38(3)	33(4)	34(4)	36(1)	36(3)
August	1966-88		24(2)	25(4)	25(3)	25(3)
	1992		33(4)	31(3)	31(3)	36(1)

measurements for the periods without observations of $\tau_{0.5}$ was studied. Simultaneous time series of 237 daily mean values of $\tau_{0.5}$ and P_2 from Mirny Observatory for the period September–March 1983–1987 have been analysed for this purpose. The parameters of the regression equation

$$\tau(0.5) = a - b \ln P_2 \tag{7}$$

were calculated. These are $a = -0.110 \pm 0.006$, $b = 0.892 \pm 0.032$. The values of a and b are significant at the 99% confidence level. The correlation coefficient, r, is equal to 0.88 and the residual standard deviation is equal to 0.012. This confirms the validity of deriving the $\tau(0.5)$ values using P_2 data for conditions under which the stratosphere is not polluted by volcanic activity.

Conclusions

Actinometric measurements of direct solar radiation from 1956 to the present have allowed the transparency variations of the atmosphere in Antarctica (Mirny Observatory) to be analysed on a range of time scales. The transmission coefficient is observed to have an annual minimum, attributed in particular to the increase of atmospheric turbidity, in December–January. The interannual variations of the atmospheric transparency are induced primarily by a dramatic increase of aerosol turbidity of the atmosphere after a strong volcanic eruptions. This conclusion is confirmed by direct measurements of aerosol optical depth. There has been no systematic decrease of transmission coefficient since 1956 in Antarctica, in contrast to the observed decrease since the late 1950s in spring in the Arctic. The results presented in this paper demonstrate that the Antarctic atmosphere is free of man-made aerosol contamination.

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References

Bodhaine, B.A. 1983. Aerosol measurements at four background sites. *Journal of Geophysical Research*, **88**, 10 753-10 768.

Bodhaine, B.A. *et al.* 1989. *Geophysical Monitoring for Climatic Change, No. 17. Summary Report.* Boulder: NOAA Air Resources Laboratory, 14-28.

Evnevich, T.V., & Savikovsky, I.A. 1989. Calculation of dierct solar radiationand the transmission coefficient. *Meteorology and Hydrology*, No. 5, 106-109 (In Russian).

Herber, A. Thomason, L.W., Radionov, V.F. & Leiterer, U. 1993. Comparison of trends in the tropospheric and stratospheric aerosol optical depths in the Antarctic. *Journal of Geophysical Research*, **18**, 441-447.

Kondratyev, K.Ya. (*ed.*) 1978. *Atmospheric aerosol and its effects on radiation transfer.* Leningrad: Gidrometeoizdat Publishing, 120pp. (In Russian).

Kondratyev, K.Ya. & Binenko, V.I. (*eds*). 1981. *First Global GARP Experiment V.2. Polar aerosol heavy cloudiness and radiation.* Leningrad: Gidrometeoizdat Publishing, 152pp. (In Russian).

Kuhn, M.N. 1972. Die spektrale Tranparenz der Antarktishen Atmosphäre. *Archive Meteorologie, Geophysik und Bioklimatologie, Series B*, No. 20, 299-344.

Leiterer, U., Sakunov, G. 1989. Messungen von Aerosolpartikeln im Grössenbereich 0.2 bis 4.0 μm in der Antarktis. *Zeitschrift für Meteorolgie*, **39**, 309-316.

Leiterer, U., Weller, M. 1988. Sunphotometer BAS and ABAS for atmospheric research. *World Meteorological Office TD*, **222**, 21-26.

Liljequist, G.H. 1957. Energy exchange of an Antarctic snowfield. Short-wave radiation. *Norwegian-British-Swedish Antarctic Expedition 1949-1952. Scientific Results Vol. II, Part 1A.* Oslo: Norwegian Polar Institute, 109pp.

Marshunova M.S. 1980. *Conditions of the formation of radiation climate in Antarctica and its characteristics.* Leningrad: Gidrometeoizdat Publishing 214pp. (In Russian).

Marshunova, M.S. & Radionov, V.F. 1988. Variations in the integral atmospheric tranparency in polar regions. *Meteorology and Hydrology*, No. 11, 71-80. (In Russian).

Radionov, V.F. & Marshunova, M.S. 1992. Long-term variations in the turbidity of the Arctic atmosphere in Russia. *Atmosphere-Ocean*, **30**(4), 531-549.

Rusina, Ye. N. 1977. Determining the characteristics of aerosol atmospheric turbidity from data of spectral actinometric observations. *Meteorology and Hydrology*, No. 5, 49-55. (In Russian).

Sakunov, G.G. 1975. Relative effects of water vapour and aerosol on total radiation extinction in Antarctica. *Soviet Antarctic Expedition Bulletin*, No. 89, 48-52. (In Russian).

Sakunov, G.G., Radionov, V.F. & Barteneva, O.D. 1986. Study of aerosol optical properties of Antarctic atmosphere. *In Meteorological studies in Antarctic. Proceedings of the 2nd All-Union Symposium, Leningrad, 19-22 October 1981.* Leningrad: Gidrometeoizdat Publishing, 110-114 (In Russian).

Sivkov, S.I. 1968. *Methods of Calculations of Solar Radiation.* Leningrad: Gidrometeoizdat Publishing, 233pp. (In Russian).

Sonntag, D. 1970. Ein Panzeraktinometer mit galvanisch erzeugter Thermosäule. *Zeitschrift für Meterologie*, **21**, 300-307.