Nonlinear Effects of Coexisting Surface and Atmospheric Forcing of Anthropogenic Absorbing Aerosols: Impact on the South Asian Monsoon Onset

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(Manuscript received 19 October 2012, in final form 29 January 2013)

ABSTRACT

The direct radiative effect of absorbing aerosols consists of absorption-induced atmospheric heating together with scattering- and absorption-induced surface cooling. It is thus important to understand whether some of the reported climate impacts of anthropogenic absorbing aerosols are mainly due to the coexistence of these two opposite effects and to what extent the nonlinearity raised from such coexistence would become a critical factor. To answer these questions specifically regarding the South Asia summer monsoon with focus on aerosol-induced changes in monsoon onset, a set of century-long simulations using the Community Earth System Model, version 1.0.3 (CESM 1.0.3), of NCAR with fully coupled atmosphere and ocean components was conducted. Prescribed direct heating to the atmosphere and cooling to the surface were applied in the simulations over the Indian subcontinent, either alone or combined, during the aerosol-laden months of May and June. Over many places in the Indian subcontinent, the nonlinear effect dominates in the changes of subcloud layer moist static energy, precipitation, and monsoon onset. The surface cooling effect of aerosols appears to shift anomalous precipitative cooling away from the aerosol-forcing region and hence turn the negative feedback to aerosol-induced atmospheric heating into a positive feedback on the monsoon circulation through latent heat release over the Himalayan foothills. Moisture processes form the critical chain mediating local aerosol direct effects and onset changes in the monsoon system.

1. Introduction

Field observations in recent decades have revealed an abundance of anthropogenic aerosols, particularly absorbing aerosols, over South Asia and its surrounding ocean during the dry season (Ramanathan et al. 2001). Modeling studies and observation-based analyses have indicated that the radiative effects of these aerosols can interrupt not just localized meteorological features but also large-scale climate features such as monsoons.

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Using prescribed direct radiative forcing derived from the Indian Ocean Experiment (INDOEX) measurements, Chung et al. (2002) and Menon et al. (2002) modeled the influence of the pan-Asian haze on the premonsoonal and monsoonal climatology, respectively. Despite differences in the forcing region and the time period used, both found aerosol-forced large-scale circulation changes over the targeted regions. Chung et al. (2002) further concluded that premonsoonal aerosol forcing could create a positive dynamical feedback, where aerosol-induced low-level convergence would shift the ITCZ northward, which then reinforces said convergence. Lau et al. (2006) suggested that accumulated dust and anthropogenic black carbon over the slope of the Tibetan Plateau could heat the atmosphere and hence intensify the monsoon circulation on a seasonal time scale; that is, an “elevated heat pump” effect. On the other hand, Ramanathan et al. (2005) found that the aerosol-induced reduction of surface solar insolation could decrease the meridional...
gradient of sea surface temperature over the Indian Ocean, which would weaken the large-scale monsoon circulation and counteract convection over land. With the inclusion of an ocean model in their climate model, Lau and Kim (2007) were able to examine the relative importance of the elevated heat pump effect and the temperature-gradient effect. They found that the relative importance of these effects varies temporally over monsoon season, with the former dominant in May–June and the latter in July–August. Using a six-member ensemble of atmosphere–ocean coupled models with increasing black carbon concentration on a century time scale, Meehl et al. (2008) found that aerosols cool sea surface temperature during the premonsoonal months of March–May, which could then affect monsoonal rainfall later in the monsoonal months.

The inclusion of an ocean model is desirable in modeling the climate impacts of aerosols. Aerosol forcing can remotely cause a globe-wise shift of Hadley circulation and thus the intertropical convergence zone (Wang 2007), and only when the feedback of ocean to aerosol forcing is included in the model can such a phenomenon along with other aerosol-induced alterations to the water cycle be captured. Besides the inclusion of interactive ocean components, several recent studies have also introduced more model features of aerosols to better reveal certain details regarding aerosol–monsoon interactions. Kim et al. (2008) developed a two-moment seven-mode aerosol model including a core-shell black carbon–sulfate mixture and coupled it with a climate model including ocean component. In simulations using this model they found a wet-north–dry-south bipolar structure of monsoonal precipitation change from June to August over the Indian subcontinent due mainly to absorbing aerosols. Bollasina et al. (2011) included aerosol microphysical effects in stratiform and shallow cumulus clouds besides the direct effect and found a wet–dry–wet tripolar southeast to northwest pattern in precipitation change from June to September, due to aerosol-induced meridional and zonal circulation anomalies. Ganguly et al. (2012) employed an atmospheric model including aerosol microphysical effects in stratiform clouds and a two-moment three-mode aerosol model to attribute monsoonal changes from June to September to anthropogenic as well as natural fire sources of aerosols, either local or nonlocal. They concluded that local anthropogenic aerosol emissions were primarily responsible for precipitation changes over South Asia.

A unique forcing feature of absorbing aerosols is a heating to the atmosphere resulting from their absorption of solar radiation, which coexists with a reduction of solar radiation at the surface (a cooling effect) by both scattering and absorption. Scattering aerosols such as sulfates have a radiative cooling effect on the surface but do not warm the atmosphere. The aerosol constituent and hence the type of radiative forcing is known to produce different climatic responses: for example, predominantly absorbing anthropogenic aerosols drive opposite responses in the tropical mean meridional circulations compared to the predominantly scattering natural aerosols (Allen and Sherwood 2010). In particular, even varying the quantity of the same aerosols and hence relative forcing strength can lead to different climatic responses because of cloud radiative feedback (Randles and Ramaswamy 2008).

To understand such complicated climate responses, it is important as a first step to ascertain the individual as well as the combined effect of the two opposite aerosol-induced direct forcings. To understand the mechanism behind these aerosol-induced climate impacts, it is essential to examine to what extent any particular climate impact is simply an overlap of the individual effects respectively associated with the atmospheric heating and surface cooling; to what extent there exists a significant nonlinear component; and, if the nonlinear component is significant, what the responsible interaction for it is.

This study aims to identify and investigate the role of the nonlinearity associated with the two coexisting direct effects of anthropogenic absorbing aerosols in affecting the South Asia summer monsoon during May and June, part of the aerosol-laden premonsoonal period and the time when the monsoon onsets over India. For this purpose, we have designed a series of experiments using prescribed forcings to mimic the direct effects of anthropogenic absorbing aerosols, with the exclusion of the more complicated aerosol microphysical effect (i.e., the “indirect effect”) and the feedback to aerosol loading from the aerosol-induced climate response (i.e., “interactive aerosols”). The prescribed forcings include the atmospheric heating and surface cooling over the Indian subcontinent based on previous aerosol-climate model simulations (Wang et al. 2009b). They were applied separately and then combined in three different simulations, in order to identify the climate response to each of them as well as to their combination.

In this paper we first describe the methodology in detail (section 2). The results of this study are then presented in section 3, beginning with forcing-induced changes in the climatological monsoon onset (section 3b) and how these changes in monsoon progression can be traced to changes in the distribution of the subcloud layer moist static energy (section 3c). Possible mechanisms, particularly the nonlinear ones, causing such climate responses are discussed in section 4, before the conclusions in section 5.
2. Methodology

a. Model setup

The model used in this study is the Community Earth System Model, version 1.0.3 (CESM 1.0.3; Gent et al. 2011). We used the fully coupled configuration of CESM (the B preset), including component models for atmosphere [Community Atmosphere Model, version 4 (CAM4)], ocean [Parallel Ocean Program (POP2)], land [Community Land Model (CLM4)], and land ice as well as sea ice [Community Ice Sheet Model (CISM) and Community Ice Code (CICE4)]. The atmosphere and land models were run on the 1.9° × 2.5° latitude–longitude grid, while the ocean and ice models were run on the 1° horizontal resolution (gx1v6). All these component models were run as active components to evolve freely for 100 yr from 2000 to reach equilibrium under the year 2000 forcings given in the model. Note that, while there are regions where local equilibrium might not be achieved by the end of the 100-yr integration period and thus transient features might still exist, over the Indian monsoon domain, an equilibrium state have been established in all the current 100-yr runs based on our analysis (not shown).

We used prescribed direct forcings to mimic the influence of anthropogenic absorbing aerosols. These forcings use an atmospheric warming of 1 K day$^{-1}$ (8 W m$^{-2}$ total atmospheric heating) and a surface reduction of incoming solar radiation of 10 W m$^{-2}$ to represent the effects of both scattering and absorption of anthropogenic absorbing aerosols. These two forcings were applied over shaded areas as shown in Fig. 1 and only during daytime in May and June (i.e., pentads 25–36 or days 121–180) in order to simulate the forcing in the premonsoon period when the anthropogenic aerosol level in the region peaks (Wang et al. 2009a). The atmospheric heating is applied to the three lowermost model layers in the terrain-following vertical coordinates corresponding to 993, 969, and 925 mb, assuming 1000-mb surface pressure.

Realistically, the heating generated by absorbing aerosols can occur in higher atmospheric layers as well as over larger regions geographically. The current setup was designed to isolate the response to a localized forcing over a critical region suggested by a previous study (Wang et al. 2009b) in contrast to remote forcing. This also creates a physically consistent scenario since, under the prescribed forcing setup, circulation changes in response to aerosol forcing do not feed back into the aerosol spatial distribution. Despite this, our preliminary experiments with different ocean models show notable differences in climate responses between the run using full ocean model POP2 and that using the slab ocean model (SOM), indicating that extremely low-frequency responses from ocean feedback are still important. The strength of the forcing was consistent in magnitude but slightly weaker than commonly used: for example, the surface and atmospheric radiative forcing of $-23 \pm 2$ W m$^{-2}$ and $16 \pm 2$ W m$^{-2}$, respectively, measured in the INDOEX (Ramanathan et al. 2001) of $-20$ to $-30$ W m$^{-2}$ surface and up to 2 K day$^{-1}$ respectively using a satellite-based radiative transfer model (Kuhlmann and Quaas 2010). In summary, the prescribed forcing represents an idealized gentle perturbation localized over the Indus plains and surrounding regions.

Four simulations were conducted. The first is a control run with constant current values of greenhouse gas forcing but without aerosol forcing. This run is referred to as the “control” and labeled as “B2000noaer” in the figures. In addition to the control run, there are three runs including only atmospheric heating (“warm atmosphere” or “Bwarmatm”), only surface cooling (“cool surface” or “Bcoolsfc”), and both atmospheric heating and surface cooling (“total effect” or “Btotalhf”).

To isolate the aerosol-induced responses to anthropogenic components, the radiative effects of natural aerosols (dust and sea salt) were excluded in all four simulations. In addition, because of the usage of prescribed aerosol forcings, the feedbacks of circulation and precipitation changes to aerosol spatial distributions were also excluded. All the simulations were run for 100 yr to reach equilibrium.

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**Fig. 1.** The shaded region is where atmospheric heating or surface cooling was prescribed. The prescribed forcing was an idealization of the direct radiative effects of anthropogenic absorbing aerosols.
In our analysis, the climate response or the change in a given climate parameter to a given prescribed aerosol forcing is derived by

\[
\text{change(forcing run)} = \text{results(forcing run)} - \text{results(control)}.
\]

It is expected that when both forcings are included, certain effects associated with only atmospheric heating or surface cooling could interact to produce new effects. We define this nonlinearity as

\[
\text{nonlinear contribution} = \text{change(total effect)} - \text{change(warm atmosphere)} - \text{change(cool surface)}.
\]

**b. Local onset calculations**

The climatological onset pentad was calculated based on a method suggested by Wang and LinHo (2002), using the model total precipitation results in the last 40 yr of the 100-yr run: that is, the period when the model atmosphere has reached a climatological equilibrium. The local mean climatological rainfall over pentad 3–9, where the annual cycle of model rainfall is at its minimum, was considered the baseline rainfall, and the monsoon onset pentad was then defined when the Fourier reconstructed climatological pentad rainfall was greater than the baseline rainfall by 5.0 mm day\(^{-1}\).

We have tested using 4.0, 5.0, and 6.0 mm day\(^{-1}\) to define the onset. A 4.0 mm day\(^{-1}\) or lower threshold appears too low to distinguish the monsoon onset from premonsoonal rainfall in our simulations, using the rainfall climatology based on the last 40 yr. The onset pentads over Asia using a 5.0 mm day\(^{-1}\) threshold are shown in Figs. 2a and 2b. As internal variability exists in the model, in particular with the time periods of 5–6 yr and ~40 yr, we calculated the threshold using climatology based on the last 30 yr of the simulation, as well as for each of the two 20-yr periods in the last 40 yr of the simulation. The use of 30-yr, 40-yr, and 2 × 20-yr climatologies combined with two different rainfall thresholds of 5.0 and 6.0 mm day\(^{-1}\) creates eight combinations in defining onset and thus derives eight sets of different climatology onset pentads for any particular run. For any forcing run, we consider a location as experiencing earlier or later onset if more than half of the combinations show consistent earlier or later onset compared to the control run. For any forcing run compared to the control run, this can be formulated as a paired sign test with the null hypothesis of there being an equal number of earlier and later onset cases. The locations where the null hypothesis was rejected at different confidence levels are marked in Fig. 3, which shows the mean of the eight sets of onset pentads.

We chose to track the progress of the climatological monsoon by a line of maximum subcloud layer moist static energy (SL-MSE; Prive and Plumb 2007; Wang et al. 2009b). This marks the location of the monsoon trough or where the energy brought in by the large-scale circulation must be dissipated in strongest convection activities (Emanuel 1995). As shall be discussed later, horizontal advection rather than just surface moisture flux is an important factor in controlling the value of SL-MSE. Such an analysis is specifically useful because the climate model arguably captures more robustly the spatial and temporal changes in model energy due to changes in large-scale temperature and humidity distributions than it captures changes in precipitation. As the South Asian monsoon arises from an interhemispheric energy imbalance, tracking model changes in moist static energy would be a better way to define the monsoon progression than tracking precipitation changes in the model. As shown in Figs. 4a–d, the maximum SL-MSE is reflected by the daily climatology value of moist static energy at the terrain-following model level corresponding to 925 mb, assuming the reference pressure of 1000 mb at the surface (shaded contours), and is hence derived along each longitude between 60° and 90°E, after a three-point running mean (thick black line). We term the centroid of this line the “monsoon centroid” (red dot) and tracked it over time for each run (thin black line trailing the red dot). The longitudinal limits (60° and 90°E) of the maximum SL-MSE calculation are so selected since the monsoon is naturally bound in the west and east at these longitudes by steep topography inland.

**3. Results**

**a. Climatology of the monsoon onset**

The climatological monsoon onset derived from the control run is compared to that derived from the Climate Prediction Centre Merged Analysis of Precipitation (CMAP; Xie and Arkin 1997) (Figs. 2a,b). Within the analysis domain that ranges from South Asia and Southeast Asia to East Asia and the west Pacific, the model captures the observed onset rather well, particularly in west India, along the north edge of all the monsoon systems in the region, and also in most of the west Pacific. The model-derived onset over west India differs from the observation-based onset by ~2 pentads while still retaining the overall onset pattern. The model-derived onset differs by ~2 pentads over the Arabian Sea and the Bay of Bengal and mismatches observations significantly over the 110°–120°E East Asian band. Note
that the control run produces an equilibrium climate corresponding to the current climate forcings (year 2000), while the observation-based analysis is better described as the climatological mean over past several decades under transient climate forcings.

June is the month when the South Asian monsoon advances over India. The rainfall climatology in June over the entire analysis domain derived from the control run exhibits a gross similarity to the CMAP data despite certain differences in the locations of regional rainfall maxima (Figs. 2c,d). Note that, in order to match the CMAP resolution of $2.5^\circ \times 2.5^\circ$, the total climatological precipitation in June of the control run (Fig. 2c) was regridded. However, little qualitative change resulted from this regridding since the original CMAP and model resolutions are already close enough ($1.9^\circ \times 2.5^\circ$ versus $2.5^\circ \times 2.5^\circ$).

In the western region of the Indian subcontinent (marked by the thick black box in Fig. 2), both the June rainfall climatology and the onset pentad of the control run agree reasonably well with the CMAP data, except in the northeast corner of the box, where model rainfall differs from that of CMAP, and in the southeast corner, where the onset occurs far too late because of weaker model rainfall over the Bay of Bengal.

b. Monsoon onset response to prescribed forcings

As calculated from the reconstructed precipitation climatology, although the differences in the monsoon onset between the model runs with idealized forcing (Figs. 3a–c) and the control run (Fig. 2a) are small, they are spatially correlated within the region of interest. That is, an earlier or later onset caused by prescribed aerosol forcing would extend over multiple grid points with some correspondence to geographical features.

The atmospheric heating effect appears to affect the monsoon onset more significantly than the surface cooling effect. Compared to the control run, regions showing early monsoon onset around $15^\circ$N in west-central India can be seen in all three forcing runs (red dashed box 1 in Figs. 3a–c). These regions lie farther north ($14^\circ$–$20^\circ$N; Figs. 3a,c) in the warm-atmosphere and total-effect runs but farther south ($10^\circ$–$18^\circ$N; Fig. 3b) in the cool-surface run. Furthermore, early onset over northwest India is seen in both the warm-atmosphere and the total-effect runs (red dashed box 2, Figs. 3a,c) but not in the
cool-surface run (red dashed box 2 in Fig. 3b). The effect on onset date appears stronger in the total-effect run compared to the warm-atmosphere run. In the latter, the monsoon arrives up to two pentads earlier over a larger region of west-central and northwest India. A region of late onset in east-central India appears in the cool-surface run, which suggests that aerosol direct forcing at the surface (cooling) can weaken the monsoon over east-central India. The monsoon rainfall is not well captured by the model there (Figs. 2c,d), and thus this work focuses on western side of India instead. The early monsoon onset over northwest India responding to forcing by only absorbing aerosols in this experiment coincides with the CMAP-based results by Kajikawa et al. (2012). Hence, it may be possible to detect observationally the effect of absorbing aerosols on monsoon onset in northwest India.

c. Northwest shift of monsoon system energy in May and June

We find that none of the three forcing runs has displayed an earlier monsoon onset at the earliest inland

![Fig. 3. Mean onset pentad and its changes over the region of interest (large black box corresponding to that in Fig. 2). Earlier monsoon onset is seen in both peninsula India (red dashed box 1) and northwest India (red dashed box 2) for most forcing runs. (a) Mean onset pentad in the control run. Mean onset pentad changes in the (b) warm-atmosphere, (c) cool-surface, and (d) total-effect runs. The mean of onset pentads shown is calculated using different rainfall thresholds and model year subsets. Positive values denote later onset and negative values denote earlier onset. Points that are nonmonsoonal and where onset is discontinuous are shaded gray. Dots and circles indicate rejection of the null hypothesis at 95% and 90% confidence level, respectively, using a paired sign test with the null hypothesis of equal cases of earlier and later onset.](image-url)
onset site: that is, the southwestern tip of India (Figs. 3a–c). Regions where the climatological monsoon onset occurred earlier in the warm-atmosphere and total-effect runs seem to fall close to the contours of equi-onset dates on the control run. Given that the prescribed forcing has been in place since pentad 25 (forcing is during May and June: i.e., pentads 25–36 or days 121–180), these patterns of onset change suggest that the changes in calculated onset are not due to temporally or spatially localized precipitation responses to local forcing.

The progress of the monsoon from the initial onset (pentad 28) onward is now tracked by the line of maximum subcloud moist static energy (SL-MSE), an arguably more robust diagnostic of the climate model, and the newly introduced concept of monsoon centroid (see section 2b).

In the control run, the monsoon centroid barely moved during pentad 28–32 (Fig. 4a) and thereafter advanced northward by about 5° in space within 2 pentads as the monsoon moving toward inland from the west coast of India (Fig. 4b). It progressed across central and east India in another 10 days (Fig. 4c) and then slowed down as it approached the Himalayas (Fig. 4d).

Centroids of the forcing runs moved differently compared to the control run, with a greater similarity existing between the warm-atmosphere run and total-effect run than that between either of these two runs and the cool-surface run (Fig. 4e). As the monsoon progressed over west-central India between days 160 and 180 (pentads 32 and 36), the movement of centroid in three forcing runs displays a difference: the centroids in the warm-atmosphere and total-effect runs moved roughly in parallel and overtake the centroid of the control run. The centroid in the cool-surface run initially advanced the fastest but slowed down even before the prescribed forcing was halted. If prescribed

FIG. 4. Tracking the progression of the monsoon centroid reveals the warm-atmosphere and total-effect runs to be monsoon advancing, while the cool-surface run is initially monsoon advancing but later monsoon retarding. (a)–(d) For the control run at various pentad-averaged times, shading shows the moist static energy on the model level corresponding to 925 mb, assuming reference surface pressure of 1000 mb. The thick black line is maximum moist static energy for each the longitude. The red dot is the location of centroid. The line following the red dot shows previous locations of centroid. (e) The movement of the monsoon centroid across India from day 156 (pentad 31). Here, the black line is control; the green line is warm-atmosphere run; the blue line is cool-surface run; and the red line is total-effect run. The centroid location at squares is day 160 (pentad 32) and at diamonds is day 180 (pentad 36). The map of India is of higher resolution than the model. To give an indication of geographical features, regions between 300 and 600 m in height are shaded beige. Regions above 600 m are shaded gray.
forcing were continued past day 181, the centroid in the cool-surface run could even fall behind that in the control run should its track be extrapolated.

After prescribed forcings were halted on day 180 (pentad 36), the centroids in the forcing runs eventually converged with the centroid in the control run. This convergence was not immediate, indicating that the effects of the prescribed forcing persist into the monsoonal period.

Local forcing can induce a shift of energy in the monsoon system to the northwest, which sets up the environment for differential monsoon progression after the initial onset. To trace the source of this energy change, we compared the temperature and water vapor contributions to the change in SL-MSE. The influence of the prescribed forcing was mainly manifested through changes in water vapor distribution rather than temperature change in all three forcing runs (Fig. 5). In particular, when both atmospheric heating and surface cooling were present, the SL-MSE gradient in northwest India was greatly strengthened (Fig. 5d) by the increase and decrease of lower-tropospheric water vapor content in the northwest and southeast, respectively (Fig. 5i). The nonlinear contribution to this gradient change (Fig. 5j) was more important than the linear contributions (Figs. 5g,h). In terms of magnitude, this nonlinear contribution was as strong as the linear contribution.

The changes in the lower-tropospheric column water vapor may be traced to advection (Figs. 6b–d) rather than local evaporation (not shown). When both atmospheric heating and surface cooling were present, a stronger westerly water vapor flux from the Arabian Sea together with a weaker flux out of India resulted in a greater flux convergence over India with a corresponding strong increase in lower-tropospheric column water vapor (Fig. 6d). A southeasterly surface wind anomaly from the Bay of Bengal brought water vapor into the normally arid northwest India (Fig. 6i). Once again, the nonlinear contributions to the changes in surface wind (Fig. 6j), lower-tropospheric column water vapor, and water vapor flux (Fig. 6e) can be calculated. Interestingly, they are quantitatively comparable to those changes caused by only atmospheric heating or surface cooling (Figs. 6g,h,b,c).
d. Monsoonal precipitation response

In response to the changes in monsoon energy induced by the prescribed forcing, increased rain intensities appear as expected to accompany the earlier onset over the western regions of India in both premonsoonal and monsoonal periods. That the effects of the prescribed forcing in May and June persist into the monsoon can be seen in the June–August precipitation changes. The changes in June–August seasonal precipitation climatology derived from both the warm-atmosphere (Fig. 7a) and the total-effect (Fig. 7c) runs show a decrease in the east Arabian Sea and an increase across west-central India. To the north and south of this band, decreased precipitation (Fig. 7a) or at least no change or weak increase in precipitation (Fig. 7c) was present in the warm-atmosphere and total-effect runs. Such a banded structure is qualitatively similar to several previously findings in precipitation responses: for example, the black carbon aerosol case of Menon et al. (2002) and Lau et al. (2006), the absorbing aerosol case of Wang et al. (2009b), and the low aerosol optical depth case of Randles and Ramaswamy (2008).

The banded structure was not present in the cool-surface run. Instead, total precipitation is generally decreased over the domain of interest, except for a “C”-shaped band of increased precipitation about peninsular India (Fig. 7b) that will be discussed in section 4b. Such a pattern is more similar to precipitation responses in several previous studies: for example, both black carbon and sulfate cases of Meehl et al. (2008), the white aerosol case of Randles and Ramaswamy (2008), and the postindustrial Asia and black carbon cases of Ganguly et al. (2012).

In particular, we note that, despite the small spatial coverage of the forcing used in this study, the presence of banded structure and the increased precipitation over northwest India in the total-effect run but not in the cool-surface run is similar to the contrast between the weak black carbon and scattering aerosol cases of Randles and Ramaswamy (2008) where Asia-wide forcing was...
applied. This suggests that the precipitation signal in northwest India is a response to forcing over a local critical region rather than to remote forcing.

The nonlinear interaction between coexisting atmospheric heating and surface cooling played a significant role in causing such a precipitation change (Fig. 7d versus Figs. 7a–c). This nonlinearity induced an L-shaped pattern of increased precipitation that resembled the L-shaped pattern from the interactive aerosol model more than any of the three idealized-forcing models (Fig. 7d versus Fig. 7e). On the other hand, this nonlinear factor led to wetter conditions over the west Arabian Sea, whereas the west Arabian Sea appeared to be drier according to the result of the interactive aerosol-climate model; only the atmospheric heating effect showed wetter conditions over west Arabian Sea. It appears that the total-effect run did not capture the amplitude of either the atmospheric heating effect or the nonlinear effect required to reproduce the result of the interactive aerosol-climate model. This is likely due to the particular shortwave forcing profile used in our simulations.

4. Mechanisms behind the monsoon response to direct aerosol forcing

As pointed out in section 3c, the nonlinear interaction between the atmospheric heating and the surface cooling increases the lower-tropospheric water vapor over northwest India through advection. We now propose the mechanism behind this interaction through three scenarios corresponding to the three forcing runs.

a. Atmospheric heating: Water vapor as proxy agent

When atmospheric heating is applied alone to the model, a strengthened monsoon circulation is observed. In particular, the stronger zonal circulation cell between the Arabian and the Indian subcontinents leads to increased adiabatic warming and reduced clouds in the descending air over the Arabian subcontinent, creating a dipole temperature anomaly with a warm west and a cold east present in the lower troposphere over the Indian subcontinent. This in turn fuels low-level convergence and cyclonic anomalies over the northwestern region of the Indian subcontinent (Fig. 6g). The cyclonic anomaly strengthens the low-level monsoonal westerlies and hence water vapor flux from the Arabian Sea (Fig. 6b), first through an enhanced flow itself and second through increased evaporation over the Arabian Sea in response to the stronger surface wind. As a result, increased convective rainfall not only occurs over the deep convection zone at the foothills of the Tibetan Plateau but over central India as well. In turn, the increased evaporative cooling of precipitation and latent heat release in the lower and upper troposphere respectively over central India perpetuate the dipole temperature anomaly as seen in Figs. 6g and 6l.

We thus propose that, as depicted in Figs. 8a and 8b, the aerosol-induced atmospheric heating increases the water content of the atmosphere by increasing convective activity over the forcing region as shown in Fig. 9a, where two significant anomalous centers of upward motion can be seen over India, one at the foothills of the Tibetan Plateau and the other over central India. The latter has a three-dimensional structure that is broader at the western edge of the forcing region (Fig. 9e), consistent with the semicircular shape of the prescribed atmospheric heating and supporting the schematic depicted in Fig. 8b.

On the other hand, there is an associated increase in precipitation over the forcing region that can act as a
negative feedback, by resulting in an evaporative cooling and thus weakening the prescribed heating in the lower troposphere. Atmospheric heating and evaporative cooling do not necessarily overlap all the time because the prescribed aerosol forcing only appears in the daytime, permitting a net positive convective effect and a net cooling at surface (Fig. 6g).

It is possible that the initial local convection induced by the prescribed aerosol atmospheric heating might have sparked off the initial increase in latent energy in the monsoon and also westerly convergence from the Arabian Sea, as well as sustained large-scale circulation change. The increase in atmospheric water content would be stronger when there is a prevailing westerly circulation to bring water vapor to the foothills of the Tibetan Plateau, which may explain why the monsoon centroid of the warm-atmosphere run only diverged north of 13°N from the control run after the initial monsoon onset at the southern tip of India (Fig. 4e).

b. Surface cooling: Convection inhibition and temporal varying influence

The most notable feature of onset climatology in the run with only prescribed surface cooling is a strong, negative temperature anomaly over India (Fig. 6h), which extends into the lower atmosphere downstream of the prescribed forcing region. This anomaly is not surprising, as prevailing surface westerlies would advect the cold air mass downstream in the absence of other feedbacks. A low-level anticyclonic anomaly over the
Bay of Bengal is induced by the cool anomaly (Fig. 6h). This anticyclonic anomaly drives an anomalous southeasterly water vapor flux from the Bay of Bengal around peninsula India (15°–20°N; Fig. 6c). C-shaped anomalies about peninsular India of increased precipitation (10°–20°N; Fig. 7b), increased cloud (not shown), and upper-tropospheric warming (Fig. 6m) are associated with this anticyclonic anomaly. The location and strength of this anomaly is important in the adjustment of the monsoonal circulation, as will be seen in case of the total-effect run (section 4c).

An initial faster monsoon advance followed by a slowdown of the advance in the cool-surface run was aforementioned in sections 3b and 3c. We propose the following explanations for the changes in monsoon advance speed. The effect of the surface cooling during initial period of the monsoon is depicted in Figs. 8c and 8d. The cold air mass associated with this cooling suppresses convection (Fig. 9b) and convective precipitation over its location: that is, slightly downstream of the forcing region. Instead, buildup of lower-tropospheric column water vapor (Fig. 6c) and enhanced precipitation occurs farther downstream of the cold air mass (east side of Fig. 7b). This gives the impression that the monsoon has advanced quickly, whereas in reality convection was just suppressed by the cold air mass.

The process depicted in Figs. 8e and 8f explains the slowdown in monsoon advance later into the monsoon season: after monsoon onset occurs over a region, the surface water vapor flux from rain-moistened surface would normally supply water vapor to feed the monsoon system. When the surface is continuously cooled after the initial onset, this surface water vapor flux would be weakened, thus retarding the monsoon.

c. Coexisting effects and their nonlinear interaction

When only atmospheric heating is present, a low-level cyclonic anomaly over the Arabian Peninsula brings an anomalous water vapor flux from the Arabian Sea to India (section 4a). When only surface cooling is present, a low-level anticyclonic anomaly brings an anomalous water vapor flux from the Bay of Bengal to India: that is, weakening monsoonal westerly water vapor flux out of India. When both effects are present, the northwestern cyclonic anomaly and the southeastern anticyclonic anomaly would still appear (Fig. 6i); however, the anomalous water vapor flux from the Arabian Sea is redirected into a different region than that in the warm-atmosphere run: one part goes toward the southeast along the west coast of India where experiences increased precipitation (Fig. 7c) and the other is redirected into northwest India by the presence of the anticyclonic anomaly, which weakens the water vapor flux across the eastern side of India. This creates a strong anomalous movement of water vapor into northwest and north India (Fig. 6d). The presence of strong anomalous upward motions at the foothills compared to either atmospheric heating or surface cooling alone cases can been seen (Fig. 9g).

Nonlinear effects are dominant in extending increases in both lower-tropospheric column water vapor (Fig. 6e) and precipitation (Fig. 7e) over northwest India. As discussed previously, evaporative cooling could compete with the prescribed atmospheric warming, resulting in a net lower-tropospheric cooling over India. This negative feedback to the prescribed atmospheric warming would be ameliorated with the addition of prescribed surface cooling. The nonlinear response in the lower troposphere shows a warming over India and a cooling only farther north (Fig. 6j). This suggests that the precipitation and
resultant evaporative cooling has been redirected away from the region of prescribed forcing. An increase in convection to the northwest is apparent when examining the nonlinear contribution. However, comparing to the warm-atmosphere run, the two centers of anomalous upward motion have moved from the north of India to the northwest of India (Fig. 9h). The strong upper-level warming and anticyclonic anomaly north of India both indicate enhanced latent heat release and upper-level divergence associated with deep convection, which is consistent with this interpretation (Fig. 6o).

5. Conclusions

The climate impact of multiple forcings can be simply a linear sum of the individual effects associated with each forcer, or it could contain a significant nonlinear component because of interaction between the individual forcers. The direct radiative effects of anthropogenic aerosols include an atmospheric heating component due to absorption of solar radiation and a surface cooling component due to both absorption and scattering. We examine in this study the climatic response of the South Asian summer monsoon system to such coexisting effects in May and June, with focus on monsoon onset and precipitation. To examine the individual effects of these two opposite direct forcings and, most importantly, to identify the nonlinear effects due to coexistence of them, we carried out century-long simulations using the CESM 1.0.3 with idealized forcings prescribed over India. These forcings include only atmospheric heating (warm-atmosphere run), only surface cooling (cool-surface run), or both of them (total-effect run).

We find that the model produced monsoon onset over most regions of South Asia compared favorably with that from observation-based CMAP. We also find that the direct aerosol forcings can cause changes over Indian subcontinent in both climatological onset pentad and associated rainfall. Regarding these changes, the warm-atmosphere and total-effect runs share more similar features including an advanced climatology local onset over both west-central and northwest India by 1–2 pentads compared to the control run. The monsoon in the warm-atmosphere and total-effect runs also progresses faster across the India subcontinent than in the control run. In contrast, the monsoon in the cool-surface run initially advanced fast but was later retarded. All these are reflected in the aerosol-forcing-induced changes in SL-MSE, which can be attributed to the change in latent energy distribution rather than in temperature. Advection of water vapor was the major contribution to the changes in latent energy compared to local increases in evaporation. Interestingly, the above effects of direct aerosol forcing applied in May and June were retained in the monsoon season after the forcing was halted. Future studies considering aerosol forcing during monsoon season in addition to premonsoon season would advance knowledge of the aerosol-induced changes in monsoonal circulation and rainfall in the monsoon season.

The nonlinear component due to coexistence of the two direct forcings has caused the response of the monsoon system to differ from a simple overlap of the two corresponding responses to these forcings individually. It is found to be the most important factor in determining the changes in subcloud layer moist static energy and thus in moisture processes over northwest India. We propose that such an importance of nonlinear effect could be established as follows: When both atmospheric heating and surface cooling coexist, the surface cooling effect delays the otherwise enhanced precipitation over central India responding to the atmospheric heating, so that the anomalous water vapor flux from the Bay of Bengal, caused by the atmospheric heating, would be redirected into the north and northwest, thus ameliorating the negative feedback through evaporative cooling when only atmospheric warming is present. This increases the water vapor content in the north and northwest of India substantially, creating an environment conducive to the convection.

Acknowledgments. This research was supported by the Singapore National Research Foundation (NRF) through a grant to the Center for Environmental Sensing and Modeling (CENSAM) of the Singapore-MIT Alliance of Research and Technology (SMART), by the U.S. National Science Foundation (AGS-0944121), and by US Environmental Protection Agency (XA-83600001-0). This research was also supported by the A*STAR Computational Resource Centre of Singapore (http://www.acrc.a-star.edu.sg) through the use of its high performance computing facilities. We like to thank the two anonymous reviewers for their many constructive comments and suggestions, which have led to several important improvements of the manuscript.

APPENDIX

Statistical Testing Procedure

a. Onset pentad

For any subset of the last 40 yr of the model run and rainfall threshold, a pair of onset pentad values can be calculated for the control run and a forcing run. Of the eight pairs of onset dates calculated from the 30-yr, 40-yr, and 2 × 20-yr climatologies using rainfall thresholds
of 5.0 and 6.0 mm day\(^{-1}\), some pairs have later forced onset while others have earlier forced onset. Most onset pentad changes were not large, about 1–2 pentads, and the distribution of onset dates in most locations failed normality. Hence, for each forcing run, we performed a paired sign test of at each location with the null hypothesis of there being equal number of earlier and later onset cases. Locations where the null hypothesis was rejected at 95% and 90% confidence level are shown as dots and open circles in Fig. 3.

b. Moist static energy, vertically integrated water vapor, and temperature

40-size samples were obtained from the last 40 yr of the pentad 32–36 mean of moist static energy at the 925-mb equivalent model level assuming 1000-mb surface pressure, temperature and humidity components of the moist static energy, vertically integrated water vapor in the lower troposphere up to the 470-mb equivalent model level, temperature at the 925-mb equivalent model level, and temperature at 300 mb. Once again, gridpoint variables failed normality at many regions in the domain of interest. Hence, for each forcing run, we performed a two-sample Kolmogorov–Smirnov test for the two samples of control and forcing runs, with the null hypothesis of both samples being drawn from the same distribution.

c. Total precipitation

Four decadal mean samples were obtained from the last 40 yr of the June–August total precipitation. Decadal means were used as the precipitation signals in all three forcing runs were weak compared to interannual variability in the model and could not be detected. However, the signals were detectable compared to decadal variability but weakly at 85% confidence level in most regions. As previously done, we performed a two-sample Kolmogorov–Smirnov test for the two samples of control and forcing runs, with the null hypothesis of both samples being drawn from the same decadal distribution.

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