Interaction between the Black Carbon Aerosol Warming Effect and East Asian Monsoon Using RegCM4

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ABSTRACT

Black carbon aerosol (BC) has a significant influence on regional climate changes because of its warming effect. Such changes will feed back to BC loadings. Here, the interactions between the BC warming effect and the East Asian monsoon (EAM) in both winter (EAWM) and summer (EASM) are investigated using a regional climate model, RegCM4, that essentially captures the EAM features and the BC variations in China. The seasonal mean BC optical depth is 0.021 over East Asia during winter, which is 10.5% higher than that during summer. Nevertheless, the BC direct radiative forcing is 32% stronger during summer ($1.85 \text{ W m}^{-2}$). The BC direct effect would induce a lower air to warm by 0.11–0.12 K, which causes a meridional circulation anomaly associated with a cyclone at 20°–30°N and southerly anomalies at 850 hPa over East Asia. Consequently, the EAM circulation is weakened during winter but enhanced during summer. Precipitation is likely increased, especially in southern China during summer (by 3.73%). Relative to BC changes that result from EAM interannual variations, BC changes from its warming effect are as important but are weaker. BC surface concentrations are decreased by 1%–3% during both winter and summer, whereas the columnar BC is increased in southern China during winter. During the strongest monsoon years, the BC loadings are higher at lower latitudes than those during the weakest years, resulting in more southerly meridional circulation anomalies and BC feedbacks during both winter and summer. However, the interactions between the BC warming effect and EAWM/EASM are more intense during the weakest monsoon years.

1. Introduction

Black carbon aerosols (BCs) can affect regional and global climates due to their strong absorption of solar shortwave radiation, leading to changes in the air temperature, atmospheric circulation, and hydrologic cycle (e.g., Menon et al. 2002; Forster et al. 2007; Ramanathan and Carmichael 2008; H. Zhang et al. 2009; Zhuang et al. 2011, 2013). The BC warming effect, which is thought to be second after that of carbon dioxide (CO$_2$; Jacobson 2002), could substantially offset the cooling effects by scattering of radiation (Kiehl and Briegleb 1993) and aerosol indirect effects (Zhuang et al. 2013). East Asia has undergone rapid economic and population growth over the past three decades; this growth has been accompanied by high aerosol emissions (Cao et al. 2006; Q. Zhang et al. 2009) and substantial air pollution (Zhang et al. 2008, 2012). The BC emissions in East Asia due to human activities account for at least one-quarter of the BC total global emissions (Streets et al. 2001). In
urban areas of China, BC concentrations exceed 10 μg m⁻³, and the absorbing aerosol optical depth (AAOD) is above 0.06 during high pollution episodes (Zhang et al. 2008, 2012; K. Li et al. 2016). Therefore, the BC warming effect on East Asian climates is significant.

East Asia is under the monsoon climate during both winter and summer (Wang and LinHo 2002). Changes in the East Asian winter monsoon (EAWM) would result in northerly and surface air temperature anomalies (Jhun and Lee 2004), while anomalies in the East Asian summer monsoon (EASM) would induce local floods or droughts (Ding and Chan 2005; Wang et al. 2008). The East Asian monsoon (EAM) could be affected to different extents by the anomalies of Pacific Ocean sea surface temperature, Eurasia’s snow cover, and Tibetan Plateau warming (Ding and Chan, 2005; Wang et al. 2008), as well as by human activities (Guo et al. 2013; Zhou et al. 2014; Song et al. 2014; Jiang et al. 2017). The changes in EAWM or EASM due to both aerosols and natural forcings might further redistribute air pollutants and lead to even poorer regional air quality (Zhu et al. 2012; Z. Q. Li et al. 2016; Wu et al. 2016; Mao et al. 2017).

In recent decades, most of the studies on BC mainly focused on its annual-scale direct/indirect radiative forcing as well as the corresponding climate effects based on observations and climate model simulations (e.g., Boucher and Rodhe 1994; Menon et al. 2002; Liao and Seinfeld 2005; Lau et al. 2006; Forster et al. 2007; IPCC 2013; Bond et al. 2013; Zhuang et al. 2014a; K. Li et al. 2016). The annual mean direct radiative forcing due to BC in urban areas of the Yangtze River Delta in China, East Asia, and worldwide is +4.5 (clear-sky; Zhuang et al. 2014a), from +0.81 to +1.22 (all-sky; Zhuang et al. 2013; K. Li et al. 2016), and +0.40 W m⁻² (all-sky; IPCC 2013), respectively, at the top of the atmosphere. Bond et al. (2013) indicated that the BC direct radiative forcing based on previous studies might be underestimated by a factor of 3. Overall, the strongest BC forcing is found in polluted urban areas, followed by national, regional (East Asia), and global means. Because of the BC warming effect, the atmospheric stability might be affected (Randles and Ramaswamy 2008). Additionally, the variations in precipitation over China in the last 40–50 years of twentieth century, with increased rainfall in the south and drought in the north, were thought to be induced by the BC absorption of solar radiation (Menon et al. 2002). Bollasina et al. (2008) suggested that BC absorption could lead to reductions in cloud amounts and precipitation and increases in surface air temperature in May and, thus, an increase in the June monsoon over South Asia. The amount of clouds could decrease by 1.33%, leading to an increase in the total absorbed solar radiation of 7–9 W m⁻² and a surface warming of 0.3 K in some regions of East Asia (Zhuang et al. 2013). These studies have illustrated the importance of black carbon aerosols in the energy balance of the Earth–atmosphere system and in regional and global climates, suggesting that BC might have substantial influences on the behaviors in modulating the EAM during both winter and summer.

In recent years, the influence of aerosols on Asian monsoon climates has gained attention (e.g., Lau et al. 2006; Lau and Kim 2006; Meehl et al. 2008; Manoj et al. 2011; H. Zhang et al. 2012; Guo et al. 2013; Jiang et al. 2013; Song et al. 2014; Wang et al. 2015; S. Li et al. 2016; Jiang et al. 2017). In South Asia, high absorption aerosol loadings over this region would facilitate the Indian monsoon transition between active and inactive periods (Manoj et al. 2011), which might lead to a delay in the onset of the South Asian summer (Mahmood and Li, 2013). Additionally, Lau et al. (2006) indicated that the BC in northern India during late spring would intensify the Indian summer monsoon. More recently, the effects of the aerosols on the East Asian summer or winter monsoon have also been studied (e.g., Sun et al. 2012; Guo et al. 2013; Jiang et al. 2013; Song et al. 2014; Wang et al. 2015; S. Li et al. 2016; Z. Q. Li et al. 2016). Most of these studies focused on changes in precipitation, circulation, and temperature due to aerosol effects using regional and global climate models. Researchers found that local droughts over East Asia might partially result from aerosol effects. They also suggested that the aerosol cooling effects could decrease the land–sea thermal contrast to a certain degree during summer, subsequently leading to a weaker EASM circulation. Jiang et al. (2017) indicated that the BC warming effect in the Tibetan Plateau would intensify the EAWM. In addition to local aerosols, the effects of non-Asian aerosols on the Asian monsoon have also been investigated (Ramanathan et al. 2005; Kim et al. 2007; Cowan and Cai 2011). Uncertainties with regard to the aerosol effects on the EAM remain, even though a considerable number of studies have examined this issue. For example, Kuhlmann and Quaas (2010) did not find strong elevated heating, which could influence large-scale monsoon circulations as seen in CALIPSO satellite data. Therefore, the effects of aerosols on monsoon climates in East Asia remain an issue of the scientific frontier.

Changes in monsoon activities could also significantly affect aerosol loadings over Asia (Zhu et al. 2012; Wang et al. 2015; Mao et al. 2017). Changes in large-scale atmospheric circulations would lead to shifts in the source.
regions of Asian dust (Zhang et al. 1997). A weaker zonal wind at 850 hPa would cause higher aerosol loadings over East Asia (Bao et al. 2008). Yan et al. (2011) and Liu et al. (2011) noted that the interannual variation in the mean wind intensity in July was a major factor in determining the spatial pattern of aerosols over eastern Asia. Zhang et al. (2010a, b) found that aerosol outflows from South Asia were small during the strong Southeast Asian summer monsoon years, and vice versa. Mao et al. (2017) indicated that the BC concentration over northern China was 3%–11% and 5%–8% higher during weak EASM and EAWM years, respectively, than were those during stronger years. Wang et al. (2015) suggested that changes in the EASM due to the combined effects of all the aerosols would favor the accumulation of aerosols in the boundary layer. These studies suggested that the EAM has a substantial influence on aerosol loadings. Subsequently, changes in monsoon climates due to aerosols might also redistribute the aerosols over East Asia.

Among the studies on the aerosol–monsoon relationship, few have addressed the interactions between BCs and the EAM (including EAWM and EASM), as well as the interactions occurring during the stronger or weaker monsoon years. In consideration of the high BC loadings in Asian areas and the active EAM, the interactions between BC and EAWM/EASM are investigated in this study using a new version of the regional climate model RegCM4 (Giorgi et al. 2012), which is combined with a recent emission inventory from the Multiresolution Emission Inventory for China (MEIC) of Tsinghua University, China (http://www.meicmodel.org). This study could enhance our understanding of the interactions between BC aerosols and the EAM. The model description and methods are described in section 2. The simulation results are presented and discussed in detail in section 3. The conclusions are provided in section 4.

2. Methods

a. Description of the regional climate model RegCM4

Regional climate models are known to have higher resolutions and can reproduce more realistic, small-scale features in a “perfect prognosis” approach than global climate models can (Denis et al. 2002). Here, an updated version of the regional climate model RegCM4 (Giorgi et al. 2012), which was developed by the Abdus Salam International Center for Theoretical Physics, is applied because of its extensive use in regional climate studies (Giorgi et al. 2002; Li et al. 2009; Zhuang et al. 2010; Solmon et al. 2012; Yin et al. 2015; S. Li et al. 2016). RegCM4 has much more comprehensive modes for describing natural and anthropogenic processes than does the previous version RegCM3. To address trace gases, a gas phase chemistry module with the Carbon Bond Mechanism, version Z (CBMZ), was added (Shalaby et al. 2012). Inorganic aerosols such as sulfate and nitrate are calculated by coupling the “ISORROPIA II” thermodynamic equilibrium model (Fountoukis and Nenes 2007; S. Li et al. 2016). The gas-particle partition process is addressed by implementing a volatility basis set model (Yin et al. 2015). Thus, the current version of RegCM4 could simulate nearly all natural (dust and sea salt) and anthropogenic aerosols (or primary and secondary aerosols) as well as the direct and indirect effects of aerosols on climate (S. Li et al. 2016). Additionally, RegCM4 also involves new schemes for the land surface boundary layer. The radiative transfer package from the National Center for Atmospheric Research Community Climate Model, version 3 (Kiehl et al. 1996), is used in RegCM4 to investigate the aerosol direct radiative forcing. The shortwave radiation transfer follows the δ–Eddington approximation, including 19 spectral intervals (Kiehl et al. 1996).

b. Optical depth of BC aerosols

In RegCM4, the wavelength-dependent BC optical depth \( \tau(\lambda) \) is calculated based on a given BC mass concentration and wavelength-dependent specific or mass extinction coefficient as follows (Kasten 1969):

\[
\tau(\lambda) = M \beta_{\lambda i} (1 - RH)^{-\kappa_i},
\]

where \( i \) is an index for the carbonaceous aerosols [including hydrophilic BC and organic carbon (OC) and hydrophobic BC and OC]. For hydrophobic and hydrophilic BC, coefficients \( \kappa_i \) are equal to 0 and 0.25, respectively. Also, \( M \) is the mass concentration of aerosol \( i \) (g m\(^{-3}\)), \( \lambda \) is the wavelength, \( \beta_{\lambda i} \) is the wavelength-dependent mass extinction coefficient (m\(^2\) g\(^{-1}\)) of aerosol \( i \) at wavelength \( \lambda \), and RH is the relative humidity. Coefficient \( \beta_{\lambda i} \) is a function of the density, size distributions, and the refractive indices of the aerosols (Huang et al. 2007).

c. Emissions and simulation schemes

An updated BC emission inventory from 2010 (http://www.meicmodel.org) is applied. This inventory is from the MEIC, which was compiled by Tsinghua University (Li et al. 2017). Tsinghua University also developed an earlier trace gas and aerosol emissions inventory in 2006 to support the Intercontinental Chemical Transport
Experiment phase B based on a series of improved methods (Q. Zhang et al. 2009). The BC emissions from 2010 are mainly from four sectors, including residential (dominant), industry, energy, and transportation. Figure 1 shows the seasonal mean BC emission rates (in grams per grid per second) from all the sectors of East Asia during both winter [December–February (DJF)] and summer [June–August (JJA)]. Figure 1 suggests that both India and China have high BC emissions. In China, the BC emission rates are high mostly over southwest (Sichuan basin), central to north, and northeast China, as well as the Yangtze River Delta (YRD) and Pearl River Delta (PRD), with a maximum exceeding 150 grams per grid per second (~4.17 × 10^-8 g m^-2 s^-1). The emissions are higher during winter than during summer and the seasonality is higher over China than over India. The seasonal BC emission within the simulated domain shown in Fig. 1 is 0.97 Tg during winter and 0.70 Tg during summer, accounting for approximately 29% and 22%, respectively, of the annual emissions in 2010.

The study domain shown in Fig. 1, centered at 29.5°N, 106.0°E, covers most parts of East Asia and parts of South Asia with a horizontal resolution of 60 km. The model is run with 18 layers, ranging from the surface to 5 hPa, in the vertical direction. The weekly mean sea surface temperature (SST) from NOAA Optimum Interpolation SST V2 (Reynolds et al. 2002) is employed and maintained as fixed values (using the same SST field in all experiments) to exclude the effects of SSTs when investigating the BC climate effects. Similarly, the CO₂ concentrations are also fixed at a preindustrial value of 284.7 ppm to exclude the warming effect of CO₂. The National Centers for Environmental Prediction (NCEP) reanalysis data (NNRP2), which have a horizontal resolution of 2.5°, are used to drive RegCM4. The climatological BC lateral boundary conditions are from a global chemical transport model [Model of Ozone and Related Chemical Tracers (MOZART)] (Horowitz et al. 2003; Emmons et al. 2010). RegCM4 is run from 1987 to 2009, with the first year as a spinup period. Only the simulations of winter (DJF) and summer (JJA) during the 21-yr period were analyzed. To understand the interactions between the BC direct effect and EAM, two numerical experiments were conducted. One experiment is a control run that does not consider BC effects. The other experiment considers the direct effects of BC. The differences between these two experiments are defined as the climate effects. Student’s t tests are used to assess the statistical significance of the differences.

3. Results and discussion

3.1 Model validations

To validate the performance of RegCM4 in simulating regional climate over East Asia, the simulated seasonal mean air temperature, specific humidity, and wind field from the control run during winter and summer were compared with those from the NCEP reanalysis data from the surface to the middle troposphere (Fig. 2), supplementing the previous validations for this model from monthly to annual scales over East Asia (Wang 2009).
FIG. 2. Comparisons between (a),(c) the NCEP reanalysis and (b),(d) simulated meteorological variables, which include seasonal mean air temperature (shaded; K), specific humidity (contours; g kg\(^{-1}\)), and wind (arrows; m s\(^{-1}\)) from the near-surface level to 550 hPa over Asia during (top),(top middle) winter and (bottom middle),(bottom) summer.
et al. 2010; Sun et al. 2012; Zhou et al. 2014; Yin et al. 2015; S. Li et al. 2016). Figures 2a and 2c show the air temperature (shading), specific humidity (contours), and wind field (vectors) from the reanalysis data during winter and summer, respectively, and Figs. 2b and 2d show the parameter values from the simulations, accordingly. The simulated air temperature and humidity are lower in the lower troposphere, while the wind speed is stronger over East Asia during winter and summer. Additionally, the simulated southwesterly winds over East Asia in the lower atmosphere are more westerly during summer than they are in the reanalysis data. Overall, RegCM4 could generally capture the main features of the distributions and magnitudes of the atmospheric thermodynamic fields and humidity during both winter and summer over East Asia, consistent with the validation from Wang et al. (2010), Sun et al. (2012), and Zhou et al. (2014); however, some differences between the simulations and reanalysis data remain.

The simulated seasonal mean surface concentrations of BC are compared against observations over 15 sites in China, including rural and urban areas from south to north and from west to east China, to validate the ability of RegCM4 to reproduce the BC loadings. The site information is listed in Table 1. The BC concentrations from 14 sites are from Zhang et al. (2008, 2012), who conducted 24-h samplings every 3 days in 2006 and 2007. These data are widely used as representative values for these sites (Zhuang et al. 2011; K. Li et al. 2016; Mao et al. 2017). The BC in Nanjing was measured online by aethalometer from 2012 to 2014 (Zhuang et al. 2014b). Thus, these values are comparable. Figure 3 shows a comparison of the seasonal mean surface BC concentrations between the simulations and observations during winter and summer. The seasonal means of the simulated BC values averaged over these sites are 5.49 \( \mu g m^{-3} \) during winter and 3.27 \( \mu g m^{-3} \) during summer, which are 2.39 and 1.03 \( \mu g m^{-3} \) smaller, respectively, than those of the observations. Overall, the simulated BC is underestimated at most sites, especially during winter. However, both the spatial distribution and seasonality are essentially captured over East Asia by RegCM4. The linear correlation coefficients between the simulated and observed BC are 0.63 and 0.85 during winter and summer, respectively, and both are statistically significant at the 95% confidence level. High BCs mainly appear in northern (Gucheng), southwestern (Chengdu), and central (Xi’an and Zhengzhou) China, especially in urban

<table>
<thead>
<tr>
<th>Site</th>
<th>Name</th>
<th>Location ('N, 'E)</th>
<th>Type</th>
<th>Reference</th>
</tr>
</thead>
<tbody>
<tr>
<td>CD</td>
<td>Chengdu</td>
<td>30.65, 104.04</td>
<td>Urban</td>
<td>Zhang et al. (2008, 2012)</td>
</tr>
<tr>
<td>DL</td>
<td>Dalian</td>
<td>38.90, 121.63</td>
<td>Urban</td>
<td></td>
</tr>
<tr>
<td>DH</td>
<td>Dunhuang</td>
<td>40.15, 94.68</td>
<td>Rural</td>
<td></td>
</tr>
<tr>
<td>GLS</td>
<td>Gaolanshan</td>
<td>36.00, 105.85</td>
<td>Rural</td>
<td></td>
</tr>
<tr>
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<td>Gucheng</td>
<td>39.13, 115.80</td>
<td>Rural</td>
<td></td>
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<td>Jinsha</td>
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<td>Rural</td>
<td></td>
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<td>Lhasa</td>
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<td>Urban</td>
<td></td>
</tr>
<tr>
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<td>Linan</td>
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<td>Rural</td>
<td></td>
</tr>
<tr>
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<td>Longfengshan</td>
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<td>Rural</td>
<td></td>
</tr>
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<td>Nanning</td>
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<td>Urban</td>
<td></td>
</tr>
<tr>
<td>PY</td>
<td>Panyu</td>
<td>23.00, 113.35</td>
<td>Urban</td>
<td></td>
</tr>
<tr>
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<td>Taiyangshan</td>
<td>29.17, 111.71</td>
<td>Rural</td>
<td></td>
</tr>
<tr>
<td>XA</td>
<td>Xian</td>
<td>34.43, 108.97</td>
<td>Urban</td>
<td></td>
</tr>
<tr>
<td>ZZ</td>
<td>Zhengzhou</td>
<td>34.78, 113.68</td>
<td>Urban</td>
<td></td>
</tr>
<tr>
<td>NJ</td>
<td>Nanjing</td>
<td>32.05, 118.78</td>
<td>Urban</td>
<td>Zhuang et al. (2014b)</td>
</tr>
</tbody>
</table>
areas, which are linked to the higher emissions; these results are shown in Fig. 1.

b. Optical depth and effective direct radiative forcing of BC

The distributions of the seasonal mean BC optical depth (BC AOD) and effective direct radiative forcing (EDRF; the net solar shortwave flux difference between perturbed and control experiments) are shown for winter and summer (Fig. 4). High BC AODs appear in southwestern (Sichuan basin) and central to northern China over East Asia during both seasons, which is mostly consistent with the spatial pattern of the emissions. Furthermore, the magnitudes and spatial distributions of the AOD both have substantial seasonality. As a result of a larger emission rate (Fig. 1), a higher BC AOD is found during winter, with a maximum exceeding 0.09, which is at least 1.2 times greater than that during summer. Figure 4 and Fig. 2 indicate that atmospheric circulation could affect the spatial pattern of the AOD. The prevailing winds in the lower troposphere are northerlies in winter, which is the opposite of those during summer over East Asia. Thus, the BC AODs during winter are generally slightly more southerly than those during summer in these regions, although the BC emissions show weaker seasonality in
these regions. These results suggest that the atmospheric circulation might significantly influence aerosol loadings, which was also suggested by Liu et al. (2011) and Mao et al. (2017). Thus, the aerosol effects on regional climate changes might differ considerably between winter and summer and between different monsoon years.

The instantaneous (from experiment 1, not shown) and effective direct radiative forcing (DRF in Fig. 4) of BC show a similar spatial distribution to AOD over East Asia, especially for the former (Zhuang et al. 2013). Different from EDRF, the instantaneous DRF, which is defined as the forcing during clear-sky conditions, excludes the influence of cloud cover and climate feedbacks. Because of the absorption of solar radiation, BC could exert a positive direct radiative forcing at the top of the atmosphere (TOA) but a negative forcing at the surface. The TOA-effective DRF is concentrated in the middle to low latitudes during winter but in the middle to high latitudes during summer. The DRF is stronger than +2.5 W m\(^{-2}\) in most of East Asia, and its maximum exceeds 6 W m\(^{-2}\). The BC DRF during summer is as strong as or even stronger than that during winter in some regions, although the BC AOD is lower during this season because of a higher solar altitude angle as implied in Zhuang et al. (2014a). Additionally, during both winter and summer, BC DRFs are strong over the western Qinghai Tibet Plateau region (QTP) but weak in the oceans (such as the Yellow and Bohai Seas around northern China) because of different surface albedo (Zhuang et al. 2014a). The differences between the instantaneous DRF and effective DRF are also related to the changes in cloud cover induced by the BC warming effects. Increases in the cloud amount would directly decrease the solar radiation reaching the surface, and vice versa.

The regional mean BC AOD and DRF values over northern China (NC; 30\(^\circ\)–45\(^\circ\)N, 108\(^\circ\)–122\(^\circ\)E), southern China (SC; 20\(^\circ\)–30\(^\circ\)N, 108\(^\circ\)–122\(^\circ\)E), and East Asia (20\(^\circ\)–45\(^\circ\)N, 100\(^\circ\)–130\(^\circ\)E) are summarized for both winter and summer (Table 2), showing the substantial seasonality of BC AOD and DRF. The regional mean BC AOD over East Asia is 0.021 during winter and 0.019 during summer, leading to a corresponding positive TOA forcing of 1.33 and 1.44 W m\(^{-2}\) for instantaneous DRF and 1.36 and 1.85 W m\(^{-2}\) for effective DRF for both seasons. The efficiency, defined as ratio of instantaneous TOA DRF to BC AOD, is 68, 80, and 75 W m\(^{-2}\)/AOD over SC, NC, and East Asia, respectively, during summer, all of which are approximately 10 W m\(^{-2}\)/AOD larger than that during winter, although differences exist among these regions. Therefore, both the TOA and surface DRFs are substantially stronger during summer than during winter.

The BC AOD and DRF have different spatial distributions in different seasons. A very high AOD is found in the lower latitudes in winter whereas it appears in higher latitudes in summer as a result of the EAWM and EASM influences, respectively. The instantaneous TOA DRF is much stronger in NC than in SC, possibly resulting from the difference in surface albedo between the two regions (Zhuang et al. 2014a).

The TOA EDRFs over East Asia are as strong as the instantaneous DRFs during both winter and summer, while the surface EDRFs are much weaker. During winter, the EDRF is nearly 1 W m\(^{-2}\) stronger in SC than in NC, which might be related to the feedbacks of the column cloud cover due to the BC warming effects. Changes in the cloud amount during winter are +0.20% in NC, approximately 3 times that in SC.

There have been many modeling and observational studies on the optical properties and direct radiative forcing from the total BC emissions over East Asia. Wu et al. (2008), Zhuang et al. (2013), and K. Li et al. (2016) showed that the annual mean BC DRF at the TOA is approximately +0.32, +0.81 and +1.46 W m\(^{-2}\) in all-sky conditions, as derived from different emission inventories (1.01, 1.81, and 1.84 Tg yr\(^{-1}\), respectively) over East Asia (100\(^\circ\)–130\(^\circ\)E, 20\(^\circ\)–50\(^\circ\)N). The clear-sky BC DRF at the TOA is approximately 1.37 times that of the all-sky BC DRF (Zhuang et al. 2011). From this study, the BC DRF over East Asia is at the same order of magnitude as that in the literature. In urban areas, Zhuang et al. (2014a) showed that the observed absorption aerosol TOA DRF was about +4.5 W m\(^{-2}\) in 2011 and 2012. In this study, the seasonal mean BC DRF at the TOA

<table>
<thead>
<tr>
<th>Types</th>
<th>BC AOD</th>
<th>IDRF_TOA</th>
<th>IDRF_SRF</th>
<th>EDRF_TOA</th>
<th>EDRF_SRF</th>
</tr>
</thead>
<tbody>
<tr>
<td>SC</td>
<td>0.029</td>
<td>1.69</td>
<td>-6.05</td>
<td>2.45</td>
<td>-1.00</td>
</tr>
<tr>
<td>NC</td>
<td>0.030</td>
<td>2.14</td>
<td>-5.63</td>
<td>1.52</td>
<td>-1.36</td>
</tr>
<tr>
<td>East Asia</td>
<td>0.021</td>
<td>1.33</td>
<td>-4.05</td>
<td>1.36</td>
<td>-0.88</td>
</tr>
</tbody>
</table>

Table 2. Regional mean values of the BC optical depth and direct radiative forcing (W m\(^{-2}\)) in southern and northern China and over East Asia during winter and summer. SC is southern China (108\(^\circ\)–122\(^\circ\)E and 20\(^\circ\)–30\(^\circ\)N), NC is northern China (108\(^\circ\)–122\(^\circ\)E and 30\(^\circ\)–45\(^\circ\)N), and East Asia spans 100\(^\circ\)–130\(^\circ\)E and 20\(^\circ\)–50\(^\circ\)N. SRF indicates “at the surface.”
is 3.09 W m$^{-2}$ during winter and +2.47 W m$^{-2}$ during summer in Nanjing; these values are smaller than the observations. However, the simulated BC DRF is larger than the global means (IPCC 2013).

c. Interactions between the BC direct effects and EAM during both winter and summer

Because of its absorption of solar radiation, BC can affect the atmospheric thermodynamic fields and hydrologic cycle. The regional climate responses to the BC effects are assessed from the differences between experiments 1 and 2. The figures in this section all illustrate the net effect, including the feedbacks of the climate system to the BC DRFs.

Figure 5 presents the changes in vertical shortwave heating rate (SWHR) and meridional circulation, which are averaged from 105° to 125°E, where high BC loadings appear during both winter and summer. The BC mainly concentrates near the lower troposphere, and its concentrations decrease with height. During summer, a stronger vertical exchange would bring more BC to the upper troposphere than that during winter. Because of the presence of black carbon in the atmosphere, the SWHR is considerably increased, especially in high BC loading layers, all of which exceed $3.5 \times 10^{-6}$ K s$^{-1}$ (~0.3 K day$^{-1}$). Changes in SWHR are more northerly and extend higher into the atmosphere during summer than during winter. Increases in SWHR would directly heat the air and induce a change in atmospheric circulation, which shows a meridional circulation anomaly in the middle to lower latitudes. Thus, ascending motion is enhanced in the middle latitudes, where the SWHR increases the most, during both winter and summer. During winter, the meridional circulation response to the BC warming effects is confined to the subtropics (15°–31°N, from the surface to 400 hPa) and is more southerly. However, the response in summer is much stronger and could extend up to the top of the troposphere. Changes in atmospheric circulation due to the BC warming effect would be in favor of EASM circulation development but unfavorable to EAWM circulation development. The BC warming effect on the meridional circulations is the opposite of the effects of total aerosols (Wang et al. 2015).

An increase in SWHR would also lead to a warmer atmosphere below 850 hPa in most of East Asia, especially during winter (Fig. 6). The warmer regions are mostly found in SC during winter and in regions from southern to NC during summer. Changes in the air temperature are weaker in southeast China. The air temperature (TA) responses during winter and summer are all larger than 0.17 K. In addition to the direct heating effect of BC, changes in cloud cover (CA) can also cause changes in TA (Zhuang et al. 2013). The spatial TA response during summer (Fig. 6) is well correlated to the changes in cloud amount (not shown) in the lower troposphere in some regions, with negative CA anomalies corresponding to positive TA changes and vice versa. For example, an increase in CA (by approximately 1%) in the middle latitudes (25°–30°N)}
somewhat weakens the BC warming effect in corresponding regions. Overall, because of the BC warming effect, the temperature gradient between the land and ocean over East Asia is enhanced during summer but weakened during winter, which to some extent would affect the EASM and EAWM circulations. Figure 7 presents the changes in horizontal winds at 850 and 230 hPa during winter and summer. The BC warming effects could induce a cyclone (or convergence) anomaly at 850 hPa in southwest to central China during winter, whereas during summer the anomaly occurs from southwest to northeast China. Thus, the southerly wind anomaly appears over southern East Asia during both winter and summer. In the upper troposphere, the changes in wind fields are opposite those at 850 hPa and exhibit an anticyclone anomaly. The divergence region covers nearly the entire East Asia region during winter and covers western China and the Japan Sea during summer. Thus, the northerly wind anomaly is formed in the upper troposphere in southern East Asia during both seasons. Additionally, easterly anomaly is found at approximately 40°N in the upper troposphere due to the BC warming effect. The responses of winds at 850 hPa to the BC warming effect can be advantageous to the EASM development, which is consistent with the results of Wang et al. (2015). Changes in the horizontal wind fields are well linked to the vertical direction anomaly, as shown in Fig. 5. The responses to the BC effects are much stronger during summer (>1.5 m s⁻¹ at 850 hPa) than during winter (<1.0 m s⁻¹ at 850 hPa). Because of the BC warming effect, the precipitation during summer is increased to a certain degree in southern China, the lower reaches of the Yangtze River, parts of northern China, and northeastern Asia, and the largest change is at least 0.2 mm day⁻¹ (Fig. 8). In the north Hetao region (near 40°N, 112°E), the summer precipitation is reduced. The response of rainfall to the BC effects during winter is much weaker than that during summer. Local BC-induced floods mostly occur due to an increase in moisture transport and vice versa, as suggested by the comparison between Fig. 7a and Fig. 8. A convergence anomaly in East Asia during summer favors an increase in rainfall in southern and eastern China. However, the semidirect effect of BC might weaken the precipitations (Zhuang et al. 2013). The rainfall changes in the figure reflect the net effect resulting from all these affecting factors. Previous studies have suggested that variations in summer precipitation, with increasing rainfall in southern East Asia and droughts in northern East Asia over the last five decades, may be caused by variations in BC emissions (Menon et al. 2002). In contrast, Lau and Kim (2006) and Meehl et al. (2008) noted that the total precipitation is increased over East Asia during summer due to aerosols. This study is more consistent with that of Menon et al. (2002), although local droughts in NC are not found.

As discussed, BC can alter the atmospheric thermal fields, dynamic circulations, and the hydrological cycle to a certain degree during both winter and summer, which subsequently leads to a redistribution of the BC
loadings. Consequently, a change in atmospheric circulation and precipitation due to the BC warming effect may decrease the surface BC concentrations (Fig. 9a). During winter, a relatively large response of BC concentrations to the BC warming effect appear in the Sichuan basin of southwest China and the coastal regions of east China, where precipitation is enhanced. During summer, a considerably large decrease in BC is found in the south, southwest, and east to northeast China, where the rainfall is also increased. The largest decrement exceeds 0.25 μg m⁻³ during both winter and summer. Negative changes in BC loading could extend up to 700 hPa, especially during the warmer season (Fig. 9b). In addition to precipitation, the atmospheric circulation could also affect the BC loadings, as shown in Fig. 9b, showing the concurrent response of BC with the meridional circulation anomaly in the vertical direction. The BC in the lower troposphere might be transported to higher altitudes due to the ascending motion anomaly during both winter and summer, leading to an accumulation of BC in the higher altitudes of SC at 105°E (not shown here). Because of a stronger response of the atmospheric circulation during summer (shown in Fig. 5), the largest increase in BC concentrations appears at 550 hPa, whereas during winter BC only peaks at 850 hPa at 105°E. Nevertheless, the absolute changes in BC loadings in the vertical direction during winter have the same order of magnitude as those during

**Fig. 7.** Changes in wind fields (arrows; m s⁻¹) at (a) 850 and (b) 230 hPa from the BC warming effect over East Asia during (left) winter and (right) summer. The purple shading indicates the 95% confidence level from Student’s t test.
In addition, the BC accumulation induced by the vertical circulation anomaly nearly covers all SC during winter. However, this is not the case during summer, possibly due to a substantial flood anomaly, as shown Fig. 9b, in which the responses of BC concentration and meridional circulation in the vertical direction are averaged from 105° to 120°E. The negative changes in BC loading are obviously greater at lower altitudes than at higher altitudes and more southerly during winter than during summer (Fig. 9b). Consequently, the columnar BC is increased in southern China during winter, leading to an increase in BC AOD and TOA instantaneous direct radiative forcing in clear-sky conditions (IDRF) exceeding 0.0005 and 0.05 W m⁻², respectively (Fig. 9c). Changes in columnar BC are less significant than surface BC concentrations during summer, which is opposite of those during winter (Figs. 9a,c).

Table 3 summarizes the regional mean absolute change in TA below 850 hPa, relative changes in precipitation (PR), surface BC concentration (SBC), BC AOD, and BC IDRF at the TOA in southern and northern China and East Asia during both winter (DJF) and summer (JJA). Table 3 clearly shows the interactions between the BC warming effect and East Asian climates. Overall, the BC direct effect could lead to regional warming and flooding during both winter and summer, consequently decreasing surface BC loadings. The BC AOD and IDRF at TOA are increased over East Asia during winter due to the considerable accumulation of column BC in south China. The effects of BC on TA in the lower troposphere over East Asia are significant during both winter and summer. Changes in rainfall and BC feedbacks are more substantial during summer than during winter. Over China, the responses to the BC warming effect show substantial seasonal and spatial variations. Generally, changes in the PR, SBC, BC AOD, and BC IDRF are stronger during summer than during winter in both southern and northern China. The warming in the lower troposphere due to BC is much stronger in southern China than that in northern China during winter, which is opposite of that during summer. Feedbacks of BC AOD and TOA IDRF are both larger in southern than in northern China during both seasons.

d. Interactions between BC and the EAM during strong and weak monsoon years

FIG. 9. Changes in (a) BC concentration (µg m⁻³) at the surface (shaded), (b) vertical direction of the altitude–longitude section averaged from 105° to 120°E, and (c) BC optical depth (shaded in the upper panel; ×10⁻²) and instantaneous direct radiative forcing (shaded in the lower panel; W m⁻²) from the BC warming effect over East Asia during (left) winter and (right) summer. The black dots in (a) and (c) and purple shadings in (b) indicate the 95% confidence levels from Student’s t test.
meteorological features are considerably different from each other between the strong and weak monsoon years. Mao et al. (2017) further indicated that the interannual variations in both EAWM and EASM have a substantial influence on aerosols loadings, direct radiative forcing, and their spatial distributions. Thus, the interaction between the BC warming effects and EAWM/EASM is further investigated during both strong and weak monsoon years. The strongest and weakest EAM years derived from Mao et al. (2017) are directly applied in this study.

Figure 10 shows the differences in the 850-hPa wind fields, BC AOD, and IDRF between the strongest and weakest EAM years during both winter and summer. Substantially, the differences in the 850-hPa wind fields between the strongest and weakest EAWM years appear mostly in eastern East Asia, including the northeast and southeast regions. Northwest winds in northern China and northeast winds in southern China are more anomalous during stronger EAWM years than during the weakest EAWM years, which causes decreases in BC AOD in north and northeast China but increases in BC AOD along the Yangtze River basin during winter, with a maximum exceeding 0.004. Consequently, the BC IDRF is at least 0.1 W m\(^{-2}\) stronger in the middle latitudes over East Asia but 0.15 W m\(^{-2}\) weaker in northeastern East Asia. Substantial differences in the wind fields at 850 hPa between the strongest and weakest EASM years are mainly found in southern and northeastern China. Relative to the weakest EASM years, the northeast winds in SC and northwest winds in northeast China are anomalous during the strongest EASM years. Thus, the BC AOD is decreased in northeastern East Asia by 0.0015 but increased in southern China by 0.004. The changes in BC IDRF between the strongest and weakest EASM years are similar to those of AOD, becoming weaker in northeast China by at least 0.15 W m\(^{-2}\) and stronger in southern China by over 0.3 W m\(^{-2}\). In this study, the changes in wind fields at 850 hPa between the strongest and weakest monsoon years are consistent with those of Mao et al. (2017). In the vertical direction (not shown), a clockwise meridional circulation difference appears from latitudes 10\(^{\circ}\) to 32\(^{\circ}\)N between the strongest and weakest EAWM years, while an anticyclonic meridional circulation difference is found from 5\(^{\circ}\) to 20\(^{\circ}\)N between the strongest and weakest EASM years; this finding is consistent with the dominant circulations of EAWM and EASM, respectively. The differences in meteorological factors (Fig. 10a) between the strongest and weakest monsoon years also suggest that the warming effect of the Asian BC might be unfavorable to the development of EAWM circulation but favorable to the development of EASM circulation. Because of the BC warming effect, positive southerly and southwesterly anomalies at 850 hPa appear at low latitudes (<20\(^{\circ}\)N) in both winter and summer as shown in Fig. 7a. These anomalies are opposite to that in winter but consistent with that in summer in Fig. 10a.

Regional mean changes in BC AOD and TOA IDRF due to the interannual variations in EAWM and EASM are listed in Table 4. Both the strongest EAWM and EASM would result in larger BC AOD and stronger TOA IDRF in southern China, which is the opposite of those in northern China. Consequently, the interactions between BC and the EAM might be different, which will be discussed in the following section. The changes in BC AOD and IDRF due to the interannual variations in EAWM and EASM have the same order of magnitude, albeit slightly larger, as those caused by the BC warming effect (Tables 3 and 4).

1) WINTER

As discussed, the BC warming effects would induce an anomaly of meridional circulation during both winter and summer. However, differences exist between the strongest and weakest EAWM years (Fig. 11a). The sinking motion of the anomalous circulation could extend to lower latitudes (near 10\(^{\circ}\)N) during the strongest EAWM years, while it appears at approximately 15\(^{\circ}\)N, possibly because of a higher BC AOD and subsequently the stronger warming effect in southern China during the strongest EAWM years. The air temperature in the lower troposphere during different monsoon years responds to the BC warming effect quite differently in East Asia (Fig. 11b). A higher AOD and stronger IDRF

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**Table 3.** Regional means of the changes in air temperature (K) below 850 hPa and relative changes (%) in precipitation, SBC, BC optical depth, and direct radiative forcing due to the BC warming effect in southern and northern China and over East Asia during winter and summer (SC, NC, and East Asia are defined as in Table 2).

<table>
<thead>
<tr>
<th>Types</th>
<th>TA (K)</th>
<th>PR (%)</th>
<th>SBC (%)</th>
<th>BC AOD (%)</th>
<th>BC IDRF (%)</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>DJF</td>
<td>JJA</td>
<td>DJF</td>
<td>JJA</td>
<td>DJF</td>
</tr>
<tr>
<td>SC</td>
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<td>0.11</td>
<td>-0.80</td>
<td>3.73</td>
<td>-0.75</td>
</tr>
<tr>
<td>NC</td>
<td>0.09</td>
<td>0.14</td>
<td>1.93</td>
<td>2.81</td>
<td>-0.96</td>
</tr>
<tr>
<td>East Asia</td>
<td>0.12</td>
<td>0.11</td>
<td>1.16</td>
<td>3.38</td>
<td>-1.17</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td>0.75</td>
<td>2.94</td>
<td>0.45</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td>2.35</td>
<td>-3.98</td>
<td>-0.49</td>
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<tr>
<td></td>
<td></td>
<td></td>
<td>2.07</td>
<td>+2.37</td>
<td>-4.13</td>
</tr>
</tbody>
</table>
FIG. 10. Differences (a) in the wind fields at 850 hPa (arrow; m s\(^{-1}\)) and (b) in the BC optical depth (shaded in the upper panel; \(\times 10^{-2}\)) and instantaneous direct radiative forcing (shaded in the lower panel; W m\(^{-2}\)) between the strongest and weakest monsoon years during (left) winter and (right) summer. The purple shading in (a) and black dots in (b) indicate the 95% confidence levels from Student’s \(t\) test.
in southern China and a lower AOD and weaker IDRF in northern China (Table 4) could lead to positive changes in air temperature, which are more concentrated in the lower latitudes of East Asia during the strongest EAWM years than those during the weakest EAWM years. In the weakest EAWM years, a cyclone anomaly induced by BC at approximately 30°N is more easterly and stronger, whereas an anticyclone anomaly is more southeasterly and weaker than that during the strongest EAWM years (not shown here). Changes in EAWM due to BC would subsequently be in favor of BC accumulation in the vertical direction. The largest increases in BC concentrations appear between 850 and 700 hPa during both the strongest and weakest EAWM years. However, the peak is more southerly during the strongest EAWM years due to a stronger BC warming effect in southern China (not shown here).

2) SUMMER

Similar to winter, the BC AOD during summer is higher in southern China but lower in central to northern China during the strongest EASM years than that during the weakest EASM years (Table 4). A substantial increase in the BC AOD could extend west from southern China to India (Fig. 10b). The BC warming effects would induce an anomaly of meridional circulation during both the strongest and weakest EASM years (Fig. 12a). However, the circulation anomaly during the strongest EASM years is more southerly and slightly weaker than that during the weakest years. Higher BC loadings in northern China would lead to a more substantial ascending motion anomaly in the higher latitudes (40°–50°N) during the weakest EASM years. The BC during summer would lead to regional warming in most of East Asia but regional cooling in northeast China during both strong and weak monsoon years (Fig. 12b). The response of air temperature in the lower troposphere during regional warming or cooling exceeds 0.3 K. However, the air temperature during the strongest EASM years responds to the BC’s direct effects in a slightly weaker, more westerly, and more concentrated manner. Therefore, a cyclone anomaly induced by the BC at 850 hPa is found in central to eastern China during the strongest EASM years, whereas it appears over large areas from central to northeast China during the weakest EASM years (not shown). Consequently, the southwest wind anomaly, which is part of the cyclone, is more northerly and weaker during the strongest EASM years. The difference in wind fields at 850 hPa between the strongest and weakest monsoon years during summer would result in different precipitation because the wind field anomaly at 850 hPa would lead to a moisture transport anomaly. Therefore, local floods induced by the BC are found in the middle and lower reaches of the Yangtze River, southwest China, and parts of northeast China during the strongest EASM years, whereas floods occur in nearly all of south China and parts of the Japan Sea during the weakest EASM years (Fig. 12c). The BC loadings also show different feedbacks (not shown). Different from during winter, BC loadings due to their own warming effect generally decrease during summer, although there is a slight accumulation in the upper troposphere transport of the ascending motion anomaly. However, the difference in the BC loading between strong and weak monsoon years during summer is similar to that during winter.

e. Brief discussion

The results above indicate that the BC aerosols might have a significant influence on the regional climate over East Asia during both winter and summer, which is consistent with the results of Li et al. (2007) and Wang et al. (2015). The monsoonal climates might also substantially affect the aerosol loadings (e.g., Corrigan et al. 2006; Bao et al. 2008; Zhu et al. 2012; Wang et al. 2015; Mao et al. 2017). Bao et al. (2008) found that a weak westerly at 850 hPa can lead to higher aerosol loadings over East Asia, which is also suggested in this study. Zhu et al. (2012) further indicated that increases in aerosol loadings over eastern China were associated with the decadal-scale weakening of the EASM. Our study suggests that the AOD is increased by 8.27% in southern China during the weakest EASM years, which supports the assessment from Zhu et al. (2012) to some extent.

Recently, there have been a substantial number of studies on the BC DRF and its associated warming effects (e.g., Menon et al. 2002; Lau and Kim 2006; H. Zhang et al. 2009; Zhuang et al. 2013; Jiang et al. 2017). All these studies indicated that the BC warming effect could lead to local warming, and hence the stabilizing effects of scattering aerosols would be offset and the land–ocean temperature gradient would be strengthened during summer. Changes in the EASM circulation can further affect regional or local cloud formation and rainfall. This study also shows similar BC effects. In

<table>
<thead>
<tr>
<th>Types</th>
<th>BC AOD (%)</th>
<th>BC IDRF (%)</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>DJF</td>
<td>JJA</td>
</tr>
<tr>
<td>SC</td>
<td>0.05</td>
<td>8.27</td>
</tr>
<tr>
<td>NC</td>
<td>1.98</td>
<td>-7.41</td>
</tr>
<tr>
<td>East Asia</td>
<td>-1.87</td>
<td>-4.68</td>
</tr>
</tbody>
</table>
addition, in this study, the interactions between the BC warming effect and EAWM were investigated further. This study shows that interactions vary with the interannual variations in the EAM.

There are some limitations in this study. First, to exclude the influence of SST and CO₂, the two variables are fixed in the simulations, which might lead to some bias. For example, the fixed SST can lead to biases in the temperature gradient between the land and oceans, which subsequently leads to monsoonal climate changes (Lau et al. 2006). Second, RegCM4 is driven by reanalysis data, which are limited by the errors in forcings from the lateral boundary conditions (LBCs) (Menéndez et al. 2001; Misra et al. 2003). Third, the BC indirect

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**FIG. 11.** Changes in (a) vertical shortwave heating rates (contours; $10^{-6}$ K s⁻¹) and meridional circulation (arrows) of the altitude–longitude section averaged from 105° to 125°E and (b) air temperature in the lower troposphere (shaded; K) averaged from the surface to 850 hPa from the BC warming effect over East Asia during the (left) strongest and (right) weakest EAWM years. For the reference arrow scale in (a), 1 unit represents the wind anomaly in the horizontal wind (m s⁻¹) and vertical motion ($-5 \times 10^{-3}$ Pa s⁻¹). The purple shading in (a) and black dots in (b) indicate the 95% confidence level from Student’s $t$ test.
FIG. 12. As in Fig. 11, but for EASM. Also shown are (c) the changes in total precipitation (shading; mm day$^{-1}$) from the BC warming effect over East Asia during the (left) strongest and (right) weakest EASM years. For the reference arrow scale in (a), 1 unit represents the wind anomaly in the horizontal wind (m s$^{-1}$) and vertical motion ($-5 \times 10^{-5}$ Pa s$^{-1}$). The purple shading in (a) and black dots in (b) and (c) indicate the 95% confidence levels from Student’s t test.
effects, which include the first and second indirect effects, are not considered in the simulation. Finally, in addition to BC, scattering aerosols might also have a substantial influence on the EAM, which is not considered in this study. These issues are important and should be addressed in future studies.

4. Conclusions

In this study, a new version of the regional climate model RegCM4 is applied to investigate the interactions between the BC warming effect and the East Asian monsoon during both winter and summer and in different East Asian winter and summer monsoon years. RegCM4 could capture the basic characteristics of the EAWM and EASM during the study period from 1988 to 2009. The model can also well simulate the magnitude and seasonal and spatial variations in SBCs over China when comparing the simulations with observations.

Our results indicate that high BC loadings are mainly concentrated in southwest and northern China with substantial seasonality. A higher BC optical depth is likely found during cold seasons in China or over East Asia. The seasonal mean BC AOD is 0.021 over East Asia during winter, which is approximately 10% higher than the BC AOD during summer. However, both the IDRF and EDRF are stronger during summer than winter at both the TOA and the surface, although the seasonality of the forcing is the opposite in southern China. The BC warming effect could exert a positive TOA EDRF of 1.36 W m\(^{-2}\) during winter and 1.85 W m\(^{-2}\) during summer over East Asia.

Black carbon aerosol is mainly concentrated at the lower troposphere, and its loadings generally decrease with altitude, although differences exist among seasons. The BC in the atmosphere directly heats the air and induces an increase in air temperature by 0.11–0.12 K in the lower troposphere over East Asia during both cold and warm seasons. Subsequently, a meridional circulation anomaly in the vertical direction at middle to lower latitudes is formed, which is concurrently associated with a cyclone and southerly anomalies at 850 hPa. The anomalies are larger during summer than during winter. In other words, the BC warming effect, to some extent, would be unfavorable to EAWM circulation development but in favor of EASM circulation development. Precipitation is increased in most regions of East Asia especially during summer, which could be changed by 3.73% in southern China. Both EAWM and EASM change due to BC effects, in turn redistributing the BC loadings. The regional mean SBC over East Asia is increased by 1.17% during winter and 3.89% during summer. However, the columnar BC is increased during winter in southern China, which leads to a higher BC AOD and stronger IDRF of 0.7% and 0.45%, respectively, over East Asia during the cold season.

There are substantial interannual variations in the EAWM and EASM, which could also affect the BC loadings over East Asia. The northwest wind in northern China and the northeast wind in southern China are more anomalous during stronger EAWM years, which results in a decrease in BC AOD in northern and northeastern China and an increase along the Yangtze River basin (and then in southern China) during winter. The east wind in SC and northwest wind in northeast China are anomalous during the strongest EASM, which leads to a decrease in BC in the northeast region of East Asia and an increase in southern China. Changes in BC loadings from the BC warming effect are significant but are weaker than those from the interannual variations of the EAM.

Interactions between the BC warming effect and EAM during stronger monsoon years are different from those during the weaker monsoon years in both winter and summer. The BC loadings are higher at lower latitudes and lower at higher latitudes during the strongest EAWM and EASM years, which would lead to more southerly meridional circulation anomalies. The BC loadings in the vertical direction show similar feedbacks. The wind anomalies at 850 hPa are more westerly during winter and concentrated in eastern China in summer during the strongest monsoon years. Air temperature responses in the lower troposphere are more southerly during winter and more concentrated and westerly during summer in the strongest EAWM and EASM years, respectively. Overall, changes in atmospheric circulation are larger during the weakest monsoon years in summer. Thus, the BC concentration feedbacks could extend to the higher altitudes. Local floods induced by BC are found in the middle and lower reaches of the Yangtze River and western areas of southern China during the strongest EASM years, which covers most of southern China during the weakest EASM years.

Acknowledgments. This work was supported by the National Key R&D Program of China (2017YFC0209803, 2014CB441203, 2016YFC0203303) and the National Natural Science Foundation of China (41675143, 91544230, and 41621005). Data used in this study are from the references of Zhang et al. (2008, 2012).

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