Seasonal variation of columnar aerosol optical properties and radiative forcing over Beijing, China

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\textbf{HIGHLIGHTS}

- Long-term seasonal characteristics of aerosol optical properties and radiative forcing of urban Beijing were analyzed.
- Both AOD and AE showed significant but reasonable monthly, seasonal and annual variation.
- The urban aerosol in Beijing leaded positive ARF the atmosphere with an average magnitude of 57 W/m\textsuperscript{2}.
- The HYSPLIT trajectories were originating from multiple source locations and had different effects on aerosol characteristics.

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\textbf{A B S T R A C T}

Long-term seasonal characteristics of aerosol optical properties and radiative forcing at Beijing (during March 2001–March 2015) were investigated using a combination of ground-based Sun/sky radiometer retrievals from the AERONET and a radiative transfer model. Aerosol optical depth (AOD) showed a distinct seasonal variation with higher values in spring and summer, and relatively lower values in fall and winter. Average Angstrom exponent (AE) in spring was lower than other seasons, implying the significant impact of dust episodes on aerosol size distribution. AE mainly distributed between 1.0 and 1.4 with an obvious uni-peak pattern in each season. The observation data showed that high AODs (>1.0) were clustered in the fine mode growth wing and the coarse mode. Compared to AOD, seasonal variation in single scattering albedo (SSA) showed an opposite pattern with larger values in summer and spring, and smaller ones in winter and fall. The highest volume size distribution and median radius of fine mode particles occurred in summer, while those of coarse mode particles in spring. The averaged aerosol radiative forcing (ARF) at the top of the atmosphere (TOA) in spring, summer, fall and winter were $-33 \pm 22$ W m\textsuperscript{-2}, $-35 \pm 22$ W m\textsuperscript{-2}, $-28 \pm 20$ W m\textsuperscript{-2}, and $-24 \pm 23$ W m\textsuperscript{-2} respectively, and these differences were mainly due to the SSA seasonal variation. The largest positive ARF within atmosphere occurred in spring, implying strong warming in the atmosphere. The low heating ratio in summer was caused by the increase in water vapor content, which enhanced light scattering capacity (i.e., increased SSA).

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1. Introduction

Atmospheric aerosols and their impact on climate change have drawn considerable attention during past a few decades (IPCC, 2013). Aerosol can affect the radiation energy budget of the Earth-atmosphere system directly by scattering and absorbing solar radiation (Charlson and Pilat, 1969; Charlson et al., 1992).
Indirectly, aerosols can alter cloud particle size, amount, and lifetime, and, thus, lead to a profound influence on global precipitation and the hydrologic cycle (Twomey, 1977; Albrecht, 1989). Massive studies have been carried out to study aerosol climatic effects, whereas our scientific understanding on the uncertainties on aerosol radiative effects is still limited compared with that on long-lived greenhouse gases (IPCC, 2007; 2013). This is mainly due to aerosol’s short lifetime, complex chemical composition, and interaction in the atmosphere that results in high spatial-temporal heterogeneities. The IPCC report (2013) indicates that global estimation of aerosol direct radiative effect varies from $-0.85$ to $+0.15$ W m$^{-2}$ with an uncertainty of $1$ W m$^{-2}$. Therefore, long-term, detailed global measurements from satellite and ground-based networks are urgently required to better understand aerosol optical properties and to evaluate their effects on climate change.

Aerosols in Beijing have attracted considerable attention due to their high mass concentration and variety of sources, such as vehicle exhaust emissions, coal combustion, and long-range transport dust and biomass burning, which influence aerosol optical and radiative properties. With the increasing urbanization and industrialization, the aerosol loading increases continuously leading to high aerosol optical depth (aerosol extinction integrated in the vertical column) and large reductions in the solar radiation at the surface. For example, according to the statistics from Beijing Municipal Environmental Protection Bureau, Beijing recorded 42 days (13% of entire days of the year) of heavy air pollution in 2015. The annual average PM$_{2.5}$ mass concentration exceeded 80 $\mu$g m$^{-3}$ in 2015, which is over 2 times larger than the Chinese Grade standard value (35 $\mu$g m$^{-3}$) (http://www.bjepb.gov.cn/bjepb/323265/397983/4387804/index.html) and 8 times larger than the standard recommended by the WHO (10 $\mu$g m$^{-3}$) (WHO, 2006). For Beijing, early, Bergin et al. (2001) measured the chemical compositions and aerosol radiative properties in surface atmosphere. Also, Eck et al. (2005) and Xia et al. (2006) analyzed aerosol optical properties in columnar atmosphere over Beijing and estimated aerosol radiative effects. He et al. (2009) conducted two-year ground-based observations of aerosol optical properties at an urban site in Beijing. These results suggested that the urban aerosols have moderate absorption as compared to the dust aerosols. Xu et al. (2015) studied the long-term variation of aerosol optical depth based on the hourly accumulated direct solar radiation over 14 cities during 1993–2002. They found that yearly mean AOD showed a significant decreasing trend and the overall mean AOD at 750 nm was 0.53 in Beijing. Yu et al. (2016a) found that the maximum value of daily mean AOD at 340 nm was up to 2.21 during dense haze-fog episode in January 2013. The aerosol optical and radiative forcing under different weather conditions were investigated by the several authors (Wang et al., 2010b; Bi et al., 2015; Yu et al., 2016b; Gharibzadeh et al., 2017). They concluded that a significant amount of absorption (i.e. a positive heating effect on the atmosphere) occurred under haze event in Beijing. Wang et al. (2009) indicated that more absorbing aerosol particles existed under foggy and hazy weather conditions than those under dusty weather conditions. However, the knowledge of aerosol physical and chemical properties and its radiative impact in this region is still insufficient. It is difficult to explain aerosol climate effect from the view of climatology. Therefore, the studies on aerosol optical and radiative properties using the long-term (over 10 years) are essential for evaluating regional climate change.

This study investigates aerosol optical properties based on ground-based observations from 2001 to 2015 (see Fig. 1) over urban Beijing, and numerically reveals its radiative effects. The paper is organized as follows. The measurements and models used are introduced in Section 2, and Section 3 discusses the corresponding results. Section 4 concludes this work.

## 2. Site, measurement and methodology

### 2.1. Site and measurement

The AERONET is an international network measuring aerosol optical properties with ground-based sun photometers, and is widely used to validate aerosol properties from satellite retrievals (Holben et al., 1998). The CIMEL CE-318 automatic sun/sky radiometer, a part of AERONET Beijing instruments, was installed on the roof (above 30 m height) of the Institute of Atmospheric Physics (IAP) building before March 2004, and, then, moved to the top of a building 2 km away from the original location. Beijing is located at the northwestern border of the Great North China Plain and surrounded by mountainous area in the west, the north, and the northeast (Fig. 2). In spring, the dust occasionally uplifted from the Kumutage and Taklimakan deserts in western China and Mongolian

![Fig. 1. Time series during study period.](Image)

![Fig. 2. Moderate Resolution Imaging Spectroradiometer (MODIS)-derived aerosol optical depth at 550 nm (2001–2015). The black star indicates the location of the Beijing site.](Image)
deserts can be transported to large part of northern China (Sun et al., 2001). As shown in Fig. 2, high aerosol loading occurred on the southern regions of Beijing.

The instruments observe direct solar radiation within a 1.2 full field-of-view at nine channels. The band width of 1640, 1020, 940, 870, 675, 500, 440 nm is 10 nm, and 2 nm in 380 nm and 340 nm. A detailed description of sun radiometer and its products has been presented by Holben et al. (1998). Measurement at the 940 nm channel can be used to retrieve water vapor content. Ångstrom exponent was derived by the AODs at the 440, 670 and 870 nm channels. Single scattering albedo, refractive index (real and imaginary parts) and aerosol size distribution were retrieved from the diffuse radiance measurements and AODs.

This study uses the AERONET level 1.5 and 2.0 data (http://aeronet.gsfc.nasa.gov) from both direct sun and inversion products between 2001 and 2015. The fine and coarse modes AOD at 500 nm use level 1.5 data, other parameters (AOD, SSA, RI and volume size distribution, and so on) use level 2.0 data. The uncertainty in AOD retrieval under cloud free conditions was $\pm 0.02$ for the channel with central wavelength of 440 nm, and $\pm 0.01$ for other channels, which is less than the $\pm 5\%$ uncertainty for the retrieval of sky radiance measurements (Dubovik et al., 2000). Retrieval errors in volume size distribution typically do not exceed 15–35% (depending on the aerosol type) for each particle radius bin within the 0.1–70 $\mu$m range. The errors for small (0.05–0.1 $\mu$m) or large particles (7–15 $\mu$m) reach 35–100% for a given particle radius bin. The SSAs were expected to have an uncertainty of 0.03–0.05 depending on aerosol type and loading (Dubovik et al., 2000).

2.2. Methodology

2.2.1. Aerosol optical depth

According to the Beer-Lambert-Bouguer law

$$\tau_l(\lambda) = \tau_R(\lambda) + \tau_O(\lambda) + \tau_{O_3}(\lambda) = \frac{1}{m} \ln(I/I_0),$$

The atmospheric optical depth $\tau_l$ is the sum of aerosol optical depth $\tau_R$, Rayleigh scattering depth $\tau_R$ and ozone absorption depth $\tau_{O_3}$. $I$ is direct sun intensity at Earth surface, $I_0$ is solar intensity above the atmosphere and $m$ is the relative path length through the atmosphere.

Rayleigh scattering optical depth is calculated using the following method:

$$\tau_R(\lambda) = 0.00864 \times \lambda^{-1.391} \times 0.074 \times 0.005/1013.25 \times \frac{p}{1013.25},$$

where $p$ is the actual atmospheric pressure and $\lambda$ presents wavelength.

The optical depth of ozone absorption is calculated according to:

$$\tau_{O_3}(\lambda) = a_{O_3}(\lambda) \times O_3 \times m_{O_3},$$

where $a_{O_3}(\lambda)$ is the absorption coefficient of ozone, $O_3$ is the quantity in atmosphere with unit of Dobson, and $m_{O_3}$ is the air mass of ozone.

2.2.2. Ångström exponent ($\alpha$)

Parameter $\alpha$ is computed from a linear fit of log (AOD) versus log (wavelength) according to the classical equation of Ångström (1964)

$$\tau_\lambda(\lambda) = \beta \lambda^{-\alpha},$$

where $\alpha$ is Ångström wavelength exponent, which can reflect the size characteristics of particle diameter. A smaller $\alpha$ represents dominant coarse aerosol particles while a larger $\alpha$ represents dominant fine particles. $\beta$ is the turbidity coefficient and $\lambda$ is the wavelength.

2.2.3. Volume size distribution

The volume particle size distribution $dV/dlnr (\mu m^3/\mu m^2)$ is retrieved in the range of sizes $0.05 \mu m \leq r \leq 15 \mu m$. For each mode the lognormal distribution is defined as

$$dV = \frac{C_v}{\sigma \sqrt{2\pi}} \exp \left[ -\frac{1}{2} \left( \frac{\ln(r/r_f)}{\sigma} \right)^2 \right].$$

where $dV/dlnr$ is the volume size distribution, the volume concentration $C_v$ is the columnar volume of particles per unit cross section of atmospheric column, $r$ is the particle radius, $r_f$ is the volume geometric mean radius, $\sigma$ is the geometric standard deviation.

2.2.4. Single scattering albedo (SSA)

The single scattering albedo is defined as the ratio of the scattering coefficient and the extinction coefficient and is a key variable in assessing the climatic effects of aerosols (Jacobson, 2000).

$$\omega = \frac{\sigma_{\text{scatt}}}{\sigma_{\text{abs}}},$$

where $\omega$ presents single scattering albedo, $\sigma_{\text{scatt}}$ is scattering coefficient and $\sigma_{\text{abs}}$ is absorption coefficient. The value of SSA is mostly dependent on the shape, size distribution and concentration of aerosol particles. The main source of error in the derived single scattering albedo is due to calibration of the sky data, and is estimated to be $\pm 0.03$.

2.2.5. Refractive index

The real $n(\lambda)$ (1.33 $\leq n(\lambda) \leq 1.6$) and imaginary $k(\lambda)$ parts of the complex refractive index (0.0005 $\leq k(\lambda) \leq 0.5$) are retrieved for the wavelengths corresponding to sky radiance measurements.

2.2.6. Estimation of aerosol radiative forcing and heating rate

The aerosol radiative forcing (ARF) at the top of atmosphere (TOA) is defined as the difference in the net solar flux (in Wm$^{-2}$, downward flux density minus upward one) with and without aerosol, i.e.,

$$\Delta F = (F_{a1} - F_{a}) - (F_{01} - F_{0}) \quad (1)$$

where $\Delta F$ denotes the ARF (in Wm$^{-2}$), and $F_{a}$ and $F_{0}$ denote the global irradiances with aerosol and without aerosols, respectively. Similarly, the ARF can be defined at the Earth surface (i.e., BOA). Furthermore, the resultant net atmospheric forcing ($\Delta F$), representing the amount of solar energy absorbed by aerosols in the atmosphere, is obtained by the difference between ARFs at the TOA and BOA. This study used the Santa Barbara Discrete-ordinate Atmospheric Radiative Transfer (SBDART) model to compute the net flux at the TOA and BOA for atmosphere with and without aerosols (Ricchiazzi et al., 1998).

The SBDART is a public implementation for radiative transfer simulations within the atmosphere under clear and cloudy conditions (Ricchiazzi et al., 1998). The crucial input parameters for the ARF included the AOD, AE, SSA, ASY, surface albedo, and column ozone concentration, all of which can be obtained from the sun-sky radiometer of AERONET Beijing site. Other input parameters used by the model include solar zenith angle, which is calculated by specifying a particular date, time, latitude, and longitude.
The atmospheric heating rate due to aerosol absorption (i.e. \( \Delta F \)) is calculated from the first law of thermodynamics and hydrostatic equilibrium (Liou, 2002) as

\[
\frac{\partial T}{\partial t} = \frac{g}{C_p} \times \frac{\Delta F}{\Delta P}
\]  

where \( \frac{\partial T}{\partial t} \) is the atmospheric heating rate (HR) in Kelvin per day (K day\(^{-1}\)), \( \frac{g}{C_p} \) is the lapse rate with \( g \) and \( C_p \) being the Earth’s acceleration due to gravity and the specific heat capacity of air at constant pressure (i.e. 1 006 J kg\(^{-1}\K\)\(^{-1}\)), and \( \Delta P \) is the atmospheric pressure difference between top and bottom boundaries of the atmosphere.

### 3. Results and discussion

#### 3.1. Aerosol optical depth and Ångström exponent

Annual variation of AODs at 440 nm and AE\(_{440-870}\) with standard deviation over Beijing from 2001 to 2015 are shown in Fig. 3. Average AOD at 440 nm was about 0.62 during the period of 2008 and 2015, whereas from 2001 to 2007 was over 29% larger with an average value of 1.06. This decrease trend before and after 2008 can be attributed to rigorous air pollution control. The annual average AODs between 2001 and 2007 indicated the significant aerosol burden over this region, and Beijing experienced several serious haze events during the period. Correspondingly, these pollution events resulted in high aerosol loading level. The annual mean AOD in Beijing was 0.71 at 440 nm, which was much higher than background sites, such as Longfengshan (0.35 at 440 nm) (Wang et al., 2010a). It was close to measurement in Shanghai (0.74, 440 nm) (Cheng et al., 2015) and Taihu (0.75, 440 nm) (Xia et al., 2007). However, it was lower than measured in some sites near urban or industrial area, such as Xianghe (0.82, 500 nm) (Li et al., 2007), Hefei (0.84, 500 nm) (Wang et al., 2014), and Nanjing (0.88, 440 nm) (Li et al., 2015), which indicated that the heavy aerosol loading could be caused by industrial pollution, urban transportation and large population. Meanwhile, the annual Ångström exponent (AE) didn’t show significant change during the entire period with values around 1.1, and this suggested the dominance of fine-mode particles. Higher AOD and lower AE occurred in 2001, because aerosol data was only provided in spring when dust events are frequently present in Beijing.

Fig. 4 shows monthly and seasonal average AOD at 440 nm, AE at 440–870 nm and water vapor content (WV) with a corresponding number of screened measurement days over Beijing from March 2001 to March 2015. Seasons of Beijing are defined as March-April-May (spring), June-July-August (summer), September-October-
November (fall), and December-January-February (winter). The maximum monthly AOD occurred in June (1.01) followed by October (0.85). It is noted that the minimum monthly AOD occurred in January (0.54). From seasonal pattern, AOD was higher in summer and spring but relatively lower in fall and winter. The results were in good agreement with the measurements in Beijing by Xia et al. (2006) based on 33 months’ aerosol data and in Shanghai by Cheng et al. (2015). However, this seasonal variation of AOD was quite different from that of surface aerosol mass concentration (Xia et al., 2006). This is possibly attributed to the seasonal variation of pollution boundary layer height and relative humidity (aerosol hygroscopic growth). Monthly AE values were normally larger than 1.14 in summer, fall and winter with the highest monthly mean of 1.32 in August. The lowest AE appeared in spring, and implied the significant impact of dust episodes on aerosol size distribution. The seasonal water vapor content showed distinct variation with the largest water vapor content in summer. The lowest water vapor content occurred in winter and was about 0.34 cm.

Fig. 5 shows seasonality of trajectory clusters arriving at 500 m and the mean AOD corresponding to these clusters. The 72-h HYSPLIT air mass back trajectory clusters (cluster medians) at 500 m ending at Beijing site (indicated by the star) were calculated using HYSPLIT 4 model (Draxler and Rolph, 2013; Rolph, 2013) based software-TrajStat (Wang et al., 2009). Air masses are mainly from south and east directions in summer and more complex in other seasons. In general, AOD were significantly higher when air masses move slowly from south directions (AOD > 1.0) and from northeast directions (AOD > 0.7), originating from inland and crossing polluted industrial areas such as Tianjin, Shandong, Henan and Hebei provinces, while AOD were obviously lower when air masses move quickly from north and northwest directions (AOD < 0.5), originating from clean remote continent such as Mongolia or Russia. However, it is noted that high AOD (>0.8) appeared with quick air mass from north direction, which was related to the long-range transport dust.

Fig. 6 presents the contributions of fine and coarse modes on the total AOD at 500 nm for different seasons. Obviously, the AODs were dominated by fine mode AOD in all four seasons, especially during summer. This phenomenon indicated that aerosol extinction was dominated by fine particles. The maximum value of fine
mode AOD occurred in summer, which was closely related to hygroscopic growth of fine particles. The coarse mode AOD in spring was over two times larger than that of other seasons because of the presence of dust events.

The frequency distributions of AOD at 440 nm of the four seasons over Beijing are given in Fig. 7. The frequency histogram of AOD showed a wide range of AOD from 0.1 to 4.2 and commonly an uni-modal structure with peak values between 0.1 and 0.2. The relative frequency of AOD <0.5 were 43%, 34%, 50% and 60% from spring to winter. The frequencies of AOD in the range of 0.5–1.0 were 27%, 28%, 25% and 22% for each season. For high AOD (>1.0), the frequencies were 30%, 38%, 25% and 18% in spring, summer, fall and winter. Especially in summer, about 7% of AOD occurred in the range of 2.0 and 4.0. These results implied that high AOD occurred all year in Beijing. The maximum of AOD occurred in summer was also related to the hygroscopic growth of aerosols and air stagnation.

Fig. 8 shows the seasonal frequency distributions of AE at 440–870 nm in Beijing. Similar to the AOD frequency distribution, the histograms of AE also presented an uni-modal structure centered at 1.1–1.4 for all seasons. The values of AE in the four seasons fluctuated in the range of 0.1–2.0, indicating that there was different size of aerosol particles in the atmosphere. For AE < 0.5, the relative frequencies of AE were about 10%, 1%, 3% and 3% in spring, summer, autumn and winter respectively. High percentage of low AE values in spring indicated the atmosphere influenced by coarse dust particles. The frequencies of AE > 1.0 were 59%, 91%, 86%
and 84% for each season.

3.2. Aerosol types

Following the method shown by Gobbi et al. (2007), Fig. 9 presents the classifications of aerosol properties as a function of $\alpha$ ($440\mathrm{~nm}$) and AE difference $\delta\alpha$ ($\alpha(440,675)-\alpha(675,870)$) for bimodal and lognormal size distributions with a refractive index $m = 1.4 - 0.001i$. Different color codes in Fig. 9 represented the values of AOD at 675 nm. The black solid lines are for a fixed size of the fine mode ($R_f$) and the blue dashed lines are for a fixed fraction contribution ($\eta$) of the fine mode to the AOD at 675 nm. Only cloud-screened data with AOD (675 nm) $>0.15$ were used in this study.

The long-term data in Beijing showed that high AODs ($>1.0$) were clustered in the fine mode growth wing (AODs $>1.0$, $\delta\alpha<0$, $\eta>60\%$) and the coarse mode (AODs $>1.0$, $\delta\alpha>0$, $\eta<30\%$). In spring (Fig. 7a), it was obvious that higher AODs ($>1.0$) were associated with the presence of more fine mode aerosols but negligible variation of AE (except for some dust cases in spring). This pattern was partly similar to that of Shanghai (Cheng et al., 2015) and Nanjing (Li et al., 2015). The main feature revealed that increased AOD was strongly associated with the presence of more fine mode aerosols (so the decrease in AE difference from positive to negative) and with a slight increase in size of fine mode (meaning aging processes of fine particles).

3.3. Seasonal variations of single scattering albedo, volume size distribution, and complex refractive index

Fig. 10 presents the seasonal averages of retrieved SSA, real (Re) and imaginary (Im) parts of refractive index, and aerosol volume size distribution over Beijing during 2001–2015. Seasonal variation of SSA was distinct with high monthly SSA in summer and spring and relatively low values in winter and fall. The averaged SSA values at $440\mathrm{~nm}$ were $0.89 \pm 0.03$, $0.92 \pm 0.03$, $0.88 \pm 0.04$, $0.86 \pm 0.04$ in spring, summer, fall and winter respectively. The low values meant enhancement in absorbing particles like black carbon, which were due to the long-range transport of biomass burning during the fall harvest season and the local urban aerosols during heating period in winter (Liu et al., 2017). However, high SSA occurred in summer...
and spring can be explained by the increase of water content and dust cases. The averaged SSA in spring of Beijing was higher than that of measured in Lahore with averaged value between 0.83 and 0.85 for the pre-monsoon season (Alam et al., 2012, 2014a), but much lower than the results on dusty days (0.92–0.99) (Alam et al., 2014b). The daily SSA at 440 nm ranged from 0.69 to 0.98 during the study period, which suggested that there were quite different types of aerosols in the atmosphere at Beijing (varying from strong absorbing aerosols to strong scattering aerosols). The multi-year average SSA in Beijing (0.89 at 440 nm) was similar to the mean value of columnar atmosphere in Shanghai (0.91 at 440 nm) (Cheng et al., 2015), Taihu (0.90 at 440 nm) (Yu et al., 2011) and Lanzhou (0.90 at 440 nm) (Bi et al., 2011), but much lower than that measured in Hefei (0.95 at 440 nm) (Wang et al., 2014). The difference in SSA was due to the difference in aerosol types, meteorological conditions and geographical locations. Noted that mean SSA of columnar atmosphere was smaller than that of observation in surface. For example, SSA measured at the ground in Beijing was 0.82 at 525 nm (Jing et al., 2015). This may attribute to the variability of aerosol vertical distribution.

The aerosol column-averaged size distribution can be described by a bimodal lognormal distribution (fine and coarse modes). It is evident that the volume size distributions in the coarse mode were higher in spring and lower in winter; the higher values in spring were due to the dust activities. During the spring season, aerosol size distribution showed about 40–50% increase in the volume concentration in the coarse mode as compared to other seasons. The volume size distribution and radius in the fine mode reached the highest in summer, followed by fall, spring and winter, which was consistent with the water vapor content presented in Fig. 4. The radius of fine-mode particles peaked at 0.19 µm in summer, 0.15 µm in fall, winter and spring. The increase in magnitude of the fine mode in summer is related to the process of hygroscopic particle growth and coagulation particles with high temperature and relative humidity.

The real parts of refractive index reached a maximum in spring and minimum in summer, the averages were 1.52 ± 0.05, 1.47 ± 0.05, 1.49 ± 0.05 and 1.51 ± 0.05 in spring, summer, fall and winter, respectively, between 440 nm and 1020 nm. Similar trends have also been found in Wuhan, China, which may be attributed to the dust episodes in spring and winter and frequent rainfall in summer (Wang et al., 2015). But, different seasonal variation occurred in the Indo-Gangetic plains with higher values of real parts during the summer and pre-monsoon (Bibi et al., 2016). The
imaginary parts of refractive index decreased drastically in the wavelength range 440–670 nm, and then increased slightly from 670 nm to 1,020 nm in fall and winter and almost kept the same value in spring and summer. The imaginary parts were higher in fall and winter than those of in spring and summer. Similar seasonal variation has also observed in Hefei by Wang et al. (2014), while is inconsistent with the results in Shanghai (Cheng et al., 2015). Overall, the averages of imaginary parts at the four wavelengths were 0.0079 ± 0.0034, 0.0079 ± 0.0039, 0.0126 ± 0.0058, 0.0135 ± 0.0066 in spring, summer, fall and winter, respectively. These results indicated that the aerosol particles in fall and winter seasons in Beijing were more absorptive to solar radiation than in spring and summer.

3.4. Aerosol radiative forcing and heating rate

Based on the AERONET retrieved aerosol optical properties, the SBDART is used to simulate the ARF in Beijing, and results are compared with values provide by the AERONET. Fig. 11 compares monthly SBDART-simulated and AERONET-retrieved ARF at the TOA and BOA. There is good agreement between the SBDART-simulated and AERONET-retrieved ARF at the BOA and TOA, with the
correlation coefficients and mean differences of 0.99 and 2.16 W m⁻² at the TOA, and 0.97 and 2.51 W m⁻² at the BOA. The good correlation between SBDART and AERONET forcing meant that the main input parameters were appropriate. Uncertainties in our ARF calculation may have arisen because of uncertainty in the retrieval of AOD, SSA, water vapor content, and surface albedo. The overall uncertainty in the estimated radiative forcing has been found to be between 10% and 15% (Alam et al., 2011; Prasad et al., 2007).

The seasonal averaged heating rate and ARF variations at the surface, TOA, and within the atmosphere using SBDART simulation during 2001–2015 at Beijing site are shown in Fig. 12. The monthly ARF at the surface ranged from −77 W m⁻² to −101 W m⁻², while at the TOA they were found to be between −21 W m⁻² and −37 W m⁻². Likewise, the monthly ARF within the atmosphere was between 44 W m⁻² and 68 W m⁻². The ARF at the surface, TOA and atmosphere showed a weak seasonal variation. The averaged ARFs at the surface in spring, summer, fall and winter were −97 ± 59 W m⁻², −87 ± 53 W m⁻², −89 ± 60 W m⁻², and −80 ± 53 W m⁻², while at the TOA they were −33 ± 22 W m⁻², −35 ± 22 W m⁻², −28 ± 20 W m⁻², and −24 ± 23 W m⁻², giving rise to averaged atmospheric forcing of 63 ± 41 W m⁻², 52 ± 36 W m⁻², 60 ± 45 W m⁻², and 55 ± 39 W m⁻² for each season. The differences in TOA forcing in the four seasons was largely because of seasonal variation of SSA, which varied in the range of 0.86–0.92 in the four seasons. The largest negative ARF at the surface and positive ARF within atmosphere were found in spring over Beijing, implying strongest cooling at the surface and warming in the atmosphere. These results were induced by relative large AOD and small SSA. The large ARF difference between TOA and surface demonstrated that solar radiation is being absorbed within the atmosphere, and, as a result, heats the atmosphere and reduces surface temperature (Ge et al., 2010; Miller and Tegen, 1999). This can change atmospheric vertical temperature gradient and influence the dynamic system of the atmosphere (Li et al., 2010).

The monthly average heating rate varied from 1.18 to 1.85 K day⁻¹ with the seasonal average values of 1.78 ± 1.13, 1.47 ± 0.98, 1.70 ± 1.21 and 1.55 ± 1.06 K day⁻¹ during the spring, summer, fall and winter respectively. These values were higher than the results reported by Bibi et al. (2016) on the Indo-Gangetic plains, but it was much lower than those of observed in Central India during the dust and biomass burning episodes (Jose et al., 2016). Similarly, a high heating ratio was found in spring, which can be attributed as enhanced absorption by possible mixing of dust with local atmosphere. However, low heating ratio was observed in summer suggested an increase in water vapor content which enhanced light scattering capacity (high SSA).

4. Conclusions

The long-term characteristics of aerosol optical and radiative properties were investigated based on the AERONET measurement and a numerical radiative transfer model (i.e., SBDART) over urban Beijing between 2001 and 2015. The AOD was higher in summer and spring but relatively lower in fall and winter mainly due to the hygroscopic growth in the wet season and dust events in spring. The seasonal variation of AOD was quite different from that of surface aerosol mass concentration. This is partly attributed to the variation of atmospheric boundary layer height. The frequencies were up to 30%, 38%, 25% and 18% in spring, summer, fall and winter respectively when AOD >1.0. About 7% of AOD in summer ranged between 2.0 and 4.0. Mean AE values were commonly larger than 1.1 in summer, fall and winter, while it was lower than 1.0 in spring due to the presence of dust episodes. The values of AE in the four seasons fluctuated in the range of −0.1–2.0, indicating that there was different size of aerosol particles in the atmosphere. The observation data showed that high AODs (>1.0) were clustered in the fine mode growth wing and the coarse mode.

The volume concentration and radius in the fine mode reached the highest in summer, followed by fall, spring and winter, which was consistent with the water vapor content. The seasonal averages of SSA are presented with mean values of 0.92 ± 0.03 (summer), 0.89 ± 0.03 (spring), 0.88 ± 0.04 (fall), and 0.86 ± 0.04 (winter) at 440 nm. The real part of refractive index reached a maximum in spring, while the imaginary part was highest in winter. The averaged ARFs at the surface in spring, summer, fall and winter were −97 ± 59, −87 ± 53, −89 ± 60, and −80 ± 53 W m⁻², while at the TOA they were −33 ± 22 W m⁻², −35 ± 22 W m⁻², −28 ± 20 W m⁻², and −24 ± 23 W m⁻². The positive ARF within atmosphere in the four seasons indicated heating of the atmosphere. The monthly average heating rate varied from 1.18 to 1.85 K day⁻¹ with the seasonal average values of 1.78 ± 1.13, 1.47 ± 0.98, 1.70 ± 1.21 and 1.55 ± 1.06 K day⁻¹ during the spring, summer, fall and winter respectively. A low heating ratio in summer suggested an increase in water vapor content which enhanced light scattering capacity.

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