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Physical processes influencing the Asian climate due to black carbon emission over East and South Asia

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Abstract

Many studies have shown that black carbon (BC) aerosols over Asia have significant impacts on regional climate, but with large diversities in intensity, spatial distribution and physical mechanism of regional responses. In this study, we utilized a set of Systematic Regional Aerosol Perturbations (SyRAP) using a reduced complexity climate model, FORTE2, to investigate responses of the Asian climate to BC aerosols over East Asia only, South Asia only, and both regions at once, and thoroughly examine related physical processes. Results show that regional BC aerosols lead to a strong surface cooling, air temperature warming in the low-level troposphere, and drying over the perturbed areas, with seasonal differences in magnitude and spatial distribution. Atmospheric energy budget analysis suggests that reductions in local precipitation primarily depend on the substantial local atmospheric heating due to shortwave absorption by BC. Increases in dry static energy (DSE) flux divergence partly offset the reduced precipitation over north China in summer and most of China and India in the other three seasons. Decreases in DSE flux divergence lead to stronger reduction in precipitation over south China and central India in summer. Changes in DSE flux divergence are mainly due to vertical motions driven by diabatic heating in the middle and lower troposphere. BC perturbations also exert non-local climate impacts through the changes in DSE flux divergence. This study provides a full chain of physical processes of the local climate responses to the Asian BC increases, and gives some insights to better understand the uncertainties of model responses.

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1. Introduction

Black carbon (BC) aerosol, a short-lived pollutant and climate forcer, is emitted from the incomplete combustion of biomass and fossil fuels, and exerts significant effects on global and regional climate (Ramanathan and Carmichael, 2008; Bond et al., 2013; Stjern et al., 2017; IPCC, 2021; Li et al., 2022). Alongside rapid economic developments of China and India over the past few decades, East and South Asia have become the highest BC emissions hotspots in the world. Despite BC emissions from China decreasing substantially in the past decade, East and South Asia are expected to remain the highest BC loadings globally in the coming decades (Lund et al., 2019). Hence, the climate impacts of BC emissions from East and South Asia have been extensively investigated (e.g. Li et al., 2016; Lou et al., 2019; Xie et al., 2020; Westervelt et al., 2020; Herbert et al., 2022; Yang et al., 2022). Although many model studies have shown that Asian BC aerosols are of great importance for local climate (especially the Asian monsoonal systems), considerable uncertainty exits regarding the intensity and spatial distribution of the Asian climate responses to BC forcings, as well as the related physical mechanisms.

Menon et al. (2002) found that during summer, a large BC forcing induced a “southern flooding and northern drought” (SFND) precipitation pattern in China, and moderate cooling in China and India based on the Goddard Institute for Space Studies (GISS) global climate model (CGCM). However, many subsequent studies were not able to reproduce the SFND pattern, in part due to the poor representation of the Asian summer monsoon in models (Wilcox et al., 2015). Gu et al. (2006) found that BC forcing acted to suppress precipitation in southern and eastern China due to warming in the middle to high latitudes based on an atmospheric general circulation model (AGCM). Zhang et al. (2009) showed that the effects of direct radiative forcing due to increased BC aerosols can decrease precipitation and increase surface temperature in southern China and India, but cause the opposite responses in northern China, by comparing CAM3 AGCM simulations with the all-aerosol types and without carbonaceous aerosol. Liu et al. (2018) found a similar dipole precipitation pattern with a decrease in southern China and an increase in the north mainly due to the fast

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adjustments (simulations with fixed SST) of BC forcing, and a surface cooling over Asia except over the Himalayan region in the 10x modern Asian BC emissions or concentrations experiments in the Precipitation Driver Response Model Intercomparison Project (PDRMIP) (Myhre et al., 2017).

Wang et al. (2017) conducted CESM1 simulations with increased BC emissions from preindustrial to present-day level, and proposed that the fast adjustment strengthened the EASM, while slow adjustment (response to aerosol-induced SST change) dominated the spatial pattern of precipitation response, which showed a tripolar precipitation pattern with wetting-drying-wetting from north to south China, although the responses in most regions were statistically insignificant. The fast/slow responses in Wang et al. (2017) are different with those in Liu et al. (2018). Xie et al. (2020) used the PDRMIP ensemble to show that the extremely high global/Asian BC forcing can cause the similar tripolar precipitation pattern over eastern China in summer, and they stressed that these responses mainly result from the enhanced upper-level atmospheric temperature over Asia instead of low-level thermal feedbacks. Mahmood and Li (2014) showed that South Asian BC also can induce the tripolar precipitation pattern over East China via a propagating wave train along the Asian upper tropospheric jet based on ensemble sensitive experiments in GFDL AM2.1.

Persad et al. (2017), based on GFDL AM3, have shown that surface solar dimming dominates the reduction of the East Asian summer precipitation due to the decreased land-sea contrast, whilst atmospheric heating from absorbing aerosols partially offset the reduction, which leads to a smaller reduction when both effects are considered simultaneously. Jiang et al. (2013) and Guo et al. (2013) indicated that there was no statistically significant change in precipitation over East China in response to global BC forcing during summertime, based on the CAM5 and HiGAM AGCM experiments, respectively. Although there are many differences across studies, some similarities can still be found in most of studies, such as surface cooling in the perturbation area, wet response in North China.

In addition, some AGCM/CGCM studies suggested that increased BC aerosols can lead to a weakened South Asian summer monsoon (SASM) (Lau and Kim, 2007; Meehl et al., 2008), while others found that there should be a strengthened SASM via increasing the atmospheric meridional land-sea thermal gradient, or an elevated heat pump effect (Lau and Kim et al., 2006; Xie et al., 2020; Westervelt et al., 2020). On the other hand, Westervelt et al. (2018) conducted aerosol removal experiments using three CGCMs, and found that Indian BC decreases lead to no change or a small decrease in precipitation in India.

By comparing the sum of Asian summer climate responses to individual responses over East and South Asia with the responses to simultaneous forcing, the regional linearity of BC forcing has been investigated. Chen et al. (2020) and Herbert et al. (2022) suggested that the responses were highly nonlinear due to the interaction of atmospheric circulation changes, based on the regional climate model RegCM4 and the Intermediate General Circulation Model 4 (IGCM4) simulations, respectively. In contrast, Reccia and Lucarini (2023) and Stjern et al. (2024) found that the responses were almost linear in most of Asian regions. The difference may be related to the different spatial extend of the aerosol perturbation in the simulation design (Stjern et al., 2024).

For the East Asian winter climate, Jiang et al. (2017) found that BC forcing can lead to an intensification of the East Asian winter monsoon (EAWM) northern mode via heating Tibetan Plateau using the CAM5 model. On the contrary, Lou et al. (2019) suggested that BC emitted from the North China can weaken the EAWM through ocean, sea ice, cloud feedbacks based on CESM. BC aerosol also can impact on spring and autumn precipitation in China (Guo et al., 2013; Hu and Liu et al., 2014; Deng et al., 2014).

These inconsistent results, and the large uncertainty in the simulated response of the Asian climate to BC changes, are partly related to differences in the modeling approach (e.g., AGCMs versus CGCMs/ESMs) and also the magnitude and location of the BC perturbation. Atmosphere-only GCMs lack SST feedbacks, which are crucial in influencing the Asian monsoon (Dong et al., 2019), while CGCMs or ESMs involving more and complex physical processes make it difficult to identify the key physical processes of impacts of regional BC aerosols on Asian climate. The inconsistency may also be associated with model-specific differences. Different models that include

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different physical processes, combined with different experiment designs that can influence the atmospheric circulation response, mean that understanding the causes of differences between studies is very difficult. Hence, reduced complexity models, such as FORTE2 (Fast Ocean Rapid Troposphere Experiment version 2), provide an alternative and useful tool for such studies, given that such models not only include all the main mechanisms of aerosol-climate interactions, but also allow fast speed of integration and longer simulations with lower cost. Stjern et al. (2024) have utilized FORTE2 to perform a series of Systematic Regional Aerosol Perturbations (SyRAP) simulations with absorbing and scattering aerosol species in East Asia only, South Asia only, and both regions simultaneously. Their results have shown that SyRAP-FORTE2 is a helpful framework to understand and decompose the local and remote climate effects of regional aerosol emissions.

Therefore, based on the simulations of the regional BC perturbations in SyRAP-FORTE2, this study aims to address the following two questions: (1) what are responses of Asian climate to either East Asian or South Asian BC emissions, or both regions at once, respectively? (2) What are key physical processes involved in these responses?

The rest of the paper is organized as follows: section 2 describes the FORTE2 model, the SyRAP simulations, and the analysis methods and datasets; section 3 briefly evaluates the climatology of SyRAP simulations, examines responses to the regional BC perturbations in SyRAP-FORTE2, and investigates the underlying physical processes involved in these responses; section 4 compares the results of the atmospheric energy budget in SyRAP-FORTE2 with those in the PDRMIP models; Finally, the summary and discussion are provided in section 5.

2. Methods

2.1 The FORTE2 model

FORTE2 is an intermediate-complexity coupled atmosphere-ocean general circulation model (Blaker et al., 2021). The atmospheric component is the Intermediate General Circulation Model 4 (IGCM4) with a horizontal resolution of approximately 2.8° (T42), and 35 sigma levels extending up to 0.1 hPa (Joshi et al., 2015). IGCM4

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includes schemes for radiation, land-surface properties, convection, precipitation, and clouds (Zhong and Haigh, 1995; Betts and Miller, 1993). The oceanic component is the Modular Ocean Model-Array (MOMA) with a horizontal grid spacing 2°×2°, and 15 z-layer levels increasing in thickness with depth from 30m at the surface to 800m at the bottom (Webb, 1996). Sea ice is represented by a barrier to heat fluxes between the ocean and atmosphere components. FORTE2 runs without flux adjustments (Blaker et al., 2021). Blaker et al. (2021) have thoroughly evaluated the skill of FORTE2 in simulating the atmosphere, ocean and major climatic modes. Hence, given the advantages of FORTE2 in terms of running speed, flexibility and economy, it is a useful tool to study a wide range of climate questions. For the SyRAP-FORTE2 simulations, FORTE2 was updated to include a parameterization of aerosol-cloud interactions, where cloud droplet effective radius in low- and mid-level clouds is reduced from 15μm to 10μm in regions where optical depth is greater than 0.07. However, this parameterization is only applied for scattering aerosol, and is not used in the BC experiments described in the following section. The semi-direct effect of BC is included.

2.2 The SyRAP-FORTE2 simulations

Stjern et al. (2024) performed a set of SyRAP experiments using the FORTE2 model. The main simulations include: (1) baseline simulations forced by GHG concentrations at different climate states (i.e. preindustrial, present-day and future CO_2 levels) and no aerosol; (2) perturbation simulations forced by added absorbing (BC, and organic carbon, OC) or scattering (sulfate, SO_4) aerosols over East Asia only, South Asia only, and over both regions simultaneously with only aerosol-radiation interactions (ARI), and GHG concentrations at different climate states; (3) Aerosol-cloud interactions (ACI) simulations forced by added SO_4 in the combined East Asia and South Asia region in which the ACI were turned on. This study focusses on the effects of adding regional BC perturbations, but the SyRAP-FORTE2 design allows the impacts of different aerosol species to be compared in a consistent framework.

FORTE2 did not use aerosol gas emissions/concentrations as most CMIP6 models use. Instead, the global gridded monthly aerosol optical depths (AOD) and vertical distributions were used from the Copernicus Atmosphere Monitoring Service reanalysis (CAMSRA) during 2003-2021 (Inness et al., 2019). Aerosols are not transported in FORTE2. The application of aerosol distributions from a reanalysis means that the simulations include a realistic aerosol spatial distribution, but the lack of aerosol transport means that there are no feedbacks between the climate response and the aerosol distribution (e.g. increased precipitation leading to increased aerosol removal). All simulations were run for 200 years, with years 51-200 used for analysis.

In this study the baseline simulation (piC) and the BC perturbation simulations of three different regions at the pre-industrial climate conditions (280 ppmv) were performed to explore potential physical processes of Asian BC aerosols influencing the local climate (Table 1). There is no significant difference in Asian climate responses to BC aerosols at different background climate states in the SyRAP-FORTE2 simulations (Stjern et al., 2024). The regional annual mean BC AOD perturbation is about 0.015 for East China, and about 0.01 for India, respectively (Figure 1a in Stjern et al., 2024). Note that only climate impacts due to ARI were considered here.

2.3 Analysis methods and Datasets

The response to a particular regional forcing is estimated by the mean difference between the perturbation simulation and the baseline simulation. Statistical significance of the response is assessed using a two-tailed Student's t-test.

The atmospheric energy budget is applied to understand the precipitation responses (Muller and O'Gorman, 2011; Richardson et al., 2016; Liu et al., 2018). The energy associated with precipitation can be separated into a thermodynamic component with only changes in the diabatic cooling (Q), and a dynamic component with only changes in the dry static energy (DSE) flux divergence (H), as shown in Eq. (1):

$$L_c\delta P = \delta Q + \delta H \quad (1)$$

where L_c is the latent heat of condensation, P is precipitation, δ denotes a perturbation. Then,

$$\delta Q = \delta LWC - \delta SWA - \delta SH \quad (2)$$

where LWC is atmospheric longwave cooling, SWA is atmospheric shortwave

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absorption, and SH is sensible heat flux from the surface.

δH is calculated as a residual between $L_c\delta P$ and δQ. Furthermore, H can be seen as the sum of the changes in mean (H_m) and eddy (H_{trans}) components. δH_m can be decomposed into four components associated with dynamic and thermodynamic effects on vertical and horizontal advection of DSE, as shown in Eq. (3):

$$\delta H_m = \delta H_{Dyn\_v} + \delta H_{Thermo\_v} + \delta H_{Dyn\_h} + \delta H_{Thermo\_h}$$

$$= \int \delta\overline{\omega}\frac{\partial\overline{s}}{\partial p}\frac{dp}{g} + \int \overline{\omega}\delta(\frac{\partial\overline{s}}{\partial p})\frac{dp}{g} + \int \delta\overline{u} \bullet \nabla\overline{s}\frac{dp}{g} + \int \overline{u} \bullet \delta(\nabla\overline{s})\frac{dp}{g} \quad (3)$$

where ω is vertical velocity, *s* is DSE, *p* is pressure, *g* is the gravitational acceleration, *u* is horizontal wind vector, ∇ is the horizontal gradient, and an overbar indicates climatological monthly means. Therefore, H_{Dyn_v} is related to changes in vertical velocity, H_{Thermo_v} is related to changes in vertical DSE gradients, H_{Dyn_h} is related to changes in horizontal winds, and H_{Thermo_h} is related to changes in horizontal DSE gradients. δH_{trans} is calculated as a residual between δH and δH_m.

To evaluate the ability of FORTE2 to represent the observed climate monthly precipitation, surface temperature (Ts), and horizontal wind components were used from the NOAA-CIRES-DOE 20th Century Reanalysis V3 (20CR) on a 2° × 2° grid spanning 1836 to 2015 (Compo et al., 2011). The 20CR is only based on surface observations of synoptic pressure of NOAA's physical Sciences Laboratory. Monthly sea level pressure (SLP) data were from the Hadley Centre (HadSLP) in a horizontal resolution of 5° × 5° (Allan & Ansell, 2006).

To compare the response of FORTE2 to BC perturbations to the responses from CMIP class models we used the PDRMIP 10 times the modern Asian BC concentrations/emissions perturbation simulations and the baseline simulations with modern aerosol concentrations/emissions and greenhouse gases with the year 2000, based on the 5 GCMs (CESM1-CAM5, GISS-E2-R, HadGEM3, MIROC, and NorESM1; Table S1). More details of the PDRMIP design can be found in Myhre et al. (2017), while an overview of the monsoon response is given in Liu et al. (2018).

3. Results

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3.1 Evaluation of baseline climate in SyRAP-FORTE2

Blaker et al. (2021) present a detailed overview of the FORTE2 climatology. The model simulates the Asian climate well, and the more focused evaluation presented by Stjern et al. (2024) demonstrated that FORTE2 is an appropriate tool to study the Asian climate effects of local aerosol perturbations. Here, for completeness, we present an overview of the skill of FORTE2 in simulating the Asian climate. The seasonal evolutions of precipitation and surface temperature (Ts) in East China and India, and the climatology of lower tropospheric circulation (SLP and 850 hPa horizontal wind) in the baseline experiment (piC) are compared with those in the reanalyses from 1851 to 1896. FORTE2 reproduces the seasonality of precipitation in East China reasonably well, but slightly overestimates the averaged magnitude with about 1 mm/day in summer (Fig. 1a). However, the model underestimates the magnitude of South Asian Summer Monsoon precipitation by approximately 4 mm/day during June-September relative to the 20CR reanalysis. Dry biases in Indian summer precipitation of similar magnitudes are found in many CMIP5 and CMIP6 models (Sperber et al., 2013; Wilcox et al., 2020; Liu et al., 2024). Differences between reanalyses and observational datasets can also have similar magnitudes (Wilcox et al., 2020). Hence, we conclude that the FORTE2 representation of monsoon precipitation is suitable for our study. The model performs fairly well in the seasonality of temperature and the magnitudes in both regions (Fig. 1b).

The simulated SLPs are generally lower than that in HadSLP in all four seasons, especially the Siberian High in winter, the western North Pacific Subtropical High (WPSH) and the Indian Low in summer (Fig. 1c-j). Compared to the reanalyses, the simulated Indian Low is too strong, and its eastern fringe and the westerly from the Indian Ocean extend too far east into the western North Pacific, so does the Indian summer monsoon trough (Fig. 1g and h). This corresponds the model dry bias over the Indian subcontinent (Fig. 1a and S1e-f). Meanwhile, the WPSH is too weak to expand sufficiently far west, corresponding to the very weak easterly along the southern fringe of the WPSH, which lead to relatively less rainfall over the Philippines (Fig. S1e-f) and underestimate of the effects of the WPSH on the East Asian summer monsoon (Fig. 1g-

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h). Despite these deficiencies, the model captures the essential features of lower tropospheric circulation, precipitation and temperature over Asia, which is consistent with Stjern et al. (2024).

3.2 Temperature and Precipitation responses

Figure 2 shows spatial patterns of Ts responses to increased Asian BC aerosols in four seasons. Firstly, there is a substantial land cooling over the perturbed regions in all four seasons, but with distinct seasonal differences in distributions and values. Under BC_CHI, a cooling can be seen in most of China in winter and spring (Fig. 2a-b), with the area-averaged values of -0.9±1.2 K (mean value ± 1 standard deviation) and -1.1±0.9 K, respectively (Fig. 3a). The large standard deviations indicate the large spread of distributions of responses (Fig. S2). There is a slight warming of sea surface temperatures near China in spring. In summer and autumn, a cooling is seen mainly in the region to the north of the Yangtze River valley, especially in North China, while there is a weak warming to the south (Fig. 2c-d). The area-averaged values therefore are relatively smaller than those in winter and spring, showing about -0.7±0.7 K for summer and autumn (Fig. 3a).

Under BC_IND, the strongest cooling occurs in the whole India in spring with the area-averaged value of -1.4±1.4 K, but a weak warming in the tropical Indian Ocean (Fig. 2f and 3b). The significant cooling is concentrated in northern India in the other three seasons, with the area-averaged Ts decreased by 0.9±1.3 K for winter and autumn, 0.6±1.1 K for summer (Fig. 2e, g, h and 3b). The ocean surrounding the Indian subcontinent shows a smaller cooling in the three seasons.

Secondly, the BC perturbations not only cause the local responses, but also non-local responses through atmospheric circulations. The increased BC over East China leads to a dipolar pattern in India with a warming to the north and a cooling to the south in summer and autumn, and a significant warming in Central Asia in summer and Southeast Asia in autumn (Fig. 2c-d). BC emissions over India can induce a cooling in Southeast Asia in all seasons, and in summer the responses are the strongest with the averaged value of about 1.15 K. Meanwhile, there is a significant cooling in central and

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southern China in summer and autumn, with decreases of 0.3±0.7 K and 0.2±0.7 K, respectively (Fig. 2g-h and 3a). These added impacts lead to an East Asian cooling response under BC_CHI+IND that is stronger than that under BC_CHI in summer and autumn (Fig. 2k-l and 3a). Moreover, by comparing the responses to both regions at once to the sum of responses to the two separate regions, it is found that the differences are almost insignificant, suggesting that the impact of Asian BC aerosol on the Asian Ts is linear regionally in all four seasons (Fig. S3).

In contrast to Ts responses, the responses in air temperature (Ta) at 850 hPa show a significant warming over much of the perturbed regions (Fig. 3c-d and 4). A cooling occurs over a few perturbed regions, such as central China under BC_CHI and BC_CHI+IND in winter (Fig. 4a, i). The vertical distribution of the temperature response is characterized by cooling at the surface and warming in the lower troposphere. This is a result of atmospheric absorption of solar shortwave radiation (SW) by BC aerosols, and is consistent with many previous studies (e.g., Li et al., 2016). Outside the perturbed regions, the Ta responses are in line with the Ts response, for example a cooling over China and Southeast Asia under BC_IND in summer and autumn (Fig. 3c and 4g, h).

Increased BC over East China primarily induces robust drying over most of China in all four seasons (Fig. 3e and 5a-d). Specifically, the spatial distribution in summer shows a substantial decrease over southern China (up to 40%), a relatively weaker decrease over north China and a weak increase over northwest China, with the area-averaged precipitation decreased by 0.5±0.5 mm/day (Fig. 3e, 5c and S4c). In spring and autumn, the decreased precipitation is mainly located in south and northeast China by up to about 40% (Fig. 5b and d). Although the absolute change is the weakest in winter (-0.2±0.3 mm/day), the relative change in central China decreases by up to about 60% (Fig. 3e, 5a and S4a). Under BC_IND, the response over the Indian subcontinent is only statistically significant in summer, with a decrease of 0.5±1.1 mm/day (~21%) (Fig. 3f and 5g).

Additionally, under BC_CHI there is increased rainfall over India in summer and autumn, which can counteract the local decrease due to Indian BC forcing (Fig. 3f and

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5c-d). Hence, when considering the simultaneous BC forcing in the two regions (BC_CHI+IND) in summer and autumn, there is no significant change in the regional mean, showing a dipolar pattern with more rainfall over northern India and less rainfall over southern India (Fig. 3f and 5k-l). On the other hand, BC_IND can induce a significant precipitation increase over China and Southeast Asia in spring and summer (Fig. 5f-g), and partly offset the decrease in response to BC_CHI over China. Hence, there are weaker responses under BC_CHI+IND for China than those under BC_CHI (Fig. 3e). For regional linearity, the responses are almost linear (Fig. S5). Only few regions exhibit nonlinear response, such as a decrease over south Thailand in winter, an increase over northeast India in spring and summer. The above analysis regarding summer part has been included in Stjern et al. (2024), but is included here as a precursor to the detailed mechanistic analysis that follows.

3.3 Energy balance response

The following analysis mainly focuses on summer for clarity; the results in the other three seasons are in the supplement (Figures S7-S10, S12-S16). The spatial patterns of net TOA and surface energy responses to regional BC aerosols in summer are illustrated in figure 6. As expected, there are increases in net downward TOA shortwave (SW) with area mean responses of 7.3~9.6 W/m^2 associated with decreases in convective clouds over the perturbed regions (Fig. 6a-c and S6j-l). Decreases in net surface downward SW can be seen with area mean responses of -22.6~-27.8 W/m^2 due to SW absorption by BC aerosols (Fig. 6d-f). Significant increases in low and middle clouds also contribute to the reduction in surface SW (Fig. S6d-i). Hence, there is warming in the troposphere and cooling at surface in these perturbed regions. The enhanced net surface upward longwave (LW) has a small contribution to the local surface cooling with area mean responses from -3.2 to -5.8 W/m^2 (Fig. 6g-i), which is related to the decreases in convective clouds (Fig. S6j-l). On the contrary, the positive changes in downward sensible heat (SH) (8.1~9.8 W/m^2) and latent heat (LH) (9.5~12.9 W/m^2) cause a warming effect partly offsetting the cooling in the perturbation regions (Fig. 6j-o). The decreased SH is caused by the vertical temperature differences between the surface and lower atmosphere, and the decreased LH associated with heating aloft and precipitation. Outside the perturbation regions, the negative changes in SW, LW and LH are responsible for the cooling in Southeast Asia and south China induced by the increased BC in India (Fig. 6e, h, n). Similar results can be seen in other three seasons (Table S2).

Figure 7 shows area-averaged atmospheric column energy budget terms (see Eq. 1) over East China and India in summer. For the drying in East China/India under BC_CHI/BC_IND, the substantial reductions of the diabatic cooling (δQ) are the prime driver of the decreases in the energy of precipitation ($L_c\delta P$), while the increases in the DSE flux divergence (δH) offset the effect of δQ to a large extent.

The enhanced precipitation in China under BC_IND, and in India under BC_CHI, is the result of increased δH (blue and green bars in Fig. 7 are almost identical), with negligible contribution from δQ. Hence, the local precipitation response to local BC increases is largely driven by δQ, while the remote precipitation response is largely driven by δH. Consequently, it can be found that the larger changes in δH lead to the smaller responses in $L_c\delta P$ over China and India under BC_CHI+IND, relative to the changes under BC_CHI and BC_IND. It should be mentioned that the relatively large uncertainties (error bars in Fig. 7) of $L_c\delta P$ mainly depend on δH. The other three seasons show the similar results (Fig. S7).

Spatially, δQ shows a significant decrease throughout the entire perturbed regions in the three simulations, with large decreases located over North China and northern India (Fig. 8d-f). Under BC_CHI, δH is characterized by a dipolar pattern with positive changes in north of the Yangtze River and negative in south (Fig. 8g). Hence, δH and δQ cancel each other out in North and Northeast China, and combine with each other in south of the Yangtze River. As a result, there is the strongest decrease in precipitation in south China, and relatively weak decrease in north China (Fig. 8a). Under BC_IND, significant positive changes in δH can be found along the southern edge of Himalayas and the southern tip of the Indian subcontinent, and weak and nonsignificant negative changes in central India (Fig. 8h). δH can therefore offset the negative changes in δQ in north and south India. The substantial $L_c\delta P$ reduction is concentrated in central India

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mainly due to δQ (Fig. 8b).

In addition, δH is the dominating factor for $L_c\delta P$ beyond the perturbation regions, for example positive changes in northeast India under BC_CHI, positive changes from Southeast Asia to the tropical western Pacific and central China under BC_IND (Fig. 8g-h). Under BC_CHI+IND, it can be found that the extent and magnitude of δH are larger than those in the simulations of individual regions (Fig. 8i), which indicates more balance between δH and δQ, corresponding to relatively weaker precipitation over East China and India (Fig. 7 and 8c). Relative to summer, δQ and δH has negative and positive changes in the whole perturbation regions in the other three seasons, respectively (Fig. S8-10). In winter and spring, there is a marked seesaw pattern of δH between Asia and the tropical Indian Ocean and maritime continent under BC_CHI+IND, leading to less precipitation in the latter regions (Fig. S8c, i and S9c, i). Overall, the combined effect of δH and δQ shape the spatial pattern of precipitation responses to the regional BC.

The reductions in δQ (see Eq. 2) are dominated by the strong atmospheric heating due to SW absorption by BC aerosols (−δSWA), and also contributed by the small decreases in the atmospheric longwave cooling (δLWC) (Fig. S11a-f). The sensible heat flux from the surface (−δSH) plays a role in increasing δQ, but with relatively small values (Fig. S11g-i).

3.4 Dynamic processes responsible for responses

Due to storage constraints, 3D atmospheric output from FORTE2 was archived on three pressure levels, 250 hPa, 500 hPa, and 850hPa, to capture three key aspects of the tropospheric circulation response. While this precludes a quantitative analysis of the component terms of δH_m (Eq. 3), it is sufficient to identify to main contributing term. We now examine the four terms of δH_m (see Eq. 3), including the dynamic components with changes in vertical and horizontal atmospheric circulations (δH_{Dyn_v} and δH_{Dyn_h}), and thermodynamic components with changes in vertical and horizontal DSE gradients (δH_{Thermo_v} and δH_{Thermo_h}). Figure 9 displays spatial patterns of the four components

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in summer in the three simulations. In general, the δH_{Dyn_v} highly resembles the δH in the three simulations, and the magnitudes in δH_{Dyn_v} are far greater than those in the other three terms, suggesting that dynamic effect of vertical circulation is the primary contributor to δH (Fig. 9a-c). δH_{Thermo_v} and δH_{Dyn_h} are small in all regions for all experiments. Larger anomalies are seen in δH_{Thermo_h}, where negative anomalies offset some of the influence of δH_{Dyn_v} over the Indo-China peninsula in the BC_IND and BC_CHI+IND experiments. However, these anomalies are not sufficient to influence the sign of δH, which is still primarily driven by δH_{Dyn_v} in this region. In the other seasons, δH_{Dyn_v} remains the most important factor (Fig. S12-14), although it is more strongly offset by δH_{Dyn_h} and δH_{Thermo_h} in winter. The effects of horizontal circulation are relatively weak in spring and autumn.

Based on the above analysis, we conclude that vertical movement is the most important contributor to δH_m. As expected, the spatial patterns of responses in Omega (vertical velocity) at 500 hPa correspond well to those in δH_{Dyn_v} (Fig. 10a-c). Anomalous ascent corresponds to the increase in δH_{Dyn_v}, leading to more precipitation, which offsets the precipitation reduction driven by decreased δQ. Anomalous descent suppresses precipitation, adding to the precipitation reduction driven by the reduction in δQ.

Why does the vertical velocity exhibit such changes? It seems to be related to the temperature responses in the troposphere, reflected by a good corresponding relationship between the Omega and Ta responses at 500 hPa (Fig. 10a-f). The warm anomalies favor a divergence in the middle troposphere, which in turn are associated with anomalous ascent. The cold anomalies are associated with a convergence and descending motion. The above-mentioned relationship also exists in the other seasons, and it is more pronounced at 850 hPa (Fig. 4 and S15).

From the transects of zonal mean diabatic heating over the perturbation regions, changes to tropospheric heating can be seen more clearly (Fig. 11). Under BC_CHI, the

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responses in diabatic heating show a meridional dipolar structure through the whole troposphere over East China, with a cooling over the region to south of 32°N (the Yangtze River basin), and a warming to north (Fig. 11a). The dipolar pattern corresponds well to the meridional distributions of precipitation, vertical velocity and Ta at 500 hPa. The cooling center located at the middle troposphere is due to the reduced latent heat release caused by the substantial decrease in precipitation over south China. The heating center at the lower troposphere in the north mainly results from SW absorption by increased BC aerosol. The difference between south China and north China is associated with the larger AOD perturbation imposed north of the Yangtze basin in SyRAP-FORTE2 (Stjern et al., 2024). In BC_CHI+IND, there is a similar dipolar pattern, except for a warming at the lower troposphere around south of 30°N (Fig. 11b). The warming is related to the increased precipitation over south China because of the BC aerosols over India.

For India in BC_IND, there is also a cooling center associated with the reduced precipitation in the middle troposphere (Fig. 11c), corresponding to a cold anomaly and descending motion at 500 hPa (Fig. 10b, e). In the lower troposphere, a warming can be seen at south of the Qinghai-Tibet Plateau. Compared with BC_IND, the cold anomaly is weaker, but the warm anomalies are strengthened under BC_CHI+IND (Fig. 11d). Hence, a significant ascending motion can be found in northeast India resulted from the effect of increased BC over East Asia (Fig. 10c), which is consistent with that in Herbert et al. (2022). In the other three seasons, however, unlike in summer, there is no cooling center in the middle troposphere, and the heating centers are situated at the lower troposphere (Fig. S16). Overall, the diabatic heating induced by the increased BC aerosols at the lower troposphere leads to an ascending motion explaining the increased δH over the perturbation regions.

Considering that the dynamic and thermodynamic effects of horizontal atmospheric circulations have some contributions to δH, we look at the changes in lower tropospheric horizontal circulation in response to changes of regional BC aerosols (Fig. 10g-i). Under BC_CHI, the cyclone anomaly over East China leads to anomalous easterly wind over North China with cutting off the moisture supply from

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south (Fig. 10g). Under BC_IND, the westerly anomalies associated with the cyclonic circulation over India favor to strength the Indian summer monsoon, which corresponds to the increase in δH_{Dyn_h} (Fig. 10h). The responses in the circulations to both regions at once can be seen as the sum of responses to the two separate regions (Fig. 10i). Additionally, there is a cyclonic circulation over East China and an anticyclonic circulation over central China in winter under BC_CHI (Fig. S17), leading to anomalous northerly wind across central China and then suppress precipitation over there, which is in agreement with the decrease in δH_{Dyn_h} (Fig. S10g). The changes in horizontal circulations are related to the changes in Ta and omega in the lower troposphere (Fig. 4 and S15).

4 Energy budget analysis in other coupled models

To evaluate the precipitation response and the mechanisms in FORTE2, we compare the results of energy budget analysis (see Eq. 1) in the PDRMIP simulations forced by 10 times the present-day Asian BC concentrations/emissions in five CMIP-class models to those in the SyRAP-FORTE2 BC_CHI+IND experiment. Spatial patterns of summer energy budgets in the PDRMIP models are illustrated in Figure 12. There are significant decreases in δQ over most of Asia in all of the PDRMIP models, which is generally consistent with the results under BC_CHI+IND (Fig. 12f-j). Three models (CESM1-CAM5, GISS-E2-R and NorESM1) have similar distributions of δQ to BC_CHI+IND, showing a maximum center in North China and northern India. δH increases significantly in India and most of East China in these models (Fig. 12k, l, o), again roughly resembling the changes of BC_CHI+IND (Fig. 8i), while the other two models show a significant increase in north India and North China with weaker magnitudes (Fig. 12m-n). The PDRMIP multi-model mean changes in δH [figure 7 in Liu et al. (2018)] are also similar to the changes in BC_CHI+IND. In spite of these model differences in the individual terms, models are broadly consistent in their total $L_c\delta P$ responses. Precipitation generally increases over India and North China, and decrease over South China in the PDRMIP models and FORTE2 (Fig. 12a-e). There is

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no significant increase in δH and precipitation in Southeast Asia in the PDRMIP models.

Figure 13 shows the regional means of each energy budget term over East China and India in summer under the PDRMIP models. There is a weak and insignificant reduction in $L_c\delta P$ from -0.6 (HadGEM3) to -7.2 W/m^2 (NorESM1) for East China, which are comparable to the value of -3.8 W/m^2 under BC_CHI+IND (Fig. 13a). Hence, the result under BC_CHI+IND in FORTE2 is agreement with the PDRMIP models, suggesting that the increased BC perturbations over Asia lead to slight decreases in precipitation over East China in all of the models. The regional means range from -1.2 (MIROC) to 13.7 W/m^2 (NorESM1) for India (Fig. 13b). Except for MIROC, $L_c\delta P$ in the other models have stronger increases than that under BC_CHI+IND, about 0.6 W/m^2, which may be related to the larger drying bias of Indian summer precipitation in FORTE2.

The effect of δQ decreases $L_c\delta P$ with a large range from -4.7 (HadGEM3) to -29.8 W/m^2 (GISS-E2-R) for East China, while δH has an opposite effect from 2.1 (MIROC) to 24.4 W/m^2 (GISS-E2-R) (Fig. 13a). Similarly, δQ changes from -5.7 (HadGEM3) to -25.8 W/m^2 (GISS-E2-R) for India, and δH from 4.9 (MIROC) to 36.5 W/m^2 (GISS-E2-R) (Fig. 13b). The magnitudes in δQ and δH under the PDRMIP models are much smaller than those in FORTE2 in both the two regions, except for GISS-E2-R. The negative effect of δQ and the positive effect of δH also can be seen in the PDRMIP multimodel mean for the whole Asian region (Liu et al., 2018). Despite large difference in the magnitudes of their responses, which are to be expected from their large range of aerosol radiative forcing and climatological precipitation, the results of these models are overall consistent qualitatively.

5 Conclusion and Discussion

In this study, we have investigated the Asian climatic responses to adding BC aerosols to the separate regions (East China and India), and both regions at once, and examined the associated physical processes, with the SyRAP simulations based on the reduced-complexity climate model FORTE2. Our main findings are as follows.

i. BC increases over East Asia or South Asia lead to a local strong surface cooling

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and lower tropospheric air temperature warming in all four seasons, with seasonal differences in magnitude and spatial distribution. The responses in temperature are dominated by the substantial decreases in surface SW radiation due to SW absorption by BC aerosols. BC over East Asia causes significant drying in south and northeast China in spring, summer and autumn. In winter, there is a significant reduction in central China. BC over South Asia induces a substantial decrease in rainfall in India in summer. Also, South Asian BC induces significant decreases in temperature and precipitation in Southeast Asia in summer and autumn.

ii. Responses in temperature and precipitation to Asian BC forcing are mostly linear regionally in all four seasons. There are relatively smaller decreases in precipitation responses to adding BC over both regions simultaneously, compared to the local reductions in precipitation responses to BC increases over East Asia and South Asia separately. This is because BC over East Asia (BC over India) increases precipitation in northeast India, while BC over South Asia increases precipitation over southern and central China.

iii. Using an energy budget analysis, we find that reductions in the energy of local precipitation ($L_c\delta P$) over the perturbation regions result from decreases in net atmospheric diabatic cooling (δQ). The increases in the dry static energy (DSE) flux divergence (δH) play a role in offsetting the effects of δQ to a large extent. Consequently, the responses in precipitation to Asian BC can be considered as the result of interactions between thermodynamic and dynamic processes. For δQ, the reductions are mainly due to the strong atmospheric heating ($-\delta SWA$). For δH, the increases depend mainly on the positive changes in the dynamic processes associated with vertical atmospheric circulations (δH_{Dyn_v}). We find that δH_{Dyn_v} patterns correspond well to vertical velocity change patterns at the middle and lower troposphere. Anomalous ascent is primarily triggered by the warming in the middle and lower troposphere over north China in summer and in most of Asia in the other seasons. However, there is anomalous descent in southern China and central India in summer, which is a result of cool anomalies in the middle troposphere due to the reduced latent

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heat release caused by the substantial decrease in precipitation. The difference in diabatic heating at the middle and lower troposphere is related to the difference in spatial distributions of AOD in the different seasons.

It is well known that the EASM and SASM underwent weakening trends during the second half of the 20th century (Wang et al., 2001; Bollasina et al., 2011). Although the variations of ASM have been attributed to many factors including internal variability and external forcing, the strong increases in BC emissions from East and South Asia (Lund et al., 2019) could play a role in weakening the ASM over the past decades according to this study. The increased BC also could alleviate the enhanced precipitation over south China due to GHG increase since the mid-1990s (Tian et al., 2018). Since the early 2010s, anthropogenic aerosols (including BC and sulfate) have been decreasing in East China, while they have continued to rise in India; trends which are expected to continue over the coming decades (Lund et al., 2019; Samset et al., 2019). Hence, there is a new dipole pattern characterized by decreasing aerosols over East China and increasing aerosols over India. Given that responses to Asian BC forcing are linear regionally in the SyRAP-FORTE2 simulations, the impacts of the dipole pattern on the Asian climate can be roughly estimated by the sum of responses to BC over China multiplied by -1 and responses to BC over India. The result shows that there are warm anomalies in north China and cold anomalies in south China, southeast Asia and most of India, and positive precipitation anomalies over most of China (especially south China) and southeast Asia, and negative anomalies over India (Fig. S18). It is overall consistent with the result in Xiang et al. (2023), although their result involves the combined effect of BC and sulfate.

Large differences in the magnitude and spatial pattern of precipitation responses to BC can be across models. The smaller precipitation responses over India in FORTE2 relative to PDRMIP may partly be due to much larger BC perturbation in PDRMIP. Liu et al. (2024) have proposed that the responses of Asian summer rainfall to Asian aerosols are strongly modulated by regional precipitation biases. Some other factors or mechanisms may play a role in causing differences in responses to Asian BC, such as bias of atmospheric circulations, land-atmosphere interaction. Further work to

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understand the mechanisms behind model differences in the response to BC would help to reduce uncertainties and would improve the confidence in future Asian climate change projections.

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Data availability. The NOAA-CIRES-DOE 20th Century Reanalysis V3 (20CR) datasets are obtained from https://psl.noaa.gov/data/gridded/data.20thC_ReanV3.html. The HadSLP2r is provided by the UK Met Office Hadley Centre and can be downloaded from at http://www.metoffice.gov.uk/hadobs/hadslp2/. The PDRMIP data can be accessed through the World Data Center for Climate (WDCC) data server at https://doi.org/10.26050/WDCC/PDRMIP_2012-2021. Data of the SyRAP-FORTE2 experiments reported in this paper are available without restriction on reasonable request from Camilla W. Stjern at CICERO Center for International Climate Research.

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Table 1. Summary of SyRAP-FORTE2 simulations used in the study

Experiment[a]	Name	Aerosol	Region	GHG	Years
Baseline	piC	No aerosol	--	Preindustrial climate conditions (280 ppmv)	200
Perturbation	BC_CHI	Added BC[b]	East China (95-133°E, 20-53°N)		
	BC_IND		India (65-95°E, 5-35°N)		
	BC_CHI+IND		both East China and India region		

a. only ARI effect is considered
b. CAMSRA monthly climatology of BC AOD for 2003-2021

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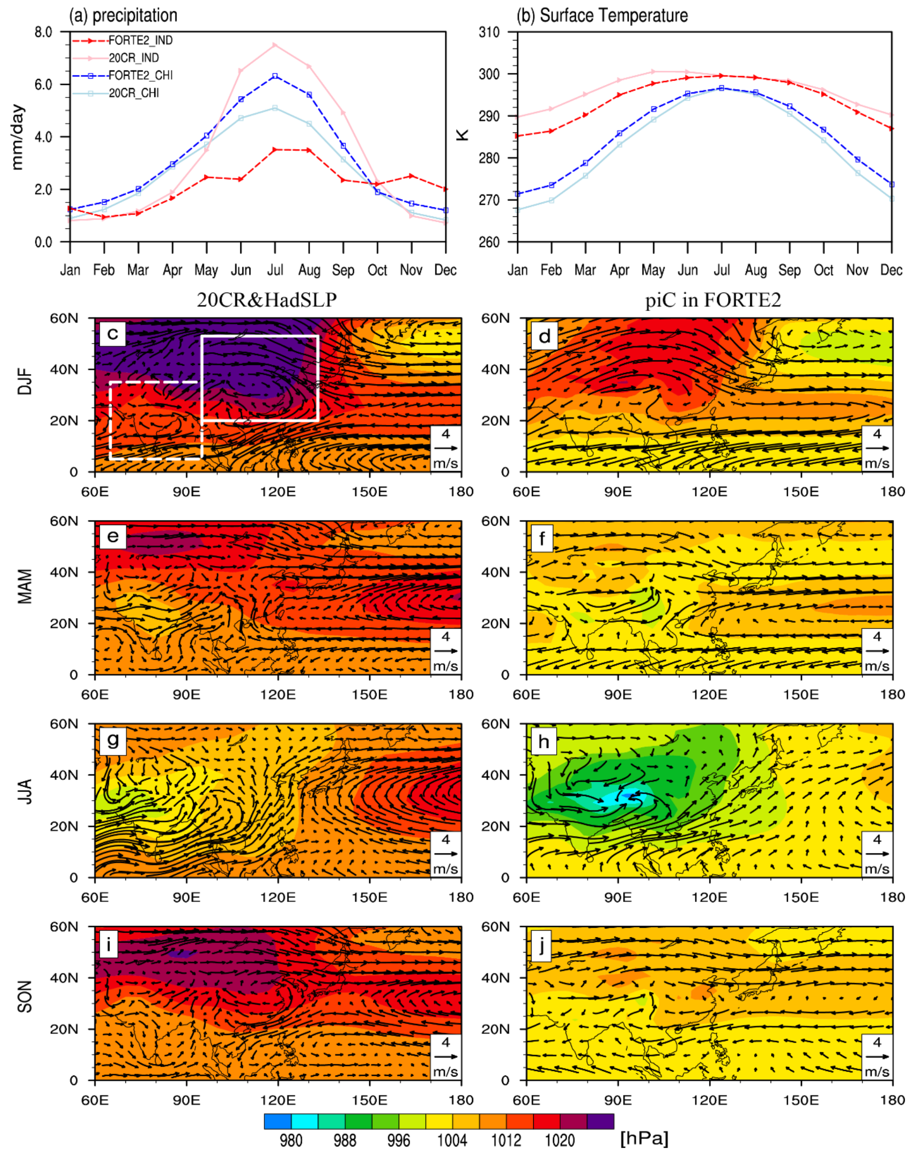

Figure 1. Seasonal evolutions of (a) the regional mean precipitation (unit: mm/day) of 20CR (solid lines) and the baseline simulation of FORTE2 (dashed lines) for East Asia (95°E-133°E, 20°N-53°N, the solid, white box in (c)) (light blue and blue lines), and India (65°E-95°E, 5°N-35°N, the dashed, white box in (c)) (pink and red lines). (b) same as (a), but for surface temperature (unit: K). Climate state of SLP (unit: hPa) and 850 hPa horizontal winds (unit: m/s) in (left) 20CR and HadSLP and (right) the baseline simulation of FORTE2 in four seasons (c-j).

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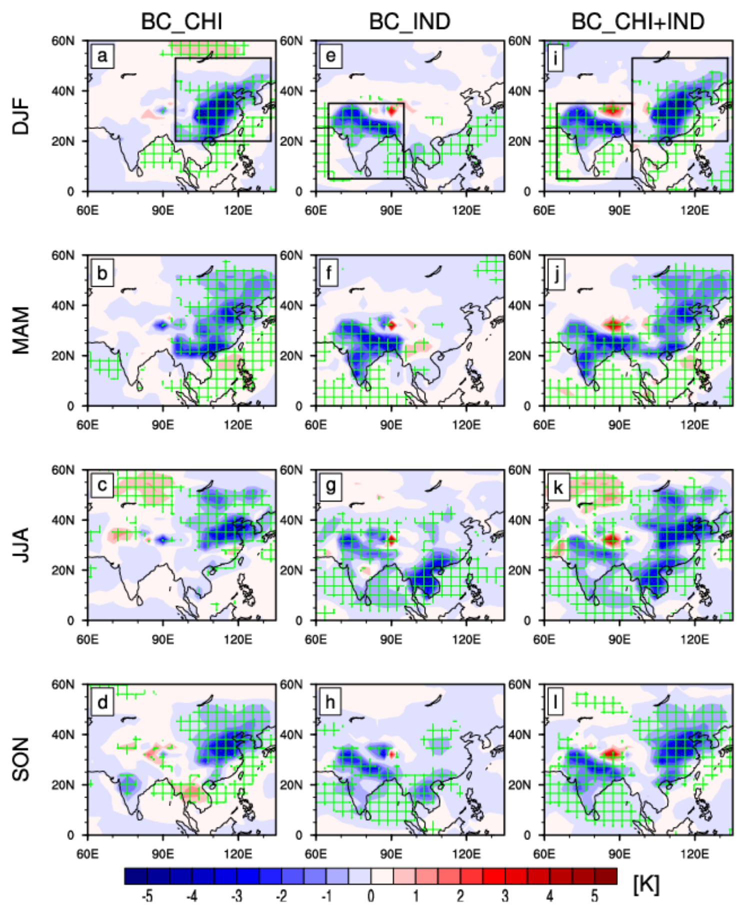

Figure 2. Spatial patterns of Ts responses in (a-d) BC_CHI, (e-h) BC_IND, and (i-l) BC_CHI+IND for four seasons. The green gridlines indicate the regions where the responses are statistically significant above 95% level based on a two-tailed Student's t-test. The black boxes in (a), (e) and (i) highlight the region where BC aerosols are perturbed. Unit: K

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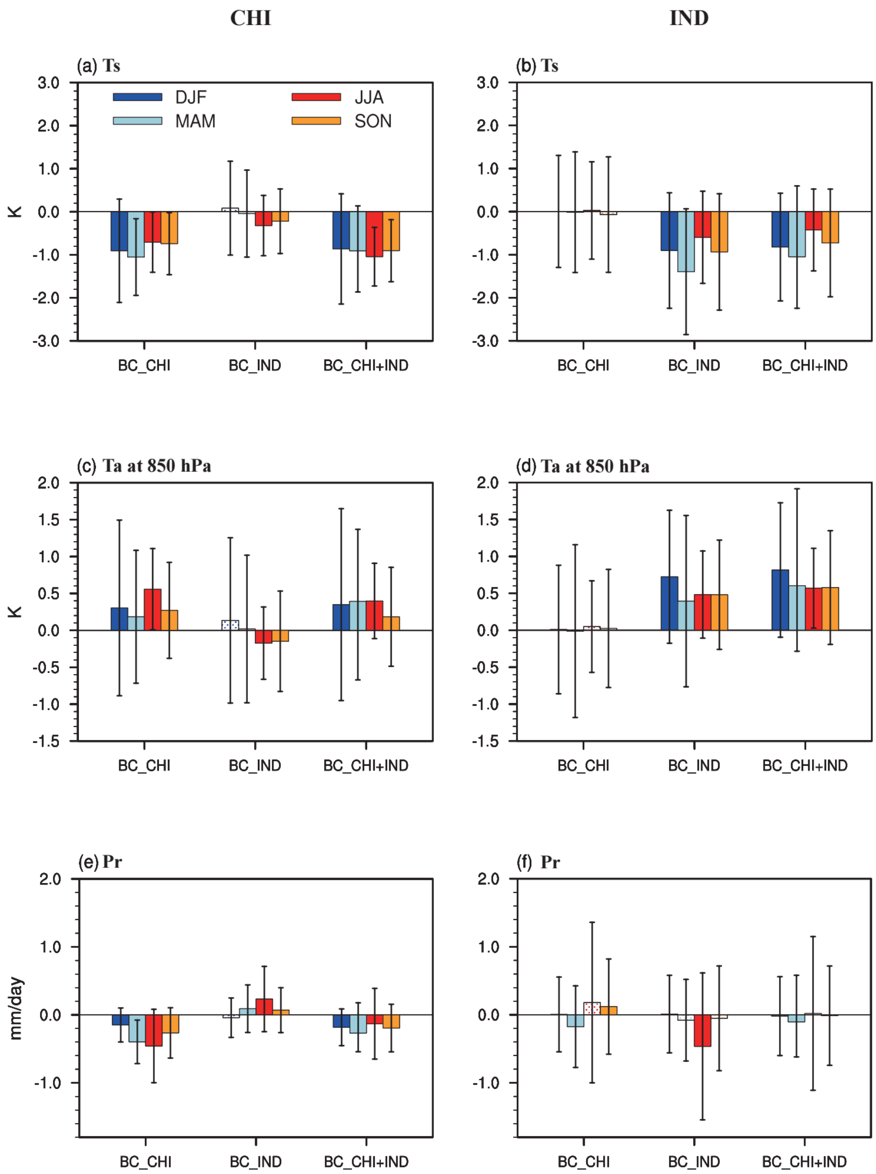

840

841 **Figure 3.** Area-averaged land responses of (a-b) Ts, (c-d) Ta at 850 hPa, and (e-f) precipitation over East China (CHI: 95°E-133°E, 20°N-53°N) and India (IND: 65°E-95°E, 5°N-35°N) for four seasons (DJF: blue bars, MAM: light blue bars, JJA: red bars, and SON: yellow bars). Solid bars indicate the responses are statistically significant above 95% level based on a two-tailed Student's t-test. Error bars represent ±1 standard deviations of the response.

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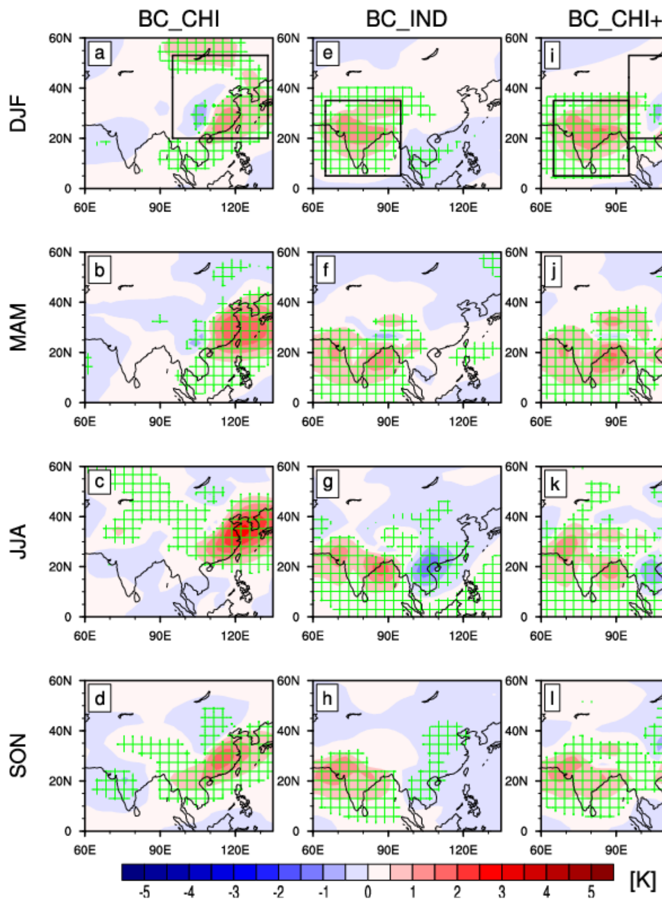

Figure 4. Spatial patterns of Ta responses at 850 hPa in (a-d) BC_CHI, (e-h) BC_IND, and (i-l) BC_CHI+IND for four seasons. The green gridlines indicate the regions where the responses are statistically significant above 95% level based on a two-tailed Student's t-test. The black boxes in (a), (e) and (i) highlight the region where BC aerosols are perturbed. Unit: K

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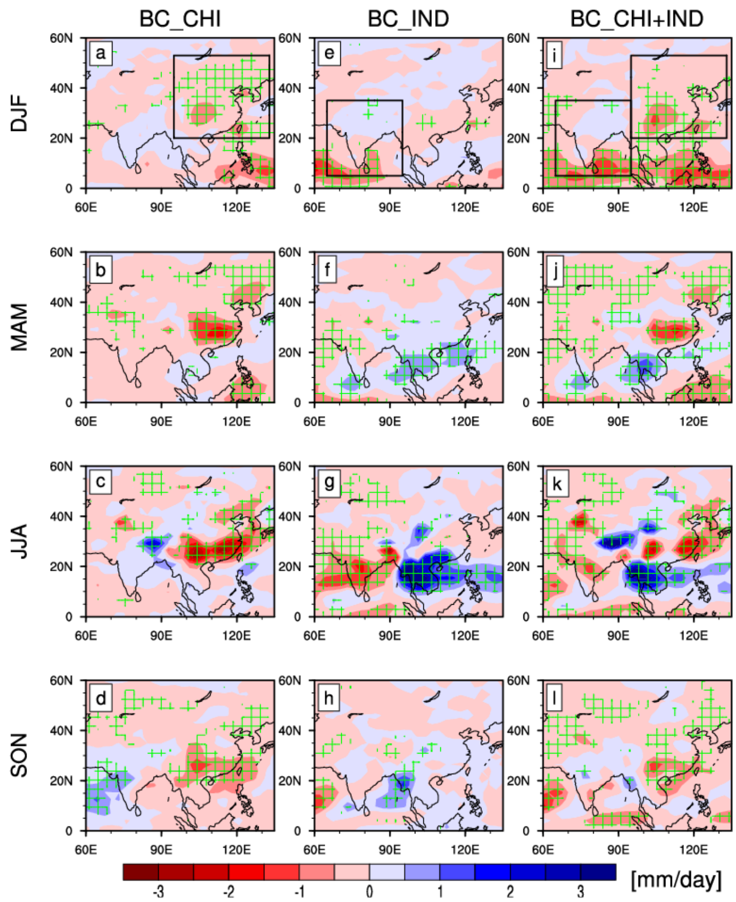

853

854 **Figure 5.** Spatial patterns of precipitation responses in (a-d) BC_CHI, (e-h) BC_IND, and (i-l) BC_CHI+IND for four seasons. The green gridlines indicate the regions where the responses are statistically significant above 95% level based on a two-tailed Student's t-test. The black boxes in (a), (e) and (i) highlight the region where BC aerosols are perturbed. Unit: mm/day

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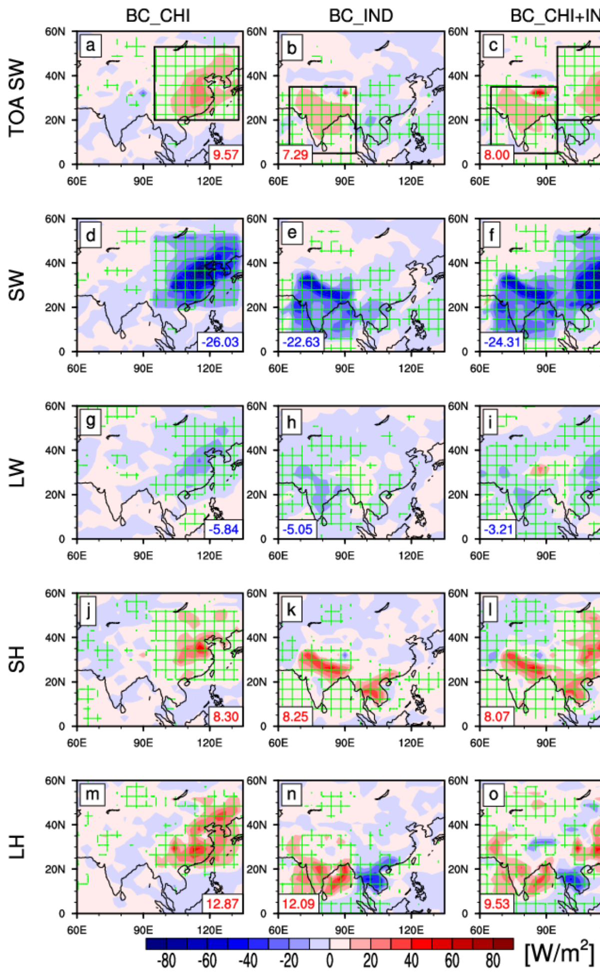

Figure 6. Spatial patterns of net TOA and surface energy responses in summer in BC_CHI, BC_IND, and BC_CHI+IND, respectively. (a-c) TOA SW, (d-f) surface SW, (g-i) surface LW, (j-l) surface SH, and (m-o) surface LH. Positive values mean downward for radiation and flux changes. Area-averaged values over East China and India are given in the lower right corners and lower left corners, respectively. The green gridlines indicate the regions where the responses are statistically significant above 95% level based on a two-tailed Student's t-test. The black squares highlight the regions where BC are perturbed. Units: W/m^2

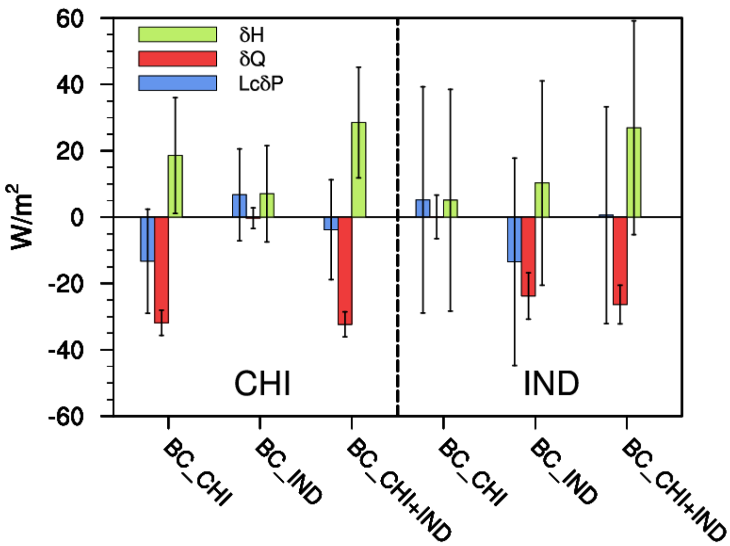

Figure 7. Summer area-averaged responses of the atmospheric energy budget terms over East China (CHI: 95°E-133°E, 20°N-53°N) and India (IND: 65°E-95°E, 5°N-35°N) in BC_CHI, BC_IND, and BC_CHI+IND. Error bars represent ±1 standard deviations of the response. Unit: W/m^2

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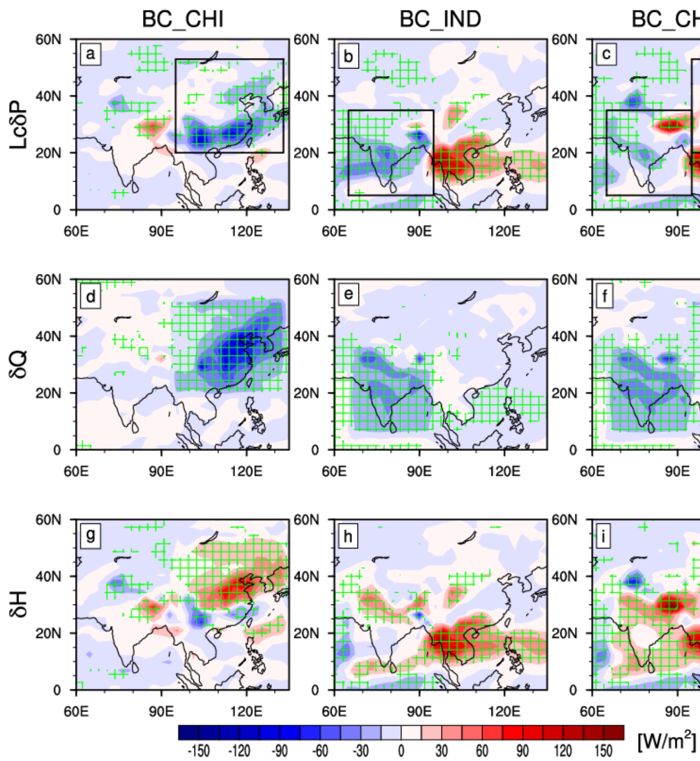

Figure 8. Summer spatial patterns of responses of the atmospheric energy budget terms in BC_CHI, BC_IND, and BC_CHI+IND. (a-c) $L_c\delta P$, (d-f) δQ and (g-i) δH. The green gridlines indicate the regions where the responses are statistically significant above 95% level based on a two-tailed Student's t-test. The black squares highlight the regions where BC are perturbed. Unit: W/m^2

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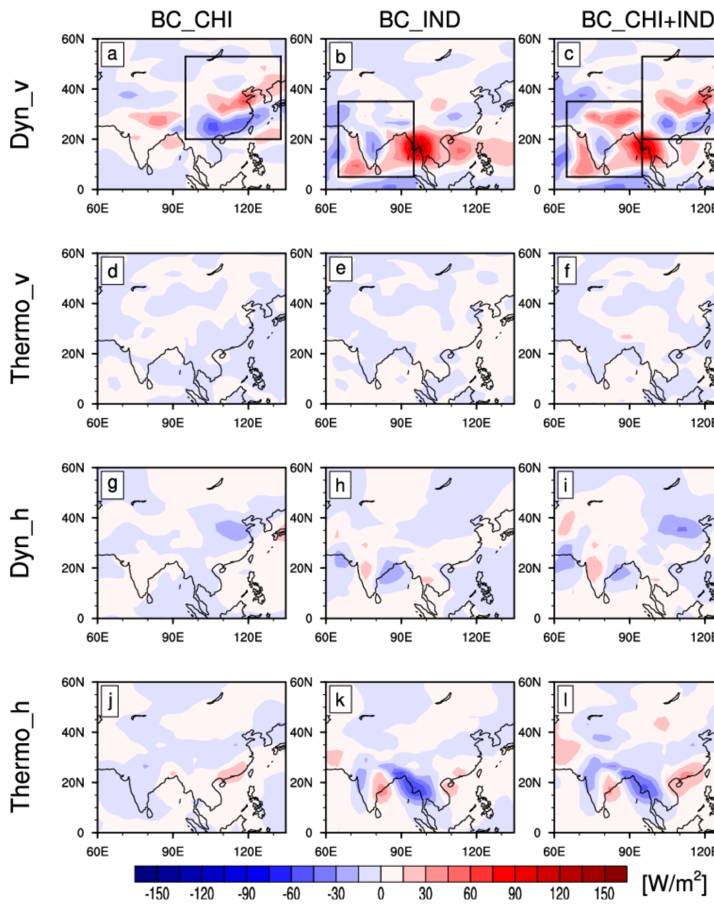

Figure 9. Summer spatial patterns of responses in the four terms decomposed by δH_m in BC_CHI, BC_IND, and BC_CHI+IND. (a-c) the dynamic components with changes in vertical atmospheric circulations (δH_{Dyn_v}), (d-f) the thermodynamic components with changes in vertical atmospheric circulations (δH_{Thermo_v}), (g-i) dynamic components with changes in horizontal DSE gradients (δH_{Dyn_h}), and (j-l) thermodynamic components with changes in horizontal DSE gradients (δH_{Thermo_h}) The black squares highlight the regions where BC are perturbed. Unit: W/m^2

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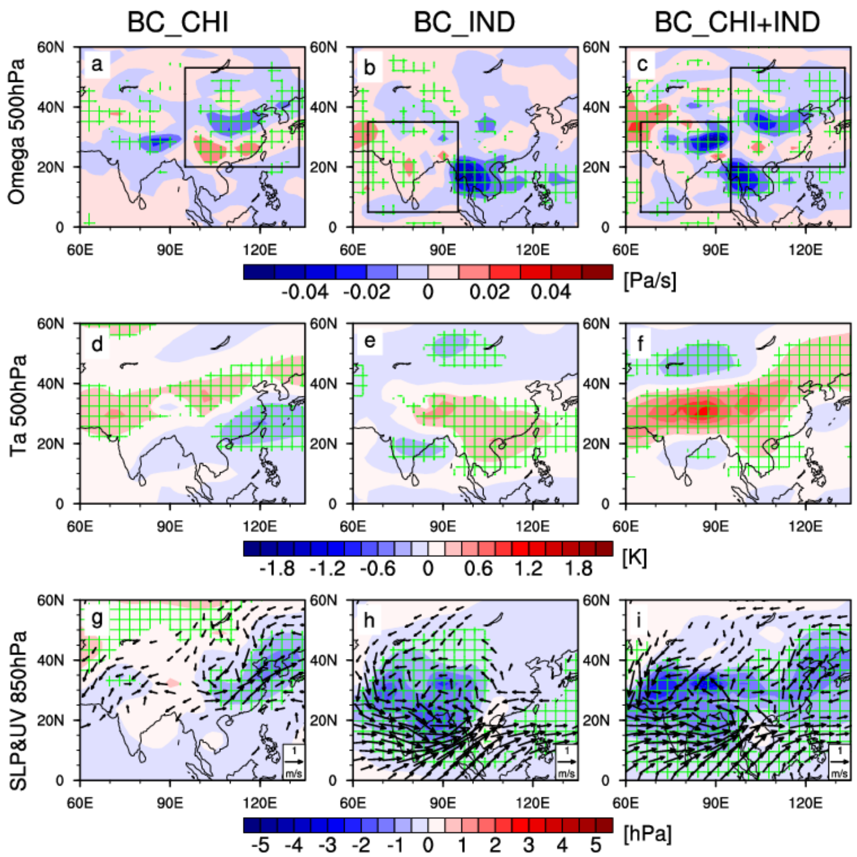

Figure 10. Summer spatial patterns of responses in (a-c) Omega at 500 hPa (Unit: Pa/s), (d-f) Ta at 500 hPa (Unit: K), and (g-i) SLP (Unit: hPa) and horizontal wind at 850 hPa (Unit: m/s) in BC_CHI, BC_IND and BC_CHI+IND. The green gridlines indicate the regions where the responses are statistically significant above 95% level based on a two-tailed Student's t-test. Wind vectors are only shown for grid boxes where at least one component of the wind significant above the 95% level are shown. The black squares highlight the regions where BC are perturbed.

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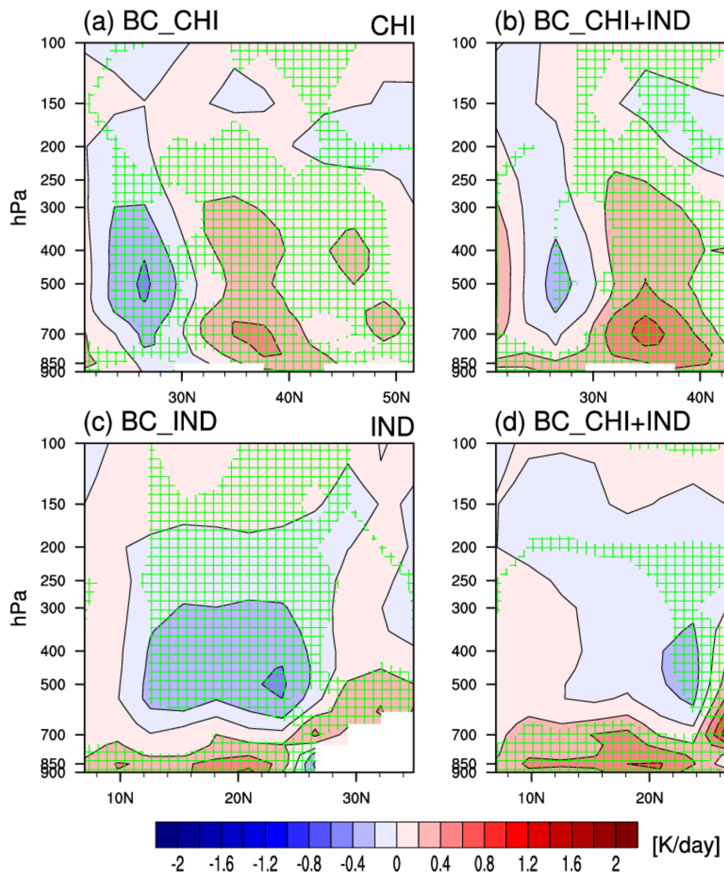

Figure 11. Zonal mean of diabatic heating responses averaged over (a-b) East China (95°E-133°E, the black square in Fig.10a) for BC_CHI and BC_CHI+IND, and over (c-d) India (65°E-95°E, the black square in Fig.10b) for BC_IND and BC_CHI+IND in summer. The green gridlines indicate the regions where the responses are statistically significant above 95% level based on a two-tailed Student's t-test. The white part indicates a symbol of topography. Unit: K/day

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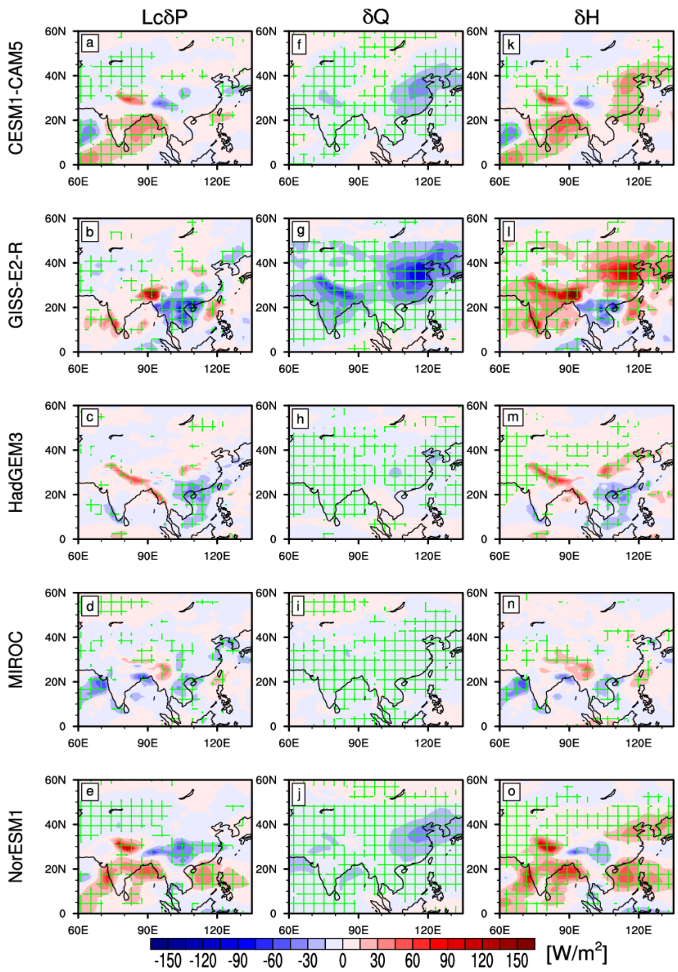

Figure 12. Summer spatial patterns of responses of the atmospheric energy budget terms in the five PDRMIP models. (a-e) $L_c\delta P$, (f-j) δQ and (k-o) δH. The green gridlines indicate the regions where the responses are statistically significant above 95% level based on a two-tailed Student's t-test. Unit: W/m^2

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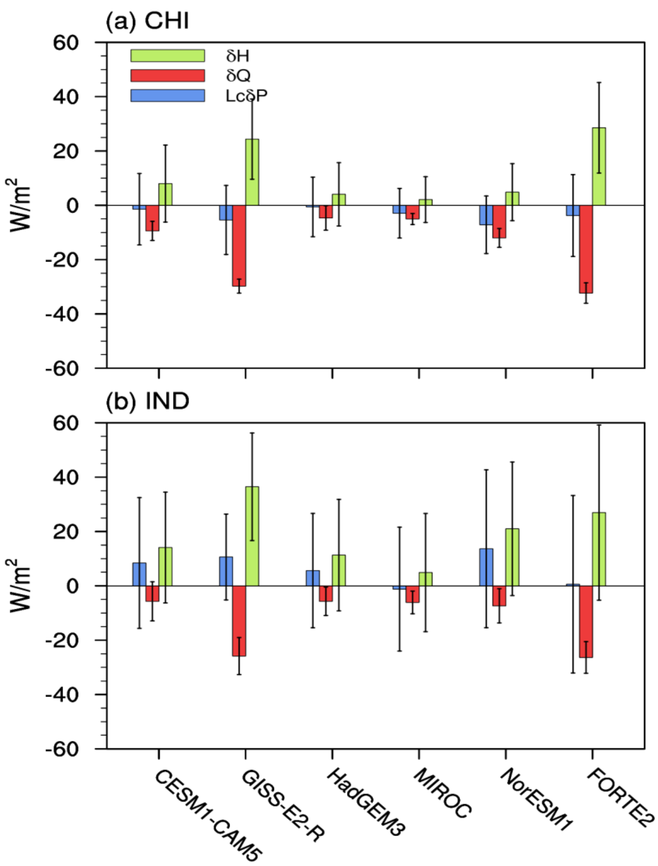

910 **Figure 13.** Summer area-averaged responses of the atmospheric energy budget terms
911 over (a) East China (CHI: 95°E-133°E, 20°N-53°N) and (b) India (IND: 65°E-95°E,
912 5°N-35°N) in the five PDRMIP models and the BC_CHI+IND simulation in FORTE2.
913 Error bars represent ±1 standard deviations of the response. Unit: W/m^2