Seasonal Responses of Indian Summer Monsoon to Dust Aerosols in the Middle East, India, and China

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ABSTRACT

The seasonal responses of the Indian summer monsoon (ISM) to dust aerosols in local (the Thar Desert) and remote (the Middle East and western China) regions are studied using the WRF Model coupled with online chemistry (WRF-Chem). Ensemble experiments are designed by perturbing model physical and chemical schemes to examine the uncertainties of model parameterizations. Model results show that the dust-induced increase in ISM total rainfall can be attributed to the remote dust in the Middle East, while the contributions from local and remote dust are very limited. Convective rainfall shows a spatially more homogeneous increase than stratiform rainfall, whose responses follow the topography. The magnitude of dust-induced increase in rainfall is comparable to that caused by anthropogenic aerosols. The Middle East dust aerosols tend to enhance the southwesterly monsoon flow, which can transport more water vapor to southern and northern India, while the anthropogenic aerosols tend to enhance the southeasterly monsoon flow, resulting in more water vapor and rainfall over northern India. Both dust and anthropogenic aerosol-induced rainfall responses can be attributed to their heating effect in the mid-to-upper troposphere, which enhances monsoon circulations. The heating effect of dust over the Iranian Plateau seems to play a bigger role than that over the Tibetan Plateau, while the heating of anthropogenic aerosols over the Tibetan Plateau is more important. Moreover, dust aerosols can decrease rainfall over the Arabian Sea through their indirect effect. This study addresses the relative roles of dust and anthropogenic aerosols in altering the ISM rainfall and provides insights into aerosol–ISM interactions.

1. Introduction

Mineral dust and anthropogenic aerosols can affect Earth’s climate by heating the atmosphere and cooling the surface because of their absorption and scattering of solar radiation (direct effect) as well as interacting with clouds by acting as cloud condensation nuclei and ice nuclei (indirect effect). Unlike greenhouse gases, which are well mixed in the atmosphere and tend to be largely evenly distributed around the globe, mineral dust and anthropogenic aerosols tend to concentrate near their source regions (Tegen et al. 2002) and have large temporal variability resulting from their short residence time (i.e., from days to weeks) in the atmosphere (Zhao et al. 2013). Therefore, the climatic effects of mineral dust and anthropogenic aerosols are more important at regional scale than at global scale (Wang 2004).

Asia is one of the world’s heaviest aerosol-laden regions because of both desert dust and anthropogenic emissions (Hsu et al. 2012). Dust storms are quite active in the Middle East (e.g., the Arabian Desert), South Asia (e.g., the Thar Desert), and western China (e.g., the Taklimakan Desert and Gobi Desert) during the boreal spring and summer. On the other side, anthropogenic emissions in South Asia contribute significantly to the total aerosol loading and have increased dramatically since the 1950s as a result of a rapid increase in population and energy consumption in this region (Ohara et al. 2007; Burney and Ramanathan 2014). Increasing evidence from observational and modeling studies shows that both mineral dust and anthropogenic aerosols in Asia can affect the Indian summer monsoon (ISM) on various time scales (i.e., from weekly to decadal) through the aerosol radiative effects (Ramanathan et al. 2005; Lau et al. 2006;...

There are four physical mechanisms underlying the aerosol–ISM connection. The first pathway is the “solar dimming” effect (Ramanathan et al. 2005; Meehl et al. 2008; Bollasina et al. 2011, 2013, 2014). The anthropogenic-aerosol-induced surface cooling effect in the Indian subcontinent can reduce the south–north ocean–land thermal contrast and increase the atmospheric stability, which results in the weakened ISM circulations and therefore less monsoon rainfall on decadal time scales. This effect is caused by both scattering (e.g., sulfate and sea salt) and absorbing aerosols (e.g., mineral dust and black carbon).

The distinction and similarity of atmospheric responses to scattering and absorbing aerosols were studied by Xu and Xie (2015). The second pathway, in contrast, is absorbing anthropogenic aerosols, especially black carbon, which have been shown to enhance the ISM rainfall by increasing the subcloud moist static energy or the convective available potential energy on the decadal time scale (Chung and Zhang 2004; Wang et al. 2009). The third pathway is the “elevated heat pump” (EHP) effect (Lau et al. 2006).

The dust aerosols from the Taklimakan Desert and black carbon aerosols from India can stack up around the Tibetan Plateau (TP) and heat the mid-to-upper troposphere, which can result in an anomalous low pressure system over the TP and strengthen the ISM on interannual time scales. It is worth pointing out that the EHP effect is still subject to debate, mainly because of the lack of observational support (Kuhlmann and Quaas 2010; Lau and Kim 2006, 2011; Nigam and Bollasina 2010, 2011; Wonsick et al. 2014). The fourth pathway is associated with the Middle East dust-induced atmospheric heating over the Iranian Plateau (IP) and the Arabian Sea (AS) (Vinoj et al. 2014; Jin et al. 2014, 2015; Solmon et al. 2015). The Middle East dust aerosols can contribute to an enhanced southwesterly monsoon flow over the AS resulting from the dust-induced heating in the troposphere, which increases the ISM rainfall on weekly time scales. Note that despite all the studies related to the fourth pathway showing a positive response of the ISM rainfall to Middle East dust aerosols, the rainfall responses display quite different or even opposite spatial distributions (Jin et al. 2015). These fast (e.g., seasonal to interannual) and slow (e.g., decadal to climate change) responses of the Indian summer monsoon to aerosols were studied and reviewed by Sanap and Pandithurai (2015).

The above studies well documented the radiative effects of mineral dust in Asia on the ISM system; however, they have not differentiated the impacts of dust aerosols emitted by local (i.e., Thar Desert) versus remote (i.e., the Arabian Desert and the Taklimakan Desert) regions on the ISM system. Given the large spatial extent of the ISM and various aerosol types and source regions in the ISM and the surrounding regions, it is important to 1) quantify the effects of mineral dust aerosols over different geographical locations on the ISM rainfall and 2) compare different roles of mineral dust and anthropogenic aerosols in modulating the ISM rainfall. Toward this goal, a regional climate model with online chemistry, which has been used in our previous study (Jin et al. 2015), is adopted by isolating individual regions of mineral dust and anthropogenic emissions in the Middle East and Asia. Comparing these simulations will help highlight the radiative, thermodynamic, and hydrological responses of the ISM system to the direct and indirect effects of various aerosols in the Middle East and Asia.

2. Model and dataset

a. Model

This study uses the Weather Research and Forecasting (WRF) Model coupled with online chemistry (WRF-Chem), which is developed collaboratively among the community (Barnard et al. 2010; Fast et al. 2006; Grell and Pincus 2007) and has been widely used in atmospheric research, including the ISM. The model is equipped with the WRF-Chem module, which includes the chemical transport and photochemical reaction processes of trace gases and aerosols together as well as interactions between aerosols and meteorological fields. The Modal Aerosol Dynamics Model for Europe (Ackermann et al. 1998) coupled with the Secondary Organic Aerosol Model (Schell et al. 2001) is employed, simulating primary aerosol types and their mixtures (e.g., sulfate, nitrate, ammonium, black carbon, mineral dust, sea salt, organic carbon, and water) and several secondary organic aerosols formed by transformation from biogenic and anthropogenic emissions. Dust emission is simulated by the Goddard Chemistry Aerosol Radiation and Transport (GOCART) scheme (Ginoux et al. 2001) through the use of surface winds and soil erodibility (Prospero et al. 2002; Zender et al. 2003), as shown in Fig. 1a.

Covering the Middle East and the ISM region with its center over the AS, the model domain has a horizontal resolution of 54 km with 160 × 120 points in the zonal and meridional directions, respectively, and 30 vertical layers up to 50 hPa. The model domain encompasses four major deserts: the Arabian Desert, the Thar Desert, the Taklimakan Desert, and the Gobi Desert, along with some other small deserts in IP, Afghanistan, and Turkistan. All simulations are run for a 104-day period from 20 May 2008 to 31 August 2008. The analyses focus on simulation results for only JJA 2008 with the first 12 days as spinup. The model meteorological boundary (i.e., both the lower
and lateral) and initial conditions are extracted from the global European Centre for Medium-Range Weather Forecasts interim reanalysis dataset (Dee et al. 2011). The anthropogenic emissions at the model lower boundary are from the combination of the reanalysis of the tropospheric chemical composition emissions inventory (http://retro-archive.iek.fz-juelich.de/data/emissions/sectorized/anthro/0.5x0.5/2000; last accessed 21 July 2016) and the emission database for global atmospheric research (http://themasites.pbl.nl/tridion/en/themasites/edgar/emission_data/index-2.html; last accessed 27 July 2016). For a detailed model configuration, please refer to Jin et al. (2015).

b. Representation of aerosol direct effect

Aerosol direct effect is calculated based on aerosol optical properties. First, the mass of each aerosol type is calculated by the corresponding aerosol emission scheme. Second, the total volumes and particle numbers for all aerosol types at various particle size modes or sections are determined according to their mass and densities. At the same time, the overall refractive indices of various aerosol types at 4 wavelengths for shortwave (SW) radiation (i.e., 0.3, 0.4, 0.6, and 0.999 \mu m) and 16 wavelengths for longwave (LW) radiation (from about 3.5 to 514.3 \mu m) are computed as the effective refractive indices of various aerosol types weighted by their volumes (Jin et al. 2016). Third, the Mie code uses the volume-averaged numbers, radii, and refractive indices of aerosols to calculate the aerosol optical properties, such as aerosol optical depth (AOD), asymmetry parameter, single scattering albedo, extinction coefficients, and backscattering coefficients for SW radiation as well as AOD and extinction coefficients for LW radiation. Last, the first three aforementioned aerosol optical properties for SW radiation and AOD for LW radiation are employed to calculate the radiative effects of aerosols. Because the Rapid Radiative Transfer Model for GCMs (RRTMG) and Goddard SW radiation schemes used in this study have 14 (from 0.23 to 8.02 \mu m) and 11 (from 0.20 to 6.14 \mu m) SW wavelengths, respectively; the aerosol SW optical properties at 4 wavelengths are converted to more wavelengths using Angstrom exponent for AOD and extrapolation for asymmetry parameter and single scattering albedo. For the detailed description of the representation of aerosol direct effects, please refer to Fast et al. (2006) and Barnard et al. (2010).

c. Representation of aerosol indirect effect

Aerosols can form cloud condensation nuclei and ice nuclei after being activated, thereby increasing the concentration but decreasing the effective size of cloud droplets, which, in turn, influences cloud albedo, liquid water content in cloud, cloud fraction, and cloud lifetime. In WRF-Chem, the concentration of cloud droplet number is parameterized in the double-moment microphysics scheme (e.g., Lin microphysics scheme) following Ghan et al. (1997):

$$\frac{\partial N_k}{\partial t} = - (\mathbf{V} \cdot \nabla) N_k + D_k + C_k + E_k + S_k,$$

where $N_k$ is the mean of cloud droplet number mixing ratio in layer $k$ in a grid cell; $\mathbf{V}$ is the horizontal wind vector; $D_k$ is vertical diffusion; $C_k$ is droplet loss resulting from collision or coalescence; $E_k$ is droplet loss due to cloud evaporation; and $S_k$ is the cloud droplet source due to nucleation. The aerosol indirect effect is taken into account by parameterizing the source term $S_k$ through aerosol activation process.
Aerosol activation, a process in which aerosol particles form cloud droplets, is the only source of cloud droplet number in Eq. (1). In WRF-Chem, aerosols are activated based on the maximum supersaturation of the air parcel, which is determined by the updraft velocity and the aerosol number concentration (Abdul-Razzak and Ghan 2002). Not all aerosol particles can form cloud droplets. Aerosol particles that are not activated to form cloud droplets will remain suspended in the interstitial air and are referred to as interstitial aerosols. When cloud dissipates, cloud droplets evaporate and cloudborne aerosols are resuspended.

The effective cloud particle size is calculated based on cloud droplet number and cloud water mixing ratio in the SW radiation schemes (first indirect effect). The aerosol-induced changes in cloud droplets mass and number are treated in the double-moment Lin and Morrison microphysics scheme (second indirect effect). It should be pointed out that these processes are only applied in the grid cell (Arpe et al. 1998; Chapman et al. 2009) so that they may not have much impact on subgrid processes, such as convection, when the model horizontal resolution (54 km) is larger than about 10 km. The total rainfall calculated by WRF-Chem consists of the stratiform and convective rainfalls. As mentioned above, only stratiform rainfall (i.e., gridcell rainfall) is affected by aerosol indirect effects.

d. Experimental design

One reference and five perturbed experiments are conducted with sea surface temperature prescribed. The reference experiment (REF) considers all aerosol types (e.g., mineral dust, anthropogenic emission, sea salt, biogenic emission, and biomass burning). The perturbed experiments are otherwise identical to REF, except for the exclusion of dust emission in the whole domain (NWH), the Middle East (NME), local regions (NLC), and western China (NWC) or the exclusion of anthropogenic emission in the whole domain (NAN). The differences between the REF and perturbed experiments are used to examine the responses of the ISM system to mineral dust aerosols emitted from various geographic locations and anthropogenic aerosols in the entire model domain as well as the nonlinearity of various aerosol effects.

To understand the influence of the model uncertainties on the results, 16 (4 × 2 × 2) ensemble simulations are carried out in each of the six experiments, following the method of Jin et al. (2015). These ensemble simulations are created using the combinations of various physical and chemical parameterizations: 1) four aerosol chemical mixing rules, 2) two aerosol turbulent mixing in the planetary boundary layer (PBL), and 3) two shortwave radiation schemes. The PBL scheme constrains the vertical diffusion of dust emissions in the boundary layer. A deeper PBL promotes a strong mixing between dust particles and the air, a longer residence time of dust aerosols, a long-distance transport, and therefore larger radiative effects on the climate. Various PBL schemes use different parameterizations of vertical diffusion coefficients, resulting in large (up to an order of one) differences in dust loadings in the air. Some radiation schemes do not consider subgrid clouds and have various numbers of bands in the visible spectrum, which can result in differences in the radiation simulations. The aerosol mixing rules actually play a very significant role in determining aerosols’ radiative effects. In our simulations only volume-average and Maxwell Garnett assumptions were used, which respectively assume an internal mixing and a random distribution of black carbon with other aerosol types. The core–shell assumption, which is not used in our simulations because of its high cost of computational time and not being fully tested in WRF-Chem, applies a different method from the above two assumptions to calculate aerosol optical properties and would show a large difference in simulations of aerosol optical properties. For details of these schemes and their differences, please refer to Table 2 of Jin et al. (2015).

e. Datasets

The AOD datasets include the Moderate Resolution Imaging Spectroradiometer (MODIS; Hsu et al. 2004) on board the National Aeronautics and Space Administration (NASA) Aqua satellite and the Multispectral Imaging SpectroRadiometer (MISR; Diner et al. 1998) on board the NASA Terra satellite. Both MODIS and MISR level-3 monthly datasets are used for model evaluation. In addition, the Monitoring Atmospheric Composition and Climate (Benedetti et al. 2009) aerosol reanalysis dataset is compared for consistency with satellite datasets.

3. Results

a. AOD in simulations

Figure 2 shows AOD spatial patterns in observations, reanalysis, and model simulations for JJA 2008. Observational AOD values from satellites and the Monitoring Atmospheric Composition and Climate reanalysis illustrate high AOD values over the AS, the Arabian Peninsula, Pakistan, and northern India, with a magnitude of about one, as shown in Figs. 2a–d. The simulated AOD in experiment REF (Fig. 2e) demonstrates similar spatial patterns to the observational AOD but with
FIG. 2. Spatial patterns of AOD (unitless) from (a) MISR (0.5° × 0.5°), (b) MODIS Aqua (1° × 1°), (c) MODIS Terra (1° × 1°), (d) the Monitoring Atmospheric Composition and Climate (0.5° × 0.5°), (e) REF, (f) NWH, (g) NME, (h) NLC, (i) NWC, (j) NAN, (k) REF − NWH, (l) REF − NME, (m) REF − NLC, (n) REF − NWC, and (o) REF − NAN averaged for JJA 2008. MISR AOD is retrieved at 558 nm, and all other AOD is shown at 550 nm. WRF-Chem AOD is the ensemble mean of 16 ensemble members in each experiment. Missing values are masked in gray color in (a)–(c).
underestimates over the AS and overestimates over the Arabian Peninsula. The potential contributors to AOD biases are inadequate representations of the aerosol humidification effect and dust particle size distribution in WRF-Chem, as discussed by Jin et al. (2015). The AOD spatial patterns in experiments NME, NLC, and NWC are as expected: AOD values are smaller in the areas where dust emissions are removed (Figs. 2f–i). Figure 2j displays AOD from dust emissions, with higher AOD values in southern AP, high values in the Thar Desert, and low values in the Taklimakan Desert. Note that AOD values in Pakistan and northern India in experiment NME are lower than in experiment REF resulting from the eastward transport of dust from the Arabian Peninsula. Figures 2k–o illustrate the AOD differences between REF and the perturbed experiments. The spatial pattern of AOD differences in each experiment can greatly influence the meteorological fields.

b. Rainfall responses to dust and anthropogenic aerosols

The rainfall responses to aerosols are analyzed in terms of total (i.e., the sum of convective and stratiform rainfall), convective, and stratiform rainfall. Figures 3a–e illustrate the ensemble means of total rainfall responses to dust and anthropogenic aerosols averaged for JJA 2008. Figure 3a shows that the ISM total rainfall increases in northern Pakistan, northern and central India, and coastal southwest India (CSWI) because of dust aerosols from the entire study domain (REF – NWH). The area-averaged rainfall response in the WHI region (red box in Fig. 1b) is 0.44 mm day\(^{-1}\), about 10% of the climatological rainfall in this region. Figure 3b demonstrates the similar spatial patterns of the total rainfall response to Middle East dust aerosols (REF – NME) to rainfall response in Fig. 3a, but with a relatively smaller magnitude and fewer significant areas. The area-averaged total rainfall response over WHI is 0.34 mm day\(^{-1}\) in REF – NME, accounting for 77% of the total rainfall response in REF – NWH. The total rainfall response to dust aerosols in local areas (i.e., the Thar Desert; REF – NLC) is not significant, as shown in Fig. 3c. On the other hand, the total rainfall response to western China dust aerosols (REF – NWC) in Fig. 3d is significant but in rather limited areas in central India. Figure 3e indicates that the rainfall increases in northern Pakistan and central and northern India, with quite weak decreases in southern India due to anthropogenic aerosols (REF – NAN). The area-averaged rainfall increase over WHI is 0.31 mm day\(^{-1}\) in REF – NAN, which is 70% of rainfall response to dust aerosols in REF – NWH. Note that the total rainfall responses to dust and anthropogenic aerosols generally follow the topography with rainfall increases in CSWI, central and northern India, and mountain regions in Pakistan.

Figures 3f–j show the convective rainfall responses calculated by model convection scheme. Figure 3f demonstrates that convective rainfall increases significantly in Pakistan and all of India except southeastern India because of dust aerosols in the entire domain, with an area-averaged value of 0.32 mm day\(^{-1}\), which accounts for 73% of the total rainfall response. The convective rainfall responses to Middle East dust (Fig. 3g) are quite similar to those in REF – NWH but with less significant areas in northern India compared to those in REF – NWH. The Middle East dust-aerosol-induced convective rainfall increase is about 0.24 mm day\(^{-1}\), contributing to about 75% to the convective rainfall increase in REF – NWH. The dust aerosols in local and remote western China do not show significant impacts on the ISM convective rainfall, as shown in Figs. 3h and 3i, respectively. Because of anthropogenic aerosols, the convective rainfall increases in northern Pakistan and northern India with a magnitude of 0.20 mm day\(^{-1}\) (Fig. 3j), which is 63% of convective rainfall response to dust aerosols in REF – NWH. Note that the convective rainfall responses are spatially much more homogeneous than the total rainfall responses, which follow the topography.

Figures 3k–o show the spatial patterns of the stratiform rainfall responses, which are similar to the total rainfall responses but with a smaller magnitude and much fewer significant areas in all experiments. The stratiform rainfall increases by 0.12 mm day\(^{-1}\), contributing to 27% of the total rainfall response in REF – NWH experiments (Fig. 3k). Middle East dust-aerosol-induced stratiform rainfall increase has a magnitude of 0.10 mm day\(^{-1}\), accounting for 83% of stratiform rainfall responses in REF – NWH (Fig. 3l). Local dust aerosols show little impact on stratiform rainfall (Fig. 3m), but remote dust aerosols in western China cause increased stratiform rainfall in some areas of central India (Fig. 3n). The anthropogenic aerosols cause stratiform rainfall increases in central and northern India, with a magnitude of 0.11 mm day\(^{-1}\) (Fig. 3o), which is 93% of rainfall response to dust aerosols in REF – NWH. One interesting phenomenon is the significant stratiform rainfall decrease over the AS in Fig. 3k, which is also seen in Fig. 3a.

The area-averaged rainfall responses over WHI region in all experiments are summarized in Fig. 4a. The convective rainfall responses contribute to about two-thirds of total rainfall responses in all experiments
Fig. 3. The ensemble mean responses of (a)–(e) total, (f)–(j) convective, and (k)–(o) stratiform rainfall (mm day$^{-1}$) in 16 pairs of WRF-Chem simulations to dust emissions in the entire model domain (REF NWH), the Middle East (REF NME), the local areas (REF NLC), western China (REF NWC), and the anthropogenic emission (REF NAN) averaged for JJA 2008. The dotted areas show a 90% confidence level based on a one-sided Student’s $t$ test.
except for REF – NLC and REF – NWC, which are reasonable because during monsoon season deep convection is quite strong and frequent and rainfall is mainly from convection. On the other hand, the stratiform rainfall responses, which are greatly affected by the large-scale moisture transport, follow the topography. Furthermore, remote Middle East dust aerosols dominate the total rainfall responses to dust aerosols in the entire domain, while local dust aerosols show little impact and remote dust aerosols in western China display limited impacts on the ISM rainfall. Although anthropogenic aerosols have smaller AOD than dust aerosols, they play a role as important as Middle East dust aerosols in modulating the ISM rainfall because of their stronger ability to absorb solar radiation. Note that the total rainfall response in REF – NWH is larger than the sum of those in REF – NME, REF – NLC, and REF – NWC as a result of the positive feedback between dust aerosols and their impact on the shamal wind, which will be discussed later.

The above analyses are based on the ensemble means of 16 members; further analysis of rainfall response in each ensemble member helps explain the model uncertainties in simulating aerosol–monsoon interactions and the robustness of the above conclusions on rainfall response. Figure 5 shows the rainfall responses in the 16 ensemble members, the ensemble means of the 16 ensemble members, and 7 ensemble means of subgroups of ensemble members selected based on seven options for various physical and chemical processes. The rainfall responses are area averaged over all of India (WHI; marked by the red box in Fig. 3a) for JJA 2008. Figure 5a displays positive rainfall responses to dust aerosols in the entire domain in all of the 16 ensemble members except member 2. More ensemble members simulate negative rainfall responses than those with negative (3 members) responses in REF – NME. Figures 5c and 5d illustrate that more than half and about half of the ensemble members have negative rainfall responses in REF – NLC and REF – NWC, respectively. Rainfall responses to anthropogenic aerosols are positive in most of ensemble members (Fig. 5e). Note that there are several members that always have positive rainfall responses in all experiments, such as numbers 1, 3, 4, 5, and 12. Overall, the variations in rainfall responses induced by model parameterization are larger (i.e., larger noise-to-signal ratio) in REF – NLC and REF – NWC than in REF – NWH, REF – NME, and REF – NAN, which is responsible for the larger areas with insignificant rainfall changes in REF – NLC and REF – NWC than in other experiments. Figures 5f,g,j show all positive rainfall responses in various ensemble means, indicating that signals from dust in the Middle East and anthropogenic aerosols in India are large enough to influence the ISM rainfall and that multimodel ensemble simulations can reduce the variations in the results because of the uncertainties in model parameterization. However, Figs. 5h,i display some negative ensemble means of rainfall responses as well as very low ensemble means of the 16 ensemble members, implying signals from dust in local areas (i.e., the Thar Desert) and western China are too low to exert an impact on the ISM rainfall. Based on Fig. 5, we
conclude that our results are, to some extent, not sensitive to these physical and chemical schemes selected in this study, but various schemes do influence the magnitude of rainfall response. Therefore, an ensemble method using multiple parameterizations is a relatively robust way to study aerosol–monsoon interactions.

c. The physical mechanisms

The proposed hypothesis of the physical mechanisms responsible for the aerosol–ISM rainfall is based on the aerosol-induced heating in the atmosphere and the consequent changes in thermodynamics, dynamics, and water vapor transport.

FIG. 5. Rainfall responses (mm day\(^{-1}\)) in five experiments for (a)–(e) 16 ensemble members and their ensemble mean (EM) and (f)–(j) 7 ensemble means of 3 subgroups of ensemble members based on seven various combinations of options of physical and chemical processes and their overall ensemble mean [as in the ensemble mean in the (a)–(e)] for JJA 2008. The rainfall responses are area averaged over all of India (marked by the red box in Fig. 3a). The Goddard and the RRTMG shortwave radiation schemes are indicated by sw2 and sw4, respectively. The Yonsei University and Bougeault-Lacarrère PBL schemes are indicated by pb1 and pb8, respectively. The groups indicated by op1, op2, op3, and op4 represent four aerosol chemical mixing rules.
1) AEROSOL RADIATIVE EFFECTS

The aerosol-induced changes in the atmosphere through aerosol direct effect are determined by both the cooling effect on Earth surface and the warming effect in the atmosphere. Figure 6 demonstrates the ensemble means of net radiative effect (i.e., the sum of SW and LW radiation) at the surface for all-sky conditions due to dust and anthropogenic aerosols, averaged for JJA 2008. Downward is defined as positive, which means the surface absorbs energy and becomes warmer. Figure 6a shows a negative radiative effect in the entire domain except for some sparse and small areas with positive radiative effect. The strongest negative radiative forcing is located in AS, the Persian Gulf, and the Caspian Sea, followed by the Red Sea, the Arabian Peninsula, and western India. The area-averaged radiative forcing is about $-7.05 \text{ W m}^{-2}$ as shown in Fig. 4b. The radiative forcing induced by Middle East dust aerosols shows similar spatial patterns to those in REF – NWH but with a smaller magnitude and fewer significant areas as shown in Fig. 6b. The area-averaged value is $-3.48 \text{ W m}^{-2}$ in REF – NME, accounting for about half of the radiative forcing in REF – NWH. Note that the surface negative radiative forcing is stronger over ocean than over land in Figs. 6a,b, even though dust aerosols have higher AOD over land than over ocean, which is due to much lower surface albedo over the ocean ($\approx 0.08$) than the deserts ($\approx 0.4$). Therefore, the spatial distribution of aerosol-induced surface cooling effect is largely influenced by the land–ocean distribution. Figures 6c and 6d show negative radiative forcing in Pakistan and western China as a result of the dust aerosols in local areas and western China with area-averaged values of $-0.76$ and $-0.41 \text{ W m}^{-2}$, respectively. The anthropogenic-aerosol-induced negative radiative forcing (Fig. 6e) is strongest in the source regions (e.g., northern India), with an area-averaged value of $-4.34 \text{ W m}^{-2}$, which is stronger than the radiative forcing due to dust aerosols in the Middle East, local regions, and western China but much smaller than that due to dust aerosols in the entire domain.

Figures 6f–j show the net radiative forcing in the atmosphere. Here positive values mean absorbing of solar radiation by the atmosphere. The spatial patterns in Figs. 6f–j are quite similar to those in Figs. 6a–e. However, the spatial patterns of radiative forcing in the atmosphere do not show a clear land–ocean difference (Figs. 6f,g) as from surface radiative effect (Figs. 6a,b). This is because the aerosol concentrations dominate the radiative forcing in the atmosphere. The radiative forcing at the top of the atmosphere (TOA) is shown in Figs. 6k–o, showing a quite strong land–ocean difference because of surface albedo. Note that the anthropogenic aerosols cause a stronger warming effect at the top of the atmosphere in Fig. 6o than dust aerosols in Figs. 6k,l, indicating a much stronger absorbing ability of solar radiation than dust aerosols.

The area-averaged radiative forcing is summarized in Fig. 4b. First, by comparing the radiative forcing in the atmosphere and at the surface and top of the atmosphere resulting from dust aerosols in various source regions, we can see that Middle East dust aerosols play a dominant role in altering the radiative forcing induced by dust aerosols in the entire domain, contributing half of the radiative forcing. Adding the radiative forcing induced by dust aerosols in REF – NME, REF – NLC, and REF – NWC, we find that their sums account for only 67% and 59% of radiative forcing of dust aerosols in REF – NWH at the surface and in the atmosphere, respectively, which indicates a nonlinearly additive feature of dust-induced radiative forcing in various regions. This nonlinearly additive feature is also seen in the dust-induced heating profiles in Fig. 7. One of the potential reasons for this feature is the positive feedback between dust emissions and the local shamal winds proposed by Jin et al. (2014, 2015), which is to be discussed later.

2) IMPACTS ON MONSOON THERMODYNAMICS

Figure 7 shows the vertical profiles of aerosol-induced atmospheric heating rates, area averaged in an area with high aerosol loadings marked by the blue box in Fig. 6f. Figure 7a illustrates the vertical profiles of the heating rates due to LW and SW radiation effects, sensible and latent heating, and the net heating effect induced by dust aerosols emitted from the entire study domain. SW radiation and latent heat are two warming sources with maximum magnitudes of 0.35 and 0.05 K day$^{-1}$, respectively, while LW radiation and sensible heat are two cooling sources with maximum magnitudes of $-0.20$ and $-0.40$ K day$^{-1}$, respectively. All of them are at about 900 hPa. However, the net heating rate shows a different vertical profile from each individual heating source, with a maximum heating rate of 0.1 K day$^{-1}$ at 600 hPa extending up to around 300 hPa. Note that a positive feedback exists between latent heat and monsoon rainfall. The vertical heating rates induced by the Middle East dust (Fig. 7b) are similar to those due to dust aerosols in the entire domain (Fig. 7a) but with a smaller magnitude. The vertical heating rates induced by the local and western China dust aerosols are very limited (Figs. 7c and 7d, respectively). Figure 7e demonstrates that anthropogenic aerosols induce weaker low-level heating and cooling from individual sources,
Fig. 6. Spatial patterns of aerosol radiative effect (W m$^{-2}$) (a)–(e) at the surface, (f)–(j) in the atmosphere, and (k)–(o) at TOA for all-sky conditions averaged during JJA 2008 for REF – NWH, REF – NME, REF – NLC, REF – NWC, and REF – NAN. Radiative effect is calculated from the ensemble mean differences between 16 REF and perturbed experiments of WRF-Chem. Downward radiation is defined as positive at the surface and TOA; therefore, positive (negative) value means absorbing or warming (irradiating or cooling) effects. Net radiative forcing is the sum of SW and LW radiative forcing. The dotted areas show a 95% confidence level for the radiative forcing based on a one-sided Student’s t test. The box in (f) indicates the aerosol region (8°–45°N, 40°–90°E) used for further analysis.
but the net vertical heating rate has a similar magnitude to that induced by the Middle East dust aerosols.

The heating in the mid-to-upper troposphere has been shown to play a more important role than that in the lower troposphere in driving the ISM (Dai et al. 2013). Figure 8 shows the spatial patterns of temperature responses at 500 hPa averaged in two “wet” and “dry” subgroups of the ensemble members, which contain four ensemble members each and are selected based on the rainfall responses in India (Fig. 5a). Figure 8 demonstrates strong atmospheric warming at 500 hPa over IP, Iraq, Afghanistan, and Pakistan in the wet subgroup with a magnitude of about 0.3–0.5 K in all experiments, while a much weaker atmospheric warming or even cooling is in the dry subgroup. The atmospheric heating is overall weaker in REF – NLC and REF – NWC than those in REF – NWH, REF – NME, and REF – NAN in both wet and dry cases. Moreover, the dust-induced warming effect is also seen over TP, where the original EHP effect is mainly located, but it has generally much smaller spatial coverage and weaker magnitude than the warming over the Middle East. Note that these temperature responses associated with the wet and dry rainfall responses could be directly attributed to dust warming effect or latent heat release from increased monsoon rainfall, but originally the temperature responses are caused by the presence of dust aerosols.

To better quantify the relationships between the ISM rainfall and the aerosol-induced atmospheric heating effect at various vertical layers in various regions, the Spearman’s rank correlation between rainfall responses in WHI and aerosol-induced temperature responses in the troposphere in IP and TP are calculated in each experiment, as shown in Fig. 9. The response of ISM rainfall shows strong correlations with the dust-induced atmospheric heating over IP between about 800 and 400 hPa as shown in Figs. 9a–d. However, the correlations between the ISM rainfall
responses and dust-induced heating over TP are not significant (Figs. 9a,b,d) except for local (i.e., the Thar Desert) dust-aerosol-induced heating in the upper troposphere around 400 hPa (Fig. 9c). Figure 9e shows the correlation between the ISM rainfall response and anthropogenic-emission-induced heating with the maximum correlation at 400 hPa over TP. Figures 9c,e indicate that dust aerosols from the Thar Desert and anthropogenic-emission-induced heating in the mid-to-upper troposphere around 400 hPa can strengthen the ISM rainfall, supporting the original EHP hypothesis. However, for dust aerosols the EHP effect over IP seems to play a bigger role than that over TP, while for anthropogenic emission the EHP is stronger over TP. Upper-level heating also appears over the Arabian Peninsula and India (Fig. 8), but we found
that their relationship with the Indian subcontinent rainfall is weak (not shown).

3) IMPACT ON CIRCULATION AND WATER VAPOR TRANSPORT

Figure 10 shows the ensemble means of responses of geopotential height and winds, both at 850 hPa, to dust and anthropogenic aerosols averaged for JJA 2008. Figure 10a shows a low pressure system centered over the AS, the Persian Gulf, Iraq, Turkmenistan, and western India due to the dust aerosols in the whole region, which is associated with an anomalous convergence zone centered over AS and western India. The spatial patterns of responses of geopotential height and winds, as shown in Fig. 10b, are akin to that in Fig. 10a but with a smaller magnitude. The dust aerosols in local regions and western China have little impact on the Indian monsoon dynamics, as shown in Figs. 10c and 10d, respectively. The anthropogenic-aerosol-induced circulation changes are weak in southern India but strong in northern India, as shown in Fig. 10e.

The dust-induced changes in circulations can alter the water vapor transport. Figure 11 shows the ensemble means of responses of precipitable water and water vapor flux due to dust and anthropogenic aerosols averaged for JJA 2008. Figure 11a demonstrates that the precipitable water increases in all of India as a result of the dust aerosols in REF NWH, with the strongest increases in the entire Pakistan and south and northern India. The precipitable water increases are caused by a positive anomaly of water vapor transport by the enhanced southwesterly and southeasterly monsoon flows over AS and northern India, respectively. In Fig. 11b, the precipitable water increases only in some regions of southern India because of Middle East dust aerosols. The dust aerosols in local areas and western China show...
little impact on water vapor transport (Figs. 11c and 11d, respectively). The anthropogenic-aerosol-induced precipitable water increases are mainly located in Pakistan and northern India because of the enhanced water vapor transport in these regions (Fig. 11e). Note that although the dust aerosols in local areas or remote dust aerosols in western China alone do not exhibit any significant impacts, they can play significant roles when taking into account Middle East dust aerosols, which can be seen by comparing Figs. 11a and 11b. This can be attributed to the aforementioned positive feedbacks between dust aerosols and the shamil winds.

The maximum subcloud moist static energy has been proved a good indicator for the boundary of monsoon system (Privé and Plumb 2007a,b). Wang et al. (2009) and Jin et al. (2015) studied the impacts of anthropogenic and mineral dust aerosols, respectively, on the poleward shifts of the ISM rainfall using moist static energy. The ensemble means of the moist static energy responses in REF − NWH, REF − NME, and REF − NAN averaged for JJA 2008 show positive values in Pakistan, most parts of India, with a magnitude of 1–2 kJ kg$^{-1}$. As these spatial patterns are very similar to the spatial distributions of precipitable water, they are not shown in this paper.

4) AEROSOL INDIRECT EFFECTS

Besides the aerosol direct effect analyzed in the above context, the aerosol indirect effect also shows impacts on rainfall in this study. As mentioned earlier in section 3b, the rainfall decreases over the AS, which is mainly attributed to the stratiform rainfall response, as shown in Figs. 3a,k. The stratiform decrease in rainfall over the AS cannot be attributed to large-scale circulation changes because the precipitable water increases over the AS (Fig. 11a). One potential contributor is the aerosol indirect effects through changing cloud microphysics. The aerosol indirect effects on rainfall generally depend on the concentrations of activated aerosol particles and the liquid water supply in clouds. Aerosols tend to decrease cloud droplet radius when the cloud water increase is smaller than the cloud droplet number increase (Albrecht 1989; Gong and Barrie 2002).

Figure 12 shows the spatial patterns of responses of vertically integrated cloud microphysical properties: cloud water, cloud droplet number, and effective cloud droplet radius. The left and center panels in Fig. 12 demonstrate a significant increase in cloud water and cloud droplet number over the Arabian Sea, India, and
Bay of Bengal due to dust in the Middle East and anthropogenic aerosols (REF – NWH, REF – NME, and REF – NAN). Consequently, the effective cloud droplet radius displays a decrease over the Arabian Sea and Bay of Bengal in REF – NWH, REF – NME, and REF – NAN. However, the stratiform rainfall decreases only over the AS in REF – NWH, indicating that the decreased cloud droplet radius does not necessarily prevent rainfall, which is reasonable because rainfall is prevented only when very heavy aerosol loadings are present and cloud droplet radius is reduced in a great magnitude. Note that although dust AOD in REF – NME is much higher than anthropogenic aerosol AOD in REF – NAN (Figs. 2l,o) over the Bay of Bengal, the increases of cloud droplet numbers due to dust in REF – NME are much weaker than those caused by anthropogenic aerosols in REF – NAN. McFiggans et al. (2006) and Petters and Kreidenweis (2007) demonstrated that the hygroscopic property of aerosols is an important factor determining the critical supersaturation for aerosol activation, directly influencing the cloud droplet source $S_k$ in Eq. (1). Therefore, the differences in cloud droplet numbers between REF – NME and NAN could be attributed to the different hygroscopic properties between the fresh dust aerosols over AS (~0.1) and the polluted dust or anthropogenic aerosols (~0.5) over the Bay of Bengal.

4. Discussion and conclusions

In this study, we have examined the radiative (heating rate; Figs. 6 and 7), thermodynamic (tropospheric temperature; Fig. 8), dynamic (circulation and moisture transport; Figs. 10 and 11), and microphysical (cloud droplet number; Fig. 12) responses of the ISM system to direct and indirect effects of dust in the Middle East, India, and western China versus anthropogenic aerosols in the entire domain.

The Middle East dust aerosols dominate the total AOD, followed by local dust aerosols in the Thar...
Fig. 12. The ensemble mean responses of vertically integrated (left) cloud water (g m$^{-2}$), (center) cloud droplet number ($10^9$ m$^{-2}$), and (right) effective cloud droplet radius ($\mu$m) in the entire atmospheric column to dust emission in (a)–(c) REF−NWH, (d)–(f) REF−NME, (g)–(i) REF−NLC, and (j)–(l) REF−NWC as well as (m)–(o) anthropogenic aerosols in REF−NAN averaged for JJA 2008. The dotted areas show a 95% confidence level based on a one-sided Student’s $t$ test.
Desert, with the least contribution from dust aerosols in western China. The dust aerosols over the Arabian Peninsula and Iran can be transported southward by the northwesterly winds to the Arabian Sea. The local and remote dust aerosols in India and western China, respectively, are mainly constrained near the source regions with quite limited dust aerosols transported to the downwind regions. The spatial patterns of anthropogenic aerosols are determined by their source locations and rainfall due to wet scavenging and mainly located in northern India.

The dust-induced rainfall increases are located in northern Pakistan, northern and central India, and CSWL, which are mainly (77%) attributed to Middle East dust aerosols. Although dust aerosols over the Thar Desert and western China alone do not show significant impacts on the ISM rainfall, they can play important roles when taking into account Middle East dust aerosols, which are attributed to the positive feedback between dust aerosols and the northwesterly winds over the Arabian Peninsula. The anthropogenic aerosols can also enhance the ISM rainfall, with a similar magnitude to the Middle East dust aerosols. However, Middle East dust aerosols tend to enhance rainfall in both southern and northern India, while anthropogenic aerosols tend to enhance rainfall only in northern India. For the total rainfall increase due to dust aerosols, the convective rainfall response contributes 73% while the stratiform rainfall contributes the remaining 27%. Furthermore, the convective rainfall responses induced by both dust and anthropogenic aerosols illustrate spatially more homogeneous patterns than the stratiform rainfall responses, whose spatial patterns follow the topography with large rainfall responses in the mountainous regions in India and Pakistan (Tahir et al. 2015).

Both dust- and anthropogenic-aerosol-induced ISM rainfall changes can be largely attributed to their direct radiative effects. Middle East dust aerosols heat the atmosphere over AS, the Arabian Peninsula, and IP, which in turn result in a low pressure anomaly in this region. This low pressure anomaly is associated with the enhanced southwesterly monsoon flow, which transports more water vapor from AS to the Indian subcontinent, enhancing the monsoon rainfall in the entire Indian subcontinent except southeastern India. Similarly, the anthropogenic aerosols in India and China heat the atmosphere in northern India and the Tibetan Plateau and cause an enhanced southeasterly monsoon flow in northern India, which transports more water vapor from southern India and the Bay of Bengal to northern India, resulting in more monsoon rainfall in central and northern India and Pakistan. Anthropogenic aerosols have an impact comparable to Middle East dust on ISM rainfall, with a smaller loading but stronger absorption from black carbon. Further analysis shows that the dust- and anthropogenic-aerosol-induced atmospheric heating is more important in the high level (e.g., 500 hPa) than in the low level (e.g., 800 hPa) and stronger over IP than over TP. Moreover, dust aerosols over the Arabian Sea can result in a decrease in rainfall by working as cloud nuclei, consequently reducing the effective cloud droplet size.

It should be pointed out that this study focuses only on the period of JJA 2008, during which the magnitudes of both AOD and the ISM rainfall are greater than their climatology by about one standard deviation. The magnitude and pattern of the ISM rainfall response to dust in our domain may depend on both aerosol concentrations and meteorological conditions. Therefore, multiyear simulations are needed to evaluate the robustness of the impacts of various aerosols on the ISM rainfall on longer time scales. Also note that the lack of fully representing the aerosol indirect effect in the model simulations might induce uncertainties in our results. In the current WRF-Chem, the aerosol indirect effect is included only in the microphysics scheme at the grid scale but not in the cumulus parameterization at subgrid scale (Arpe et al. 1998; Chapman et al. 2009). Therefore, the simulations at 54-km grid spacing cannot resolve convective clouds, nor can they fully simulate the aerosol indirect effect. This problem may become more important particularly for monsoon applications in which convective rainfall events are active and dominate the total rainfall. Increasing model resolution helps resolve convective clouds and better represent the aerosol indirect effects, but it will dramatically increase the simulation time and render long-term simulations difficult. Therefore, we suggest that aerosol indirect effects be included in cumulus parameterization for simulations with horizontal resolutions larger than 10 km in future model development.

Another issue of this study is the opposite ISM rainfall responses to the local anthropogenic aerosols found in this study compared to some previous studies. Negative responses of the ISM rainfall to anthropogenic aerosols in India have been found in various studies (Bollasina et al. 2011, 2014; Ganguly et al. 2012; Meehl et al. 2008; Ramanathan et al. 2001, 2005), while positive responses have also been reported in other studies (Chakraborty et al. 2014; Chung and Seinfeld 2005; Lau et al. 2006; Menon et al. 2002; Wang et al. 2009). This discrepancy could be related to different time scales considered in the modeling studies and the large uncertainties in the anthropogenic emission inventories. Most of the negative responses of ISM to aerosols in the above literature were obtained by using coupled ocean–atmosphere...
general circulation models on decadal to climate change time scales, while positive responses were found by employing atmosphere-only models, focusing on seasonal to interannual time scales. On the other hand, the anthropogenic emissions in India, such as black carbon and sulfate, have been demonstrated to have very large uncertainties among various emission inventories (Granier et al. 2011; Lu et al. 2011). For example, the black carbon emission can be 2 times as much in the mitigation of air pollution and greenhouse gases dataset as that in the representative concentration pathways dataset (Granier et al. 2011). These uncertainties can be reduced by linearly tuning the anthropogenic emissions by multiplying it with a constant to match the modeled total AOD with the satellite observed AOD in the anthropogenic-emission-dominated regions, such as in northern India and the Sichuan basin. However, the ratio of the absorbing aerosols (e.g., black carbon) to scattering aerosols (e.g., sulfate), which determines the magnitude or even the sign of aerosol-induced atmospheric heating and in turn affects the thermodynamic and hydrological cycles of the ISM system (Menon et al. 2002), can have a large range for a certain total AOD. Xu et al. (2013) shows that observationally constrained inventory can improve model simulation of aerosol absorption optical depth and AOD. Therefore, adopting better inventory estimation of the anthropogenic emissions of each aerosol type in future simulations can reduce the uncertainties not only in the results of the NAM experiments but also in the results of other dust-related experiments due to the nonlinear interactions between dust and anthropogenic aerosols (Bollasina et al. 2014).

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