A Method to Determine Atmospheric Aerosol Optical Depth Using Total Direct Solar Radiation

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ABSTRACT

Based on sensitivity of total direct solar radiation (DSR) detected by a pyrheliometer for aerosol physical–optical properties, a method is proposed to retrieve the 0.75-μm aerosol optical depth from the radiation, and an iteration inversion algorithm and a parameterized DSR model are developed. A key question of the method is the effect of aerosol size distribution uncertainty on the depth solution. As shown in inversion simulations using Junge size distributions and LOWTRAN7 aerosol models, a depth accuracy better than 5% can generally be expected for a solar zenith angle less than 75° if a Junge distribution with exponent \( n = 3 \) is selected for the retrievals. In general, the smaller the depth and solar zenith angle, the higher the accuracy. In addition, it is important for improving solution accuracy to exactly determine the DSR and water vapor absorptance. If errors in DSR and the vertical water vapor amount are within ±2% and ±0.2 cm, resultant depth errors are usually within ±0.02 and ±0.013, respectively. The method is tested for comparative observations using a pyrheliometer and sunphotometer. There are 1267 sets of comparative aerosol optical depths measured over 25 days. Results show that the 0.75-μm aerosol optical depths measured by the pyrheliometer conform well with those of the sunphotometer. Standard deviation of the total 1267 sets of optical depth is 10.2%, and the difference between two average optical depths is only 1.2%.

1. Introduction

Atmospheric aerosol optical depth is an important parameter for studies of atmospheric pollution, aerosol radiation–climate effects, atmospheric correction in remote sensing from space, and so on. Based on light scattering by the atmosphere, optical methods for measuring aerosol physical–optical parameters may be broadly classified into two categories: passive and active (Deirmendjian 1980). As far as the aerosol optical depth measurement is concerned, the passive spectral extinction method is one of the most reliable. The extinction method generally employs a sunphotometer to detect narrowband direct solar radiation and then derive atmospheric optical depth from the radiation. It was introduced by S. P. Langley and used by C. G. Abbot and colleagues at the Smithsonian Astrophysical Observatory (Abbot et al. 1902–1932). In recent years, the method has been used considerably to measure total column aerosol optical depth by ground-based sunphotometers (Murai 1967; Bulrich et al. 1969; Quenzel 1970; Shaw et al. 1973; King et al. 1978; Nakajima et al. 1983; Qiu et al. 1985; Qiu and Jinhui 1994). The method has also been used to measure global-scale stratospheric aerosol (including its optical depth) by spaceborne sunphotometers, and of particular significance in this respect are SAM II and SAGE observations (McCormick and Verga 1992; Kent and McCormick 1984; Lenoble and Brogniez 1985; Yue et al. 1991). The focus of this paper is on a technique based on similar principles as in the method, that is, a technique to determine aerosol optical depth (or atmospheric turbidity), for total direct solar radiation detected using ground-based pyrheliometers.

The technique is based on total extinction of solar radiation by the atmosphere. In fact, pyrheliometer measurements of the atmospheric turbidity have been made for a long time (Ångström 1964; Kondratyev 1969; WMO 1983), and there are different turbidity coefficients derived from total or wideband direct solar radiation, such as the Linke turbidity factor and the Ångström and Schüepp turbidity coefficients. The Linke factor, introduced by F. Linke in 1922, has been widely used in meteorological observatories (WMO 1981, 1983). It is defined as the ratio of the average (extinction equivalent for the whole solar spectrum) total atmospheric optical depth (including aerosol, molecular scattering, and water vapor) to the average molecular scattering optical depth. For characterizing the total extinction by the atmosphere, the factor is an appropriate measure (Kasten 1980), but it cannot be used as a quantitative indicator of column aerosol content or its optical depth because it depends not only on aerosol...
content, but also on other atmospheric parameters such as water amount. Ångström adopted a power law to analyze dependence of aerosol optical depth on wavelength and defined the depth at a wavelength of 1 μm as the turbidity coefficient (Ångström 1929, 1964). The Schüepp (1949) turbidity coefficient is related to the 0.5-μm aerosol optical depth. Both coefficients are usually used in measurements of wideband (not total) direct solar radiation, and they are derived usually under some assumptions on the relationship between aerosol optical depth and wavelength. There are many meteorological observatories around the world having long-term pyrheliometer observations of total direct solar radiation (DSR). So, it is very significant to develop a suitable method to determine aerosol optical depth from the observation data. In this paper, based on sensitivity of total DSR to aerosol physical–optical parameters, a method is proposed to determine the 0.75-μm aerosol optical depth using the DSR, and its accuracy is analyzed according to numerical simulations and comparative measurements by sunphotometer and pyrheliometer.

2. Sensitivity of DSR to aerosol physical–optical parameters

The λ-wavelength direct solar radiation \( S(\lambda) \) reaching the surface is

\[
S(\lambda) = S_0(\lambda)e^{-\tau_r(\lambda)+\tau_m(\lambda)m(\theta_0)},
\]

where \( S_0(\lambda) \) is the extraterrestrial solar irradiance at wavelength \( \lambda \), \( \theta_0 \) is solar zenith angle, \( m(\theta_0) \) is airmass, \( \tau_r(\lambda) \) is the λ-wavelength column optical depth of the atmospheric aerosol, and \( \tau_m(\lambda) \) is total molecular optical depth. At \( \theta_0 < 75^\circ \), \( m(\theta_0) \approx \sec(\theta_0) = 1/\mu_0 \), where \( \mu_0 \) is the cosine of solar zenith angle.

Derived from Eq. (1), the formula of the spectral extinction method to determine total column aerosol optical depth is

\[
\tau_r(\lambda) = \log[S_0(\lambda)/S(\lambda)]/m(\theta_0) - \tau_m(\lambda).
\]

Assuming no forward scattered light and a plane-parallel atmosphere, DSR \( S \) detected by a pyrheliometer on the ground can be expressed as

\[
S = \int_{\lambda_1}^{\lambda_2} S_\lambda T_a(\lambda, \theta_0) T_m(\lambda, \theta_0) d\lambda
= \int_{\lambda_1}^{\lambda_2} S_\lambda e^{-\tau_r(\lambda)m(\theta_0)} T_m(\lambda, \theta_0) d\lambda,
\]

where \( \lambda_1 \) and \( \lambda_2 \) are lower and upper spectral limits of the pyrheliometer, and \( T_a \) and \( T_m \) are aerosol and molecular spectral transmittances. In this paper, \( T_a \) as for LOWTRAN aerosol model and \( T_m \) are calculated using LOWTRAN7 with a step of 20 cm⁻¹ (Kneizys et al. 1988). In the case of no LOWTRAN data, \( T_m \) is calculated using the expression, \( \exp[-\tau_m(\lambda)m(\theta_0)] \), where \( m(\theta_0) \) is determined by Kasten’s (1966) empirical formula.

The pyrheliometer usually has a spectral response range from 0.3 to 4 μm with 97.8% of the solar radiation within this range (Thekaekara 1973). Most of the solar radiation for \( \lambda < 0.3 \mu m \) or \( \lambda > 4 \mu m \) is absorbed by trace gases before reaching the ground, so the radiation detected by the pyrheliometer can generally be regarded as total direct solar radiation. In this paper, \( \lambda_1 = 0.3 \mu m \) and \( \lambda_2 = 4 \mu m \) are adopted.

Define

\[
R = \int_{\lambda_1}^{\lambda_2} S_\lambda(\lambda) d\lambda/S_0,
\]

\[
t_m = \int_{\lambda_1}^{\lambda_2} S_\lambda(\lambda) T_m(\lambda, \theta_0) d\lambda/\int_{\lambda_1}^{\lambda_2} S_\lambda(\lambda) d\lambda,
\]

\[
G = G(\lambda_0, T_m(\lambda, \theta_0), T_m(\lambda, \theta_0)) = e^{-\tau_r(\lambda_0)m(\theta_0)} S \int_{\lambda_1}^{\lambda_2} S_\lambda T_m(\lambda, \theta_0) d\lambda,
\]

where \( S_0 \) is the solar constant, \( R \) is the ratio between the solar radiation in the range of \( \lambda_1 \leq \lambda \leq \lambda_2 \) and \( S_0 \), and \( t_m \) stands for the average molecular transmittance in the range.

Combining Eqs. (3)–(6), we have

\[
S = RS_0 t_m G \exp[-\tau_m(\lambda_0)m(\theta_0)].
\]

Now \( S \) is transformed as a product of five factors. The first three factors, \( R, S_0, \) and \( t_m \), are independent of the aerosol. The factor \( \exp[-\tau_m(\lambda_0)m(\theta_0)] \) is the aerosol transmittance at \( \lambda_0 \) wavelength. Only \( G \) depends on aerosol optical depths \( \tau_r(\lambda) \) in the range of \( \lambda_1 \leq \lambda \leq \lambda_2 \). The dependence of \( \tau_r(\lambda) \) on wavelength is related to aerosol size distribution and its refractive index, especially the distribution. So it is impossible to retrieve wavelength-dependent aerosol optical depth \( \tau_r(\lambda) \) only from the direct solar radiation \( S \). However, if there is a constant wavelength, \( \lambda_0 \), which can make the function \( G \), defined in formula (6), independent or weak-dependent on \( \tau_r(\lambda) \), then according to Eq. (7) the \( \lambda_0 \) wavelength aerosol optical depth \( \tau_r(\lambda_0) \) can be derived from \( S \) in the case of known molecular property.

Next, the sensitivity of \( G \) to aerosol size distribution is studied. In the following study six atmospheric models defined in LOWTRAN7 are used

1) Model 1: tropical atmosphere
2) Model 2: midlatitude summer
3) Model 3: midlatitude winter
4) Model 4: stratospheric summer
5) Model 5: stratospheric winter

In these six models, total column \( H_2O \) amount changes from 0.328 cm (Model 5) to 3.32 cm (Model 1) and \( O_3 \) amount from 0.277 to 0.376 cm.
According to midvalue theorem, Eq. (3) can be changed into

\[ S = \int_{\lambda_1}^{\lambda_2} S_0(\lambda) [1 - \tau_0(\lambda, \theta_0)] T_m(\lambda, \theta_0) \, d\lambda \]

\[ = \int_{\lambda_1}^{\lambda_2} S_0(\lambda) T_m(\lambda, \theta_0) \, d\lambda, \]  

(8)

where \( \tau_0(\lambda, \theta_0) \) can be regarded as radiation-equivalent mean aerosol optical depth and \( \lambda \) represents the wavelength corresponding with depth.

Combining Eq. (6) with Eq. (8) leads to

\[ G = \exp[\tau_0(\lambda_m, \theta_0) - \tau_0(\lambda, \theta_0)]. \]  

(9)

So, if \( \lambda_0 = \lambda_m \), \( G = 1 \). If \( \lambda_m \) is close to a constant \( \lambda \) for any atmospheric condition, \( G \) in the condition of \( \lambda_0 = \lambda \) could approximately be equal to unity, and then \( \lambda_0 \)-wavelength aerosol optical depth can be determined using Eq. (7).

Next, the variation range of \( \lambda_m \) for the Junge size (power law) distribution and constant aerosol refractive index is estimated according to Table 1. Under this condition the aerosol distribution \( n(r) \) and its optical depth are given by

\[ n(r) = C r^{-(\nu^* + 1)}, \]  

(10)

\[ \tau_0(\lambda) = \beta \lambda^{-\nu^*}, \]  

(11)

where \( r \) is aerosol particle radii, and \( \beta \) is known as the Ångström turbidity coefficient being equal to the aerosol optical depth at \( \lambda = 1 \) \( \mu m \). A typical value of the Junge distribution parameter \( \nu^* \) is 3 (Junge 1963). As shown later, if aerosol size distributions used in LOWTRAN7 are fitted to the Junge distribution, their \( \nu^* \) generally vary between 2 and 3.5. In this paper, \( \nu^* \) ranging from 2 to 3.5 is emphasized.

For each of the six atmospheric models, 480 sets of values \( S \) and \( \lambda_m \) are simulated to yield Table 1. Input parameters on Junge distributions and \( \theta_0 \) in the 480 simulations are \( \theta_0 = 0^\circ, 50^\circ, \) and \( 75^\circ; \) \( \nu^* = 1.9 + 0.1n, n = 1, \cdots, 16; \beta = 0.05m, m = 1, \cdots, 10. \) This implies that \( \nu^* \) changes from 2 to 3.5 with a step of 0.1 and \( \beta \) from 0.05 to 0.5 with a step of 0.05. In Table 1, \( \overline{\lambda}_m \) and \( \Delta \lambda_m \) are the mean value and the standard deviation of \( \lambda_m \), which are given by

\[ \overline{\lambda}_m = \frac{480}{\sum_{i=1}^{480} \lambda_0^i / 480}, \]  

(12)

\[ \Delta \lambda_m = \frac{\sqrt{\sum_{i=1}^{480} (\lambda_0^i - \overline{\lambda}_m)^2 / 480}}, \]  

(13)

where \( \lambda_0^i \) is the \( i \)-th \( \lambda_m \) among the 480 cases.

As shown in Table 1, mean values of \( \lambda_m \) change from 0.738 to 0.783 for six atmospheric models in LOWTRAN7, and standard deviations \( \Delta \lambda_m \) are all \( \leq 0.054 \). About half of the solar radiation is within the wavelength range of 0.732 \( \mu m \) (Thekaekara 1973). It is of interest that \( \lambda_m \) is quite close to the 0.732-\( \mu m \) wavelength. In the case of no aerosol, the attenuation of the direct solar radiation reaching the surface in the range of 0.732 \( \mu m \) is caused mainly by molecular scattering and ozone absorption, and water vapor absorption contributes dominant attenuation to the solar radiation in the range of 0.732 \( \mu m \). For the \( U.S. Standard Atmosphere \) (Model 6), these two attenuation contributions are 9.6% and 12.8% for \( \theta_0 = 0^\circ, 13.3\% \) and 14.7% for \( \theta_0 = 50^\circ, \) and 23.9% and 19.4% for \( \theta_0 = 75^\circ, \) respectively, both attenuations being basically matched. Aerosol extinction of the direct solar radiation in the range of 0.732 \( \mu m \) is usually larger than that of 0.732 \( \mu m \) because of a decreasing trend of aerosol optical depth with wavelength. Therefore, it can be understood that \( \lambda_m \) is generally a little larger than 0.732 \( \mu m \). The smaller water vapor amount is, and the larger \( \beta \) and \( \nu^* \) are, the larger the value \( \lambda_m \). As a result, \( \lambda_m \) is usually within 0.75 \( \mu m \pm 0.1 \mu m \), and the 0.75-\( \mu m \) aerosol optical depth is approximately equal to the radiation-equivalent mean aerosol optical depth. Latter, function \( G(\lambda_m, T_a, \theta_0, T_{s_a}(\lambda, \theta_0)) \) for \( \lambda_0 = 0.75 \mu m \) is called \( G_o \).

For the Junge distribution, as long as the 0.75-\( \mu m \) aerosol optical depth and the value \( \nu^* \) are known, \( T_{s_a}(\lambda, \theta_0) \) can be determined for any \( \lambda \) and \( \theta_0 \), so, if \( G_o \) is not sensitive to \( \nu^* \), it is possible to retrieve the depth from DSR. Clearly, the key problem is sensitivity of \( G_o \) to \( \nu^* \). Next, sensitivity of \( G_o \) to \( \nu^* \) is analyzed according to Fig. 1.

Figure 1 illustrates curves of \( G_o \) versus \( \nu^* \) for the \( U.S. Standard Atmosphere \) 1976 and \( \theta_0 = 0^\circ, 50^\circ, \) and 75\(^\circ\). Here Figs. 1a–d are corresponding to four cases of \( \beta = 0.05, 0.1, 0.5, \) and 1.0, respectively. For the case of \( \beta = 0.05 \) shown in Fig. 1a, as \( \nu^* \) ranges from 2 to 4, \( G_o \) changes within 1.0 \pm 0.034 for all three solar zenith angles. For \( \beta = 0.1, \) as \( 2 \leq \nu^* \leq 3, \) \( 0.992 \leq G_o \leq 1.014; \) as \( 3 \leq \nu^* \leq 3.5, \) \( 0.975 \leq G_o \leq 1.015; \) and as \( 3.5 \leq \nu^* \leq 4, \) \( 0.946 \leq G_o \leq 1.022. \) Obviously, when \( \beta = 0.1, G_o \) is insensitive to variation of \( \nu^* \) and close to unity, and the smaller \( \nu^* \) is, the closer to unity \( G_o. \) For \( \beta = 0.5 \) shown in Fig. 1c, as \( \theta_0 = 0^\circ \) and \( 50^\circ, \) \( G_o \) is also insensitive to \( \nu^* \) and close to unity, especially in the range of \( 2 \leq \nu^* \leq 3, \) but for \( \beta = 0.5 \) and \( \theta_0 = 75^\circ, G_o \) is more sensitive to \( \nu^* \) and is systematically larger than unity, especially as \( \nu^* > 3.5, \) and only as
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Fig. 1. Function $G_0$ vs value $\nu^*$ for (a) $\beta = 0.05$, (b) $\beta = 0.1$, (c) $\beta = 0.5$, and (d) $\beta = 1.0$.

For $\beta = 1.0$ shown in Fig. 1d, as $\theta_0 = 0^\circ$, $G_0$ is insensitive to $\nu^*$ and $G_0 \approx 1$, whereas at $\theta_0 = 50^\circ$ and $75^\circ$, $G$ is more sensitive to $\nu^*$ and is systematically larger than unity, especially in the range of $\nu^* > 3$. Even in the case of $\theta_0 = 75^\circ$ and $\beta = 1$, as long as $\nu^* = 2$, then $G_0 \approx 1$. As $\nu^* > 2$, $\tau_\nu(\lambda)$ decreases with the wavelength, and then, according to Eq. (9), $\lambda_m > 0.75 \mu m$ can result in $G_0 > 1$. Therefore, at $\beta \approx 0.5$ and $\theta_0 = 75^\circ$, $G_0 > 1$ implies that $\lambda_m > 0.75 \mu m$.

In other words, $G_0$ is not sensitive to the Junge distribution parameters and is close to unity for $\beta \leq 0.5$ and $\theta_0 \leq 50^\circ$. This implies that dependence of DSR on aerosols is mainly determined by the 0.75-μm aerosol optical depth, so it is possible to derive the depth from DSR. The larger $\theta_0$ is, the more sensitive to $\nu^*$ the function $G_0$. It can be seen from Eq. (7) that error of the depth retrieved from DSR, which is caused by an error in $G_0$, is inverse to $m(\theta_0)$. The larger $\theta_0$ is, the larger $m(\theta_0)$, then the weaker the error effect. This characteristic is the same as in the Langley method. Therefore, as shown in later numerical simulations, the present method can be available for large $\theta_0$ even up to $75^\circ$ and $\beta > 0.5$.

3. Parameterization model and inversion algorithm

Let $\overline{\tau}_\nu = \tau_\nu(\lambda = 0.75 \mu m)$, then we have

$$S = R S_0 G_0 t_m \exp[-\overline{\tau}_m m(\theta_0)].$$  \hspace{1cm} (14)

Next, in order to simplify calculations function $G_0$ and mean molecular transmittance $t_m$ in Eq. (14) are parameterized for the Junge distribution.

The N$_2$ absorption effect to DSR $S$ is very weak. Neglecting the effect, total molecular transmittance $T_m(\lambda, \theta_0)$ can be written as

$$T_m(\lambda, \theta_0) = T_{m\text{-}N_2}(\lambda, \theta_0) T_{m\text{-}H_2O}(\lambda, \theta_0) T_{m\text{-}O_3}(\lambda, \theta_0) T_{m\text{-}air}(\lambda, \theta_0),$$  \hspace{1cm} (15)

where $T_{m\text{-}N_2}$, $T_{m\text{-}H_2O}$, $T_{m\text{-}O_3}$, and $T_{m\text{-}air}$ are spectral transmittances of molecular scattering, water vapor, ozone, and uniformly mixed gases defined in LOWTRAN (including O$_3$, CO$_2$, CH$_4$, and so on), respectively.

Define

$$t_m = \int_{\lambda_1}^{\lambda_2} S_0(\lambda) T_m(\lambda, \theta_0) d\lambda / \int_{\lambda_1}^{\lambda_2} S_0(\lambda) d\lambda,$$  \hspace{1cm} (16)

$$T_{m\text{-}N_2} = \int_{\lambda_1}^{\lambda_2} S_0(\lambda) T_{m\text{-}N_2}(\lambda, \theta_0) d\lambda / \int_{\lambda_1}^{\lambda_2} S_0(\lambda) d\lambda,$$  \hspace{1cm} (17)

$$T_{m\text{-}H_2O} = \int_{\lambda_1}^{\lambda_2} S_0(\lambda) T_{m\text{-}H_2O}(\lambda, \theta_0) d\lambda / \int_{\lambda_1}^{\lambda_2} S_0(\lambda) d\lambda,$$  \hspace{1cm} (18)

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\[
t_{m} = \int_{\lambda_{1}}^{\lambda_{2}} S_{\lambda}(\lambda)T_{m}(\lambda, \theta_{x}) \, d\lambda \left/ \int_{\lambda_{1}}^{\lambda_{2}} S_{\lambda}(\lambda) \, d\lambda \right. ,
\]
(19)

\[A_{\text{H}_{2}O} = 1 - t_{\text{H}_{2}O},\]
(20)

\[f_{1} = \frac{t_{m}}{t_{\text{H}_{2}O}} \frac{1}{1 - A_{\text{H}_{2}O}},\]
(21)

where \(t_{m}, t_{\text{H}_{2}O}, t_{\text{O}_{3}},\) and \(t_{u}\) are average transmittances (in the range of \(\lambda_{1} \leq \lambda \leq \lambda_{2}\)) of molecular scattering, water vapor, ozone, and uniformly mixed gases, respectively, and \(A_{\text{H}_{2}O}\) is average water absorptance. Thus, we have

\[t_{m} = t_{m}(1 - A_{\text{H}_{2}O})t_{\text{O}_{3}}t_{u}f_{1},\]
(22)

where \(f_{1} = 1\) and \(t_{u} = 1.\)

The three factors \(A_{\text{H}_{2}O}, t_{\text{O}_{3}},\) and \(t_{m}\) in formula (22) are parameterized using a nonlinear regression method and LOWTRAN data, the first two among which are based on the parameterized absorptance model of water vapor presented by Wang (1976) and that of ozone by Lacis and Hansen (1974). The other two factors, \(t_{u}\) and \(f_{1}\), are empirically determined based on LOWTRAN7 data. Thus, final parameterizing forms of the five factors are

\[A_{\text{H}_{2}O} = 10^{-0.9366 + 0.037 \log_{10}(U/p_{a}) - 0.0121 \log_{10}(U/p_{a})^{2}},\]
(22.1)

\[t_{m} = \exp \left[ -0.0068 + 0.089/\mu_{0} - 0.0662/\mu_{0}^{2} \right] \frac{p}{1013},\]
(22.2)

\[t_{\text{O}_{3}} = \frac{1 - 0.0345X}{\mu_{0} + 2.2X} \frac{0.0218X}{\mu_{0} + 0.042X + 0.000323X^{2}},\]
(22.3)

\[t_{u} = \frac{1 - 0.000014p/\mu_{0}}{(1 + 1.2V(U)/\mu_{0})},\]
(22.4)

\[f_{1} = 1 - \frac{0.000012pU}{1 + 1.2V(U)/\mu_{0}} + \frac{0.0011 \exp(-1013U/p)}{\mu_{0}^{2}},\]
(22.5)

where \(p\) is the surface atmospheric pressure (hPa), \(X\) is the vertical ozone amount (cm), and \(U\) is the vertical water vapor amount (cm). Here \(t_{u}\) is expressed in terms of four parameters: \(p, X, U,\) and \(\mu_{0}\).

For the Junge distribution and constant molecular parameters, \(G_{0}\) is a function of \(\tau_{\text{air}}, \nu^{*},\) and \(\mu_{0}\). The parameterizing procedure of \(G_{0}\) is as follows. 1) Use the U.S. Standard Atmosphere, 1976, to conform relationship between \(G_{0}\) and three parameters \(\tau_{\text{air}}, \nu^{*},\) and \(\mu_{0}\), by using the least square method and the quadratic approximation; 2) Use other five atmospheric models in LOWTRAN7 to make an empirical correction to the result obtained in the last step. Finally, the parameterization form is

\[G_{0} = [1 + b\tau_{\text{air}} + c\tau_{\text{air}}^{2}]f_{2},\]
(23)

\[b = (0.38 - 0.91 \mu_{0} + 0.536\mu_{0}^{2})(\nu^{*} - 2) - (0.38 - 0.64\mu_{0} + 0.34\mu_{0}^{2})(\nu^{*} - 2)^{2},\]
(23.1)

\[c = (0.87 - 2.06\mu_{0} + 1.19\mu_{0}^{2})(\nu^{*} - 2) + (1.39 - 3.\mu_{0} + 1.7\mu_{0}^{2})(\nu^{*} - 2)^{2},\]
(23.2)

\[f_{2} = 1 + 0.158(\nu^{*} - 2)(1 - 0.889U)\tau_{\text{air}}^{2}
\]
\[+ (1 + 0.3U + \mu_{0}^{2}\sqrt{U}),\]
(23.3)

where \(f_{2}\) is the correction factor. Now \(G_{0}\) is expressed in terms of four parameters: \(\nu^{*}, \tau_{\text{air}}, U,\) and \(\mu_{0}\). In the LOWTRAN7 mode, air mass correction and effects of the pressure and temperature on gas absorption have been incorporated (Kneizys et al. 1983, 1988). Since our parameterizing procedure is based on LOWTRAN7 data, the correction and effects can be incorporated in our parameterizing forms of \(t_{u}\) and \(G_{0}\), although they are expressed in terms of \(\mu_{0}\) and the actual water amount.

It can be seen from Table 2 that the parameterization model has satisfactory accuracy. In Table 2, \(M\) indicates atmospheric model, and \(\delta_{p}\) is percentage standard deviation of the parameterized \(S\) to the LOWTRAN-calculated \(S\) for 612 cases with the input parameters: \(\theta_{90} = 0^\circ, 50^\circ\) and \(75^\circ; \nu^{*} = 1.8 + 0.1t, i = 1, \ldots, 17;\) and \(\beta = 0.05i, i = 1, \ldots, 12.\) As shown in Table 2, the standard deviation is \(\leq 0.495\%\) for each of the six atmospheric models.

As stated above, \(G_{0} \approx 1\) for smaller \(\theta_{90}.\) So, for the condition of known molecular parameters an iteration algorithm to Eq. (14) for deriving \(\tau_{\text{air}}\) is proposed:

1) Let \(G_{0}^{(0)} = 1;\)
2) Take \(n = 1;\)
3) Derive \(\tau_{\text{air}}^{(n)} = [\log(RS_{90}^{(n)}G_{0}^{(n)} - \log(S)]/\log(\theta_{90});\)
4) Using \(\tau_{\text{air}}^{(n)}\) and assuming a value of \(\nu^{*},\) calculate \(G_{0}^{(n)}\) according to Eq. (23); and
5) If abs \((1 - \tau_{\text{air}}^{(n-1)}/\tau_{\text{air}}^{(n)}) < \epsilon (\epsilon = 0.001\) in this paper), stop the iteration, otherwise let \(n = n + 1\) and repeat steps 3 and 4.

The parameterization model and inversion method are developed under the assumption of a Junge size distribution. However, they are expected to be applicable in the case of non-Junge distribution because of the aforementioned fact that \(G_{0}\) is not very sensitive to variation of \(\nu^{*}\) for \(\theta_{90} \approx 50^\circ\). In the model only two aerosol parameters, that is, 0.75-\(\mu\) aerosol optical depth and the value \(\nu^{*}\), are used in characterizing dependence of

<table>
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<th>M</th>
<th>(\delta_{p}(%))</th>
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<tr>
<td>1</td>
<td>0.495</td>
</tr>
<tr>
<td>2</td>
<td>0.397</td>
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DSR on the aerosol physical-optical property. Therefore, as molecular parameters are known, only the value \( \nu^* \) is required for retrieving the depth from DSR.

4. Numerical simulations and error analysis

Molecular scattering and absorption parameters must be known to retrieve the 0.75-\( \mu \text{m} \) aerosol optical depth from DSR. The molecular optical depth can be determined according to surface pressure. Ozone and water vapor are responsible for the dominant absorption to \( S \). Therefore, the main factors resulting in an error in the depth solution are:

1) Uncertainty of aerosol size distribution,
2) Error in the water vapor amount,
3) Error in the ozone amount, and
4) Measurement error of \( S \).

Next, the four error factors are analyzed according to numerical simulations, respectively.

a. Effect of size distribution uncertainty

At first, inversion simulations for the Junge distribution are analyzed according to Table 3, where \( M \) indicates atmospheric model, \( \delta_i \) is the percentage standard error of 510 aerosol optical depth solutions to their true values, and \( \nu_0 \) is a guess of \( \nu^* \) in the retrieval. Input parameters for 510 simulations are \( \theta_0 = 0^\circ, 50^\circ \) and \( 75^\circ; \nu^* = 1.8 + 0.1i, i = 1, \ldots, 17; \) and \( \overline{\nu}_0 = 0.05i, i = 1, \ldots, 10. \)

As shown in Table 3, given \( \nu_0 = \nu^* \) (using true \( \nu^* \) in inversion), standard error \( \delta_i \) for each of the six atmospheric models is \( \leq 0.9\% \). The error is caused by an error in parameterized DSR and is very small. For \( \nu_0 = 3, \delta_i \leq 4.58\%, \) and for \( \nu_0 = 2.5, \delta_i \leq 4.97\%. \) In the last two cases solution errors are mainly caused by uncertainty of the assumed aerosol size distribution. The value \( \nu^* \) selected in these 510 numerical simulations changes from 1.9 to 3.5, but, if \( \nu_0 = 3 \) or 2.5 is used in the retrieval, the standard errors of 510 optical depth solutions are all \( \leq 5\% \) for every atmospheric model. So, solution accuracy better than 5\% can generally be obtained using the Junge distribution.

Furthermore, inversion results in the case of a non-Junge distribution are analyzed according Table 4 and Figs. 2–4.

Table 4 shows 20 LOWTRAN aerosol models with a non-Junge size distribution, which are selected in later inversion simulations. Here “No” is the index of the aerosol model, and, as in LOWTRAN, IHAZE selects the type of the boundary layer model, IVULCN selects the type of the stratospheric aerosols, ICSTL selects the season profile for both tropospheric and stratospheric aerosols, ISEASN selects the seasonal profile for both tropospheric and stratospheric aerosols, ICSTL represents the air mass character only used for the U.S. Navy maritime model (IHAZE = 3), ICLD specifies the cloud models, \( H_i \) is an initial altitude for a given problem, and \( W \) is wind speed only used as IHAZE = 10 (desert model). Considering the important contribution of the boundary layer aerosol to column optical depth, all types of boundary layer aerosol models except for the fog model (IHAZE = 8 and 9) in LOWTRAN are selected, including rural, U.S. navy maritime with ICLD = 1 (open ocean) and 10 (strong continental influence), LOWTRAN maritime, urban, tropospheric, and desert (with wind speed \( W \) of 0, 10, and 20 m s\(^{-1}\) ) models. Four types of stratospheric aerosol models are used: 1) background stratospheric, 2) moderate volcanic, 3) high volcanic, and 4) extreme volcanic, marked as IVULCN = 1, 2, 3, and 8. ISEASN = 1 (spring–summer) is selected for No = 1–19, and ISEASN = 2 (fall–winter) is only for No = 20. Indices No 11 and 12 have ICLD = 18 and 19, corresponding to standard cirrus and subvisual cirrus models, respectively. In four models, No = 16, 17, 18, and 19, \( H_i = 2, 4, 8, \) and 12 km, respectively, and IVULCN = 8 (extreme volcanic), where the volcanic aerosol has dominant contribution to total aerosol optical depth, especially for No = 19 with \( H_i = 12 \) km. These 20 aerosol models cover almost all aerosol types in LOWTRAN.

Figure 2 shows the true 0.75-\( \mu \text{m} \) aerosol optical
The 0.75-μm aerosol optical depths and values \( n^* \) for the 20 aerosol models.

Figure 3 shows errors of DSR calculated by the parameterization model under an assumption of \( n^* = 3 \) for these 20 aerosol models and five solar zenith angles of \( \theta_0 = 0^\circ, 30^\circ, 50^\circ, 60^\circ, \) and \( 75^\circ \). Here the DSR, which is regarded as exact, is actually calculated by LOWTRAN7 using exact \( \tau_a(\lambda) \) data. As shown in Fig. 3, when \( \theta_0 \leq 60^\circ \), the error is <7%, but for \( \theta_0 = 75^\circ \) and four aerosol models of \( \text{No} = 2, 6, 10, \) and 15 with a larger optical depth, the error is larger, even being larger than 20% for \( \text{No} = 2 \) and 6 having \( \tau_a(\lambda = 0.75 \mu\text{m}) \) larger than 0.8. The DSR error is mainly caused by uncertainty in \( \tau_a(\lambda) \) determined under an assumption of Junge size distribution of \( n^* = 3 \).

Figure 4 illustrates percentage errors of 100 optical depth solutions retrieved for these 20 aerosol models and five solar zenith angles of \( \theta_0 = 0^\circ, 30^\circ, 50^\circ, 60^\circ, \) and \( 75^\circ \). There are 20 solutions for every angle. In these inversion simulations, \( n^* = 3 \) is taken. For five cases
of $\theta_o = 0^\circ$, 30$^\circ$, 50$^\circ$, 60$^\circ$, and 75$^\circ$, maximum errors among 20 solutions are 7.0%, 7.8%, 4.8%, 6.3%, and 14.1%, and their standard errors are 2.7%, 4.2%, 2.2%, 3.6%, and 6.9%, respectively. Among 100 solutions, only two have an error larger than 10%, and the standard error of the 100 solutions is 4.3%. This means that error of 0.75-$\mu$m optical depth solution, which is caused by an assumption of a Junge size distribution with $n^* = 3$, is generally <5%. As shown again in Fig. 3, the parameterizing DSR has generally the smallest error for $\theta_o = 50^\circ$. As a result, when $\theta_o = 50^\circ$, the solution accuracy is best. When $\theta_o = 75^\circ$, the DSR error is larger, and the standard error of 20 solutions is 6.9%, being larger but tolerable. For No. 2 and 6 with $\tau_o (\lambda = 0.75 \mu m) > 0.8$, the DSR errors are larger and the corresponding solution errors are also larger. In the case of $\theta_o = 75^\circ$, the solution error is within $\pm 5.1\%$ for 14 models with the optical depth being less than 0.25. So, in the case of non-Junge distribution the present method is also more suitable for the smaller aerosol optical depth and $\theta_o$. In addition, some solutions are overestimated and some underestimated. Therefore, the average result would have a better accuracy.

Furthermore, as $\nu^*$ listed in Table 4 is selected instead of $\nu^* = 3$, the standard deviation of 100 solutions is 3.1% and the maximum error is 12.6%, showing better accuracy. Thus, the present method is also suitable for the non-Junge aerosol model, and the error of the 0.75-$\mu$m optical depth solution is generally <5%, if a Junge distribution of $\nu^* = 3$ is taken. In addition, selecting a suitable $\nu^*$ can improve the solution accuracy.

Up to now, the effect of wavelength-dependent aerosol refractive index has not been analyzed. In fact, in the above numerical simulations using the LOWTRAN aerosol model, aerosol refractive index varies with wavelength. Here 93.5% of solar radiation concentrates in the wavelength range of $\lambda \approx 2 \mu m$, in which variation of the aerosol refractive index with wavelength is usually smaller. So a weak effect of the wavelength dependence of the aerosol refractive index on the optical depth solution can be expected.

**b. H$_2$O-error effect**

According to Eqs. (14) and (22), the error in the optical depth caused by an error in H$_2$O amount $U$ can be estimated approximately as

$$\Delta \tau_o = -\mu_o \frac{\partial \Delta A_{H_2O}}{\partial U} \Delta U.$$  \hspace{1cm} (24)

Because $\partial A_{H_2O}/\partial U$ in Eq. (14) is positive, $\Delta U > 0$ (H$_2$O overestimated) can result in an underestimated optical depth solution. When $U = 0.5$, 1, 3, 10 cm and $\Delta U = 0.2$ cm, solution errors derived using Eqs. (22) and (24) are $-0.0129$, $-0.0083$, $-0.0041$, and $-0.0018$ for $\mu_o = 1.0$, and $-0.0081$, $-0.0052$, $-0.0025$, and $-0.0011$ for $\mu_o = 0.5$, respectively. Because there may be H$_2$O-absorbing situation, the value $\partial A_{H_2O}/\partial U$ decreases with increasing $U$ and $\theta_o$, and so, under the same $\Delta U$, the larger $U$ and $\theta_o$ are, the smaller the solution error. As $U \approx 1$ cm, the largest error of the depth is 0.0121 for $\Delta U = \pm 0.3$ cm, and 0.225 for $\Delta U = \pm 0.5$ cm. Therefore, as $\Delta U = \pm 0.2$ cm, $\Delta \tau_o$ is within $\pm 0.013$; as $U \approx 1$ cm and $\Delta U = \pm 0.5$ cm, it is within $\pm 0.023$.

**c. Ozone-error effect**

According to Eqs. (14) and (22), the error of the optical depth caused by an error in ozone amount $X$ can be estimated as

$$\Delta \tau_o = -\frac{\partial \Delta A_{H_2O}}{\partial X} \Delta X.$$  \hspace{1cm} (25)

Because $\partial A_{H_2O}/\partial X$ is given by a constant (see Table 5), $\Delta \tau_o$ is not affected by $\Delta U$. In addition, the maximum values of $\partial A_{H_2O}/\partial X$ are $2.2 \times 10^{-4}$ and $2.6 \times 10^{-2}$ for $\mu_o = 1.0$ and 0.5, respectively. Thus, $\Delta \tau_o$ varies within $\pm 0.005$ as $U \approx 1$ cm and $\Delta U = \pm 0.5$ cm, being tolerable.
The representative value of $X$ is 0.35 cm. If it has a 20% error, $\Delta X = 0.07$ cm, then according to Formula (19), the solution errors caused by it are $-0.00172$, $-0.00188$, and $-0.00165$ for $\mu_0 = 1$, 0.5, and 0.3, respectively. Obviously, the errors are smaller.

d. Measurement error effect

The error of the optical depth caused by an error, $\Delta S$, in DSR $S$ is

$$\Delta S = -\mu_0 \Delta S/S,$$

(26)

which implies that under the same $\Delta S$, the larger $\mu_0$ is, the larger the solution error. If $\Delta S$ is within $\pm 2\%$, the solution error is within $\pm 0.02$.

To sum up, as shown in inversion simulations, the error of the optical depth less than 5% can generally be expected for the usual aerosol condition and $\theta \approx 75^\circ$, if a Junge distribution of $n^a = 3$ is used instead of true Junge and non-Junge aerosol size distributions. Errors in $S$ and $H_2O$ amount within $\pm 2\%$ and $\pm 0.2$ cm can result in solution errors within $\pm 0.02$ and $\pm 0.013$, respectively.

5. Comparative observations

In this section, 0.75-μm aerosol optical depths measured by pyrheliometer and sunphotometer are compared. There are altogether 1267 sets of comparative data.

a. Measurement instruments and experiments

Comparative observations using a pyrheliometer and sunphotometer were made on the roof of an office building at our institute in Beijing. Altogether there are 25 observing days from January to June 1995. The sunphotometer has eight 1° field of view interference filters with central wavelengths at 0.4, 0.44, 0.52, 0.612, 0.67, 0.78, 0.88, and 1.03 μm. A thermostat was installed to obtain $\pm 1\%$ accuracy of temperature. A microcomputer was used to track the sun’s center, transform optical filters, and sample and store the data on time. The sunphotometer was calibrated using the Langley method at an astronomy observatory (on top of a hill) of the Chinese Academy of Sciences in north China from 23 September to 12 October 1995. Figure 5 illustrates data logS($\lambda$, $\theta_0$) for $\lambda = 0.78 \mu m$ obtained on 5 and 8 October versus airmass $m(\theta_0)$ and linear regressions between them by the least squares method. According to the regression result, the logarithms of the sunphotometer constants, log $S_0(\lambda)$, are 8.4621 and 8.4739, re-
spectively, and the difference of the two values of log $S_o(\lambda)$ is $-0.0118$. Five sets of calibration results with the correlation coefficient between log $S(\lambda, \theta_p)$ and $m(\theta_p)$ larger than 0.995% are selected. For $\lambda = 0.75\mu$m the mean value of the five log $S_o(\lambda)$ is 8.4694, and their standard deviation is 0.00687. So, $\pm0.687\%$ calibration accuracy of the 0.78-µm sunphotometer constant is estimated. The pyrheliometer with DFY-3 model name, produced in China, has a spectral response of 0.3–4 µm and a field of view of 6°12′29″. It was calibrated at the Chinese Academy of Meteorological Sciences with a first-class NIP pyrheliometer. The NIP pyrheliometer is calibrated by an H–F absolute pyrheliometer having accuracy better than 0.3%. The calibration error of our pyrheliometer is within $\pm1\%$, and its annual stability is $\pm1\%$.

b. An empirical expression on water vapor amount

For the convenience of actual use, an empirical relation between the water vapor amount $U$ (cm) and the surface water vapor pressure $e$ (hPa) was developed, based on routine observation data at Beijing’s meteorological observatory in 1988. It is given by

$$
\log U = -0.7443 + 0.9425 \log e.
$$

(27)

Figure 6 compares $U$ derived according to Formula (27) with the H$_2$O amount $U_0$ calculated using radiosonde observation data. As shown in Fig. 6, $U$ conforms well with $U_0$; their correlation coefficient is as high as 0.934, and the standard error of $U$ is 0.23 cm. As $U$ is larger, the error in $U$ is sometimes larger, having a maximum value of 0.893 cm.

c. Comparative measurement results

Figures 7–14 and Table 5 compare 0.75-µm aerosol optical depths (marked as $\tau_{ap}$) measured by pyrheliometer with those ($\tau_{as}$) by sunphotometer. In these figures, $\alpha$ is the percentage standard deviation of $\tau_{ap}$ to $\tau_{as}$. The $\tau_{as}$ are derived from the 0.78-µm optical depths measured by the sunphotometer according to the power law fitted by eight-wavelength aerosol optical depths.

In the program of retrieving $\tau_{ap}$ from $S$, the ozone amount detected by Dobson in Beijing is used, and the solar constant of 1367 (W m$^{-2}$) and the value $R$ of 0.982 are taken. Considering a larger viewing field of the pyrheliometer and sometimes a larger aerosol optical depth over the Beijing area, a correction using a single scattering model is made in order to compensate the forward
scattering component entering the pyrheliometer. The correction procedure is as follows. 1) Assume no scattering to retrieve $\tau_{ap}$; 2) use the depth and aerosol scattering phase function for the Junge distribution of $\nu^a = 3$ and the aerosol refractive index of $1.5 \pm 0.01$ to numerically calculate the scattering component and DSR as well as the ratio of the DSR plus the component to the DSR; 3) use the ratio to subtract the scattering component from radiation detected by pyrheliometer and then again to retrieve $\tau_{ap}$ from the corrected radiation.

Representative 8-day observation data are shown in Figs. 7–14, where triangles, and dots, indicate aerosol optical depths $\tau_{ap}$ and $\tau_{as}$, respectively, and $\nu^a = 3$ is selected to retrieve $\tau_{ap}$ from $S$.

As shown in Fig. 7, on 8 January in winter there is a large variation of aerosol optical depths ranging from 0.349 to 0.78, and there is a good agreement between $\tau_{ap}$ and $\tau_{as}$. Their correlation coefficient is 92.3%, and their standard deviation $\sigma$ is 9.5%. It is noted that there is a larger deviation of $\tau_{ap}$ to $\tau_{as}$ for a larger $\theta_v$.

Figure 8 shows observation results on 9 January in winter. Compared with other 24-day results, the optical depths during 0850–1500 (local time) of the day are the smallest and relatively stable, being within 0.13 $\pm$ 0.03. The difference between $\tau_{ap}$ and $\tau_{as}$ is generally within $\pm 0.02$, and $\sigma$ is 9.9%.

Figures 9–12 show 4-day observation results made in spring. As shown in Fig. 9, on 31 March both $\tau_{ap}$ and $\tau_{as}$ increase during 0800–1000 and 1440–1600, and they are basically stable during 1000–1430. Their standard error is 11.4%.

As shown in Fig. 10, on 1 April the aerosol optical depth is stable during 0910–1300, and there is a cloud layer with the maximum optical depth of about 1.4 after 1300. In the variation process, cloudiness transition good agreement between $\tau_{ap}$ and $\tau_{as}$ is obtained. Here $\sigma$ is only 8.1%.

On 4 April there is a serious haze layer in Beijing, surface visibility is $<2$ km, and a thin cloud occurs during 1250–1400. As shown in Fig. 11, in the day
aerosol optical depths vary between 0.98 and 1.6. The maximum error of $\tau_{ap}$ is 27.6% and its standard error $\alpha$ is 10.9%. So, the present method seems to be available in the case of thin cloud. In addition, $\tau_{ap}$ is systematically larger than $\tau_u$ during 0828–0948 and 1437–1633, especially when the solar zenith angle is larger, the deviation is larger with a maximum value of 27.8%. This trend is same as in Fig. 7. Considering that the aerosol distribution uncertainty can result in a relatively large solution error for the large solar zenith angle, the uncertainty is estimated to be responsible for the systematic error.

Figure 12 shows measurement results on 14 May, when $\alpha = 9.6\%$, and $\tau_{ap}$ is systematically larger than $\tau_u$ with a deviation range of 0.1%–16% except for three sets of data.

Figures 13–14 compare $\tau_{ap}$ with $\tau_u$, measured on 4 and 25 June in summer. In Fig. 13, $\alpha$ is 11.3%; $\tau_{ap}$ conforms well with $\tau_u$, except for several sets of data around 1235. Around noon, there is a quick change in the aerosol optical depth and a larger deviation of $\tau_{ap}$ to $\tau_u$. What caused the larger deviation is not clear.

As in Fig. 12, there is a systematic deviation of $\tau_{ap}$ with $\tau_u$ in Fig. 13, where $\tau_{ap}$ is systematically underestimated by 2.9%–20.3%, whereas $\tau_u$ is systematically overestimated in Fig. 11. Besides, in Fig. 14 the standard deviations of $\alpha = 14.4\%$ is the largest among all 25 days. As stated above, the error of $\tau_{ap}$ caused by the aerosol distribution uncertainty is generally <5% and is larger only for larger solar zenith angle. So the uncertainty is not a main factor resulting in the systematic error. On the 25th water vapor amount $U = 1.98$ cm is the largest for all 25 days. And as shown in Fig. 5, as $U$ is larger, error of $U$ derived by formula (26) is sometimes larger. Thus, it is estimated that the error of $U$ is a main factor resulting in the error of $\tau_{ap}$ in Fig. 14, and $\Delta U = 0.86$ cm can basically respond to it.

Table 5 gives statistic results of 1267 sets of observation data. Here $N$ is observation times; $\tau_u$ and $\tau_{ap}$ are mean values of $N$ sets of $\tau_u$ and $\tau_{ap}$ measured by sunphotometer and pyrheliometer, respectively; $\Delta\tau_{ap}$ is difference of $\tau_{ap}$ with $\tau_u$ in percentage; $\alpha$ is percentage standard deviation of $N$ sets of $\tau_{ap}$ to $\tau_u$; $\nu^\phi$ is Junge parameter fitted from aerosol optical depths measured by sunphotometer; and $\nu^\phi = 3$ is assumed to retrieve $\tau_{ap}$ from DSR. These 1267 sets of measurements cover the range of $\theta_u$ changing from 11° to 79°.

From Table 5 it can be seen that
1) Daily-average optical depth $\tau_u$ changes from 0.15 to 1.086. Mean values of total 1267 sets of $\tau_u$ and $\tau_{ap}$ are 0.415 and 0.410, respectively, being in good agreement with a difference of only 1.2%.
2) Daily-average $\nu^\phi$ varies between 2.31 and 3.25, and the total average $\nu^\phi$ is 2.65.
3) The daily standard deviation \( \alpha \) of \( \tau_{ap} \) to \( \tau_{ap} \) changes from 5.7% to 14.4%, and the total standard deviation of 1267 sets of \( \tau_{ap} \) and \( \tau_{ap} \) is 10.2%. So an error in \( \tau_{ap} \) less than 11% can usually be expected, and the result averaged over a lot of data can have much better accuracy.

Furthermore, as the fitted \( \nu^* \) (listed in Table 5) is taken instead of \( \nu^* = 3 \), total \( \alpha \) and \( \Delta \tau_{ap} \) are 9.7% and -1.4%, and as \( \nu^* = 2.5 \), total \( \alpha \) and \( \Delta \tau_{ap} \) are 8.8% and -2.2%, respectively, showing a commensurate accuracy with that as \( \nu^* = 3 \). As \( \nu^* = 3.5 \), total \( \alpha \) is 15.9%, showing worse accuracy, but total \( \Delta \tau_{ap} \) is also smaller, being only 0.3%.

As stated above, the error in \( \tau_{ap} \) determined by the present method can usually be within \( \pm 11\% \). The error mainly comes from errors in the water vapor absorbance, DSR, and the aerosol size distribution.

The errors of water vapor absorbance basically depend on uncertainty of its amount. At Beijing’s meteorological observatory there are two daily radiosonde measurements starting at 0715 and 1915 local time, respectively. During the period of our comparative observation there was no radiosonde measurement. In this paper, the H2O amount \( U \) is determined by the empirical formula (27), which may have a standard error of 0.23 cm. And according to Eq. (24), \( \Delta U = \pm 0.23 \) cm can result in an error of \( \tau_{ap} \) within 0.0146. In addition, from Fig. 6, we can note also that the error of \( U \) calculated by formula (27) varies randomly, so it would result in relatively small errors in monthly mean and annually mean aerosol optical depth measured by pyrheliometer.

Error in \( U \) calculated by formula (27) may sometimes be larger, especially for a larger \( U \). So the error can sometimes cause a large error in \( \tau_{ap} \). If \( U = 2 \) cm and \( \Delta U = \pm 0.8 \) cm, the error in \( \tau_{ap} \) is within \( \pm 0.021 \). Because there was no simultaneous radiosonde measurement data, a quantitative analysis of the error factor cannot be made.

Calibration errors of the sunphotometer and the pyrheliometer are less than 0.7% and 1%, respectively. The error can result in a systematic deviation of \( \tau_{ap} \) to \( \tau_{ap} \) less than 0.0085 for \( \mu_a = 0.5 \) (a percentage deviation of 2.05% for the total mean optical depth of 0.415).

As analyzed in the last section, according to inversion simulations, error in the aerosol optical depth solution caused by an uncertainty of aerosol size distribution is usually within \( \pm 5\% \). Thus, it is estimated that the uncertainty is responsible for about half of the error in \( \tau_{ap} \) measured by the pyrheliometer. Besides, the depth solution error caused by this uncertainty may be positive or negative, which is another reason why the average value of a large number of optical depths measured by the pyrheliometer has better accuracy.

6. Summary

Total direct solar short radiation can be expressed as a product of several factors, one of which is the 0.75- \( \mu \)m aerosol transmittance, and others are independent or weakly dependent on aerosol physical–optical parameters, especially for the smaller solar zenith angle. Based on this property, a method to determine the 0.75- \( \mu \)m aerosol optical depth using the radiation is proposed, and the corresponding iterative inversion algorithm and a parameterization model on the radiation are developed using the Junge size distribution.

A key question is the effect of the aerosol size distribution uncertainty on the optical depth solution. Numerical simulations show that a solution error within \( \pm 5\% \) can generally be expected for both Junge and non-Junge aerosol size distributions, if a Junge distribution of \( \nu^* = 3 \) is used in retrievals. In addition, it is important for improving the solution accuracy to exactly determine the S and H2O amount. Errors in S and H2O amount within \( \pm 2\% \) and \( \pm 0.2 \) cm can result in solution errors within \( \pm 0.02 \) and \( \pm 0.013 \), respectively.

The Junge size distribution is an idealized case of a true atmospheric aerosol distribution. It may not be appropriate in oceanic areas or even continental areas remote from local pollution sources. In this case, it is possible to improve the accuracy if a suitable non-Junge distribution is adopted. This problem will be studied in the future.

The method is tested in comparative observations...
with the result that aerosol optical depths measured by pyrheliometer conform well with those measured by sunphotometer. As far as the 1267 sets of comparative experiments are concerned, the standard deviation is 10.5%, and the deviation of the mean optical depth only \(-0.7\%\) for \(n_0 = 3\). The mean result has better accuracy. The error in aerosol optical depth measured by pyrheliometer is mainly due to errors in the water vapor absorption, the aerosol size distribution, and the DSR.

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