

Inconsistent Fast and Slow Responses of East Asian Summer Monsoon Precipitation Forced by Black Carbon

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ABSTRACT: The East Asian summer monsoon (EASM) region, teeming with anthropogenic black carbon (BC), lacks clarity on the fast and slow responses of EASM precipitation and other atmospheric water cycle processes to BC forcings. Through Community Earth System Model, version 1, simulations with an atmospheric water tracking (AWT) method, fast atmospheric adjustments and slow sea surface temperature (SST)-mediated responses of EASM to BC are found inconsistent, and the summertime precipitation anomaly exhibits a “positive–negative” and “negative–positive” meridional pattern, respectively. Quantitative AWT results highlight the dominant influences of BC-induced atmospheric dynamics adjustments in shaping the EASM precipitation anomaly pattern. The impacts of moisture transport by the southwesterlies, primarily from tropical Indian Ocean, outweigh those of surface evaporation changes, governing both fast and slow precipitation responses. Tropospheric warming from high BC concentrations over northern East Asia in the fast response creates a positive meridional temperature gradient anomaly at the lower level of the East Asian subtropical westerly jet (EASWJ) center. This weakens EASWJ, leading to significant upper-level convergence at 200 hPa and anomalous descending motion north of 30°N, reducing precipitation over North China. Compensatory ascending motion and low-level convergence south of 30°N enhance precipitation over South China. Conversely, BC-induced SST warming in the Indo-Pacific warm pool generates a negative meridional temperature gradient anomaly and strengthens EASWJ, leading to an opposite slow precipitation response pattern. Overall, the total EASM responses manifest as a “wetter–drier–wetter” meridional precipitation pattern and slightly strengthened circulation, aligning with the main structures of interdecadal summer precipitation anomalies over East Asia in recent years reported in previous observational studies.

SIGNIFICANCE STATEMENT: Black carbon (BC), the most influential absorbing aerosol with warming effects second only to greenhouse gases, poses significant risks to human health and the climate. East Asia, a major contributor to global BC loading, is recognized as one of the largest monsoon regions, with its summer hydrological processes being sensitive to BC forcings. Our study utilizes an Earth system model coupled with an atmospheric water tracking method and reveals the opposing effects of BC-induced fast atmospheric adjustments and slow SST-mediated climatic responses on East Asian summer monsoon precipitation. Emphasis is placed on the dominant role of moisture transport related to the BC-induced atmospheric circulation adjustments in governing both fast and slow summer precipitation responses over East Asia.

KEYWORDS: Monsoons; Precipitation; Anthropogenic effects/forcing; Water vapor; Aerosols/particulates; Air-sea interaction

1. Introduction

East Asia (EA) is not only a major contributor to the black carbon (BC) concentrations and radiative forcings (RFs) worldwide (Wang et al. 2012, 2014; Klimont et al. 2017) but also one of the largest monsoon-controlled regions in the world (Ding 1994). The warm and wet air brought by the southerly airflow of the East Asian summer monsoon (EASM) dominantly

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affects the weather and climate over EA in summer (Wu et al. 2016). However, the EASM is characterized by strong temporal and spatial variability (Ding and Chan 2005), which may trigger natural disasters such as drought and flood, seriously affecting the producing and living of countries in the monsoon region (Qian et al. 2012; Tian et al. 2018; Yang et al. 2022). The EASM changes are associated with its internal variability (e.g., Pacific decadal oscillation, Atlantic multidecadal oscillation) and external forcings (e.g., solar radiation variation, volcano eruption, and anthropogenic activity). A great deal of the EASM changes can be attributed to the variation and diversity of aerosols over EA (Li et al. 2016). BC is not only an important component of fine particulate matter but also the most influential absorbing aerosol with warming effects only second to greenhouse gases (carbon dioxide, methane, nitrous oxide, etc.), posing significant risks to both human health and the climate (Lau et al. 2010; Bond et al. 2013).

BC, mainly contributed by anthropogenic emissions in EA, may alter the EASM in terms of surface and atmospheric temperature, cloud characteristics, precipitation, atmospheric circulation, monsoon onset, and retreat through its impacts on Earth's radiation budget (Liu et al. 2009; Guo et al. 2013; Jiang et al. 2013; Li et al. 2016; Wang et al. 2016; Xie et al. 2020). The significant effects of BC forcings on the EASM are widely recognized and studied (e.g., Jiang et al. 2013; Zhuang et al. 2018, 2019; Xie et al. 2020). Zhuang et al. (2019) estimated that the BC forcings could cause a 0.2 K increase in surface air temperature and a 0.01 mm day^{-1} decrease in precipitation over EA in summer, though the local BC emissions and concentrations in summer are lower than those in other seasons. The Sixth Assessment Report (AR6) of Intergovernmental Panel on Climate Change (IPCC) estimates that the global mean annual effective RF (ERF) due to BC emissions is $+0.11$ (-0.2 to $+0.42$) W m^{-2} . This small net positive BC ERF is due to a large portion of the positive ERF from the aerosol-radiative interaction (ARI) being offset by the negative BC ERF from the aerosol-cloud interaction (ACI) and an adjustment in lapse rate and atmospheric water vapor (IPCC 2021). However, Liu et al. (2022) use observations and phase 6 of the Coupled Model Intercomparison Project (CMIP6) simulations to show that the annual mean ERF from the ARI effects of BC over China is $2.4\text{--}3.0\text{ W m}^{-2}$, exceeding previous estimates.

Depending on the time scale and the buffering by changes in sea surface temperature (SST) and sea ice, the EASM's responses to BC radiative effects can be divided into rapid adjustments to the BC radiative effects (fast response) and slow SST-mediated climate feedback process (slow response). At present, most studies focusing on the impacts of BC climate forcings upon the EASM mainly discuss the fast response of the EASM to BC (e.g., Liu et al. 2009; Jiang et al. 2013; Wang et al. 2016; Persad et al. 2017; Zhuang et al. 2018, 2019). Ganguly et al. (2012) pointed out that the slow climatic feedbacks related to the SST change due to the anthropogenic climate drivers could be more essential in shaping the total response of the monsoon system to aerosol forcings. Using the Community Earth System Model, version 1 (CESM1), Wang et al. (2017b) showed that the total response of summer

precipitation over EA to the total BC RFs was more consistent with its slow response, indicating the nonnegligibility of the BC-induced slow EASM response.

As a key manifestation of the EASM changes under the BC climate impacts and the important process of atmospheric water cycle, the precipitation response to BC (or other aerosol species) forcings over EA has been widely examined in previous observational and modeling studies (Menon et al. 2002; Gu et al. 2006; Jiang et al. 2013; Li et al. 2015; Xie et al. 2016; Wang et al. 2019). Note that surface evaporation and moisture transport are also important atmospheric water cycle processes. Numaguti (1999) pointed out that continental land evaporation is the major contributor to precipitation over the Eurasian continent in summer. Chow et al. (2008) and Xu et al. (2008) showed that the southerly moisture associated with the northwestern Pacific subtropical high and Indian summer monsoon determines the precipitation over China during summer. However, the climatic responses of evaporation and moisture transport over EA to BC forcings were rarely discussed. According to the simulated results from Pan et al. (2021), the moisture transport process determines the fast adjustments of precipitation induced by BC over EA in summer. However, few studies have focused on the SST-mediated slow responses of different atmospheric water cycle processes to BC forcings over EA.

Most of previous studies were concentrated on the influences of the total climate forcings or direct radiative forcing (or referred as the ARI effects of BC) based on sensitivity experiments in which BC emissions were fixed at preindustrial (PI) or present-day (PD) values. The climate impacts caused by BC RFs result from ARI, ACI, and surface albedo changes (SAC) in snow- and sea ice-covered regions. However, the relative contribution of fast and slow adjustments over the EASM region induced by the different components of the BC forcings were seldom examined and remains uncertain. Pan et al. (2021) found that the fast adjustments in precipitation and wind fields over EA in summer induced by BC forcings were jointly influenced by the ARI and NRI (non-ARI, including ACI and SAC) effects of BC using the CESM1 model. Dong and Zhou (2014) pointed out that the SST variation in the Indian Ocean due to aerosol RFs was mainly from the ACI effects. These findings suggest the necessity of separating the different forcing components when discussing the fast and slow climatic responses to aerosol forcings.

Here, we separated and compared the fast and slow responses of the EASM to BC forcings based on the CESM1 simulations. Considering the remaining uncertainties regarding the fast and slow responses of different processes of the atmospheric water cycle in the EASM region to BC, an atmospheric water tracking (AWT) technique was applied to the simulations to enable the quantitative identification of evaporation and moisture transport processes in the model and moisture sources of summer precipitation over EA. Given the minor effect of SAC on RF over EA during summer, the climate adjustments caused by BC over EA can be approximated as the effects of ARI and ACI. Therefore, this study isolates and discusses the climate adjustments caused by the ARI and ACI effects of BC. The primary scope of this study

is to distinguish the fast adjustments and slow climate feedback process of the EASM caused by the ARI and ACI impacts of BC, respectively, with a focus on the fast and slow performances of summer atmospheric water cycle over EA. The model and simulation settings, as well as the emission inventories, are described in [section 2](#). [Sections 3a–3c](#) present the model evaluations, the SST adjustments, the total responses of the EASM, and the corresponding fast and slow components to the BC forcings. [Section 3d](#) examines the different fast and slow responses of summer atmospheric water cycle processes over EA to the BC forcings using the AWT technique. Atmospheric circulation adjustments and their impacts on EASM precipitation are investigated in [section 3e](#). Uncertainties in the simulated results are discussed in [section 4](#). Conclusions are given in [section 5](#).

2. Methods

a. Model description

In this study, the CESM1 coupled with its atmospheric component Community Atmosphere Model, version 5 (CAM5; [Neale et al. 2012](#)), was used to examine the fast and slow responses of monsoon and atmospheric water cycle over EA to BC forcings. The three-mode version of the modal aerosol model (MAM3) simulates aerosol (including BC, sulfate, primary and secondary organic matter, mineral dust, and sea salt) mass, number concentration, and mixing state ([Liu et al. 2012](#)).

The Rapid Radiative Transfer Model for GCMs (RRTMG; [Iacono et al. 2008](#)) can be combined with aerosol schemes to demonstrate the aerosol ARI effect. RRTMG accounts for the radiative impacts of various atmospheric constituents, including cloud cover, aerosols, greenhouse gases, water vapor, and ozone, among others. For shortwave radiation (solar radiation), RRTMG uses 14 shortwave spectral bands (0.2–12.2 μm). It calculates the absorption coefficients of various gases for each band based on parameters like altitude, temperature, and pressure. It also considers the scattering and absorbing effects of clouds and aerosols. In the longwave spectrum (terrestrial radiation), RRTMG employs 16 longwave spectral bands (3.08–1000 μm) to calculate the emissivity and layer transmittance of gases, while also accounting for the emission, reflection, and absorption effects of cloud layers.

CESM1 simulates ACI effects through combining cloud parameterization schemes with aerosol modules. The model employs a stratiform cloud microphysics scheme ([Gettelman et al. 2008, 2010; Morrison and Gettelman 2008](#)), a stratiform cloud macrophysics scheme ([Gettelman et al. 2010; Park et al. 2014](#)), a shallow convection cloud parameterization ([Park and Bretherton 2009](#)), and a modified deep convection scheme ([Zhang and McFarlane 1995; Neale et al. 2008](#)). These parameterizations interact with the aerosol scheme (e.g., MAM3) to represent various ACI processes, including aerosol activation as cloud condensation nuclei, impacts on cloud droplet and ice crystal formation, effects on cloud lifetime and extent, and influences on precipitation. The model explicitly calculates aerosol activation based on particle size, composition, and ambient conditions, allowing for the simulation of the

indirect effects of aerosols on clouds. By coupling these cloud and aerosol schemes with the radiation module (RRTMG), CESM1 provides a comprehensive representation of aerosol–cloud–radiation interactions and their subsequent impacts on climate.

The Snow, Ice, and Aerosol Radiative Model (SNICAR; [Flanner et al. 2007](#)) within the CESM1 land component simulated the changes in the snow surface albedo caused by BC and dust deposition, as well as the resulting impacts on snow cover and snow grain size.

b. Model configuration and experimental design

We have conducted two sets of simulations at a horizontal resolution of $0.9^\circ \times 1.25^\circ$ and 30 vertical levels from the surface to 4 hPa. One was running with the prescribed historical SST and sea ice (from Hadley Centre Sea Ice and SST datasets; <https://www.metoffice.gov.uk/hadobs/hadisst/data/download.html>) to represent the fast EASM responses to BC forcings [referred as fast response (FR)]; another one was coupled with the single-layer ocean model [SOM; [Neale et al. 2012](#); referred as total response (TR)] to examine the total climatic response over EA. The SST in CESM1 simulations with the SOM is determined by the energy balance between radiative flux at surface, latent and sensible heat fluxes, and an internal heat source (Q flux, representing seasonal deep-water exchange and horizontal ocean heat transport). The SST-mediated slow response (SR) is the difference between the TR and FR. The assumption for the SR has been applied in previous studies using SOM ([Hsieh et al. 2013; Jiang et al. 2020](#)) or fully coupled model ([Xu and Xie 2015; Wang et al. 2017b](#)). The FR- and TR-type experiments were performed 45 and 60 years, respectively, and we focused on the results from the final 30 years for each type of simulation.

As shown in [Table 1](#), each type of experiments consists of two pairs of simulations, one with BC and other aerosol emissions set to 2000 (PD) level and the other where only the BC emissions are set to 1850 (PI) level. To further distinguish the ARI and ACI effects of anthropogenic BC, each pair of simulations includes a simulation with both ARI and ACI effects and a simulation excluded the ARI effects through turning off the BC scattering and absorption. Thus, the responses of different climate variables (referred to as Var; e.g., precipitation) to anthropogenic BC forcings could be decomposed as follows:

$$\text{Var}_{\text{TR}} = \text{Var}_{\text{TR_PDBC}} - \text{Var}_{\text{TR_PIBC}}, \quad (1)$$

$$\text{Var}_{\text{FR}} = \text{Var}_{\text{FR_PDBC}} - \text{Var}_{\text{FR_PIBC}}, \quad (2)$$

$$\text{Var}_{\text{SR}} = \text{Var}_{\text{TR}} - \text{Var}_{\text{FR}}. \quad (3)$$

Then, the responses to the ARI effects of anthropogenic BC could be isolated by following Eqs. (4)–(6). Note that the subscripts on the left-hand sides denote the different response to different BC forcings. For example, the total response to the total BC forcings is referred as TR, while the total response to the ARI effects of BC is referred as TR_ARI. And the subscripts on the right-hand sides denote the simulation we used.

TABLE 1. Summary of numerical experiments settings.

Simulation type	Simulation ID	Sea surface temp and sea ice	BC emissions	ARI effects of BC
FR	FR_PDBC	Prescribed	2000	On
	FR_PDBC_noARI			Off
	FR_PIBC		1850	On
	FR_PIBC_noARI			Off
TR	TR_PDBC	Coupled with a slab ocean model	2000	On
	TR_PDBC_noARI			Off
	TR_PIBC		1850	On
	TR_PIBC_noARI			Off

$$\text{Var}_{\text{TR_ARI}} = (\text{Var}_{\text{TR_PDBC}} - \text{Var}_{\text{TR_PDBC_noARI}}) - (\text{Var}_{\text{TR_PIBC}} - \text{Var}_{\text{TR_PIBC_noARI}}), \quad (4)$$

$$\text{Var}_{\text{FR_ARI}} = (\text{Var}_{\text{FR_PDBC}} - \text{Var}_{\text{FR_PDBC_noARI}}) - (\text{Var}_{\text{FR_PIBC}} - \text{Var}_{\text{FR_PIBC_noARI}}), \quad (5)$$

$$\text{Var}_{\text{SR_ARI}} = \text{Var}_{\text{TR_ARI}} - \text{Var}_{\text{FR_ARI}}. \quad (6)$$

Considering the negligible forcings induced by the SAC effects of BC over EA in summer, the simulation that disables only the ARI effect of BC can be assumed to retain the ACI effect of BC, allowing differences between simulations with the same disabled ARI but different BC emission levels to represent the climatic responses to the ACI effects of anthropogenic BC. The results calculated by Eqs. (7)–(9) could be approximated as the climatic responses to the ACI effects of anthropogenic BC.

$$\text{Var}_{\text{TR_ACI}} = \text{Var}_{\text{TR_PDBC_noARI}} - \text{Var}_{\text{TR_PIBC_noARI}}, \quad (7)$$

$$\text{Var}_{\text{FR_ACI}} = \text{Var}_{\text{FR_PDBC_noARI}} - \text{Var}_{\text{FR_PIBC_noARI}}, \quad (8)$$

$$\text{Var}_{\text{SR_ACI}} = \text{Var}_{\text{TR_ACI}} - \text{Var}_{\text{FR_ACI}}. \quad (9)$$

All simulations were conducted with the AWT method added in the CESM1, which was first introduced and detailed in Pan et al. (2017). This method was developed based on the physical parameterization schemes of atmospheric water substances in the model and can track the atmospheric water substances in different atmospheric water cycle processes from different source regions directly and quantitatively.

Atmospheric water substances in the CESM1 include forms such as water vapor, cloud droplets, cloud ice, rain, and snow. The global simulation area is divided into different source regions based on research objectives, each containing distinct tagged atmospheric water substances. The surface emission of the tagged water vapor within the source region equals the surface evaporation of the original water vapor at the beginning of the tracking process. In contrast, the emission of the tagged water vapor outside the source region is set to zero. Subsequently, the tagged atmospheric water substances undergo the same physical processes as the original atmospheric water substance. Ultimately, we obtain the contributions of various tagged atmospheric water substances from different source regions to the corresponding atmospheric water

substance concentrations in the study area from the simulation results. The temporal variation of the tagged atmospheric water substances in the model can be expressed as

$$\frac{\partial q_{k,\text{tg}}}{\partial t} + \frac{1}{\rho} \nabla \cdot (\rho \mathbf{u} q_{k,\text{tg}}) = \left(\frac{\partial q_{k,\text{tg}}}{\partial t} \right)_{\text{dp conv}} + \left(\frac{\partial q_{k,\text{tg}}}{\partial t} \right)_{\text{shlw conv}} + \left(\frac{\partial q_{k,\text{tg}}}{\partial t} \right)_{\text{cld macro}} + \left(\frac{\partial q_{k,\text{tg}}}{\partial t} \right)_{\text{cld micro}} + D(q_{k,\text{tg}}), \quad (10)$$

$$\frac{\partial q_{p,\text{tg}}}{\partial t} = \left(\frac{\partial q_{p,\text{tg}}}{\partial t} \right)_{\text{dp conv}} + \left(\frac{\partial q_{p,\text{tg}}}{\partial t} \right)_{\text{shlw conv}} + \frac{1}{\rho} \frac{\partial (V_q \rho q_{p,\text{tg}})}{\partial z} + \left(\frac{\partial q_{p,\text{tg}}}{\partial t} \right)_{\text{cld micro}}, \quad (11)$$

where $q_{k,\text{tg}}$ is the mass mixing ratio of tagged water vapor ($q_{v,\text{tg}}$), cloud droplets ($q_{l,\text{tg}}$), or cloud ice ($q_{i,\text{tg}}$); $q_{p,\text{tg}}$ is the mass mixing ratio of tagged rain ($q_{r,\text{tg}}$) or snow ($q_{s,\text{tg}}$); t denotes the time; $\mathbf{u}$ is the three-dimensional wind vector; ρ is the air density; D is the turbulent diffusion operator; z is the height; and V_q is the mass-weighted terminal fall speeds. The right-hand side of the equations describes the physical processes that the tagged atmospheric water substances participate in within the model. Here, $(\partial q_{k,\text{tg}}/\partial t)_{\text{dp conv}}$, $(\partial q_{k,\text{tg}}/\partial t)_{\text{shlw conv}}$, $(\partial q_{k,\text{tg}}/\partial t)_{\text{cld macro}}$, and $(\partial q_{k,\text{tg}}/\partial t)_{\text{cld micro}}$ represent the tendencies of tagged water vapor, cloud droplets, or cloud ice in deep convection, shallow convection, cloud macrophysics, and cloud microphysics processes, respectively. The net production of precipitation during deep and shallow convection processes is represented as $(\partial q_{p,\text{tg}}/\partial t)_{\text{dp conv}}$ and $(\partial q_{p,\text{tg}}/\partial t)_{\text{shlw conv}}$, respectively, with the grid-mean source/sink term denoted as $(\partial q_{p,\text{tg}}/\partial t)_{\text{cld micro}}$ in the cloud microphysics process.

Note that the tracking in the AWT method stops when the tagged water substances form precipitation and fall to the ground outside their source region, resulting in the disappearance of the tagged substances from the model. If they fall to the ground inside the source region, they are treated as new tagged water substances and enter the simulation calculations. The tracked water substances are independent and passive, which means they have no impact on dynamical and thermal fields in the model.

Previous studies have systematically validated the performance of this method on tracking the water substances in

different physical processes and quantifying the moisture sources of precipitation over EA and the Tibetan Plateau (TP; Pan et al. 2017, 2019). In this study, atmospheric water substances from 13 moisture source regions (Fig. 8) contributed to the precipitation, surface evaporation, and moisture transport over EA monsoon regions were tracked, aiming for examining the fast and slow responses of atmospheric water cycle processes to BC and quantifying the corresponding water source contributions in the EASM region.

c. Dataset description

The BC emissions in PD level are sourced from the Peking University BC Inventory (PKU-BC; Wang et al. 2014). PKU-BC is a global monthly emission inventory with a subnational fuel consumption data and updated emission factor measurements. Fang et al. (2020) have found that the bias of simulated BC concentrations in China, using PKU-BC based on the CESM1 model, ranges from -34% to $+3\%$, which performs better compared with the simulation results based on other inventories.

This study employs surface and column BC concentrations and aerosol absorption optical depth (AAOD) reanalysis data from Modern-Era Retrospective analysis for Research and Applications, version 2 (MERRA-2; Gelaro et al. 2017), for comparison with the simulated data. Compared to MERRA, MERRA-2 incorporates additional aerosol observation assimilation data. Studies have validated that the assimilation of aerosol observations in MERRA-2 can well reflect the mass concentrations level, spatial distribution, and temporal variation of BC over EA (Fang et al. 2020; Xu et al. 2020). The AAOD assimilated in MERRA-2 is based on the Advanced Very High Resolution Radiometer, Moderate Resolution Imaging Spectroradiometer (MODIS), Multiangle Imaging SpectroRadiometer, and Aerosol Robotic Network (AERONET; Gelaro et al. 2017). This study also collects observed BC concentrations from the China Meteorological Administration and previous studies. Details about the observation sites and sampled times are summarized in Table S1 in the online supplemental material.

Since this study primarily focuses on the EASM responses to BC forcings, it is essential to validate the model's ability to simulate the key characteristics of the EASM. The global precipitation observation data from the Global Precipitation Climatology Project (GPCP; Adler et al. 2017) are downloaded for this purpose. This dataset integrates satellite observations of microwave, infrared, and acoustic data with ground-based rain gauge measurements. This study also utilizes the global precipitation observation data from the Climate Prediction Center (CPC; Chen et al. 2008). Unlike the GPCP dataset, the CPC dataset covers only global land areas. The wind field reanalysis from the fifth major global reanalysis produced by the European Centre for Medium-Range Weather Forecasts (ECMWF) (ERA5; Hersbach et al. 2020) is also used for the validation of the CESM1-simulated EASM wind fields. Furthermore, as this study examines the ACI effects of BC, we utilize cloud fraction data from the MODIS *Terra/Aqua* satellite observations (Platnick et al. 2015) to compare with our simulated results. This comparison serves for the validation of the model simulated cloud distribution over EA.

3. Results

a. Results evaluations

The ability of the CESM1 model to reproduce the characteristics of BC concentrations and light absorptions needs to be evaluated before investigating the fast and slow response of the EASM to the BC forcings. Considering the short lifetime of BC (Samset et al. 2014), Fig. 1 shows the June–August (JJA) mean mass concentrations of BC at the surface over EA and its adjacent areas from the simulations, the MERRA-2 analysis, and observations. Our simulated BC concentrations show close alignment with the observed surface BC at most sites, except for some city sites over southern China (Nanning, 22.38°N , 108.13°E), Southeast Asia (Hanoi, 21.02°N , 105.81°E), and South Asia (Kathmandu; 27°N , 85°E and Srinagar, 34°N , 74°E). The coarse-resolution and grid-averaged values used in global models hardly exhibit the BC concentrations at city sites. The existing bias in emission inventory also brings the underestimation of simulated BC at these sites. The magnitudes and spatial pattern of BC simulated from both FR- and TR-type simulations in our study are well consistent with those from the MERRA-2 reanalysis and observations. Note that the spatial distributions of BC concentrations in TR and FR simulations are similar, indicating that the BC-induced SST changes hardly alter the main pattern of BC distributions, though there are relatively high variations in BC surface concentrations caused by SST changes over northwest Pacific Ocean, the north Indian Ocean, and the south Tibetan Plateau. The relative difference in BC surface concentrations over the ocean is greater than that in EA, attributed to the lower BC levels over the ocean. Therefore, small variations in BC concentrations due to SST changes result in larger relative variations.

The AAOD of absorbing aerosol associated with the column burden estimated in the model determines their simulated RFs (Bond et al. 2013). Figure 1 shows the JJA mean AAOD at 550 nm derived from the simulations and MERRA-2 reanalysis. The AAOD from both FR- and TR-type simulations generally shows similar pattern with the AAOD in MERRA-2 over EA but with higher values, especially over north and northwest China. Bakatsoula et al. (2023) reported that MERRA-2 largely underestimates the AAOD over Asian dry areas, including North China and central Asia. Besides, CESM1 produces underestimation in AAOD over sea areas compared to MERRA-2.

The CESM1 model can also reproduce the characteristics of summer precipitation and circulation over EA (Fig. 2). The intensity and distribution characteristics of the simulated EASM precipitation are consistent with the merged observational average from the GPCP and CPC datasets, but the simulated precipitation is weaker in the southeast coastal areas and stronger in inland areas, with a root-mean-square error (RMSE) of 2.56 mm day^{-1} over EA compared to the observation. The precipitation biases of the CMIP6 and CMIP5 models are quite similar, though the precipitation from the CMIP6 models shows improvement in certain regions (Gusain et al. 2020; Xin et al. 2020). Chen and Frauenfeld (2014) estimated that the RMSE of the annual precipitation for the CMIP5 multimodel mean in EA is 3.98 mm day^{-1} compared to the observation.

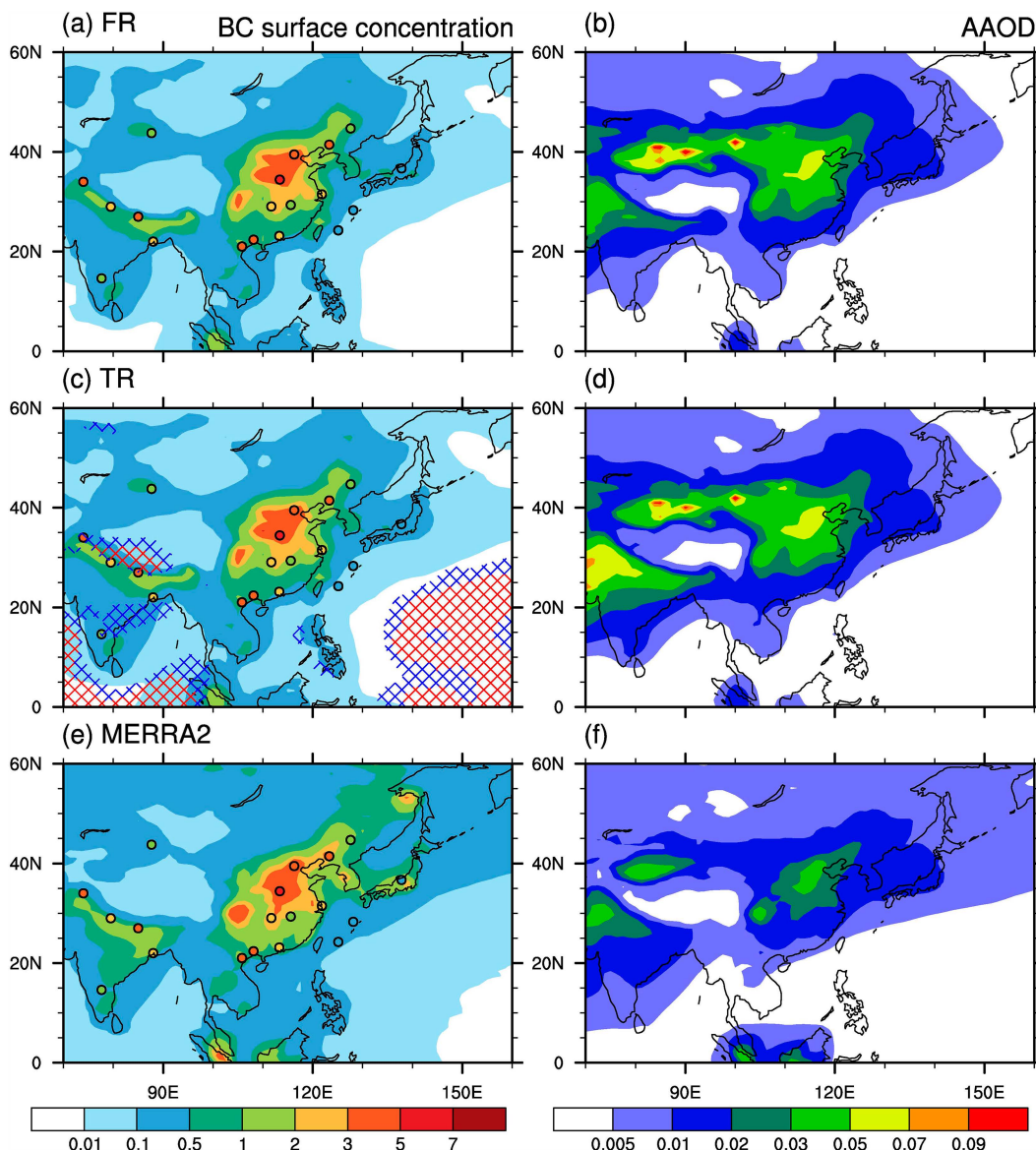


FIG. 1. Spatial distributions of the JJA mean (a),(c),(e) BC surface concentration ($\mu\text{g m}^{-3}$) and (b),(d),(f) AAOD (unit: 1) from (a),(b) FR-type simulation with emissions set to PD level, (c),(d) TR-type simulation with emissions set to PD level, and (e),(f) MERRA-2 data (1985–2014) over EA. Scatters in (a), (c), and (e) represent the observed BC concentrations (1990–2014) at different sites. Blue and red shading area indicate where the relative difference in BC surface concentration between TR- and FR-type simulations induced by SST changes is between 20% and 50% and more than 50%, respectively.

The bias of the simulated EASM precipitation in this study is lower than that of the CMIP5 multimodel means, which could be explained by the specified SST used in the FR-type simulations. Besides, the averaged summer precipitation over EA from CESM1 and merged means of GPCP and CPC datasets is 5.56 and 5.75 mm day^{-1} , respectively. Therefore, the simulated EASM precipitation is 3.3% lower than the merged means of the observations.

Through the comparison with the ERA5 reanalysis, we validate that CESM1 can show the characteristics of the low-level EASM wind fields, but the simulated horizontal wind

field at 850 hPa over EA is stronger, especially in South China. In addition, the EASM characteristics from the TR-type simulation are similar to those from the FR-type simulation (figure not shown). Note that the difference between simulation results is calculated to investigate the fast and slow EASM responses to BC forcings by assuming consistent systematic errors across different simulations, and the impacts of the model's systematic errors are thereby reduced.

The cloud fraction characteristics from CESM1 also closely resemble MODIS satellite observations (Fig. S1). However, the model underestimates cloud fraction over the Indo-China

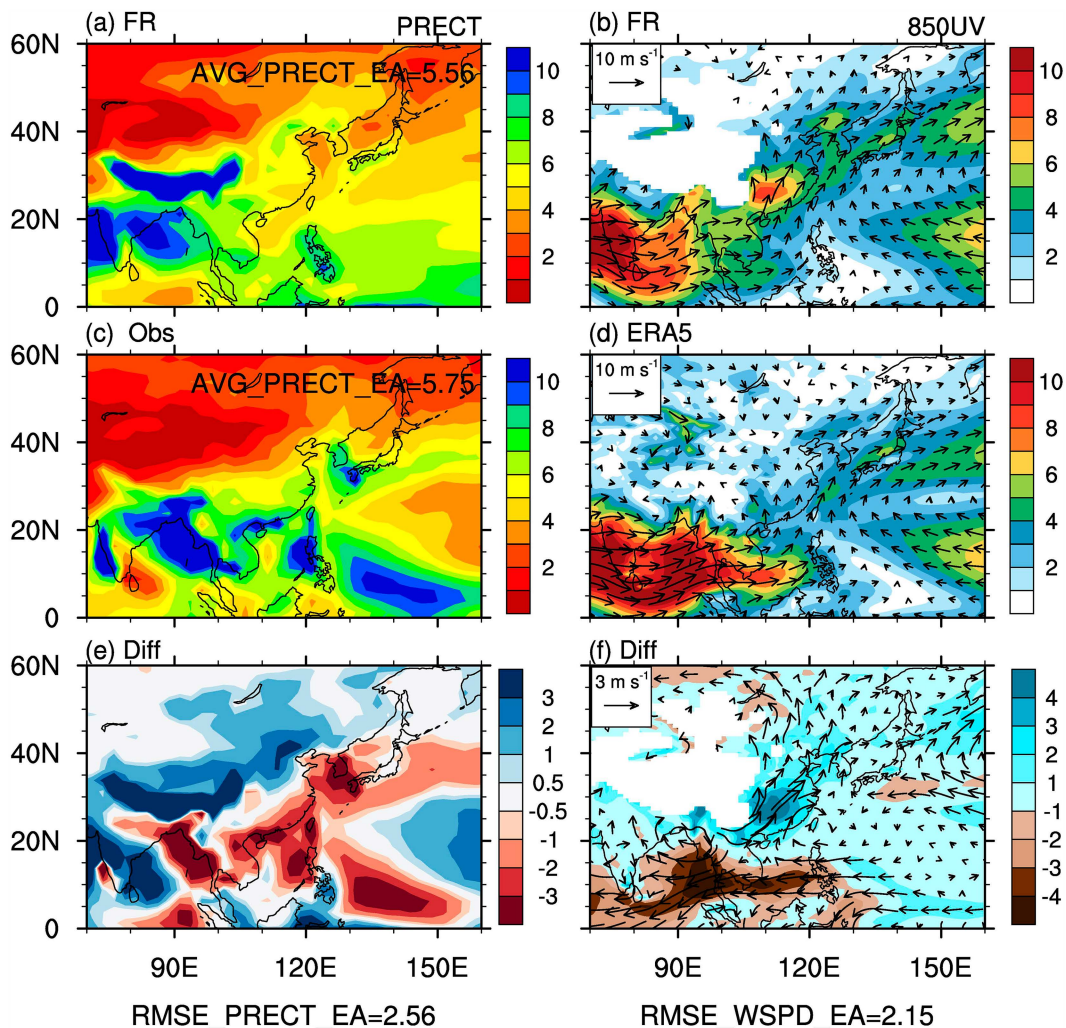


FIG. 2. Spatial distributions of the JJA (a),(c) mean precipitation (shading; mm day^{-1}) and (b),(d) 850-hPa wind fields (vector; m s^{-1}) from FR-type simulations with emissions set to (a),(b) PD level, (c) merged GPCP and CPC observational average (1985–2014), and (d) ERA5 (1985–2014) over EA. The wind speeds (shading; m s^{-1}) are also shown in (b), (d), and (f). (e),(f) The differences between the simulation and observation or reanalysis. The regional mean precipitation values over EA from the model results and observation are shown in (a) and (c), respectively. The regional mean RMSE values of precipitation and wind speed over EA are shown below the figure.

Peninsula while overestimating cloud fraction in the north-west Pacific and northern EA. Overall, the model-simulated regional mean cloud fraction over EA is lower than MODIS observations ($\text{RMSE} = 0.068$). Besides the inherent limitations of the model, the discrepancy between the simulation baseline (PD emissions) and the observation period (2003–17) also contributes to these differences.

b. BC-induced temperature changes and RFs

The global pattern of the annual mean SST changes and total RFs induced by total anthropogenic BC forcings is shown in Fig. 3. BC exhibits a pronounced absorption of solar radiation, leading to significant positive total RF in the atmosphere over Asia, central Africa, and their surrounding ocean regions (Fig. 3d). Consequently, the SST rises (Fig. 3a) despite the

reduction in shortwave radiation reaching Earth's surface (Fig. 3c). The TR-type simulations show a global annual mean total RF of $+0.01 \text{ W m}^{-2}$ at the top of the atmosphere (TOA), indicating the incorporating effects of the complex ocean feedbacks (Fig. 3b). In addition, the regional mean RF at the TOA over EA from the TR-type simulations is 1.34 W m^{-2} , exceeding the global-mean RF value. The total BC RF at TOA from FR-type simulations is 0.15 W m^{-2} (figure not shown), aligning with the IPCC AR6's estimated BC ERF of $+0.11$ (-0.2 to $+0.42$) W m^{-2} . In the MAM3 aerosol module used in this study, BC ages immediately and can be coated by other aerosol after emission, potentially enhancing its absorption through the lensing effect. Therefore, the modeled total BC RF at the TOA slightly exceeds the IPCC AR6 estimate, suggesting a possible overestimation in the simulation.

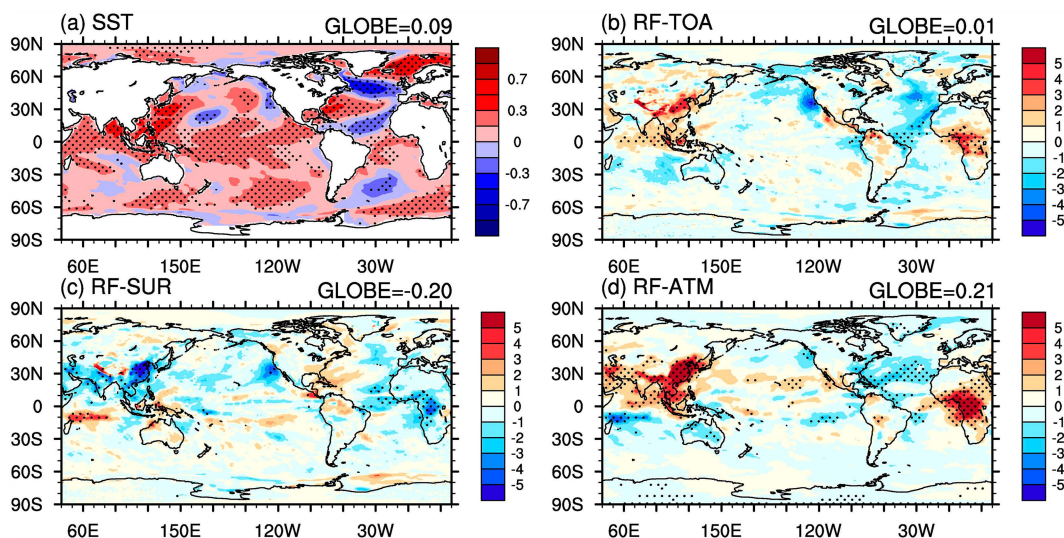


FIG. 3. Global distributions of the (a) annual mean SST change (K), (b) total RF at the TOA (W m^{-2}), (c) total RF at the surface (W m^{-2}), and (d) total RF in the atmosphere (W m^{-2}) induced by anthropogenic BC from TR-type simulations. Dotted regions denote where the change is statistically significant at the 95% confidence level. The global annual mean values of SST change and RFs are shown in the upper-right corner of the panels.

The SST changes forced by BC show consistent characteristics with those given in previous studies (Hsieh et al. 2013; Ocko et al. 2014; Wang et al. 2017b), and a fully coupled dynamic ocean or SOM was employed in their simulations: 1) The SST rises across most of the world's oceans, with the largest increases occurring in the waters adjacent to Eurasia and North America. 2) The SST response has been greater in the Northern Hemisphere (NH) compared to the Southern Hemisphere (SH), which can be attributed to the greater accumulation of anthropogenic influences in the north. 3) There are regions of negative SST anomalies in parts of the North Atlantic Ocean, particularly in the northernmost areas. Ocko et al. (2014) suggested the SST cooling in this region is related to the weakening of the Atlantic meridional overturning circulation driven by the BC-induced overall warming. Globally, the annual mean SST change is 0.09 K, with increases of 0.13 K in the NH and 0.08 K in the SH. These results are close to but slightly lower than the SST change values caused by BC given by Wang et al. (2017b; globe = 0.12 K, NH = 0.17 K, and SH = 0.07 K). This may be related to the fact that the simulation conducted by Wang et al. (2017b) used a fully coupled dynamic ocean model, and the simulation time was longer (75 years). Therefore, the slow response of the EASM to anthropogenic BC shown in this study may be weaker than the results in Wang et al. (2017b).

The significant solar absorption of BC leads to atmospheric warming over EA (Fig. 4). Temperature increase in total, fast, and slow responses to BC total forcings is dominated by the ARI effects, while the ACI effects of BC decrease the temperature over EA. In the fast response, BC-induced temperature changes over EA increase with altitude, and this increase levels off in the upper troposphere as BC concentration decreases, aligning with the simulated results from Stjern et al. (2017). However, in the slow response, BC-induced temperature

changes below 850 hPa over EA decrease with altitude. This may be attributed to the fact that atmospheric warming in the slow response is primarily achieved through the atmosphere heating of BC-induced SST warming in seas adjacent to the EA. The magnitude of slow temperature responses from different BC forcings matches that of fast responses, underscoring the importance of evaluating the slow climate response to BC forcings. Both fast and slow responses to the ARI effect of BC contribute to the overall temperature increase in the total response. The total temperature response to BC total forcings shows maximum

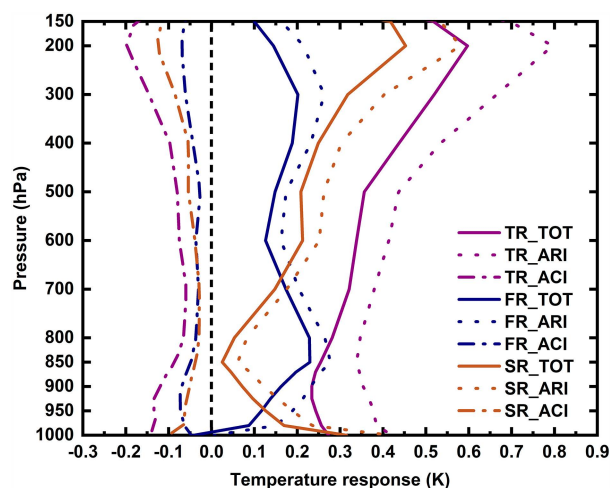


FIG. 4. Vertical profiles of the TR (purple), FR (blue), and SR (brown) responses of the JJA mean temperature (shading; K) to the BC total forcing (TOT; solid lines) and its ARI (dotted lines) and ACI (dotted-dashed line) components over EA (15° – 50° N, 100° – 140° E).

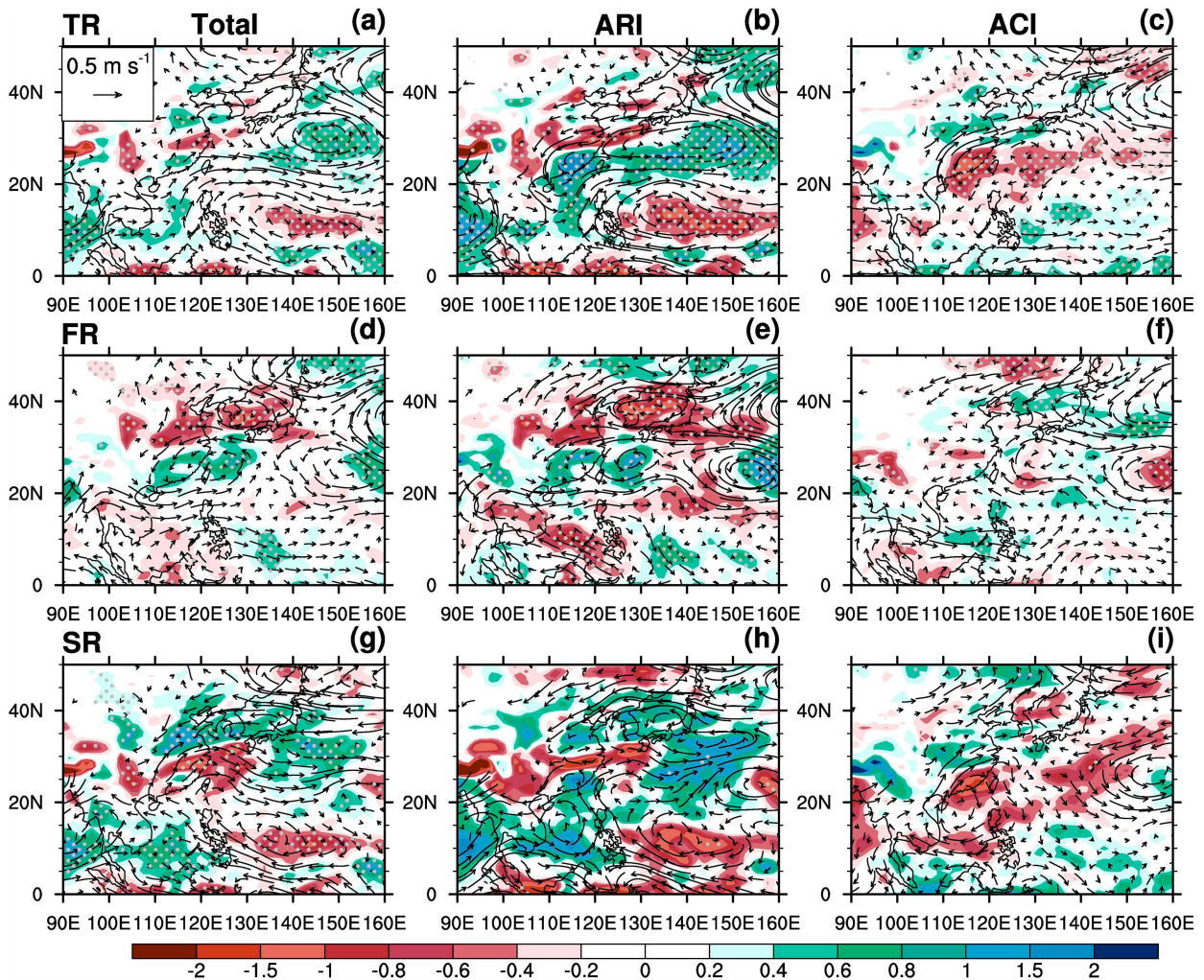


FIG. 5. Spatial distributions of the (a)–(c) TR, (d)–(f) FR, and (g)–(i) SR of the JJA mean precipitation (shading; mm day^{-1}) and 850-hPa wind fields (vector; m s^{-1}) to (a),(d),(g) BC total forcing, and its (b),(e),(h) ARI, and (c),(f),(i) ACI components over EA. Dotted regions denote where the precipitation anomaly is statistically significant at the 95% confidence level.

warming of 0.26 K below 850 hPa, 0.36 K at 850–500 hPa, and 0.60 K at 500–200 hPa.

c. Fast, slow, and total responses of the EASM

The fast, slow, and total JJA mean responses of monsoon circulation and precipitation to BC forcings over EA are shown in Fig. 5 and decomposed into the ARI- and ACI-induced components, respectively. Anthropogenic BC induces opposite fast and slow wind fields responses across the EASM region (Figs. 5d,g), with the Yangtze River Delta (YRD; $\sim 30^\circ\text{N}$) marking a division that results in nearly inverse circulation changes in North China (NC) and South China (SC). Located in the eastern part of China (east of 110°E), both NC and SC are recognized as the most densely populated areas over EA. Figure 5d indicates that BC induces a divergence (convergence) anomaly in NC (SC) and its adjacent seas with decreased (increased) precipitation in the fast responses. Although the increase in precipitation south of 30°N is not significant over the EA continent, it is notable in the

nearby ocean. The divergence (convergence) anomaly in NC (SC) leads to a southerly (northerly) wind anomaly and strengthen (weaken) the local monsoon circulation. Differing from the fast responses, BC-induced slow 850-hPa wind fields responses (Fig. 5g) show a cyclone-like (anticyclone-like) anomaly north (south) of 30°N , where precipitation is increased (decreased). With the combined forces of fast and slow responses, total BC forcings eventually contribute to a slight southerly wind anomaly and enhance the EASM circulation south of 30°N (Fig. 5a), consistent with the results shown in Wang et al. (2017b).

It has been mentioned in some previous studies that BC forcings could lead to the regional precipitation decrease in NC but an increase in SC in terms of the fast climatic response (e.g., Menon et al. 2002). Though Wu et al. (2016) pointed out that the simulation settings in Menon et al. (2002) may overestimate the absorbing impacts of BC, many simulated results based on other regional (e.g., Wang et al. 2015; Zhuang et al. 2019) or

global climate models (e.g., [Lau et al. 2006](#); [Sun and Liu 2008](#); [Liu et al. 2009](#)) with different model settings also captured this BC-induced fast precipitation change pattern. Moreover, our simulated results show the precipitation adjustments over EA due to the ocean feedbacks on the EASM ([Fig. 5g](#)) presenting a significant “drier south and wetter north” feature, contrasting the fast precipitation changes ([Fig. 5d](#)), but the magnitude of which is close to that of the fast precipitation response. The absolute value of this SST-mediated precipitation adjustments could be up to 1.5 mm day^{-1} over EA, indicating the nonnegligible impacts of slow climatic response on the EASM.

Jointly influenced by the fast adjustments and aerosol-induced SST feedbacks, BC total forcings result in a positive–negative–positive (“+ − +” pattern) meridional precipitation distribution in EA ([Fig. 5a](#)). The precipitation along the 27° – 32° N latitudinal band in East China is reduced, but the precipitation in the north and south area of this latitudinal band is enhanced. Hence, our study indicates that the total precipitation response to BC in the EASM-controlled region is synergistically affected by the rapid radiative effects and slow climate feedbacks, but the climate change amplitude of which is obviously weaker than that in fast or slow response alone. Using the multimodel means from Precipitation Driver Response Model Intercomparison Project, [Xie et al. \(2020\)](#) show the same tripole pattern of total precipitation response due to the BC forcings over eastern China, although the perturbation of their simulations is performed via multiplying the modern global BC emissions or concentrations by 10 times. Moreover, the meridional triple pattern of precipitation adjustments forced by BC in our results is in agreement with the dominant modes of summer precipitation anomalies over eastern China shown in previous observational studies ([Ding et al. 2008](#); [Huang et al. 2011](#); [Gong et al. 2017](#); [Duan et al. 2023](#); [Wang et al. 2023](#)). [Wang et al. \(2023\)](#) found a tripole pattern in decade-to-decade summer precipitation anomalies across the 20° – 40° N latitude band in EA since the 1990s, based on station observations and reanalysis data. The YRD area exhibited negative precipitation anomalies, while areas south and north of it showed positive precipitation anomalies. Their results also found that with the precipitation changes, positive SST anomalies appeared in parts of the northern Indian Ocean, western Pacific, and other sea areas. Thus, our results suggest that BC aerosol may serve as one of the forcing factors that contribute to the “positive–negative–positive” meridional triple pattern of summer precipitation anomaly over EA in recent years. [Section 3d](#) will use the AWT method to explore the main causes of the total EASM precipitation response and its fast and slow components induced by BC.

Additionally, by decomposing the EASM changes caused by different BC forcing effects, we found that the fast response of JJA mean precipitation and atmospheric circulation in the EASM region is consistent with the ARI-induced changes, indicating the dominating role of ARI effects in fast EASM changes ([Figs. 4d,e](#)), while the slow EASM response to BC forcings is controlled by both ARI and ACI effects ([Figs. 5g–i](#)). To isolate the ARI effects of BC, this study sets the aerosol optical depth (AOD) of BC to zero in the model’s radiative transfer scheme, effectively eliminating the ARI

effects of BC and the heating effects in the cloud (BC semidirect effect). The ARI effects of BC on the EASM ([Figs. 5b,e,h](#)) are the difference between the simulations with all BC effects and those neglecting the AOD of BC, thus encompassing the semidirect effect. In addition, the CAM5 model employs an ice nucleation scheme based on [Liu et al. \(2007\)](#), which considers homogeneous freezing of sulfate and heterogeneous immersion freezing on mineral dust ([Liu and Penner 2005](#)). However, the model has not treated immersion freezing on BC due to uncertainties of its efficiency. Therefore, at present, only the effects of BC–water cloud interaction are involved in [Fig. 5i](#) and control the changes in the slow responses of precipitation and atmospheric circulation.

d. Fast, slow, and total responses of summer atmospheric water cycles over EA

1) RESPONSES OF SUMMER ATMOSPHERIC WATER CYCLE PROCESSES

Water vapor evaporates from Earth’s surface, transports and holds in the atmosphere for some time, and then precipitates back to the surface. Large-scale anthropogenic activities can bring significant changes in this atmospheric water cycle process ([Eltahir and Bras 1996](#)). [Li et al. \(2015\)](#) pointed out that aerosol forcings dominate over other external forcings (e.g., the greenhouse effects) during the historical precipitation change period over the EASM-controlled region. Generally, the local evaporation and moisture transport processes are the main sources of precipitation. However, the adjustments of surface evaporation and moisture transport due to BC forcings are rarely isolated in previous studies. Here, the BC-induced fast and slow responses of surface evaporation and moisture transport are obtained with the AWT method and moisture budget analysis proposed by [Kim and Ha \(2015\)](#) to further explore the effects of BC forcings upon the summer atmospheric water cycle over EA.

The total climatic response of surface evaporation exhibits a decrease over most of the EA continent east to 110° E and an increase over the west Pacific and north Indian Ocean that surround the Asia continent ([Fig. 6a](#)). The local evaporation reduction at the EA land surface in the total response could be explained by the reduced surface solar radiation in the fast climatic response ([Figs. 6a,d](#)), while the local evaporation enhancement at the ocean surface surrounding the EA may be related to the BC-induced SST warming ([Figs. 6a,g](#)).

The moisture transport can be divided into the moisture convergence ([Fig. S2](#)), advection ([Fig. 7](#)), and eddy ([Fig. S3](#)) terms using the moisture budget analysis:

$$\begin{aligned} \bar{P} - \bar{E} = & - \left[\frac{1}{g} \int_{p_t}^{ps} (\bar{q} \nabla \cdot \bar{\mathbf{V}}) dp \right] - \left[\frac{1}{g} \int_{p_t}^{ps} (\bar{\mathbf{V}} \nabla \cdot \bar{\mathbf{q}}) dp \right] \\ & - \left[\frac{1}{g} \int_{p_t}^{ps} (\nabla \cdot \bar{\mathbf{q}}' \bar{\mathbf{V}}') dp \right]. \end{aligned} \quad (12)$$

The term on the left-hand side denotes the response of precipitation (P) minus evaporation (E). The first, second, and third terms on the right-hand side denote the moisture convergence,

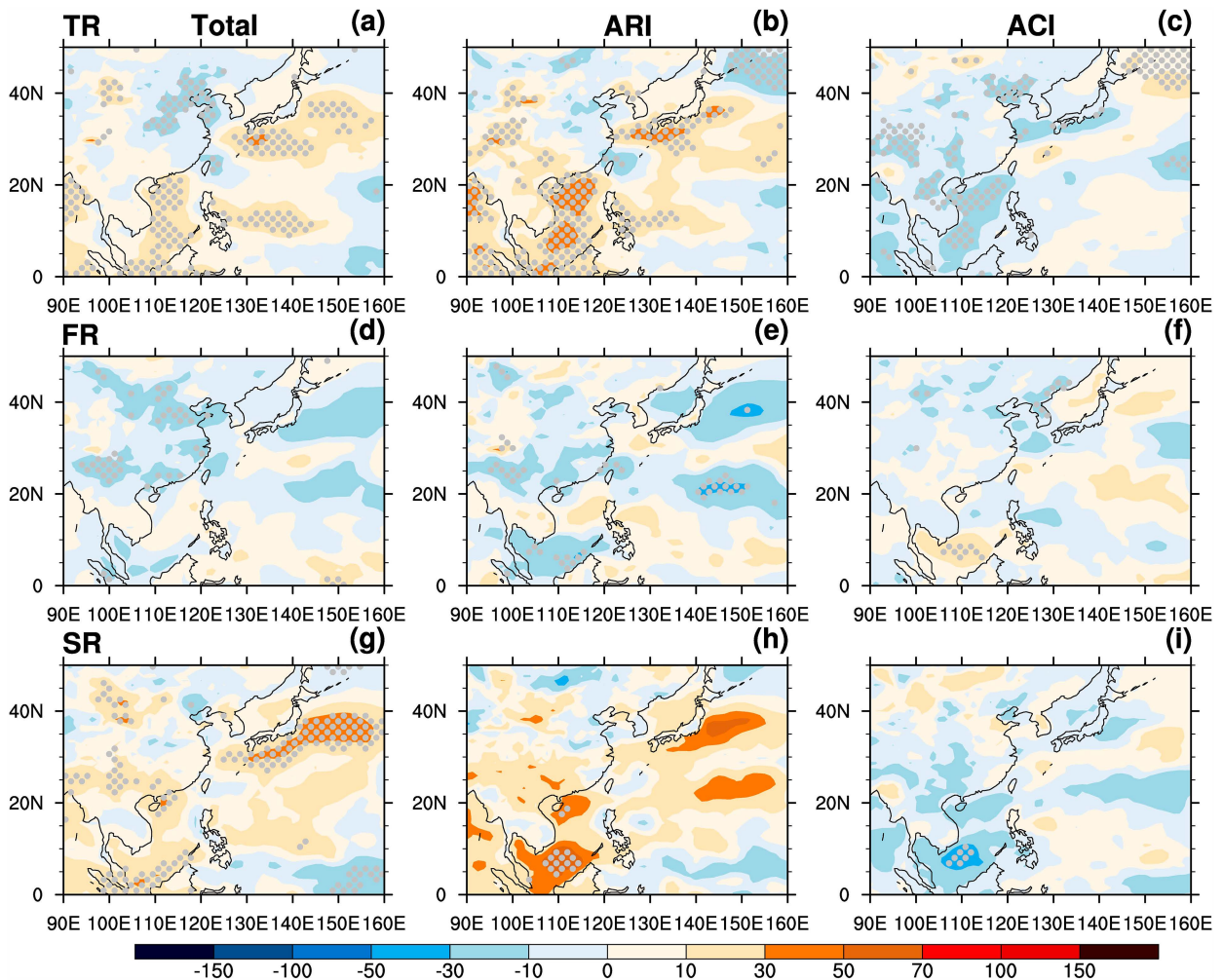


FIG. 6. Spatial distributions of the (a)–(c) TR, (d)–(f) FR, and (g)–(i) SR in the JJA mean surface evaporation ($10^{-1} \text{ mg m}^{-2} \text{ s}^{-1}$) to (a),(d),(g) BC total forcing and its (b),(e),(h) ARI and (c),(f),(i) ACI components over EA. Dotted regions denote where the surface evaporation anomaly is statistically significant at the 95% confidence level.

advection, and eddy terms, respectively. In this study, g , q , $\mathbf{V}$, p , p_t , and p_s are the gravity, specific humidity, horizontal wind vector, pressure, top pressure, and surface pressure, respectively. The superscript “ $\sim$ ” denotes climatological JJA values of different variables, and the symbol “ $\sim$ ” denotes the disturbance term of different variables.

The magnitude of changes in moisture advection (Fig. 7) is significantly larger than those in moisture convergence and eddy terms, indicating that the BC-induced moisture transport is primarily driven by the advection process in the EASM region. In addition, the BC-induced moisture advection change also outweighs the local evaporation change (Fig. 6). For the fast response of moisture advection (Figs. 7d–f), there is abundant outflow (inflow) in NC (SC), which is consistent with the precipitation decrease (increase) shown in Figs. 5d–f. The slow response of moisture advection over EA continent is contrary to the fast response, providing favorable (unfavorable) moisture condition for local precipitation. In addition, the fast adjustment of moisture advection over EA is close to the variation pattern

due to the ARI effect, while the adjustment in slow response is influenced by both ARI and ACI effects, which is same with the leading effects in precipitation change. The above results indicate that BC forcings regulate both the fast and slow adjustments of monsoon precipitation over EA through altering the moisture transported from remote source regions rather than affecting local evaporation.

2) RESPONSES OF THE CONTRIBUTIONS FROM DIFFERENT MOISTURE SOURCES

The precipitation changes in NC and SC show distinct difference either in the fast or slow EASM responses to BC forcings, and the difference is mostly contributed from the moisture transport from remote sources. Here, the regional contributions of the main water source regions to the summer precipitation changes over EA subregions were tracked and quantified with the AWT method. As shown in Fig. 8, BC induces disparate source contributions to the summer precipitation over NC and SC between fast and slow responses of the EASM. The

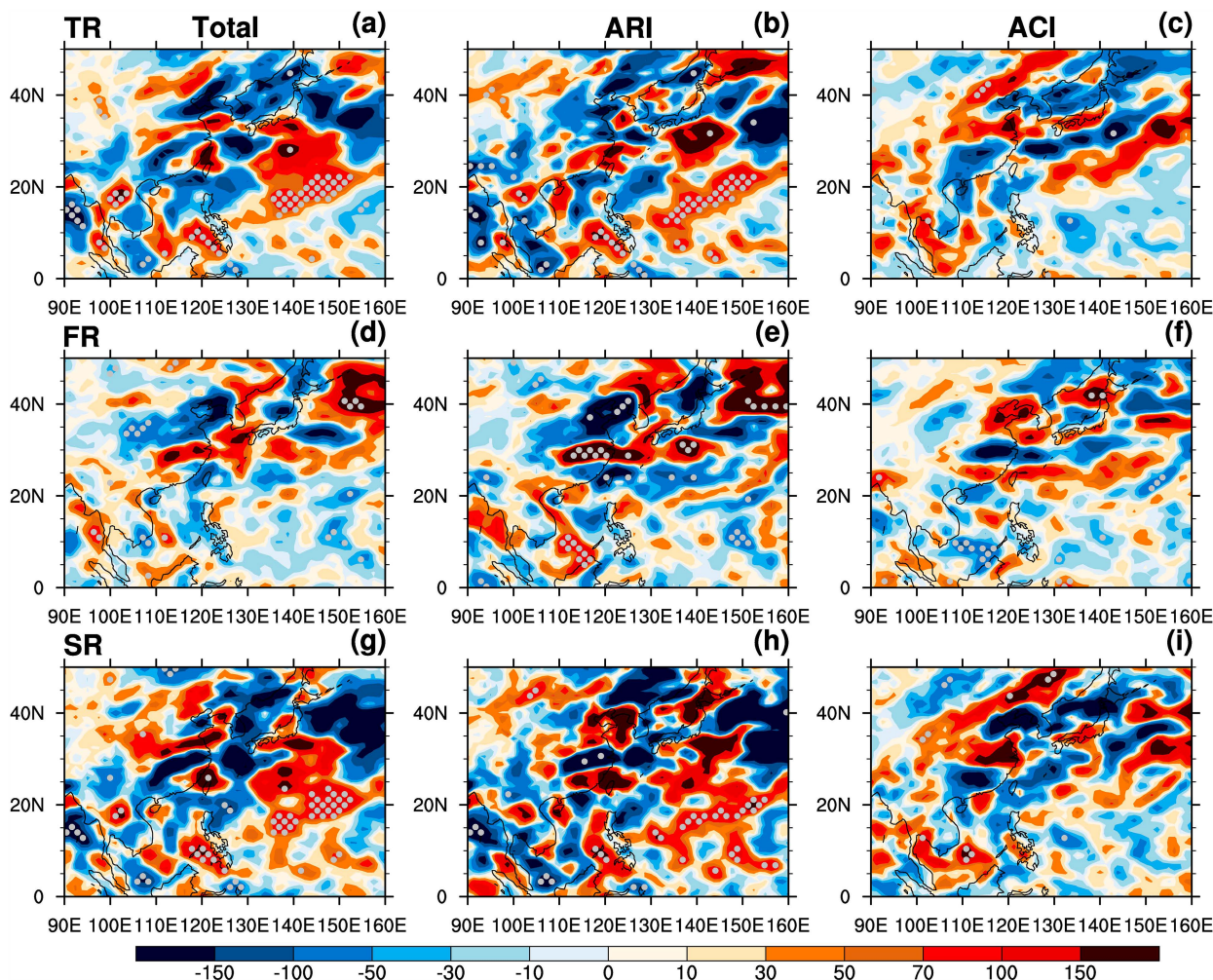


FIG. 7. Spatial distributions of the (a)–(c) TR, (d)–(f) FR, and (g)–(i) SR in the JJA mean moisture advection ($\text{mg m}^{-2} \text{s}^{-1}$) to (a),(d),(g) BC total forcing and its (b),(e),(h) ARI and (c),(f),(i) ACI components over EA. Dotted regions denote where the moisture advection anomaly is statistically significant at the 95% confidence level.

precipitation contributed from both local and remote moisture sources decreases (increases) over NC (SC) due to the fast responses, and vice versa for the slow responses of the EASM. Note that the moisture from tropical Indian Ocean (TIO) dominantly accounts for the absolute variations in precipitation over the subregions of EA, regardless the time scale of the EASM response. It supplies 20.9% (26.5%) of precipitation anomalies in NC and 46.7% (35.7%) in SC through fast (slow) responses. The spatial pattern of precipitation anomaly from the TIO source region also resembles the pattern of total summer precipitation over EA. This similarity applies to both the total response and its fast and slow components (Figs. 5a,d,g and 9a–c). The above reveals that the moisture contributed from TIO is most susceptible to BC effects and dominates the precipitation anomaly in the EA subregions, no matter in fast EASM responses or SST-mediated EASM responses, although the TIO-contributed precipitation in the BC-induced fast and slow responses of the EASM show opposite changes over EA.

In addition to the contributions from TIO, moisture from Bay of Bengal (BOB), South China Sea (SCS), and Indo-China Peninsula (ICP) also cause similar spatial pattern in the total response of precipitation over NC and SC but with much smaller extent (Fig. 10). The spatial pattern of precipitation adjustments from BOB, SCS, and ICP in the fast and slow components are also close to the TIO-contributed precipitation change (Figs. S4 and S5). The moisture originating in these source regions is borne on southwesterly winds toward EA. Zhou and Yu (2005) have summarized that the moisture for the EASM is transported through three main climatological branches. The first is by the southwesterlies, the second is from the northwestern Pacific (NWP), and the third is the cross-equator flow across 105° – 150° E. Among these, the cross-equator flow is the weakest. In our results, the NWP-contributed precipitation anomaly over EA land regions mainly occurs north of 30° N (Fig. 10d) and accounts for only 12.2% (7.1%) of fast (slow) precipitation adjustments over NC and 3.9% (16.4%) over SC (Fig. 8). Thus, the abundant moisture transported by the

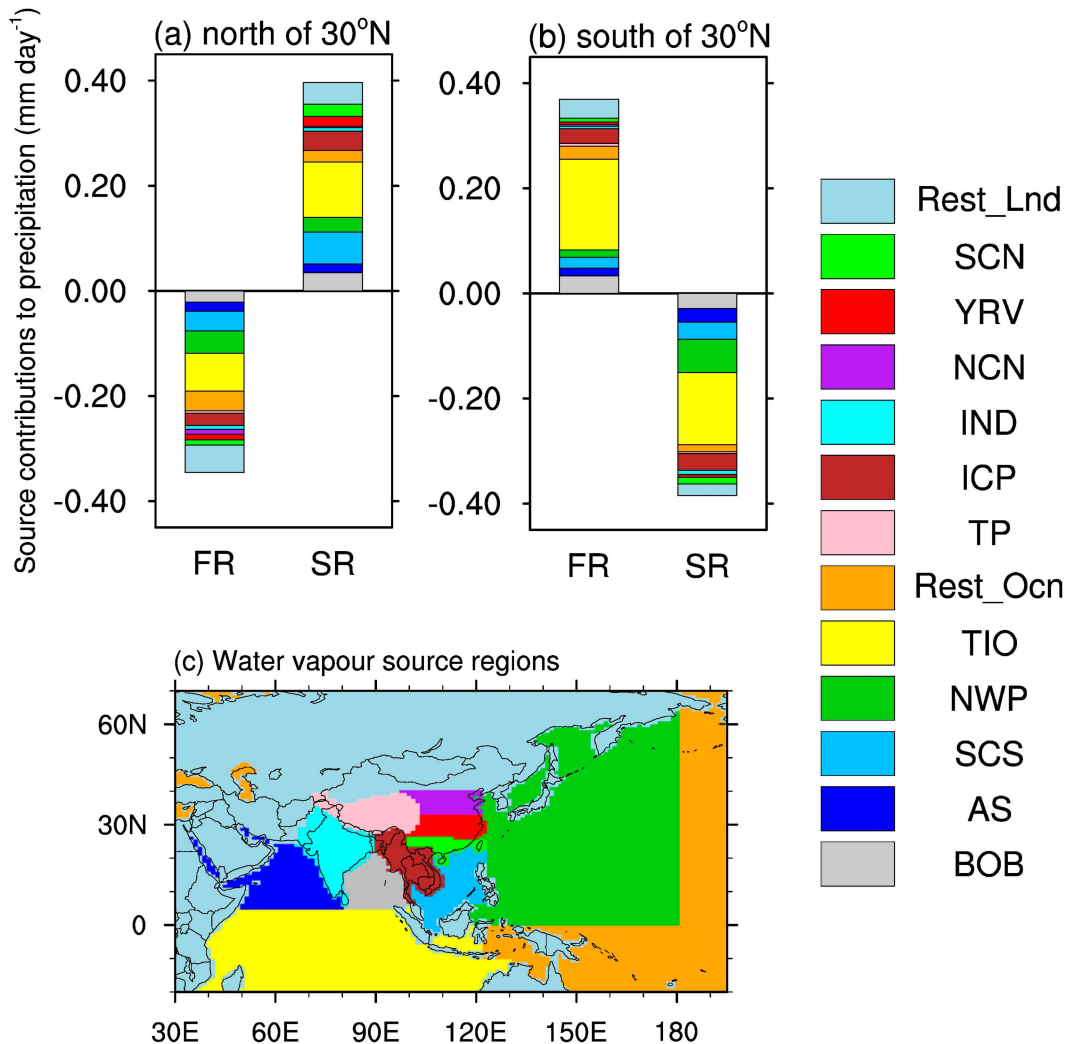


FIG. 8. Contributions (mm day^{-1}) of each regional moisture source to the JJA mean precipitation anomaly over (a) NC and (b) SC in fast and slow EASM responses to BC forcings. (c) The tagged moisture source regions [BOB, Arabian Sea (AS), SCS, NWP, TIO, Rest_Ocn, TP, ICP, India Peninsular (IND), NC, Yangtze River Valley (YRV), southern China (SCN), and Rest_Lnd].

southwesterlies from TIO and BOB, crossing the ICP and the SCS into EA, is more sensitive to the BC forcings and governs the corresponding precipitation change pattern of the EASM. Notice that the precipitation from rest ocean regions (Rest_Ocn) and rest land regions (Rest_Lnd) also show some adjustments to BC forcings over EA (Figs. S4 and S5). Pan et al. (2017) quantified that the moisture from ocean sources of South and northeast Pacific and land sources of Malay Archipelago and northeast Asia also supply nonnegligible summer precipitation over EA (accounting for about 7% and 9%, respectively). Therefore, the fast and slow adjustments of precipitation from Rest_Ocn and Rest_Lnd may be related to moisture transport from these regions.

e. Fast, slow, and total responses of East Asian upper-level jet

Considering the large contribution of regional moisture transport associated with the dynamic process to precipitation

changes over EA, it is necessary to explore the fast and slow responses of atmospheric circulation to BC forcings and the corresponding impacts on the EASM precipitation and circulation. As one of the dominant circulation features over EA, the East Asian subtropical westerly jet (EASWJ) is located in the upper troposphere and has been widely proven to have strong associations with the evolution of Asian summer precipitation (Lin and Lu 2005; Zhang et al. 2006; Xie et al. 2015). Research shows that the intensity, onset and retreat of the EASM, and summer precipitation changes could be explained by the variations of EASWJ intensity or meridional location (Jin et al. 2012; Xie et al. 2015; Wei et al. 2017). Compared with EASWJ in winter, the axis of summer EASWJ moved northward to 40°N with central values of 25–30 m s^{-1} at 200-hPa level (Yu et al. 2021), which is also shown in Fig. S6. The strongest EASWJ center at 200 hPa is located

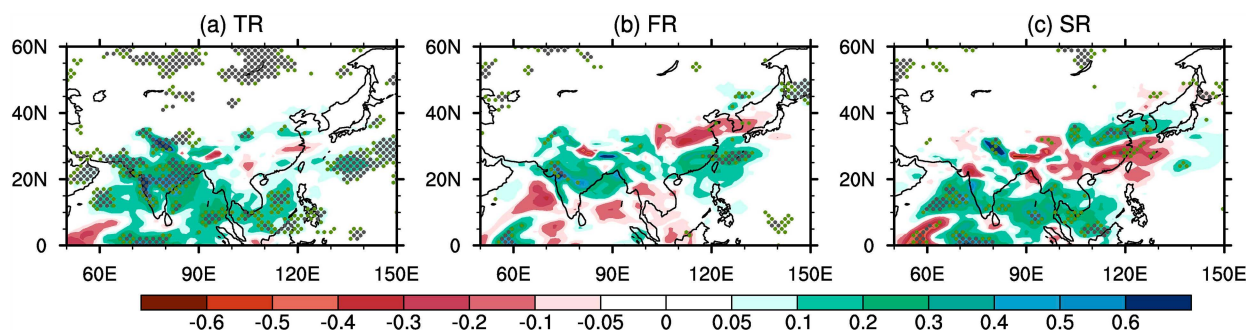


FIG. 9. Spatial distributions of the (a) TR, (b) FR, and (c) SR of the JJA mean precipitation anomaly (mm day^{-1}) transported from TIO to BC forcings over EA. Green and gray dotted regions denote where the precipitation anomaly is statistically significant at the 90% and 95% confidence level, respectively.

north of the Qinghai–Tibet Plateau, with wind speed values of about 30 m s^{-1} . In addition, a narrow westerly belt occurs across the Asian continent with wind speed values exceeding 20 m s^{-1} , extending eastward to the Japanese island and the northwest Pacific.

According to the thermal wind equation (Holton 2004), the zonal wind at the upper level is proportionally subjected by the meridional temperature gradient at the lower level. As a result, an increased poleward meridional temperature gradient at the lower level would weaken the upper zonal wind. In our study, in terms of the fast response of EASWJ to BC forcings as shown in Fig. 11b, a positive anomaly of meridional temperature gradient occurs below 40°N , the center of EASWJ, which is associated with the strong tropospheric warming. The high regional BC concentrations over northern EA (Fig. 1) induce the significant tropospheric warming, which could reach up to $0.6\text{--}0.8 \text{ K}$ over $40^\circ\text{--}50^\circ\text{N}$ (Fig. S7d). Therefore, the meridional temperature gradient has a positive anomaly of $0.04\text{--}0.06 \text{ K } 100 \text{ km}^{-1}$ south of the maximum warming region ($40^\circ\text{--}50^\circ\text{N}$), encompassing the lower layer of the EASWJ center, while a negative anomaly of -0.04 to $-0.02 \text{ K } 100 \text{ km}^{-1}$ is observed north of this warmed area. The significant positive anomaly of the low-level meridional temperature gradient between 30° and 45°N markedly weakens the corresponding high-level zonal wind speed (Fig. 11b). The center of the weakened zonal wind speed is around 200 hPa near 40°N , with the maximum wind speed reduced by more than 1.4 m s^{-1} , which weakens EASWJ.

Generally, upper-level divergence and ascending lower-level motion are generated on the south side of EASWJ, which in turn causes a convergence zone in the lower atmosphere, thereby forming a vertical circulation (Dong et al. 2010). There is a strong divergence on the south side of the upper westerly belt near 40°N , of which the divergence zone above the Qinghai–Tibet Plateau is the strongest (Fig. S6). The upper-level divergence also appears from the YRD to the Japanese island ($30^\circ\text{--}40^\circ\text{N}$). Therefore, the weakened EASWJ in the fast climatic responses to BC forcings leads to reduced upper-level divergence on its south side. Figure 12b shows anomalous upper-level convergence over the YRD to Japan ($30^\circ\text{--}40^\circ\text{N}$), which excites descending motion around $30^\circ\text{--}40^\circ\text{N}$ and strengthened ascending airflow around $20^\circ\text{--}30^\circ\text{N}$

(Fig. 12e). Eventually, anomalous divergence and convergence correspondingly appear in the lower atmosphere ($925\text{--}500 \text{ hPa}$) over $30^\circ\text{--}40^\circ\text{N}$ and $20^\circ\text{--}30^\circ\text{N}$, respectively (Fig. 12h). As a result, an anticyclonic anomaly appears at 850 hPa north of the YRD ($30^\circ\text{--}40^\circ\text{N}$; Fig. 5d), with moisture flowing out (Fig. 7d) and reduced precipitation (Fig. 5d). In contrast, moisture flows in (Fig. 7d) and precipitation increases (Fig. 5d) over $20^\circ\text{--}30^\circ\text{N}$. The meridional overturning circulation between higher and lower latitudes under the impacts of the upper westerly jet is also shown in previous model-based study (Wang et al. 2019).

In contrast, in the slow climatic responses to BC forcings over EA, a negative meridional temperature gradient appears over $30^\circ\text{--}45^\circ\text{N}$ (Fig. 11c), with a value of $-0.4 \text{ K } 100 \text{ km}^{-1}$. The warmed SST induced by BC over the north Indian Ocean, South China Sea, and western Pacific adjacent to the Asian continent (Fig. 3) further warms the air over the adjacent EA continent north of 20°N (Fig. S7g). The maximum warming amplitude is the same as that of the fast atmospheric temperature response, reaching up to $0.6\text{--}0.8 \text{ K}$, thereby causing anomalous negative meridional temperature gradients below the EASWJ center. Therefore, the westerlies above the negative meridional temperature gradient are enhanced, and the amplitude of the 200-hPa wind speed enhancement is in accordance with that of the weakened westerlies in the fast response. The westerlies increase by 1.4 m s^{-1} at 40°N , resulting in an EASWJ intensification (Fig. 11c). Therefore, upper-level divergence is strengthened south of EASWJ around $30^\circ\text{--}40^\circ\text{N}$ (Fig. 12c), with anomalous ascending motions (Fig. 12f) and low-level convergence anomalies (Fig. 12i). Meanwhile, strengthened descending motions (Fig. 12f) and low-level divergence anomalies (Fig. 12i) appear over $20^\circ\text{--}30^\circ\text{N}$. Subsequently, the moisture flows in over $30^\circ\text{--}40^\circ\text{N}$ (Fig. 7g) and precipitation increases (Fig. 5g), while the opposite occurs over $20^\circ\text{--}30^\circ\text{N}$.

Under the combined impacts of the direct atmosphere adjustment to anthropogenic BC and the slow climate feedback mediated by SST changes, there are divergence anomalies in the upper atmosphere over NC, southern northeast China, and a small part of SC (Fig. 12a), with compensatory lower-level convergence anomalies between 925 and 500 hPa (Fig. 12g). However, over the latitudinal band around 30°N covering the YRD and its adjacent sea areas, there exist lower-level divergence

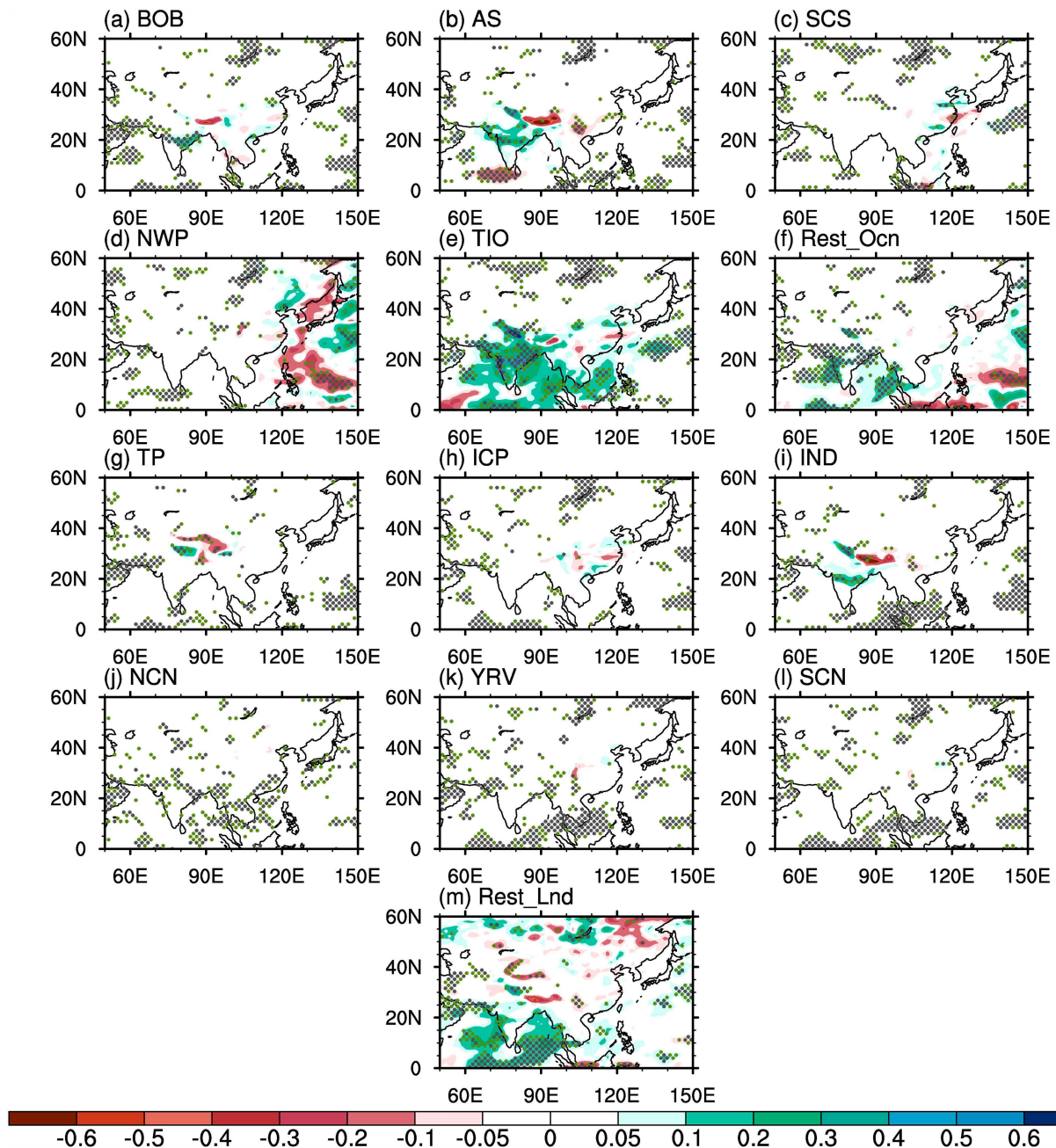


FIG. 10. Spatial distributions of the total responses of the JJA mean precipitation anomaly (mm day^{-1}) contributed by each tagged moisture source region to BC forcings over EA. Green and gray dotted regions denote where the precipitation anomaly is statistically significant at the 90% and 95% confidence level, respectively.

anomalies caused by upper-level convergence anomalies, but such changes only pass the significance test in certain regions.

4. Discussion

The simulated EASM responses to BC forcings are compared with the results of previous studies. The simulated fast EASM circulation response in our study demonstrates mechanistic

consistency with the fast precipitation response, showing anomalous cyclones at 850 hPa with increased precipitation in regions south of 30°N , characterized by abnormal southerly winds over adjacent oceans and northerly winds over land. Conversely, areas north of 30°N exhibit anticyclonic anomalies with enhanced southerly winds over the continent where precipitation decreases. In Wang et al. (2017b), the fast response of EASM circulation strengthens from south to north eastward of 110°E with

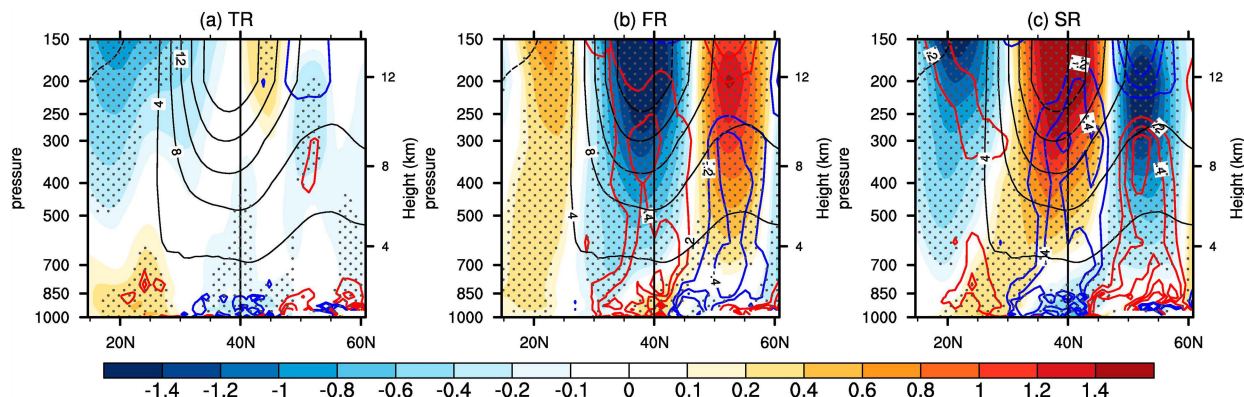


FIG. 11. Vertical distributions of the zonal-averaged (100° – 140° E) (a) TR, (b) FR, and (c) SR of JJA mean zonal wind (shading; m s^{-1} ; red denotes westerlies and blue denotes easterlies) and meridional temperature gradient (red lines; $10^{-1} \text{ K } 100 \text{ km}^{-1}$; red lines denote positive values and blue lines denote negative values) to BC forcings over EA. Black contour represent the climatological JJA mean zonal wind (m s^{-1} ; solid lines denote westerlies and dash lines denote easterlies) from TR-type simulation with emissions set to PI level. Black vertical line at 40°N represents the climatological JJA mean center of EASWJ from TR-type simulation with emissions set to PI level. Dotted regions denote where the zonal wind anomaly is statistically significant at the 95% confidence level.

increased precipitation over NC and the YRD, while Zhuang et al. (2018) highlight EASM strengthening primarily in SC, with increased precipitation across SC and part of NC. Hence, although our simulated changes in precipitation in NC and EASM circulation differ from Wang et al. (2017b) and Zhuang et al. (2018), our simulation aligns with their results regarding the increased precipitation in the area south of 30°N (including the YRD and SC). Moreover, the regional precipitation decrease in NC but increase in SC in terms of the fast EASM response to BC forcings in our study is consistent with the results of previous studies using different models (Menon et al. 2002; Lau et al. 2006; Wu et al. 2008; Liu et al. 2009; Mahmood and Li 2014; Wang et al. 2015; Zhuang et al. 2019). Our findings regarding the fast EASM circulation responses are also consistent with Zhuang et al. (2019; Fig. 9a in their study). They utilized a different model (regional climate model RegCM4) and similarly simulated anticyclonic anomalies in NC with reduced precipitation in some regions, contrasting with the cyclonic precipitation and increased precipitation in SC.

Additionally, Wang et al. (2017b) also focused on the total response of the EASM to BC forcings, which mirrors our simulation results displaying a “wetter–drier–wetter” triple meridional precipitation pattern over eastern China and enhanced EASM circulation. Similarly, Xie et al. (2020) utilized the multimodel means from the Precipitation Driver Response Model Intercomparison Project and pointed out the consistent characteristics of the total EASM response with our simulated results. Based on the consistency with previous research findings, this study highlights the distinct fast and slow EASM responses over EA.

Therefore, although there are differences in some aspects compared to the simulated results from previous studies (e.g., Wang et al. 2017b; Zhuang et al. 2018), there are also similarities in certain aspects. The differences may stem from several factors. These could include differences in initial conditions,

the specific BC emission scenarios, representation of land–sea interactions, and the complexity of atmospheric dynamics captured by the models. More climate simulations for the total, fast, and slow EASM responses to BC forcings with different models and settings are needed in the future.

In this study, we investigate the fast and slow responses of summer atmospheric circulation and atmospheric water cycle over EA to the ARI and ACI effects of BC. Scattering aerosols, such as sulfate, can also induce significant fast and slow climate responses over EA, with resulting adjustments that differ markedly from those induced by absorbing aerosols like BC. For instance, simulation results based on the CESM1 by Wang et al. (2017b) indicate that the total climate response caused by sulfate weakens EASM circulation and reduces precipitation, in contrast to the enhancement of the EASM driven by BC. Both fast and slow components of the EASM total response induced by sulfate contribute to the EASM weakening, with the slow component playing an even more substantial role. Using a different climate model (the atmospheric general circulation model of the Beijing Climate Center coupled with the China Meteorological Administration’s Unified Atmospheric Chemistry Environment for Aerosols), Wang et al. (2017a) also showed the opposite changes in the total response of the EASM precipitation due to sulfate and BC, although the changes induced by BC were smaller in magnitude. Additionally, Fang et al. (2023), utilizing the U.K. Earth System Model, version 1, pointed out that sulfate and BC also have opposing effects on the EASM duration adjustments. And the aerosol-induced changes in EASM intensity are primarily driven by scattering aerosol perturbations, while adjustments to the onset and withdrawal dates of the EASM result from the combined impacts of scattering and absorbing aerosols. Therefore, the fast and slow EASM responses induced by sulfate may offset the effects of BC. Future climate projections should focus on the fast and slow responses of the EASM and summer atmospheric water cycle process to different aerosol types and the resulting

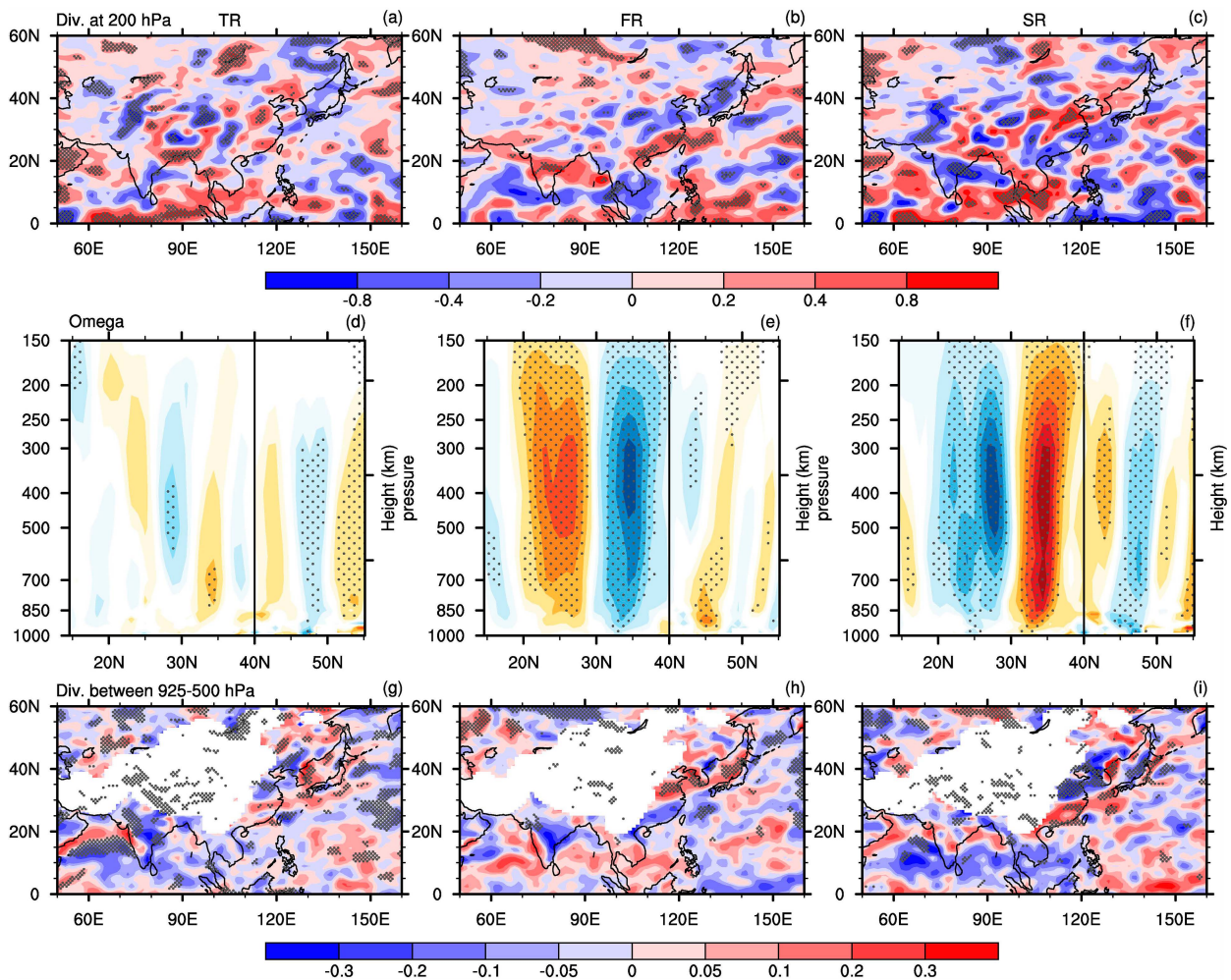


FIG. 12. Spatial distributions of the (a) TR, (b) FR, and (c) SR of JJA mean divergence field (10^{-6} s^{-1}) at 200 hPa to BC forcings over EA. (d)–(f) As in (a)–(c), respectively, but for responses of zonal averaged (100° – 140°E) vertical motions ($10^{-2} \text{ Pa s}^{-1}$). Black vertical lines at 40°N in (d)–(f) represent the climatological JJA mean center of EASWJ from TR-type simulation with emissions set to PI level. (g)–(i) As in (a)–(c), respectively, but for responses of divergence field (10^{-6} s^{-1}) between 925 and 500 hPa. Dotted regions denote where the divergence or vertical motion anomaly is statistically significant at the 95% confidence level.

adjustment scales. Given the higher efficiency of sulfate as a cloud condensation nucleus compared to the BC, studies also need to distinguish the ARI and ACI effects of sulfate on the summer atmospheric water cycle process responses over EA.

Moreover, aerosol climate effects exhibit significant regional variations across EA. This study identifies eastern China as the region experiencing the most pronounced precipitation adjustments due to BC forcings, whereas western regions show relatively minor impacts. Simulations by Wang et al. (2017b) indicate that anthropogenic sulfate demonstrates the strongest annual mean ERF in the Sichuan Basin, central China, and eastern China. Additionally, Zhang et al. (2012) emphasized considerable seasonal variation in nitrate forcing across China, peaking at -3.83 W m^{-2} during winter in central China. Liu et al. (2019) indicated that the response of surface air temperature in China to the total aerosol radiative effect ranges from -0.1 to 1.1 K , exhibiting strong spatial and seasonal variability.

Wang et al. (2015) reported that the overall changes in summer surface temperature, precipitation, and wind speed at 850 hPa over EA due to total aerosol effects are -0.31 K , $-0.29 \text{ mm day}^{-1}$, and -0.18 m s^{-1} , respectively, while the ARI effects of BC cause changes of $+0.08 \text{ K}$, $-0.07 \text{ mm day}^{-1}$, and $+0.02 \text{ m s}^{-1}$, respectively. They mentioned that the ARI effects of BC should not be overlooked in certain localized areas, although the total aerosol effect is generally large.

The varying impacts of aerosols across regions may be attributed to local climatic conditions and differences in aerosol concentration. Utilizing data from the AERONET, Zhang and Li (2019) reported that dust predominantly occurs in spring over North and northwest China, whereas scattering mixed aerosols are most common in spring and summer. Conversely, absorbing aerosols prevail during the heating season in autumn and winter. Additionally, scattering fine particulate aerosols are most frequently observed in eastern China during

summer. Therefore, future research could further identify the dominant aerosol types responsible for driving climate change across various regions of EA.

Note that the simulations in this study are based on the MAM3 aerosol module, where BC ages immediately after emission, while BC may be coated by other aerosol components in the aging process with enhanced light absorption ability (lens effect). Therefore, the ARI effects of BC in the simulation results may be overestimated. Aged BC can act as cloud condensation nuclei, and the immediate aging of BC in MAM3 will also affect the simulated ACI effects. Therefore, the fast and slow climate responses of the EASM and summer hydrological cycles due to the ARI and ACI effects of BC shown in this study need to be interpreted with caution. Liu et al. (2022) estimated that the annual mean ERF due to the ARI effects of BC over China is $2.4\text{--}3.0\text{ W m}^{-2}$ using observations and the latest CMIP6 simulation results, while our simulated annual BC-induced ERF_{ARI} over East Asia is 2.24 W m^{-2} , indicating the reasonability of the ARI impacts of BC on the EASM in this study. However, future studies still need to use the MAM4 aerosol module with the BC aging process (Liu et al. 2016) or incorporate the treatments of BC aging (e.g., Croft et al. 2005; Yu et al. 2012) into the original aerosol module to investigate the relative importance of the ARI and ACI effects of BC on the EASM and summer hydrological cycles and their underlying mechanisms.

5. Conclusions

This study has examined the inconsistent roles of fast atmospheric adjustments and slow SST-mediated responses caused by anthropogenic BC forcings in shaping the regional precipitation and atmospheric circulation adjustments over the EASM-controlled region with the CESM1 model. The responses of the summer atmospheric water cycle on two time scales to BC forcings are particularly quantitatively revealed using the model-based AWT method.

By comparing with the observations and reanalysis, we validate the ability of model to capture the spatial feature and magnitude of BC mass concentrations, light absorption capability, and EASM characteristics over EA during summertime. The consistence between the global pattern of SST changes induced by BC forcings in this study and that from previous research using fully coupled ocean reflects the reasonability of the simulated results based on the coupled SOM model.

Opposite regional climatic responses to the BC forcings appear between NC and SC (bounded by 30°N). This contrasting change over NC and SC during summertime associated with the EASM variation has been mentioned in previous studies (e.g., Yao et al. 2008; Fu et al. 2017; Li et al. 2017). Some studies involving the fast responses of the EASM due to BC impacts have shown that anthropogenic BC could result in a different variation in precipitation and regional circulation over NC and SC based on global or regional climate models (Menon et al. 2002; Liu 2008; Wang et al. 2015; Zhuang et al. 2019). In our study, the discrepancy of precipitation and atmospheric circulation between NC and SC mediated by

BC-induced SST changes is highlighted. Our results show that the fast EASM responses over EA manifest as the decrease in precipitation in NC but an increase in SC. However, slow climatic feedback over EA acts in the opposite way, manifesting as the anomalous northerly wind and increased precipitation in NC and an anomalous southerly wind and decreased precipitation in SC.

The fast and slow responses of local evaporation and moisture transport processes to BC forcings over EA are less discussed in previous studies. We found that both the fast and slow adjustments of precipitation over the EASM regions responding to BC impacts were regulated by regional transport of moisture rather than local evaporation change. The moisture contributed from tropical Indian Ocean (TIO) is most susceptible to BC forcings and dominates the precipitation anomaly over EA subregions either in the total EASM response or in its fast and slow components. In fact, the channel of moisture migrated by southwesterlies from TIO, Bay of Bengal, South China Sea, and Indo-China Peninsula triggers opposite fast and slow precipitation anomaly pattern over EA and is responsible for the different precipitation responses between NC and SC.

Our study emphasizes the dominant role of BC-induced atmospheric circulation changes in regulating the distribution of summer precipitation anomaly over EA (Fig. 13). For the fast adjustments, the atmospheric warming above northern EA caused by the local high BC concentrations results in positive anomalies of poleward meridional temperature gradient at the lower level of EASWJ center around 40°N and weakens the EASWJ intensity. The zonal wind fields adjustments at 200 hPa leads to a significant upper convergence descending action in $30^\circ\text{--}40^\circ\text{N}$, while strengthened ascending motions and low-level convergence anomalies appear in $20^\circ\text{--}30^\circ\text{N}$, thereby inducing precipitation decrease in $30^\circ\text{--}40^\circ\text{N}$ but increase in $20^\circ\text{--}30^\circ\text{N}$. However, the decomposed SST-mediated summer climatic responses forced by BC show opposite signals over EA compared with the fast response. The BC-induced strong Indo-Pacific warm pool warming heats the atmosphere over oceans and adjacent continent south of the EASWJ center, causing negative anomalies of meridional temperature gradient under the EASWJ center. As a result, EASWJ and the upper-level divergence from the YRD to the Japanese islands at the upper level is strengthened, and a corresponding lower-level convergent anomaly and increased precipitation appear over NC. However, a strengthened descending motion and the lower-level divergence weakens the precipitation over SC.

In addition, this study finds that the temperature increase over EA in total, fast, and slow responses to BC total forcings is dominated by the ARI effects, while the ACI effects of BC decrease the temperature over EA. The fast responses of the EASM precipitation and atmospheric circulation to the BC forcings are dominated by the ARI effect, while the slow responses are influenced by the cooperation of ARI and non-ARI effects of BC. Given the relatively small RF induced by SAC over EA (east to 110°E) in summer, we infer that the slow EASM responses are jointly governed by the ARI and ACI effects.

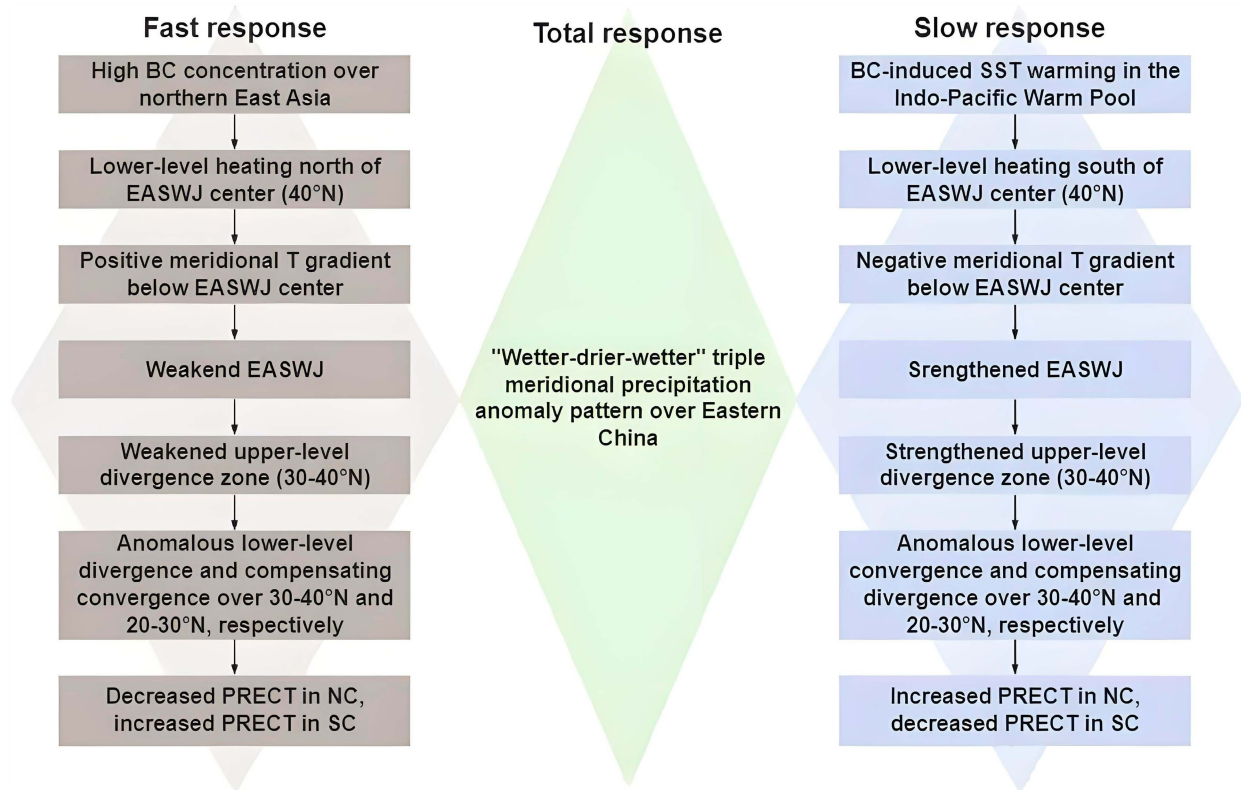


FIG. 13. The impacts of the BC-induced summer atmospheric circulation responses on the EASM precipitation anomalies. PRECT is the abbreviation for precipitation.

The “wetter–drier–wetter” triple meridional precipitation pattern is found over eastern China in terms of the BC-induced total summer atmospheric water cycle response. Observational studies (Duan et al. 2023; Wang et al. 2023) also found the meridional modes of “positive–negative–positive” pattern as the main structures of decadal-to-decadal summer precipitation anomaly over eastern China, meanwhile positive SST anomalies appeared in parts of the northern Indian Ocean and western Pacific. Therefore, the total response of the EASM to BC forcings shown in our study may hint one of the possible causes for this meridional pattern of EA summer precipitation anomaly.

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Data availability statement. The CESM1 code is available online (<https://svn-ccsm-models.cgd.ucar.edu/cesm1/>). The BC concentrations and aerosol absorption optical depth from the MERRA-2 reanalysis are available at https://disc.gsfc.nasa.gov/datasets/M2TMNXAER_5.12.4/summary?keywords=MERRA-2%20avgM (last access: 24 December 2017). The CPC unified gauge-based daily observations and the GPCP rain gauge–satellite combined precipitation dataset can be obtained

from <https://psl.noaa.gov/data/gridded/data.cpc.globalprecip.html> (last access: 2 January 2022) and <https://www.ncei.noaa.gov/data/global-precipitation-climatology-project-gpcp-daily/access/> (last access: 15 February 2022), respectively. The wind fields from the ERA5 are available at <https://doi.org/10.24381/cds.6860a573> (last access: 16 January 2020). The cloud fraction from the MODIS satellite observation is available at <https://ladsweb.modaps.eosdis.nasa.gov/search/order> (last access: 12 September 2024).

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