Research Article

Study on the Influence of Aerosol Radiation Balance in One-Dimensional Atmospheric Medium Using P_n -Approximation Method

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A numerical study, used for the influence of aerosol on the atmospheric radiation transfer, is conducted in this paper. Based on the established atmospheric radiation transfer model, we calculated the effect of species, concentration, and height distribution of aerosols on atmosphere radiation. The calculation results show that the aerosol particles affect the atmospheric radiation balance greatly and thus are the important component of the radiation balance of the earth-troposphere system.

1. Introduction

Atmospheric aerosols are suspensions of small solid or liquid particles with the diameters between 0.001 and 10 μm . Aerosol affects the Earth's radiation budget by direct and indirect effects and plays an important role in global climate systems. The direct effect is that aerosol particles scatter and absorb the solar radiation and surface infrared radiation [1]; the indirect effect is that aerosol particles act as cloud condensation nuclei and ice nuclei [2, 3]. When the climate system is in balance, its absorbing solar radiation energy is accurately equal to the infrared radiation energy that the earth and atmosphere emit into outer space. Any factor that disturbs this balance and then changes the climate is the radiative forcing factor. Its forcing on the earth-troposphere system is the radiative forcing [4, 5]. Radiative forcing (in $\text{W} \cdot \text{m}^{-2}$) is defined as the change in the net radiative flux at the tropopause due to some factors change. Aerosol is a very important correction factor in the atmospheric radiation transfer. So it is significant to study the influence of aerosols particles on the atmospheric transfer.

Many numerical methods have been developed to solve the problem of radiative heat transfer, such as the discrete ordinates method (DOM) [6, 7], finite volume method (FVM) [8], finite element method (FEM), Monte Carlo method (MCM) [9], and spherical harmonics method (SHM). The spherical harmonics method, also known as the P_n -approximation method, was originally proposed by Jeans [10] for interstellar radiative transfer and further developed by Davison [11] and Kourganoff [12] for neutron transport. Khouaja et al. [13] developed the P_n -approximation method for radiative heat transfer. Recently, Larsen et al. [14] used a double- P_n approximation for the transfer equation in neutron transport, but only to low order in the expansion. McClarren et al. [15] analyzed numerical oscillations of the P_n -approximation method and also suggested improvements based on wavelet analysis.

We established atmospheric radiative transfer calculations model based on the spherical harmonics method in one-dimensional media and analyzed the influence of aerosols on the atmospheric radiative transfer. The calculations from the species, number, and distribution height of aerosols were achieved and the results were given in Figure 1.

FIGURE 1: Radiative heat transfer model diagram of one-dimensional atmospheric layer.

2. Numerical Model

The radiative transfer equation for one-dimensional absorbing, emitting, and scattering gray medium can be expressed as

$$\mu \frac{dI(\tau, \mu)}{d\tau} = -I(\tau, \mu) + (1 - \omega) I_b(\tau) + \frac{\omega}{2} \int_{-1}^1 I(\tau, \mu') \Phi(\mu', \mu) d\mu', \quad (1)$$

where τ is the optical thickness given by $\tau = (\kappa_a + \kappa_s)L$, ω is the scattering albedo with $\omega = \kappa_s/(\kappa_a + \kappa_s)$, L is the medium thickness, and κ_a and κ_s are the absorbing and scattering coefficients, respectively. The radiative intensity $I(\tau, \mu)$ and scattering phase function $\Phi(\mu', \mu)$ can be expanded in terms of associated Legendre polynomials $P_n(\mu)$. Using recurrence relations between these polynomials, (1) becomes

$$\begin{aligned} & \frac{d}{d\tau} \left(\sum_{n=0}^N \frac{(n+1) P_{n+1}(\mu) + n P_{n-1}(\mu)}{2n+1} I_n(\tau) \right) \\ &= - \sum_{n=0}^N I_n(\tau) P_n(\mu) + (1 - \omega) I_b(\tau) \\ &+ \sum_{n=0}^N \frac{\omega g_n}{2n+1} I_n(\tau) P_n(\mu), \end{aligned} \quad (2)$$

where g_n is an asymmetry parameter. Multiplying (2) by $P_n(\mu)$ and integrating over $\mu \in [-1, 1]$, we obtain using the orthogonality properties of the Legendre polynomials $N + 1$ differential equations for coefficients $I_n(\tau)$ that comprise

the $N + 1$ P_n -approximation equations. Consider the following:

$$\begin{aligned} & \frac{1}{3} \frac{dI_1(\tau)}{d\tau} + (1 - \omega g_0) I_0(\tau) = (1 - \omega) I_b(\tau) \quad n = 0 \\ & \frac{2}{5} \frac{dI_2(\tau)}{d\tau} + \frac{dI_0(\tau)}{d\tau} + \left(1 - \frac{1}{3} \omega g_1\right) I_1(\tau) = 0 \quad n = 1 \\ & \dots \\ & \frac{n+1}{2n+3} \frac{dI_{n+1}(\tau)}{d\tau} + \frac{n}{2n-1} \frac{dI_{n-1}(\tau)}{d\tau} + \left(1 - \frac{\omega g_n}{2n+1}\right) I_n(\tau) \\ &= (1 - \omega) I_b(\tau) \delta_{0n} \quad n = 2, 3, \dots, N. \end{aligned} \quad (3)$$

For one-dimensional opaque diffuse emitting and reflecting interfaces, odd terms in the Legendre polynomials form an orthogonal set. The boundary conditions are given by

$$\begin{aligned} & \int_{\vec{n} \cdot \vec{s} > 0} I(\tau_{W=0}, \vec{s}) P_{2n+1}(\vec{s}) d\Omega \\ &= \varepsilon_W \int_{\vec{n} \cdot \vec{s} > 0} I_W(0, \vec{s}) P_{2n+1}(\vec{s}) d\Omega \\ &+ \int_{\vec{n} \cdot \vec{s} > 0} (1 - \varepsilon_W) \frac{1}{2\pi} \int_{\vec{n} \cdot \vec{s}' < 0} I(\tau_W, \vec{s}') d\Omega' P_{2n+1}(\vec{s}) d\Omega, \end{aligned} \quad (4)$$

where ε is the emissivity. The boundary conditions imposed on the spherical harmonics functions are given by

$$\begin{aligned} & \sum_{n=0}^N I_n(0) \left[\int_0^1 P_n(\mu) \mu^{2n+1} d\mu \right. \\ & \quad \left. - \rho^d \int_0^1 \int_{-1}^0 P_n(\mu') \mu^{2n+1} d\mu' d\mu \right] \\ &= I_{b0} (1 - \rho_0^d) \int_0^1 \mu^{2n+1} d\mu \\ & \sum_{n=0}^N I_n(\tau_L) \left[\int_{-1}^0 P_n(\mu) \mu^{2n+1} d\mu \right. \\ & \quad \left. - \rho^d \int_{-1}^0 \int_0^1 P_n(\mu') \mu^{2n+1} d\mu' d\mu \right] \\ &= I_{b\tau_L} (1 - \rho_0^d) \int_{-1}^0 \mu^{2n+1} d\mu \end{aligned} \quad (5)$$

$$n = 0, 1, \dots, \frac{N-1}{2}.$$

The boundary conditions [16] are applicable in the one-dimensional case:

$$\begin{aligned} & \int_{\vec{n} \cdot \vec{s} > 0} I(\tau_{W=0}, \vec{s}) P_{2n+1}(\vec{s}) d\Omega \\ &= (1 - \rho_0^d) \int_{\vec{n} \cdot \vec{s} > 0} I^0(\vec{s}) P_{2n+1}(\vec{s}) d\Omega \\ &+ \int_{\vec{n} \cdot \vec{s} > 0} \rho^d \frac{1}{2\pi} \int_{\vec{n} \cdot \vec{s}' < 0} I(\tau_W, \vec{s}') d\Omega' P_{2n+1}(\vec{s}) d\Omega, \end{aligned} \quad (6)$$

where $I^0(\vec{s})$ is the circumstance radiation, ρ_0^d and ρ^d are reflectivities of the two sides of the boundary surfaces, respectively, when considering total reflection, and the values of these depend on the medium refractive index n .

Separating the large refractive index from small, the diffuse reflectance at the boundary surfaces is given by Siegel and Spuckler [17]:

$$\begin{aligned} \rho(n) \equiv F(n) &= \frac{1}{2} + \frac{(3n+1)(n-1)}{6(n+1)^2} \\ &+ \frac{n^2(n^2-1)^2}{(n^2+1)^3} \ln\left(\frac{n-1}{n+1}\right) - \frac{2n^3(n^2+2n-1)}{(n^2+1)(n^4-1)} \\ &+ \frac{8n^4(n^4+1)}{(n^2+1)(n^4-1)^2} \ln(n). \end{aligned} \quad (7)$$

In contrast, separating small refractive index from large, the diffuse reflectance at the boundary surfaces is given by

$$\rho(n) = 1 - \frac{1}{n^2} [1 - F(n)], \quad (8)$$

where $n = (n_h/n_s)$, $n_h \geq n_s$. In treating a semitransparent interface, we can use opaque interface treatment as reference.

Considering one-dimensional unsteady coupled radiative and conductive heat transfer, the general form of the energy equation of the internal radiation and the transient coupled conductive heat transfer is defined as

$$\rho c \frac{\partial T}{\partial t} = \text{div}(k_c \text{grad} T) - \text{div} \mathbf{q}^r + \phi. \quad (9)$$

Considering unsteady state, no internal heat source and coupled radiative and conductive heat transfer, the energy equation (9) can be discretized as follows:

$$\begin{aligned} & \rho c \Delta x_i \frac{T_i^{m+1} - T_i^m}{\Delta t} \\ &= \frac{k_e (T_{i+1}^{m+1} - T_i^{m+1}) + k_w (T_i^{m+1} - T_i^{m+1})}{\Delta x_i} + \Phi_{i,m}^r. \end{aligned} \quad (10)$$

In order to make the problem have universal significance, the dimensionless form of the energy equation (10) will be evaluated. Introducing the following definitions: reference temperature T_{rf} , the dimensionless temperature $\tilde{T}_i^m = T_i^m/T_{rf}$, unit heat capacity $C = \rho c$, the dimensionless time $\Delta \tilde{t} = (4\sigma n^2 T_{rf}^3/CL)\Delta t$, $\tilde{t} = (4\sigma n^2 T_{rf}^3/CL)t$, the dimensionless length $\Delta X_i = \Delta x_i/L$, the radiative and conductive parameter $N = \lambda/4\sigma n^2 T_{rf}^3 L$, the dimensionless convective heat transfer coefficient $H = h/4\sigma n^2 T_{rf}^3$, the dimensionless radiative heat flux $\tilde{\Phi} = \Phi/(4\sigma n^2 T_{rf}^4)$, (10) becomes

$$\begin{aligned} & \Delta X_i \frac{\tilde{T}_i^{m+1} - \tilde{T}_i^m}{\Delta \tilde{t}} \\ &= \frac{N_e (\tilde{T}_{i+1}^{m+1} - \tilde{T}_i^{m+1}) + N_w (\tilde{T}_{i-1}^{m+1} - \tilde{T}_i^{m+1})}{\Delta X_i} + \tilde{\Phi}_{i,m}^r. \end{aligned} \quad (11)$$

According to [18], in considering one-dimensional non-steady coupled radiative and conductive heat transfers, the thermal radiation source term $\Phi^r(T)$ is found to be

$$\Phi^r(T) = \int_{\Delta V} -\text{div} \mathbf{q}^r dV = \int_{\Delta V} -\kappa_a [4\pi I_b(\tau) - G(\tau)] dV. \quad (12)$$

Here, the incident radiative power $G(\tau) = \int_{\Omega} I(\tau, \Omega) d\Omega$ can be transformed to $G(\tau) = 2\pi \int_{-1}^1 I(\tau, \mu) d\mu$. According to the properties of Legendre polynomials and $P_0(\mu) = 1$, the above equation is rearranged into the form:

$$G(\tau) = 2\pi \int_{-1}^1 I_0(\tau) d\mu. \quad (13)$$

Meanwhile, considering the physical properties of the inhomogeneous medium, we next develop a P_n -approximation method for a multilayer medium. We redetermine the boundary conditions of each layer. The modified boundary conditions are

$$\begin{aligned} & \sum_{n=0}^N I_n(0) \left[\int_0^1 P_n(\mu) \mu^{2i+1} d\mu - \rho^d \int_0^1 \int_{-1}^0 P_n(\mu') \mu^{2i+1} d\mu' d\mu \right] \\ &= \sum_{n=0}^N I_{n,b0} (1 - \rho_0^d) \int_0^1 P_n(\mu) \mu^{2i+1} d\mu, \\ & \sum_{n=0}^N I_n(\tau_L) \left[\int_{-1}^0 P_n(\mu) \mu^{2i+1} d\mu - \rho^d \int_{-1}^0 \int_0^1 P_n(\mu') \mu^{2i+1} d\mu' d\mu \right] \\ &= \sum_{n=0}^N I_{n,b\tau_L} (1 - \rho_0^d) \int_{-1}^0 P_n(\mu) \mu^{2i+1} d\mu, \\ & i = 0, 1, \dots, \frac{N-1}{2}. \end{aligned} \quad (14)$$

TABLE 1: The main input parameters of the atmospheric radiation calculation.

	Parameters	Values
Initial condition	Initial temperature T_0	280 K
	Reference temperature T_{rf}	1000 K
	Dimensionless time t^*	1×10^6
	Relaxation factor f_{re}	0.01
Boundary condition	Terrestrial temperature $T_{-\infty}$	280 K
	Terrestrial emissivity $\varepsilon_{-\infty}$	0.7
	Solar radiation temperature $T_{+\infty}$	5762 K
	Angle factor $A_{+\infty, 2}$	5.5844×10^{-6}
Media physical properties	Layers of height	7 layers, seen in Table 3
	Bands absorption coefficient	12 bands, seen in Table 4
	Thermal conductivity k_c	$0.023 \text{ W} \cdot \text{m}^{-1} \cdot \text{K}^{-1}$
Aerosol properties	Distribution function	Multilognormal distributions
	Height distribution	0-1 km

TABLE 2: Multilayer atmospheric radiative transfer model.

Layer number	Thickness (km)	Upper-lower limit of height (km)	Density (kg/m^3)	Dimensionless thermal conductivity	Refractive index
1	1	0~1	1.15566	$7.51e-8$	1.0
2	2	1~3	0.96517	$5.89e-8$	1.0
3	6	3~9	0.60438	$3.74e-8$	1.0
4	8	9~17	0.30815	$1.43e-8$	1.0
5	10	17~27	0.02573	$3.23e-9$	1.0
6	11	27~38	0.010076	$6.90e-10$	1.0
7	12	38~50	0.002195	$4.23e-10$	1.0

Here, $I_{n,b0}$ and I_{n,bT_λ} are the emission radiation intensities of two adjacent layers, and the initial values are obtained from the temperature field. The total emission radiation intensity is determined by an iterative calculation of intensities from successive layers, the last providing the required total. The convergence condition of the multilayer model calculation is decided by the radiative intensity I_n . After the intensity calculation has converged, the iterative calculation of the temperature field is then performed.

3. Results and Discussion

According to statistics, 98.8% of the total atmospheric mass distributes in the atmosphere within 30 km above the ground, and with the height increasing to 50 km the proportion reaches 99.92%. Most of the atmospheric radiation (involving solar radiation, absorption, and scattering of the ground infrared radiation and the emission of the atmosphere itself) occurs within certain atmospheric thickness. Compared with the earth radius (6378 km), the atmospheric thickness can be regarded as one-dimensional radiative media. Therefore, we treated the atmosphere within 50 km above the ground as a multilayer one-dimensional media distributing along the radial of the earth. The upper surface of the medium layer is near to space and is considered as a semitransparent medium. The lower interface is close to ground and is considered as opaque interface. The emissivity and radiative temperature

related to the ground are given. For the inner of the medium layer we mainly consider the gas absorption and emission of the radiation energy and the aerosol particles scattering and absorption of the radiation energy and consider the atmospheric coupled radiation and conduction in the vertical direction.

The atmospheric convection is very important for the energy transfer of the atmosphere. But we mainly discuss the influence of aerosols on the atmospheric radiative energy and investigate the atmospheric radiative balance and the variation in radiation characteristic, which is not affected by the atmospheric convection that acts only as the background. Therefore, we ignore the influence of the atmospheric convection.

The atmospheric radiation physical model is shown in Figure 1 and the atmospheric radiation transfer model is shown in Figure 2. The atmospheric internal heat source is set as $\phi = 0$.

The standard atmospheric condition is given in Tables 1–4.

Table 1 gives the main input parameters for the calculation of atmospheric radiation transfer.

Table 2 gives the parameters associated with layers divided along the vertical height. The density of each layer is determined by the calculation of the pressure. We consider $(\rho/\rho_0) = (p/p_0)$ (the standard atmosphere pressure p is

FIGURE 4: Influence of scattering phase function change on atmospheric temperature.

TABLE 3: Spectral bands for the atmospheric radiative transfer model and radiative characteristic parameters under the standard dust aerosol in the height of 0-1 km.

Band number	Wavelength range (μm)	Optical thickness	Albedo
1	0.2~0.5	92.92	0.86
2	0.5~0.8	12.30	0.88
3	0.8~1.25	4.42	0.86
4	1.25~1.59	43.28	0.80
5	1.59~2.22	4.27	0.48
6	2.22~3.45	23.65	0.07
7	3.45~5.0	93.93	0.02
8	5.0~10.0	353.22	0.00
9	10.0~11.11	12.51	0.11
10	11.11~12.99	1.62	0.45
11	12.99~20.0	35.29	0.01
12	20.0~50.0	951.33	0.00

particle size distribution function is the multilognormal distribution.

From the results in the figure we can see that as the different scattering effect of the aerosol particles, their influence on the temperature is greatly different. Compared with dust aerosol, the marine and water-soluble aerosol particles have stronger scattering effect. They will scatter upwards more radiation energy and decrease the atmospheric temperature. The soot particles have stronger absorption and their existence will cause the atmospheric temperature increasing.

3.2. The Influence of the Scattering Phase Function Variation. The aerosol particles scattering characteristics are considered

as forward and back scatter (the scattering phase functions are $\Phi_1(\Theta)$ and $\Phi_2(\Theta)$, resp.; see (16) and (17)). Other parameters are not changed. The results are shown in Figure 4:

$$\Phi_1(\Theta) = 1 + P_1(\cos \Theta), \quad (16)$$

$$\Phi_2(\Theta) = 1 - P_1(\cos \Theta). \quad (17)$$

It can be seen from Figure 4(a) that, compared with the anisotropic scattering, the variation in the scattering phase functions causes the atmospheric temperature to change. But the effect is not obvious. Figure 4(b) shows the difference value between the atmospheric temperature fields and the isotropic scattering with different scattering phase functions. From the figure we can see that the forward and back scatters have opposite effect on the atmospheric temperature fields but the effect degree is similar.

3.3. The Influence of the Aerosol Particles Concentration. The aerosol particles transport with the tropospheric movement will cause the aerosol particles concentrations to vary. In Figure 5 we consider the influence of the variation in total particles number concentrations, and the particles number concentrations are increased or decreased by 50%. So the radiation characteristics of the aerosol particles are changed, the atmospheric energy balance is affected, and then the atmospheric temperature fields are influenced. From the results it can be seen that the variation in aerosol particles concentrations can affect the atmospheric temperature but not obviously compared with other factors.

3.4. The Influence of the Height Distribution of Aerosol Particles. We investigate the effect of the medium layer height on the aerosol particles size distribution; that is, the atmospheric radiation temperature fields change with the aerosol particles

TABLE 4: Complex refractive index of different wavelength for different aerosol types [19].

λ	Dust		Water-soluble		Marine		Soot	
	n	k	n	k	n	k	n	k
0.200	1.530	0.070	1.530	0.070	1.429	$2.87E-5$	1.500	0.350
0.250	1.530	0.030	1.530	0.030	1.404	$1.45E-6$	1.620	0.450
0.300	1.530	0.008	1.530	0.008	1.395	$5.83E-6$	1.740	0.470
0.337	1.530	0.008	1.530	0.005	1.392	$1.20E-7$	1.750	0.470
0.400	1.530	0.008	1.530	0.005	1.385	$9.90E-9$	1.750	0.460
0.488	1.530	0.008	1.530	0.005	1.382	$6.41E-9$	1.750	0.450
0.515	1.530	0.008	1.530	0.005	1.381	$3.70E-9$	1.750	0.450
0.550	1.530	0.008	1.530	0.006	1.381	$4.26E-9$	1.750	0.440
0.633	1.530	0.008	1.530	0.006	1.377	$1.62E-8$	1.750	0.430
0.694	1.530	0.008	1.530	0.007	1.376	$5.04E-8$	1.750	0.430
0.860	1.520	0.008	1.520	0.012	1.372	$1.09E-6$	1.750	0.430
1.060	1.520	0.008	1.520	0.017	1.367	$6.01E-5$	1.750	0.440
1.300	1.460	0.008	1.510	0.020	1.365	$1.41E-4$	1.760	0.450
1.536	1.400	0.008	1.510	0.023	1.359	$2.43E-4$	1.770	0.460
1.800	1.330	0.008	1.460	0.017	1.351	$3.11E-4$	1.790	0.480
2.000	1.260	0.008	1.420	0.008	1.347	0.00107	1.800	0.490
2.250	1.220	0.009	1.420	0.010	1.334	$8.50E-4$	1.810	0.500
2.500	1.180	0.009	1.420	0.012	1.309	0.00239	1.820	0.510
2.700	1.180	0.013	1.400	0.055	1.249	0.0156	1.830	0.520
3.000	1.160	0.012	1.420	0.022	1.439	0.197	1.840	0.540
3.200	1.220	0.010	1.430	0.008	1.481	0.0669	1.860	0.540
3.392	1.260	0.013	1.430	0.007	1.439	0.0151	1.870	0.550
3.500	1.280	0.011	1.450	0.005	1.423	0.00717	1.880	0.560
3.750	1.270	0.011	1.452	0.004	1.398	0.0029	1.900	0.570
4.000	1.260	0.012	1.455	0.005	1.388	0.00369	1.920	0.580
4.500	1.260	0.014	1.460	0.013	1.377	0.00997	1.940	0.590
5.000	1.250	0.016	1.450	0.012	1.366	0.00957	1.970	0.600
5.500	1.220	0.021	1.440	0.018	1.333	0.00931	1.990	0.610
6.000	1.150	0.037	1.410	0.023	1.306	0.0796	2.020	0.620
6.200	1.140	0.039	1.430	0.027	1.431	0.0691	2.030	0.625
6.500	1.130	0.042	1.460	0.033	1.374	0.0294	2.040	0.630
7.200	1.400	0.055	1.400	0.070	1.343	0.0249	2.060	0.650
7.900	1.150	0.040	1.200	0.065	1.324	0.0279	2.120	0.670
8.200	1.130	0.074	1.010	0.100	1.324	0.0308	2.130	0.680
8.500	1.300	0.090	1.300	0.215	1.336	0.0336	2.150	0.690
8.700	1.400	0.100	2.400	0.290	1.366	0.0356	2.160	0.690
9.000	1.700	0.140	2.560	0.370	1.373	0.0365	2.170	0.700
9.200	1.720	0.150	2.200	0.420	1.356	0.0371	2.180	0.700
9.500	1.730	0.162	1.950	0.160	1.339	0.0368	2.190	0.710
9.800	1.740	0.162	1.870	0.095	1.324	0.0388	2.200	0.715
10.000	1.750	0.162	1.820	0.090	1.310	0.0406	2.210	0.720
10.591	1.620	0.120	1.760	0.070	1.271	0.0522	2.220	0.730
11.000	1.620	0.105	1.720	0.050	1.246	0.0731	2.230	0.730
11.500	1.590	0.100	1.670	0.047	1.227	0.105	2.240	0.740
12.500	1.510	0.090	1.620	0.053	1.208	0.190	2.270	0.750
13.000	1.470	0.100	1.620	0.055	1.221	0.223	2.280	0.760
14.000	1.520	0.085	1.560	0.073	1.267	0.271	2.310	0.775
14.800	1.570	0.100	1.440	0.100	1.307	0.292	2.330	0.790

TABLE 4: Continued.

λ	Dust		Water-soluble		Marine		Soot	
	n	k	n	k	n	k	n	k
15.000	1.570	0.100	1.420	0.200	1.321	0.297	2.330	0.790
16.400	1.600	0.100	1.750	0.160	1.407	0.331	2.360	0.810
17.200	1.630	0.100	2.080	0.240	1.487	0.341	2.380	0.820
18.000	1.640	0.115	1.980	0.180	1.525	0.341	2.400	0.825
18.500	1.640	0.120	1.850	0.170	1.536	0.339	2.410	0.830
20.000	1.680	0.220	2.120	0.220	1.560	0.324	2.450	0.850
21.300	1.770	0.280	2.060	0.230	1.568	0.318	2.460	0.860
22.500	1.900	0.280	2.000	0.240	1.579	0.316	2.480	0.870
25.000	1.970	0.240	1.880	0.280	1.596	0.313	2.510	0.890
27.900	1.890	0.320	1.840	0.290	1.612	0.320	2.540	0.910
30.000	1.800	0.420	1.820	0.300	1.614	0.320	2.570	0.930
35.000	1.900	0.500	1.920	0.400	1.597	0.383	2.630	0.970
40.000	2.100	0.600	1.860	0.500	1.582	0.561	2.690	1.000

FIGURE 5: Influence of the aerosol concentration on atmospheric temperature.

being in different height medium layers. The calculation conditions are as follows: the aerosol particles distribute in the height of 0~1 km, 1~3 km, and 3~9 km, respectively. The calculation results are in Figure 6.

From the calculation results in Figure 6 it can be seen that, with the aerosol particles transporting to the high altitude, the atmospheric temperature is affected greatly if the radiation characteristics of the aerosol particles have no changes. With the distribution height of the aerosol particles increasing, the medium temperature above the medium layer in which the aerosol particles exist increases greatly but the below temperature decreases obviously. This phenomenon shows the scattering effect of the aerosol particles on the

solar radiation energy, that is, as the scattering effect of the atmospheric aerosols, most of the solar radiation energy to the ground is scattered back to the high altitude and cannot reach lower atmosphere, which causes the lower atmospheric temperature to decrease to about 200 K.

4. Conclusion

We analyzed the influence of the aerosol particles characteristics on the atmospheric radiation transfer model. The results showed the different effect of aerosol types on the atmospheric radiation. The dust, marine, and water-soluble

FIGURE 6: Influence of temperature fields on different aerosol distribution altitude.

aerosols had negative effects on the atmospheric absorption, but the soot aerosol could increase the atmospheric absorption. The scattering phase functions affected the ground temperature. The variation of the aerosol particles distribution height could determine the atmospheric temperature; that is, the ground temperature decreased greatly with the aerosol particles existing in the high altitude stable.

Conflict of Interests

The authors declare that there is no conflict of interests regarding the publication of this paper.

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