A Study of the Radiation—Climate Effect of Aerosol over Beijing Area

Qiu Jinhuan (邱金桓) and *Wang Kaixiang* (王开祥)
Institute of Atmospheric Physics, Chinese Academy of Sciences, Beijing 100029
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ABSTRACT

In this paper, the monthly-averaged planetary albedo and equivalent blackbody temperature are calculated by using Gauss-Seidle numerical model to solve the equation of radiative transfer, based on measured aerosol data over Beijing aera. With the increase of atmospheric turbidity, the planetary albedo has different characteristics in different seasons, and there is an evident decrease in the winter season. It means that the local aerosol has an heating effect to the atmosphere in winter. The correlation feature between the surface temperature and the horizontal visibility from 1963 to 1986 is analyzed, and anticorrelation is discovered in winter. It is found that surface temperature increases with the increase of aeresol.

I. INTRODUCTION

With the developing of industry, the content of aerosol in the atmosphere increases continuously and its climate effect is gradually paid to great attention. Whether aerosol has a heating or cooling effect on the air is greatly concerned by meteorologists since 1960's. For first order approximation and with neglecting the aerosol radiation effect in the infrared, the planetary albedo's variation caused by aerosol was used to assess the aerosol radiation climate influence (Ensor, 1971; Yamamoto and Tanaka, 1972; Reck, 1975; Ackerman, 1989). Yamamoto and Tanaka (1972) calculated the planetary albedo varying with the atmospheric turbidity (aerosol optical depth) and the imaginary part of aerosol refraction index under given aerosol distribution and surface reflectivity. Their results show that the planetary albedo decreases with the increase of atmospheric turbidity, when the imaginary part of the refraction index is larger than 0.05, the aerosol has a heating effect on the atmosphere. But a cooling effect is found when the imaginary part is less. The other studies (Ensor, 1971; Reck, 1975) also express that the imaginary part of the aerosol refraction index is an important factor to influence the planetary albedo through changing atmosphere absorption to the solar radiation. However, all these studies do not discuss the effect of the aerosol size distribution. Otherwise, in recent years some scholars put the aerosol disturbance into the global climate spectral model to study its climate effect (Geleyn et al., 1983). But as Diermendjian (1980) pointed out, the question of heating or cooling effect of aerosol to the atmosphere is not solved yet, owing to lack of real observations of aerosol physical and optical properties.

According to radiative transfer equation, aerosol influence on the planetary albedo mainly depends upon its phase function, single scattering albedo and optical thickness. The former two factors can be determined if aerosol size distribution and its refractive index are known. In our paper, based on observational data (Qiu et al., 1988) of aerosol size distribution, its refraction index and atmosphere column optical thickness over Beijing area in 1983 and 1984, we calculate the planetary albedo and study the aerosol thermal effect in Beijing area by using Gauss-Seidle iteration method to solve the radiative transfer equation. In addition the statistical analysis of the correlation property between surface temperature and visibility in Beijing from 1963 to 1986 is made to discuss the aerosol climate influence.

II. CALCULATION MODEL

Under plane–parallel atmosphere, with neglecting polarization, the azimuth–independent solar radiative transfer equation is as follow:

$$\mu\frac{dI_\lambda(\tau_\lambda,\mu)}{d\tau_\lambda}=I_\lambda(\tau_\lambda,\mu)-\frac{\tilde{W}_\lambda}{2}\int_{-1}^{1}P^{(0)}(\tau_\lambda,\mu,\mu')I(\tau_\lambda,\mu')d\mu'$$
$$-\frac{\bar{F}_\lambda\tilde{W}_\lambda}{4}e^{-\tau_\lambda/\mu_0}P^{(0)}(\tau_\lambda,\mu,\mu_0)\ , \tag{1}$$

where μ is zenith cosine. $\mu_0=\cos\theta_0$, θ_0 is the solar zenith. τ_λ is the optical thickness at λ wavelength, $\tilde{W}_\lambda$ the single scattering albedo. $\pi\bar{F}_\lambda$ is the solar illumination at the top of the atmosphere at λ. I_λ and $P^{(0)}$ are, respectively, the intensity of the diffuse radiation field and the phase function independent of the azimuth.

Gauss–Seidle iteration method (Herman and Browning, 1965) is used here to solve the radiative transfer equation. We obtain the upper diffuse flux $F_\lambda^\uparrow(\mu_0)$ and diffuse reflectivity $R_\lambda(\mu_0)$ at the top of the atmosphere at λ as below:

$$F_\lambda^\uparrow(\mu_0)=2\pi\int_0^1 I_\lambda(0,\mu)d\mu\ , \tag{2}$$

$$R_\lambda(\mu_0)=F_\lambda^\uparrow(\mu_0)/\pi\bar{F}_\lambda\mu_0\ . \tag{3}$$

The same as Yamamoto et al. (1972), 0.3 μm to 2.3 μm wavelength range is considered, so the diffuse reflectivity with wavelength average $R(\mu_0)$ is

$$R(\mu_0)=\int_{0.3}^{2.3}\pi\bar{F}R_\lambda(\mu_0)d\lambda\Big/\int_{0.3}^{2.3}\pi\bar{F}_\lambda d\lambda\ . \tag{4}$$

In this paper, the monthly–averaged Beijing's local planetary albedo A_0 is expressed as

$$A_0=\int_{0.01}^{\bar{\mu}_0}R(\mu_0)\mu_0 d\mu_0\Big/\int_{0.01}^{\bar{\mu}_0}\mu_0 d\mu_0\ , \tag{5}$$

where the lower limit of integration 0.01 is the minimum solar zenith cosine considered, and the upper limit $\bar{\mu}_0$ the maximum solar zenith cosine on the 15th for every month. The solar zenith is calculated by using Spencer's approximate expression (Spencer, 1971).

The step length of solar zenith cosine is chosen as $\Delta\mu_0=\frac{\bar{\mu}_0-0.01}{10}$. Unequal wavelength step of 0.3, 0.45, 0.55, 0.7, 0.9, 1.3, 1.6, 1.9 and 2.23 μm is used. In addition, some assumptions are made in this paper as below.

(1) The diffuse phase function is independent of the height, i.e.,

$$P(\theta)=\frac{\tau_{a\lambda}P_a(\theta)+\tau_{m\lambda}P_m(\theta)}{\tau_{a\lambda}+\tau_{m\lambda}}\ , \tag{6}$$

where $P_a(\theta)$ and $P_m(\theta)$ are the scattering phase function of particles and molecules respectively, $\tau_{a\lambda}$ and $\tau_{m\lambda}$ the optical thickness.

(2) No cloud.

(3) The absorption of H_2O and O_3 in 0.3–2.3 μm wavelength range is not considered.

To further study the thermal effect of aerosol on the atmosphere–earth system, the same as Yamamoto's paper (1972), T_e (K) is assumed as the equivalent blackbody temperature of atmosphere–earth system, and it is expressed as

$$\pi R^2 S(1 - A) = 4\pi R^2 \sigma T_e^4 \ , \tag{7}$$

where R is the radius of the earth, S the solar constant, and σ the Stefan–Boltzemann constant. The value of $T_e = 254.1$ K is taken as a reference for the molecular atmosphere. From Eq.(7) we can get

$$\pi R^2 S \Delta A = 16\pi R^2 \sigma T_e^3 \Delta T_e \ , \tag{8}$$

$$\Delta T_e = \frac{SA}{16\sigma T_e^3} \ , \tag{9}$$

where ΔT_e is the variation of equivalent blackbody temperature caused by changing the planetary albedo owing to aerosol existence (related to the molecular atmosphere).

III. INPUT DATA OF AEROSOL OPTICAL PROPERTY AND SURFACE ALBEDO

The aerosol refraction index, aerosol distribution and atmosphere column optical thickness are very important physical parameters in analysing the aerosol radiative climate effect. Data assumed are commonly used to study the sensitivity of climate model to those parameters. Here we use the Beijing's measured aerosol optical parameters (Qiu et al., 1988) to study its climate effect. These data include monthly–averaged column optical thickness, seasonally–averaged aerosol size distribution and the real and imaginary part of the refraction index at wavelength 0.7 μm from November 1983 to December 1984. The refraction index is assumed not changed with the wavelength. The atmosphere column optical thickness at other wavelength is calculated according to Mie theory from the former data.

The surface albedo is taken from results measured by optical remote sensing method (Qiu et al., 1986), and a mean albedo of 0.185 at two wavelengths of 0.7 μm and 0.4 μm is used in the model.

IV. RESULT AND DISCUSSION

In this paper, December January and February are defined as winter of Beijing; March, April and May as spring; June, July and August as summer, September, October and November as autumn.

Fig.1 shows the monthly averaged planetary albedo from November, 1983 to December, 1984, where the curves connecting in circles and crosses correspond to the molecular atmosphere and the turbid atmosphere, respectively. It can be seen from Fig.1 that in the summer of Beijing, the planetary albedo of the turbid atmosphere is very close to that of the molecular atmosphere. In spring and autumn, the albedo of turbid atmosphere is less than that of the molecular atmosphere, which means that aerosol has an heating effect to the atmosphere. In winter, the planetary albedo is much smaller than that of molecular atmosphere, even with a difference of up to 0.03, the strong heating effect of aerosol on the atmosphere appears obviously.

Fig.2 shows the variation of the planetary albedo with the turbidity (aerosol optical thickness at wavelength 1 μm) in April, July, October and December. In October, the planetary albedo varies very little with increasing turbidity. In April and July, as turbidity increases from zero, the planetary albedo decreases at first and then increases. Only the albedo in

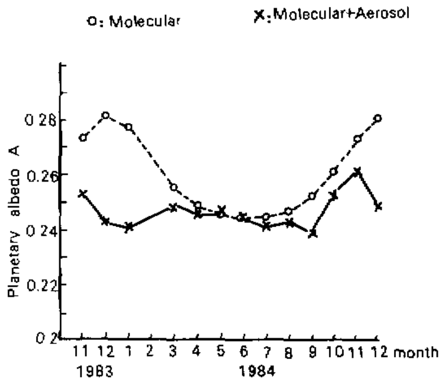

Fig.1. The variation of planetary albedo caused by aerosol.

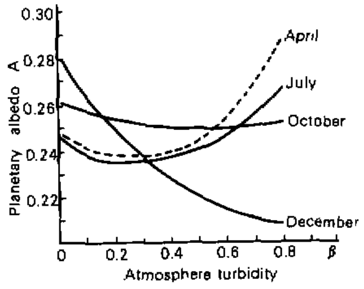

Fig.2. The changing of planetary albedo as function of turbidity.

December continuously decreases with the increase of atmosphere turbidity with the result that the albedo drops to about 25% as turbidity varies from 0 to 0.8.

Fig.3 shows the variation of atmosphere–earth system equivalent blackbody temperature T_e for the aerosol existence. It can be seen from Fig.3 that in summer T_e almost has no change and ΔT_e is near zero. In spring and autumn, ΔT_e is a little more than zero, which means that there is heating effect. But in winter, ΔT_e can be up to 3.5 (K), the heating effect of aerosol is quite obvious.

Fig.4 shows the ΔT_e of three different months as a function of atmosphere turbidity β. As can be seen that the ΔT_e of October changes very small. In April, when $\beta < 0.4$, ΔT_e is a little larger than zero, meaning heating effect. When β varies from 0.4 up, ΔT_e decreases, and when $\beta > 0.6$, $\Delta T_e < 0$, showing cooling effect. In December, ΔT_e dramatically increases with the result that ΔT_e is up to 4 K when β increases from 0 to 0.4.

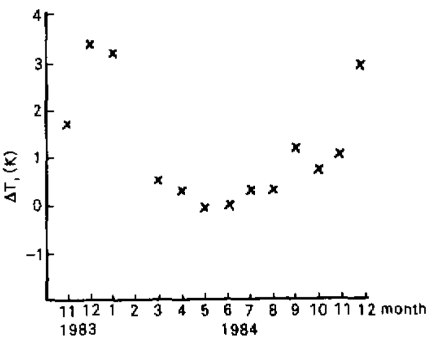

Fig.3. The variation of the equivalent blackbody temperature caused by aerosol.

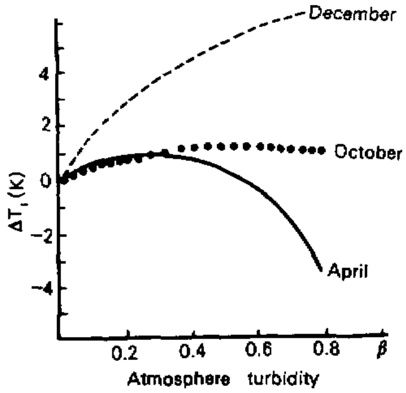

Fig.4. The changing of the equivalent blackbody temperature as function of turbidity.

As mentioned above, the aerosol not always has the cooling effect on the atmosphere in four seasons over Beijing area. Its heating effect is quite obvious in winter. The reason of this is discussed below from Table 1, Table 2 and according to the aerosol distribution presented by Qiu et al. (1988).

Table.1. The Refraction Index, the Single Scattering Albedo and the Backward Scattering Phase Function of Four Seasons

	Winter	Spring	Summer	Autumn
m_R	1.55	1.51	1.49	1.51
m_I	0.0572	0.0152	0.0128	0.0215
$\tilde{W}_a$	0.661	0.683	0.761	0.726
$\tilde{W}_a^*$	0.719	0.878	0.889	0.846
P_π	0.00847	0.0114	0.0146	0.0125
P_π^*	0.0101	0.0212	0.020	0.0177

Table.2. The Single Scattering Albedo and the Backward Scattering Phase Function of Both the Spring and Junge Size Distribution

M	Spring size distribution		Junge distribution ($\gamma^* = 3$)	
	$\tilde{W}_a$	P_π	$\tilde{W}_a^*$	P_π^*
0	1	0.0478	1	0.0369
0.01	0.732	0.0145	0.914	0.0235
0.02	0.679	0.0099	0.849	0.0174
0.03	0.651	0.0075	0.798	0.0136
0.05	0.606	0.0045	0.725	0.0093

In Table 1, m_R and M_I are the real and imaginary parts of aerosol refraction index (Qiu et al., 1988), respectively. $\tilde{W}_a$ is the single scattering albedo calculated according to the former refraction index and the seasonally-averaged aerosol size distrbution (Qiu et al., 1988). P_π is the backward scattering phase function. $\tilde{W}_a^*$ is the single scattering albedo calculated according to the above refraction index but Junge distribution of $\gamma^* = 3$, and P_π^* are corresponding backward scattering phase function. Table 2 lists the aerosol single scattering are albedo and backward scattering phase function for both the spring aerosol size distribution presented by Qiu at al. (1988) and Junge distribution with $\gamma^* = 3$, where the real part of aerosol refraction index is assumed 1.5 and the imaginary part is 0, 0.01, 0.02, 0.03 and 0.05, respectively.

Yamamoto et al. (1972) calculated the planetary albedo using Junge distribution with $\gamma^* = 3$, and it was discovered that only as imaginary part is larger than 0.05, the aerosol existence can make the planetary albedo decrease and so has heating effect under no cloud condition. It can be found from Table 1 that in winter the imaginary part of aerosol refractive index is considerably large in Beijing, and its average value is 0.0572. This may be because of the fact that there is a large amount of strong absorbing carbon particles in the atmosphere during winter warming period in Beijing (You Ronggao et al., 1983). The larger the imaginary part is, the smaller the single scattering albedo is, and so the stronger absorption to the solar radiation the aerosol has. Therefore the aerosol over Beijing in winter has a heating effect. Also from Table 1, we can see that the imaginary parts of refractive index are all less than 0.022 for the other seasons of Beijing. But it can be found From Fig.1 and Fig.4 that the aerosol existence does not make the planetary albedo increase in spring, summer and autumn seasons. This maybe is relative to the distinctive aerosol size distribution in Beijing. According to some measurement results (Qiu et al., 1988; You Ronggao et al., 1983; Ren Lixin et al., 1984), the aerosol size distribution over Beijing area is usually two–model distribution. In the range of larger than 1 μm, there is usually an obvious second peak, especially in the spring. According to Mie theory, the wider the distribution is, the more large particles there are, and the smaller the single scattering albedo is, the larger the asymmetry factor is. As shown in Table 1, for spring aerosol distribution, when m_I is 0.0152, $\tilde{W}_a$ is 0.683 and P_π is 0.0114. With the same m_I, for the Junge distribution with $\gamma^* = 3$, $\tilde{W}_a$ is 0.878 and P_π^* is 0.0212, which are obviously larger than the formers. The smaller the single scattering albedo is, the stronger the aerosol absorps the solar radiation; the smaller the planetary albedo is, the larger possibility of aerosol heating effect there is. Also, the smaller backward scattering phase function is, the larger the asymmetry factor is and so the smaller the solar radiation reflected to the space, which is corresponding to the smaller planetary albedo. The same situation is true to summer and autumn seasons, as shown in Table 1. In summer, m_I is only 0.0128, $\tilde{W}_a$ is as small as 0.761, but for Junge distribution with $\gamma^* = 3$, $\tilde{W}_a^*$ is as high as 0.889. Again as shown in Table 2, when m_I increases gradually from zero, $\tilde{W}_a$ and P_π of spring distribution decrease more rapidly than the ones of Junge distribution with $\gamma^* = 3$. For Junge distribution, when $m_I = 0.05$, $\tilde{W}_a^*$ is 0.725, but for the Beijing spring distribution, when $m_I = 0.01$, $\tilde{W}_a^*$ is as small as 0.732 which is quite near the former one. In some papers about the aerosol absorptance and its radiation climate effect (Yamamoto et al., 1972 and Pueschel et al., 1984), the function of aerosol size distribution is analyzed. In this paper, our results show that aerosol distribution and the imaginary part of its refraction index are both very important physical parameters for assessing the aerosol radiation climate effect. Just because the aerosol in Beijing area has wide distribution containing a lot of large particles, aerosol existence also makes the planetary albedo decrease under smaller imaginary part.

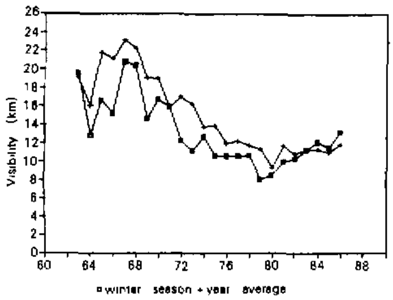

Fig.5. The visibility of year average and winter season average.

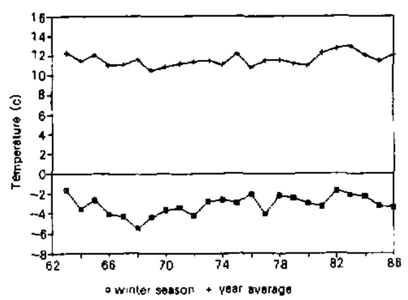

Fig.6. The temperature of year average and winter season average.

Next, the changing tendency of the aerosol concentration at surface and the correlation characteristic between it and the surface temperature are analyzed by using the meteorological data from 1963 to 1986. As far as the first order approximation is concerned, the concentration of aerosol particle is inversely proportional to the horizontal visibility. The larger the visibility is, the smaller the particle concentration is. Fig.5 shows the changing of the visibility of year average and winter season average. Either visibility or temperature used below, only the measurement values at 8:00 a.m. and 2:00 p.m. are concerned. It can be seen from Fig.5 that the visibility of winter season average is lower than year average from the beginning to the end. This illuminates that the winter visibility is the worst. From the visibility annual variation, we can find that the visibility in Beijing in 1967 and 1968 is the best, and then decreases ranging from 20 km in 1960's to 11 km in 1980's, i.e., the aerosol concentration increases nearly one fold.

Fig.6 shows the variation of surface temperature of both year average and winter average from 1963 to 1986. The lowest temperature is found in 1968, and then the temperature increases, having a negative correlation tendency with the visibility variation. Table 3 lists correlation results between the surface temperature and the surface visibility of both month average and year average from 1963 to 1986.

Table.3 The Correlation Factor between the Surface Visibility and Surface Temperature

Month	Jan.	Feb.	March	April	May	June	July
Correlation factor	−0.348	−0.419	−0.297	−0.0094	0.036	0.007	−0.026
Month	Aug.	Sep.	Oct.	Nov.	Dec.	Year average	
Correlation factor	0.046	−0.088	−0.276	−0.41	−0.567	−0.301	

It can clearly be seen from Table 3 that for the winter months of December, January and February there is a negative correlation between the surface temperature and horizontal visibility. The lower the visibility is, the higher the turbidity is and the higher the surface temperature is. This may have relation with the aerosol heating effect in Beijing winter season. Besides, the correlation of year average is also negative. In summer season, the correlation ratio is nearly zero which means that there is no relation between aerosol intensity and surface temperature.

V. CONCLUSIONS

According to above-mentioned theoretical calculation and statistical correlation analysis, we come to the following conclusions.

(1) In winter over Beijing area, aerosol has a heating effect on the atmosphere because it results in the decrease of planetary albedo.

(2) In winter of Beijing, there is a negative correlation between the horizontal visibility and the surface temperature, which means the aerosol particle concentration near the ground has positive correlation with surface temperature. The larger the aerosol concentration is, the higher the temperature is. This may be owing to the aerosol heating effect in winter.

(3) In spring and autumn of Beijing, although the imaginery part of aerosol refraction index is less than 0.022, there is no increase of the planetary albedo caused by aerosol existence. In summer, the aerosol changes the planetary albedo very little. All this may have relations with the wide aerosol size distribution. For the wider aerosol size distribution, smaller imaginary part of refraction index corresponds to less single scattering albedo and larger

asymmetry factor (or smaller backward scattering phase function). So the imaginary part of refraction index and aerosol distribution are both very important parameters to assess the aerosol radiation climate effect.

Any changes of the planetary albedo will change the radiation energy balance. The other feedback process like cloud variation can also change the planetary albedo. In the above calculation model, the other kinds of feedback processes were not dealt with. So the estimated heating and cooling effect of aerosol on the atmosphere is just a prediction of the primary tendency.

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