STRATOSPHERIC AEROSOL LAYER OVER TOMSK (56.5°N, 85.0°E) FROM THE RESULTS OF OBSERVATIONS AT THE SIBERIAN LIDAR STATION IN 1986–1997

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The paper generalizes the results of laser sensing of vertical distribution of the stratospheric aerosol at the Siberian Lidar Station of the Institute of Atmospheric Optics of the SB RAS in Tomsk (56.5°N, 85.0°E) *in 1986–1997. At the initial stage of regular monitoring of the stratospheric aerosol layer (SAL) we managed to observe the products of eruption of the Colombian volcano Del-Ruis in the stratosphere. In summer 1986 – summer 1991 the SAL state approached the background one. For this period the vertical aerosol distributions are presented in the paper separately for winters and summers. On their basis we have formulated the criteria for the background SAL state, which was observed only in 1989–1990. The experimental data on the background content of the stratospheric aerosol are presented in comparison with the model ones. After the Mt. Pinatubo eruption in June 1991, the volcanic aerosol became the dominating factor in the SAL formation. The results of measurements are considered that illustrate development and subsequent decay of the volcanic SAL. Since summer 1995 the stratospheric aerosol has been recovering its background state.*

INTRODUCTION

The state of the stratospheric aerosol layer (SAL) is affected by the natural (the influence of the atmospheric circulation on the thermodynamic conditions of aerosol generation and transport as well as sporadic volcanic eruptions[1,2]) and anthropogenic (products of fuel combustion of high-altitude aviation and emissions of gases and particles of technogenic origin into the stratosphere[3]) factors. Owing to its optical properties and global distribution, the stratospheric aerosol significantly affects the radiation balance and hence the climate of the planet. The variations of the SAL and related radiative effects become especially pronounced after powerful volcanic eruptions, whose products fill the stratosphere. This results in the considerable increase of the stratospheric aerosol mass, leads to local warming of the stratosphere at altitudes of SAL localization due to the upwelling IR radiation absorption, and, on the contrary, to cooling of the Earth's surface due to downwelling short-wave solar radiation scattering by the aerosol layer. Thus, for example, two months after the Mt. Pinatubo eruption in June 1991 the temperature in the tropical stratosphere increased by 3.5°C (at a pressure altitude of 30 hPa), whereas on the Earth's surface it decreased[4] by 0.7°C. After the eruptions we recorded the depletion of the stratospheric ozone layer, and the balance of other minor gas components was disturbed.[5,6] As a result, the radiative regime and the general atmospheric circulation changed.

The SAL variability in different periods has not only annual, but also seasonal and regional peculiarities. To study the peculiarities of vertical distribution of the stratospheric aerosol, regular SAL monitoring has been conducted at the Siberian Lidar Station (SLS) since early 1986 with the lidar operating at a wavelength of 532 nm. Till April 1992 the observations had been carried out with the lidar having the receiving telescope 1 m in diameter and a high-frequency (2.5 kHz) Nd:YAG laser.[7] Since November 1991 the SAL observations have been continued with the lidar having the receiving telescope 2.2 m in diameter and a higher-power Nd:YAG laser.[8] This paper generalizes the results of lidar observations of the SAL in 1986–1997.

The optical parameters of the stratospheric aerosol were determined from the laser sensing data on the basis of the lidar signal calibration against molecular scattering. The main aerosol parameter, derived from the laser sensing data, was the aerosol backscattering coefficient ($\beta_\pi^a(H)$), which was used to estimate other optical aerosol characteristics. With the increase of the altitude H the values of $\beta_\pi^a(H)$ and the values of the molecular backscattering coefficient $\beta_\pi^m(H)$ decreased

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by the exponential law, which was dominating in the aerosol vertical distribution. For more vivid and delicate representation of the aerosol vertical stratification, the scattering ratio $R(H)$ was used, which is connected with $\beta_\pi^a(H)$ as follows:

$$R(H) = \left[\beta_\pi^a(H) + \beta_\pi^m(H)\right] / \beta_\pi^m(H). \quad (1)$$

In the lidar equation

$$N(H) = C\left[\beta_\pi^a(H) + \beta_\pi^m(H)\right] T^2(H) / H^2 \quad (2)$$

(here, C is the instrumental constant including the characteristics of the transceiving system, $N(H)$ is the lidar signal coming from the sensing path, and $T^2(H)$ is the squared atmospheric transmission) the scattering ratio enters in an explicit form after its normalization to the molecular backscattering coefficient. This normalization is made using the values of the lidar signal, corrected for the squared distance and the molecular density, coming from the altitude range, where the main contribution to scattering comes from the air molecules. It has been shown[9] that at altitudes greater than 30 km the aerosol contribution to scattering does not exceed 2%. The specific implementation of the method of lidar return signal calibration against molecular scattering corrected for the atmospheric transmission, used by the authors, was described in detail elsewhere.[10,11,12]

The vertical distribution of the molecular scattering coefficient used for reconstruction of the profiles of $R(H)$ in 1986–1991 was calculated from the data of meteorological sensing, and in 1992–1997 it was determined for the optical-meteorological model.[13] Clearly, the use of the model values of $\beta_\pi^m(H)$ increases the error in determination of the aerosol characteristics. However, a comparison between the values of $R(H)$ for 1986–1991 determined with the model values of $\beta_\pi^m(H)$ and the aerological data have shown that their difference does not exceed 7%. For high content of the stratospheric aerosol, for example, after volcanic eruptions this difference becomes insignificant. The random error in reconstruction of the scattering ratio between 10–15 km is negligible; however, it increases with the altitude and reaches 7–10% between 25–30 km for the lidar with a receiving mirror 1 m in diameter, whereas for a receiving mirror 2.2 m in diameter it is less than 3%.

RESULTS OF LIDAR OBSERVATIONS

The general pattern of filling of the stratosphere with the aerosol is most vividly illustrated by the total aerosol backscattering coefficient $\Sigma\,\beta_\pi^a$. The temporal behavior of this SAL characteristic between 15–30 km is shown in Fig. 1 for the decade. The temporal behavior of $\Sigma\,\beta_\pi^a$ agrees well with the data obtained at the Mid-Latitude Lidar Observatories in Obninsk[14] (55°N, 38°E), Garmisch-Partenkirchen[15] (47.5°N, 11.1°E), and Tsukuba[16] (36°N, 140°E), which indicates the common character of the processes of SAL formation at mid-latitudes of the Northern Hemisphere on the global scale. Five characteristic periods can be identified in the temporal behavior of $\Sigma\,\beta_\pi^a$ shown in Fig. 1 by the values of $\Sigma\,\beta_\pi^a$, their variability, and dynamics of the vertical distribution of the stratospheric aerosol optical characteristics. Periods I and IV are characterized by the occurrence of the pronounced aerosol layers in the stratosphere after the volcanic eruptions. Periods II and V are characterized by the near-background content of the stratospheric aerosol, but with clearly pronounced seasonal differences in its vertical stratification. They can be considered as the periods of the transition quasi-background SAL state with the weakly pronounced maximum of the scattering ratio at altitudes of the Junge layer. Period III is characterized by the background SAL state. Below we consider the peculiarities and distinguishing properties of these periods in chronological order.

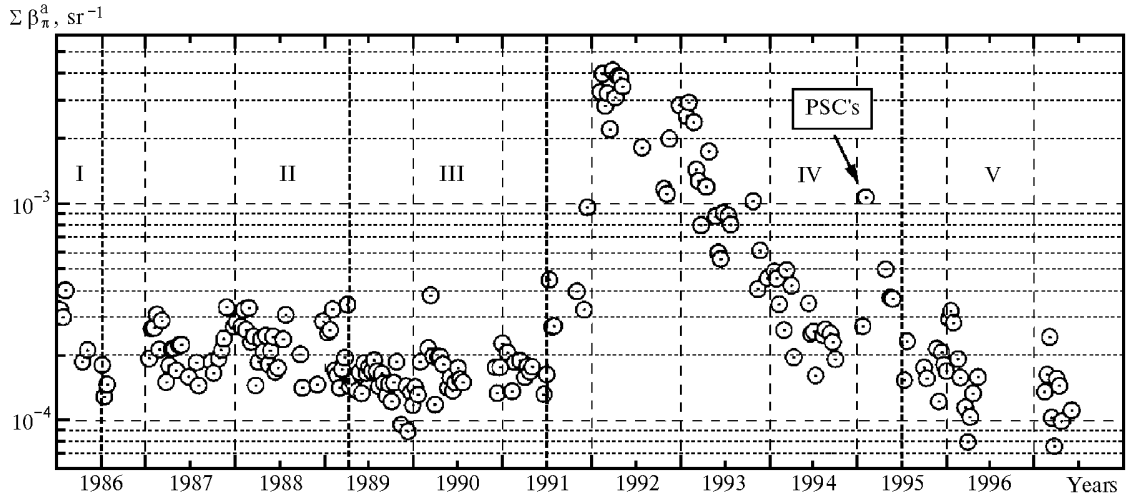

FIG. 1. Temporal behavior of the total aerosol backscattering coefficient $\Sigma\,\beta_\pi^a$ *at* λ = 532 nm *between* 15–30 km *over Tomsk* (56.5°N, 85.0°E).

Period I (January–May 1986). Aerosol layers after the volcano Del-Ruis eruption. In January 1986 the aerosol layers were observed in the stratosphere over Tomsk, which were the result of the volcano Del-Ruis eruption in December 1985 in Columbia.[17] During this period analogous aerosol layers were also observed in Hawaii,[18] Japan,[19] and Germany.[20] Over West Siberia the volcanic aerosol manifested itself through the sharply pronounced maxima in the vertical profiles of the scattering ratio, clearly distinguished against the background of the smooth stratification of the stratospheric aerosol layer. In different days these layers were observed at different altitudes, but were localized between 15–22 km. Figure 2*a* shows the vertical distribution of the stratospheric aerosol on January 24, 1986. In this case, the maximum of $R(H)$ reached ~ 1.4 and was at an altitude of 21.5 km. Note that under the background conditions the maximum of $R(H)$ at this altitude did not exceed 1.13.

The volcanic aerosol clouds could be identified already in April 1986 (see Fig. 2*b*). Moreover, the similarity of the profiles of the scattering ratio shown in the figure enabled us to assume that on January 24 and April 19, 1986 we observed the stratification of one and the same volcanic aerosol cloud, but 85 days later. Under the effect of the dynamic processes (diffusion, advection, convection, sedimentation, and so on) the cloud was transformed, its maximum was reduced, and its altitude was decreased by 1660 m. The rate of its sedimentation turned out to be equal to $2.3 \cdot 10^{-2}$ cm/s. As a whole, the eruption of the volcano Del-Ruis was weak and produces only insignificant increase of $\sum \beta_\pi^a$ (see Fig. 1).

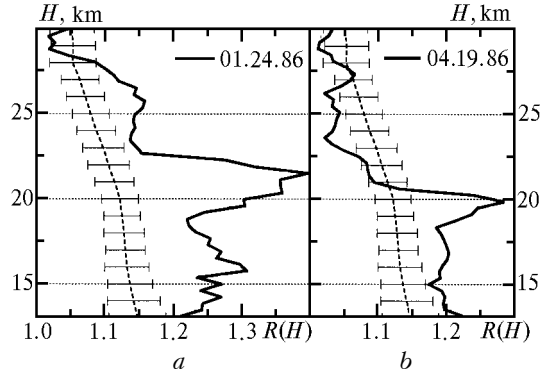

FIG. 2. Vertical profiles of the scattering ratio with the pronounced volcanic aerosol layers measured on January 24, 1986 (Fig. 2a) and April 12, 1986 (Fig. 2b). The average profile of the scattering ratio, calculated for the background stratospheric state, is also shown by the dashed curve.

Period II (summer 1986–winter 1989) and period III (spring 1989–June 1991). Relaxation of the stratosphere to the background SAL state. Although the aerosol perturbation of the stratosphere after the eruption of the volcano Del-Ruis was weak, its influence on the aerosol filling of the stratosphere was summed with the residual effect of the more powerful eruption of the Mexican volcano El-Chichon in March-April 1982. After that the stratospheric aerosol state recovered its stable background state rather slow and gradually. Since the SAL state in period II transformed smoothly into period III, we consider these periods together. The results of investigation of these periods were partly published elsewhere.[21,22]

Period II is characterized by the average values of $\sum \beta_\pi^a$ of about $2 \cdot 10^{-4}$ sr^{-1}. In this period the seasonal differences were pronounced in the aerosol filling of the stratosphere (the winter values of $\sum \beta_\pi^a$ exceeded twice its summer values) and in the peculiarities of the vertical distribution of the stratospheric aerosol. For period III the average values of $\sum \beta_\pi^a$ were somewhat lower, its minimum values reached $9 \cdot 10^{-5}$ sr^{-1}, and its seasonal variability was almost not expressed. The spread in amplitudes of maximum and minimum values of $\sum \beta_\pi^a$ for period III was less than for period II. The variance of the time series of $\sum \beta_\pi^a$ for period II exceeded that for period III 4 times.

The differences between these periods can be followed very well from an analysis of seasonal variability of the vertical distribution of the stratospheric aerosol. Figure 3 shows the profiles of $R(H)$ averaged over the summer (June, July, and August) and winter (December, January, and February) months for every year from summer 1987 till winter 1990/1991. It is well seen that considerable seasonal differences of $R(H)$ at altitudes below 20 km at the beginning of observations gradually disappeared by 1989–1990. This process is illustrated in more detail in Fig. 4. The vertical profiles of the parameter $\mu(H)$, which is defined as a ratio of the adjacent average summer profiles of $R(H)_{sum}$ to the winter profiles $R(H)_{winter}$, i.e., $\mu(H) = R(H)_{sum}/R(H)_{winter}$, are shown in this figure. It can be seen that in 1987–1990 we observed progressive smoothing of the vertical distributions from season to season. The Junge layer is less pronounced due to sedimentation of the residual volcanic aerosol, the seasonal differences are smoothed, and the parameter $\mu(H)$ approaches unity.

Thus, the absence of considerable seasonal variations in the vertical distribution of the stratospheric aerosol can be considered as one of the criteria of its background state. From the entire array of the available data, the profiles of the scattering ratio in 1989–1990 satisfy this criterion.

The average profiles of $R(H)$ for the main seasons – winter and summer – were obtained in this period. Then they were transformed into the profiles of $\sum \beta_\pi^a$. The number of profiles was 54 (33 summer profiles and 21 winter profiles). Based on these profiles, the background profile (BGP) of the aerosol backscattering coefficient over Tomsk was obtained for a wavelength of 532 nm by way of the calculation of the average profile and its standard deviation.

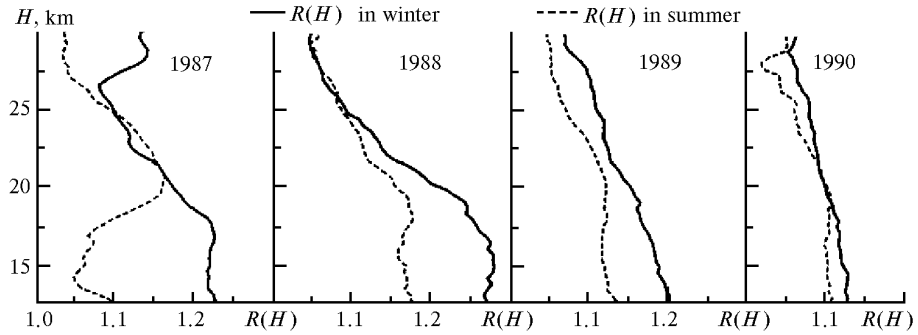

FIG. 3. Vertical winter (the solid curves) and summer (the dashed curves) profiles of the scattering ratio at λ = 532 nm *for the indicated years.*

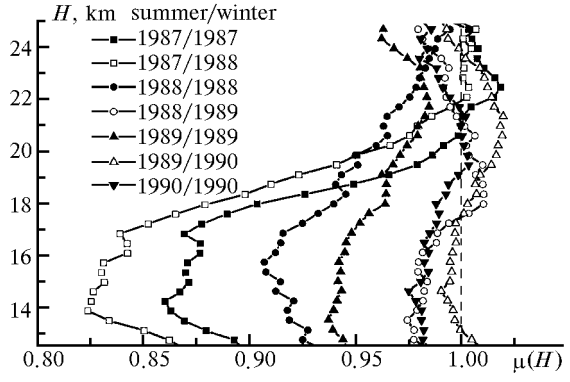

FIG. 4. Vertical behavior of the parameter $\mu(H)$ *calculated for the corresponding annual combinations of the average summer and winter profiles of R(H).*

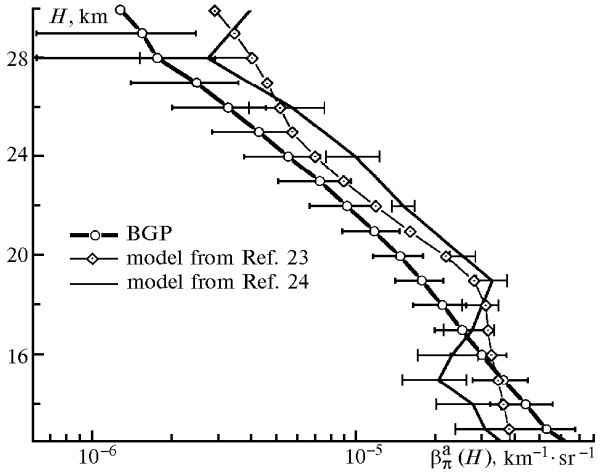

FIG. 5. Vertical distribution of the aerosol backscattering coefficients at λ = 532 nm, *obtained by averaging of the individual experimental profiles of* $\sum \beta_\pi^a$ *measured under background conditions in* 1989–1990, *in comparison with the aerosol models.*

This vertical profile is shown in Fig. 5 in comparison with the most widespread statistical background aerosol models.[23,24] It can be seen from the figure that above 17 km both models exceed our empirical values by about 25–30%, and their stratification is more complex. The model of McClatchey et al.[24] has the sharply pronounced bending point between 19–20 km. The model of Zuev and Krekov[23] has the well pronounced maximum at these altitudes. Hence, these models cannot be considered background ones. They explicitly describe the Junge layer (the residual volcanic aerosol). Between 12.5–30 km, our empirical model is well approximated by the straight line $y = A + B \cdot x$ on the logarithmic scale, where $A = -6.791 \pm 0.095$ and $B = -0.224 \pm 0.004$ (the same approximation can be written for $\beta_\pi^m(H)$ with $B = -0.1575 \pm 0.0003$); on the linear scale, we can write down $\beta_\pi^a(H) = 1.124 \cdot 10^{-3} \cdot e^{-0.224 \cdot H}$.

Thus, the smooth exponential vertical distribution of the stratospheric aerosol can be considered as the second criterion for the SAL background state.

Period IV (June 1991–spring 1995). The SAL disturbed after the Mt. Pinatubo eruption. The powerful Mt. Pinatubo eruption in the Philippines in June 1991 has resulted in the global SAL changes all over the world. In Tomsk the eruption products were first recorded[25] in July 1991. The results of the further lidar investigations of dynamics of the volcanic stratospheric aerosol over Tomsk were partly published elsewhere.[26,27]

The maximum aerosol filling of the stratosphere was observed in January-February 1992. In this period the values of $\sum \beta_\pi^a$ reached $4 \cdot 10^{-3}$ sr^{-1}. The aerosol content in the stratospheric layer between 15–30 km exceeded almost twice its background values due to filling with the volcanic aerosol. The calculated values of the optical thickness τ of this layer for a lidar ratio of 0.02 sr^{-1} (the ratio of the aerosol backscattering coefficient to the extinction coefficient), borrowed from Ref. 28, were of the order of 0.2 for the maximum filling of the stratosphere with the volcanic aerosol in January–March 1992. As a whole, the values of τ between 15–30 km exceeded 0.05 and were comparable with τ of cirrus clouds[29] for one and a half year since late 1991. Thus, in that period the disturbed SAL could significantly affect the radiative regime of the atmosphere on the global scale.

After the maximum aerosol filling of the stratosphere in January–March 1992, the process of the stratospheric volcanic aerosol layer decay lasted more than three years. To estimate the rate of decay of the SAL volcanic aerosol, as a rule, the *e*-fold decay time is calculated for the aerosol mass or any other characteristic related to it. For the total aerosol backscattering coefficient between 15–30 km, the *e*-fold decay time was calculated for the period February 1992 – December 1994. This period is $354 \pm^{41}_{33}$ days or 11.6 months. Close values were obtained for a wavelength of 532 nm from the series of lidar observations at the Mid-Latitude Stations in Obninsk[14] (55°N, 11 months), Garmish–Partenkirchen[28] (47.5°N, 12.4 months), and Hampton[30] (37.1°N, 10 months).

It should be emphasized that the decay time differed for different altitude ranges. The upper stratospheric layers decayed quicker in comparison with the lower ones. This is due to the altitude dependence of the sedimentation rate of aerosol particles and the penetration of the aerosol from the above stratospheric layers into the lower layers in the process of aerosol sedimentation. According to our measurements, between 15–20 km the *e*-fold decay time was 386 days (12.7 months), whereas between 20–25 km it was 305 days (10 months). Figure 6 illustrates this process in more detail. Here, the average aerosol backscattering coefficients are shown for four seasons: winter, spring, summer, and fall for 1992–1995 (some average seasonal points are not shown because of insufficient data statistics). The values of β_π^a, averaged over the altitude ranges 13–15, 17–19, 21–25, and 27–29 km, are denoted by $\{\beta_\pi^a\}$. It can be seen that the scattering aerosol is mostly concentrated in the lower stratosphere (below 20 km). The relaxation of the volcanic aerosol in the upper layers (above 20 km) occurred much faster and had been completed practically by late 1993. As a whole, the total aerosol backscattering coefficient between 15–30 km had recovered its background value[31] only by summer 1995.

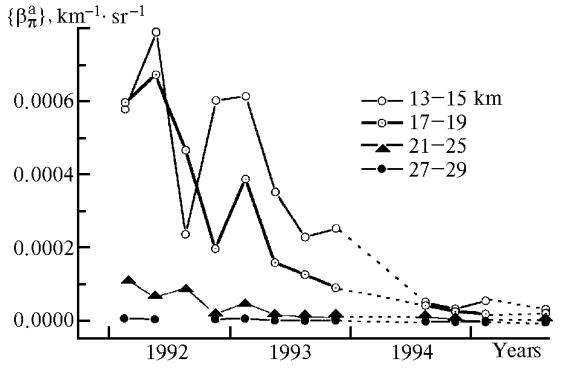

FIG. 6. Seasonal (winter, spring, summer, and fall) behavior of the aerosol backscattering coefficient $\{\beta_\pi^a\}$ *averaged over the indicated altitude ranges.*

In the period of SAL perturbation by the volcanic aerosol the annual winter peaks of the aerosol content are clearly manifested.[15,16,30] These annual cycles are due to intensification of the meridional component of the stratospheric circulation in winter and transport of the volcanic aerosol from the tropical reservoir to the mid-latitudes. The seasonal variability of the SAL in this period was also characterized by the different behavior of the average winter and summer profiles of the aerosol backscattering coefficients. Figure 7 shows these profiles in 1991–1995. It can be seen that at the beginning (in summer 1991) the volcanic aerosol, directly emitted in the stratosphere after the Mt. Pinatubo eruption, was observed only in the lower stratosphere (below 17 km). In winter 1991/1992 the content of the volcanic aerosol was much larger, and it spread over much wider altitude range (see Fig. 7). Between 15–23 km, the values of β_π^a in winter 1991/1992 exceeded the background values of β_π^a (BGP) by more than an order of magnitude.

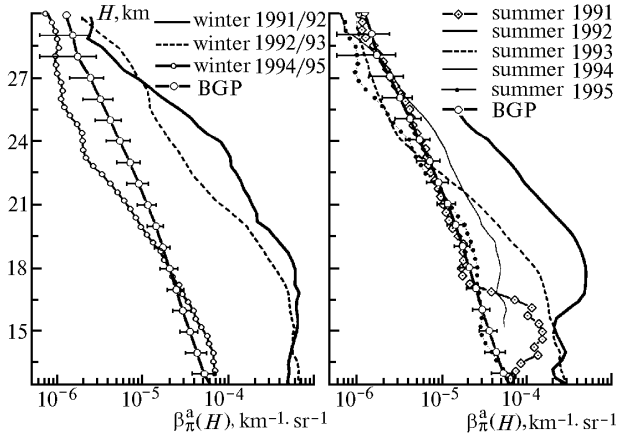

FIG. 7. Average seasonal vertical profiles of the aerosol backscattering coefficients for winter and summer at $\lambda = 532$ nm *in comparison with the average profile of* $\beta_\pi^a(H)$ *for the background period (BGP) in 1989–1990.*

In summer the penetration of the stratospheric aerosol from the tropical zone into the SAL at mid-latitudes was terminated due to the decrease of the temperature gradient in the stratosphere between the low and high latitudes. In spring–summer 1992 the stratosphere was cleaned from the volcanic aerosol particles. This was especially well pronounced below 17 km. In summer 1993 the aerosol content in the upper stratospheric layers above 20 km decreased sharply. The aerosol layer below 20 km was uniformly spread and retained relatively high aerosol content due to the penetration of the aerosol into the lower layers in the process of cleaning of the upper layers. In summer 1994 the values of β_π^a exceeded slightly its background values, and near 19 km the Junge layer

had been formed. Figure 7 shows that the profiles of β_π^a in summer 1995 and winter 1994/1995 approached the average background profile, which, according to the above-formulated criterion, means the onset of the transition period of SAL stabilization.

At the end of the examined period in winter 1995 we also observed significant but short-term SAL perturbations. The values of $\Sigma\beta_\pi^a$ increased sharply by a factor of 5–6. This SAL perturbation was connected with the polar stratospheric clouds (PSC's) formed in the cold center of the circumpolar vortex displaced in that period toward mid-latitudes. The PSC's were recorded in the stratosphere over Tomsk late in January[31] simultaneously with abnormally low temperatures of –80°C. Figure 8 shows the profile of $R(H)$ recorded on January 24, 1995 and the temperature profile corresponding to that date, measured with a weather-balloon. Here, the typical profiles of $R(H)$ obtained in September 1994 and April, May, and July 1995 are also shown for comparison. It can be seen that at the instant of recording PSC's considerable SAL perturbation was observed in a wide altitude range up to 25 km. The well-defined PSC maximum between 16–17 km practically coincided with the temperature profile minimum. At the same time PSC's were observed at the Mid-Latitude Lidar Station in Obninsk.[4] Moreover, PSC's were observed in Obninsk late in winter 1989, i.e., for the background SAL state.[32]

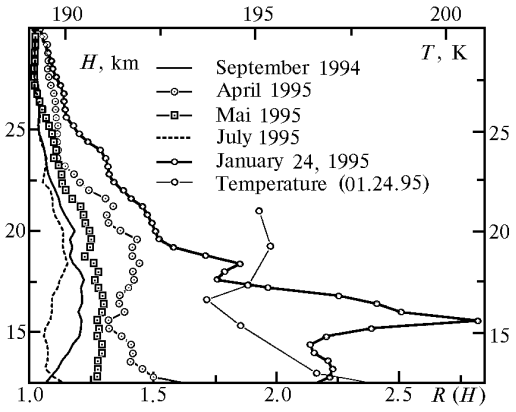

FIG. 8. Typical profiles of the scattering ratio for indicated months in comparison with the profile of the scattering ratio in the presence of PSC's (on January 24, 1995) and the temperature profile for the same date measured with the weather balloon.

Figure 8 shows that in winter and spring 1995 the aerosol filling of the SAL was higher than in fall 1994. This again indicates the annual cyclicity of the increase of the stratospheric aerosol content in winter as a result of intensification of the meridional transport. The change of the vertical structure of spring profiles of $R(H)$ in April–May 1995 is indicative of intense penetration of the stratospheric volcanic aerosol into the troposphere. In July 1995 the summer aerosol profile of $R(H)$ typical of the transition period was observed with the weak maximum near 18 km.

Period V (summer 1995–December 1997). Stabilization of the SAL. In summer 1995 the scattering ratio in the SAL maximum reached its background value of ~1.16. The average values of $\Sigma\,\beta_\pi^a$ in that period were also at pre-Pinatubo levels, and even the tendency for their decrease was observed (the values of $\Sigma\,\beta_\pi^a$ in 1997 were somewhat less than in 1996). The minimum values of $\Sigma\,\beta_\pi^a$ in 1997 reached $8\cdot10^{-5}$ sr^{-1}, which was even less than the corresponding values in 1989–1990. However, unlike background period III, in period V the SAL was unstable. The difference between the maximum and minimum values of $\Sigma\,\beta_\pi^a$ in period V was larger than in period III. Nearly the same spread in values of $\Sigma\,\beta_\pi^a$ was typical of period II. This indicates that this period is the transition one. In our opinion, after the relaxation of aerosol filling of the stratosphere to the pre-volcanic values, additional time is required in order that all the factors affecting the general circulation of the atmosphere recover their stability. This time we call the transition period or the period of quasi-background SAL state.

In conclusion of this section, we also note that the level of the total coefficient of aerosol scattering in 1996 and 1997 was comparable to its levels in 1979 and 1989–1990, when the aerosol content was very low. This was also pointed out by other authors.[15,30] Therefore, the proposed trend for the increase of the background stratospheric aerosol mass by up to 5% per year as a result of anthropogenic activity[3,33,34] is not evident at all.

CONCLUSIONS

In the present paper the results of lidar investigations of the stratospheric aerosol layer (SAL) at the Siberian Lidar Station in 1986–1997 have been generalized. Three types of the SAL state have been identified: 1) aerosol layer disturbed by the volcanic eruption, 2) transition quasi-background layer, and 3) background layer.

The first type of the SAL state encompasses the periods when the pronounced aerosol layers were observed in the stratosphere over Tomsk after eruptions of the volcanoes Del-Ruis (in Colombia in December 1985) and Mt. Pinatubo (in the Philippines in June 1991). The stratospheric volcanic aerosol from the first eruption could be identified for 3–4 months. However, as a whole, the eruption was weak and did not affect strongly the atmospheric processes. The Mt. Pinatubo eruption was the most powerful in the 20th century. The traces of the volcanic aerosol were recorded till summer 1995. The maximum values of the total aerosol backscattering coefficient increased in January–February 1992 up to $4\cdot10^{-3}$ sr^{-1}, and the aerosol optical thickness reached 0.15. The *e*-fold decay time, determined from the total aerosol backscattering coefficient between 15–30 km at a wavelength of

532 nm, was 11.9 months. The short-term SAL perturbations, including those in the background period, were observed in winter caused by formation of the polar stratospheric clouds (PSC's) as a result of intrusion of the cold center of the circumpolar vortex into mid-latitudes. The transition quasi-background SAL state was observed in summer 1986 – spring 1989 and again since summer 1995. For this SAL type the typical values of the total aerosol backscattering coefficient $\sum \beta_{\pi}^{a}$ were at a level of $2 \cdot 10^{-4}$ sr^{-1}. In this case the seasonal variability was pronounced not only in the temporal behaviorof $\sum \beta_{\pi}^{a}$, but also in the vertical stratospheric aerosol distribution with maximum values in winter.

The average values of the total aerosol backscattering coefficient for the background SAL were at a level of $1.5 \cdot 10^{-4}$ sr^{-1}. It was shown that the seasonal (winter–summer) variations in the vertical stratospheric aerosol distribution were absent for this SAL state. The vertical aerosol distribution in this period was well approximated by the exponential function. Both factors are considered as the criteria for the background SAL state.

The fact that the level of the total aerosol scattering in 1996 and 1997 was analogous to that in 1979 and 1989–1990, when very low aerosol content was observed in the stratosphere, calls into question the hypothesis about the increase of the background stratospheric aerosol mass by up to 5% per year as a result of anthropogenic activity.

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