### Effects of the South Asian Absorbing Haze on the Northeast Monsoon and Surface–Air Heat Exchange

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(Manuscript received 11 July 2001, in final form 21 Febraury 2002)

### ABSTRACT

The effects of the south Asian haze on the regional climate are assessed using the National Center for Atmospheric Research Community Climate Model version 3 (CCM3) at the T42/L18 resolution. This haze, as documented during the Indian Ocean Experiment (INDOEX) campaign (1995–2000), consists mainly of an-thropogenic aerosols, and spans over most of south Asia and the north Indian Ocean. It reduces the net solar flux at the surface by as much as  $20-40 \text{ W m}^{-2}$  on a monthly mean basis and heats the lowest 3-km atmosphere by as much as  $0.4-0.8 \text{ K day}^{-1}$ , which enhances the solar heating of this layer by 50%-100%. This widespread haze layer is a seasonal phenomenon limited to the dry period between November and May.

The imposed haze radiative forcing leads to several large and statistically significant climate changes during the dry monsoon season, which include cooling of the land surface, and warming of the atmosphere. These temperature change features lead to the stabilization of the boundary layer that results in a reduction of evaporation and sensible heat flux from the land. The dynamical response to the aerosol forcing is surprisingly large. The aerosol forcing weakens the north–south temperature gradient in the lower level, which results in an enhancement of the area mean low-level convergence and a northward shift of the ITCZ. The increase in low-level convergence leads to increased convective rainfall and latent heat release, which in turn leads to a further increase in lowlevel convergence. This positive feedback between the low-level convergence and deep convective heating increases the average precipitation over the haze area by as much as 20%. The ocean surface undergoes a suppression of evaporation. Because of this decreased evaporation accompanied by the increase in the hazearea precipitation, the precipitation over the rest of the Tropics decreases, with a large fraction of this decrease concentrated over the Indonesian and the western Pacific warm pool region. The prescribed dry monsoon haze effect affects the summertime wet monsoon too, but a detailed analysis has to await the availability of yearround aerosol data.

The major inference from this study is that the effects of absorbing aerosols on the regional climate can be quite large. The simulated surface temperature response was very sensitive to the ratio (R) of the surface aerosol forcing to the atmospheric forcing. The R itself varies from -1.5 in clear skies to about -0.5 in overcast skies over ocean, and available experimental data are not sufficient to constrain its value more narrowly.

#### 1. Introduction

The south Asian region is characterized by a monsoonal climate, in which the wind flows north to northeastward (south-southwesterly wind) from the tropical Indian Ocean during the wet summer season and south to southwestward from the continent during the dry season (see Ramage 1971 for a detailed discussion of monsoon). As documented during the Indian Ocean Experiment (INDOEX) field campaign (conducted from 1996 to 1999), a brownish haze layer spread over most of the North Indian Ocean and the South/Southeast Asian continent during the dry season. A detailed summary of this haze layer, its radiative properties and forcing is given in Ramanathan et al. (2001, hereafter referred to as R01). The haze layer has been identified every dry monsoon season since the beginning of the INDOEX campaign in 1995 (Jayaraman et al. 1998; Satheesh et al. 1999; R01).

The black carbon and dust in this haze resulted in a large solar absorption with a column-averaged singlescattering albedo (SSA) of about 0.9 at 0.5  $\mu$ m (R01). SSA is the ratio of the scattering coefficient of aerosol to the sum of the scattering and absorption coefficients. Here, we refer to SSA at the ambient relative humidity, as in R01. The highly absorptive aerosol layer led to a large increase in solar absorption in the lowest 3 km of the atmosphere, accompanied by a comparably large reduction in solar radiation absorbed by the land and sea surface. Because of the near cancellation of the

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FIG. 1. INDOEX aerosol radiative forcing: (a) Ramanathan et al.'s (2001) estimate of surface forcing  $F_s$ , atmospheric forcing  $F_A$ , and topof-atmosphere forcing  $F_{TOA}$  for clear skies over ocean; (b) forcings for average cloudy skies (Ramanathan et al. 2001); and (c) ratio R,  $F_s/F_A$ , as a function of low-level cloud fraction as in Podgorny and Ramanathan (2001).

changes between the solar heating of the surface and that of the atmosphere, the changes at the top-of-theatmosphere (TOA) were a factor of 3–5 smaller than the changes at the surface (R01). INDOEX aerosols' direct radiative forcing values for clear and average cloudy skies, as obtained from R01, are reproduced in Figs. 1a and 1b. It is seen that the TOA negative forcing is reduced substantially with the inclusion of clouds. The ratio *R* of surface to atmospheric forcing changes from -1.5 to -0.5 with the inclusion of clouds over the ocean (Fig. 1c). It is clear from Figs. 1a and 1b that the main effect of the Asian absorbing aerosol is to redistribute the solar radiation between the surface and the atmosphere.

The magnitude of the added heating to the atmosphere is also large when compared with climatological heating in the Tropics. For example, the enhancement in the daily and seasonally averaged atmospheric solar heating, between the surface and about 3 km, exceeded 0.6 K day<sup>-1</sup> over most of the Indian subcontinent and the northern Indian Ocean. This magnitude is as much as 50%-80% of the climatological solar heating of this layer. It is also comparable to the perturbation in deep atmospheric diabatic heating of about 1–2 K day<sup>-1</sup> over the tropical Pacific during ENSO events (Nigam et al. 2000), and diabatic heating of less than 1 K day<sup>-1</sup> over the tropical Atlantic during Atlantic Niño events (Ruiz-Barradas et al. 2000). The spatial extent of the haze forcing varies from year to year, and here we have taken a rather extreme case where the haze spreads south of the equator as during January 1998 and March 2000.

How does the atmosphere respond to such a heating perturbation, concentrated regionally over the Indian Ocean region? This paper reports on initial assessments of the regional effects using a general circulation model (GCM). Aerosol can affect the climate indirectly as well; for example, through changing the cloud condensation nuclei concentration and composition. Thus far, we have included only the direct effect of the haze, that is, aerosols directly affecting the radiation balance at the surface and in the atmosphere. Our major thrust is on the response of the dry monsoon to the haze. The initiated GCM studies are listed below:

- 1) Present study: It imposes a regional distribution of the heating perturbation due to the haze as documented by R01 from observed aerosol optical depths, aerosol composition, and cloud distribution. The rationale for directly prescribing the heating field is to keep the radiative forcing as close to the observed values as possible, given that the aerosol forcing depends very strongly on cloud fraction. For example, as shown in Fig. 1c, R changes from -1.5for clear skies to -0.5 for overcast conditions. Thus if the model cloud fraction is very different from the observed values, the aerosol forcing (even its sign at the TOA) will depart drastically from the observed values. In addition, if the cloud fraction changes in response to the forcing, the forcing will also change. The main objectives of this study are twofold: (i) to understand the response of the regional climate to prescribed solar heating change and (ii) to understand the dependence of the simulated climate change on the ratio R.
- 2) Kiehl et al. (2000): This study is similar to the present study except in two major aspects. First, instead of prescribing the aerosol forcing, it prescribes the observed aerosol properties and optical depths, and



FIG. 2. Prescribed aerosol radiative forcing in numerical experiments: (a) atmospheric forcing at the second-lowest model level, (b) column-integrated atmospheric forcing, (c) surface forcing for the R = -1.5 experiment, and (d) seasonal evolution of area mean of prescribed forcing. Forcings in (a)–(c) correspond to those on the 29th Julian day, and the area mean in (d) was taken over the grid points where  $F_A$  in (b) exceeds 3 W m<sup>-2</sup>. The other experiments differ from the R = -1.5 experiment only in the surface forcing  $F_s$ . In (d), each of the *x*-axis labels point to the middle of a calendar month (as in subsequent figures). In (a)–(c) values greater (less) than one contour interval are shaded dark (light).

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TABLE 1. Summary of the numerical experiments of the south Asian haze effects. Here R is  $F_s/F_A$ , where  $F_s$  is the haze forcing at the surface and  $F_A$  is the forcing within the atmosphere. All of the experiments were conducted using the NCAR CCM3 and the climatological seasonal cycle of SSTs.

Name	Design of the haze forcing	Length of model integration
Control run	No haze effect	85 yr
$R \approx -0.6 \text{ expt}$	R varies, and is about $-0.6$ on the average	100 yr
$R \approx -0.9$ expt (standard expt)	R varies, and is about $-0.9$ on the average	60 yr
R = -1.5 expt	R is exactly $-1.5$ everywhere	28 yr

allows the National Center for Atmospheric Research (NCAR) Community Climate Model Version 3 (CCM3) radiation code to generate the radiative forcing, using model-generated cloud fields. In this configuration, the spatial variations in cloud fraction increase the spatial and temporal variability of R. The other difference is that the sea surface temperature (SST) responds to the aerosol forcing, using a mixed-layer slab ocean model. This case allows them to assess the impact of sea surface temperature variation on the simulated changes. Our study utilizes the climatological seasonal cycle of SST. A comparison of these two studies will also be useful to understand the role of the ocean. An analysis of the Indian Ocean SSTs (Clark et al. 2000) shows a cooling trend near the continent and a warming trend in the equatorial region, both trends being not of large amplitude. In addition, Levitus et al.'s (2000) assessment of the Indian Ocean heat storage indicates an insignificant change in the northern Indian Ocean. Kiehl et al.'s study with a slab ocean model simulates greater SST responses (about -0.5 to -1.0 K) to the haze, and so an implication is that the greenhouse gas increase has partially cancelled the aerosol effects at the ocean surface. Thus, it is likely that the present simulation with fixed sea surface temperature



FIG. 3. Standard error of CCM3 control run: (a) Jan–Feb mean surface temperature from 85 yr of CCM3 run, (b) May–Jun surface temperature, (c) Jan–Feb precipitation, and (d) May–Jun precipitation. Standard error here was computed by treating each of T42 grid points independently and with the assumption that years are not related to each other. An ellipselike contour in each panel denotes the area where  $F_A$  in Fig. 2b exceeds 3 W m<sup>-2</sup>.



FIG. 4. Surface temperature averaged over the land area where  $F_A$  exceeds 3 W m<sup>-2</sup>: (a) daily max and min temperature changes for the  $R \approx -0.9$  experiment, (b) daily mean temperature changes for each of the three experiments (i.e.,  $R \approx -0.6$ ,  $R \approx -0.9$ , and R = -1.5), and (c) the CCM3 temperature climatology from the control run.

may not be an unreasonable sensitivity experiment of the haze effects.

Global effects of aerosols have been estimated in several studies. Haywood et al. (1997) and Tett et al. (1999) included sulfate aerosols in GCMs, and demonstrated that the aerosol cooling compensated the CO<sub>2</sub>-induced warming considerably. Our study differs from the majority of previous GCM studies, in that the aerosol forcing is obtained from the INDOEX, and hence is more realistic and inclusive of absorbing aerosols. The forcing used in this study is a synthesis of observations, and model estimates using models that have been validated with INDOEX observations. In this first attempt, we seek to simulate the steady-state response of the dry monsoon climate to INDOEX aerosols as documented in the late 1990s, rather than the transient response to increasing Asian aerosols from the preindustrial to the 1999 levels. Even with this simplification of the problem, there still exist a number of uncertainties associated with the forcing.

Apart from the temporal day-to-day variability within the winter monsoon season (which is ignored in our study), the pattern and the magnitude of the aerosol climate forcing within the atmosphere  $F_A$ , those of climate forcing at the surface  $F_S$  or the ratio  $R = F_S/F_A$ , and the vertical profile of  $F_A$  are still unclear. The ratio



FIG. 5. Low-level temperature change, averaged over the land area like in Fig. 4, between the control run and each experiment (a) at 925, (b) 850, (c) 775, and (d) 700 hPa. In (a)–(d) the short dashed line represents results from the  $R \approx -0.6$  experiment, the long dashed line from the R = -1.5 experiment, and the solid line from the  $R \approx -0.9$  experiment.

R can vary from -0.5 for the overcast-sky case to -1.5for clear skies over the ocean (see Fig. 1c). Furthermore, in clear-sky cases *R* can be even > -1.0 over land where the surface albedo is large compared to the ocean value. Thus, the lower limit for R appears to be -1.5 with INDOEX aerosols, while the upper limit is yet to be diagnosed. Given the uncertainty of the R estimate, we simulate three cases. All of the three experiments have the same  $F_A$  but differ in the magnitude of R, that is,  $F_s$ . In the standard experiment, the downward direct solar radiation reaching the surface is reduced by 1.5 times the imposed lower-atmosphere heating. The diffuse radiation reaching the surface is not affected. As a result,  $F_s$ , that is, the reduction in net solar radiation absorbed by surface, becomes much smaller when there is very little direct radiation at the surface due to cloudiness or the surface (e.g., dry land surface or snowcovered surface) reflecting a lot of solar radiation. The effective value of R averaged over land is about -0.9,



FIG. 6. Land surface temperature change during Feb–Apr for each of the three experiments. (a)–(c) An ellipselike contour denotes the area where  $F_A$  in Fig. 2b exceeds 3 W m<sup>-2</sup>. The contour interval in (a)–(c) is 0.3 K, and values greater (less) than 1 contour interval are shaded dark (light) to highlight large change.

very close to the mean value of the ratio estimated by R01. This effective *R* value was calculated by taking the ratio of the average (spatial and temporal) model surface solar flux change and the average model atmospheric solar heating change. We refer to this standard case as  $R \approx -0.9$  experiment throughout this paper. The second case is referred to as  $R \approx -0.6$  experiment in which the downward direct radiation is decreased by as much as the imposed (direct plus diffuse) atmospheric heating. In this experiment, the effective value of *R* 

under average cloudiness is about -0.6. The third case is referred to as R = -1.5 experiment in which the net surface solar flux reduction is prescribed to be exactly 1.5 times as large as the imposed atmospheric heating. We believe that the  $R \approx -0.9$  experiment is the most realistic case.

This paper is organized into six sections. In section 2, we describe the numerical experiments, and assess statistical significance of the results. The difference between the "control" run (without the INDOEX aerosol heating) and each experiment is presented in sections 3–5. In section 3, the focus is on surface and air temperature and boundary layer stability over land. The precipitation features are described in section 4. In section 5, we provide the possible dynamical mechanisms for the precipitation change pattern. A summary and discussion follow in section 6.

### 2. Numerical experiment design

Throughout this paper,  $F_A$  denotes the solar heating rate change in the atmosphere due to anthropogenic aerosols, and  $F_s$  denotes the net solar flux change at the surface. The three experiments conducted have the same prescribed  $F_A$ , and only differ in the ratio R, that is, in  $F_s$ . In prescribing  $F_A$ , we use a vertically uniform profile from surface to 700 hPa and zero above it. R01 estimated  $F_A$  for the period January-March 1999, and in their estimate,  $F_A$  peaks over the Indian subcontinent and asymptotes to zero south of the ITCZ. Figure 2b depicts the pattern we have adopted. The vertical structure of  $F_A$  was also an issue of consideration in our study. During the INDOEX field campaign, two distinctively different profiles were observed: (i) aerosols below cloud and (ii) aerosols mostly above low-level cloud. The former profile (briefly "boundary layer profile") occurs roughly 1/3 of the time, and the latter ("elevated profile") 2/3 of the time (R01). Podgorny and Ramanathan (2001) calculated the atmospheric solar heating rate corresponding to each of the two, and showed that the heating in case of elevated profile peaks at slightly above 3 km, and that in the case of boundary layer profile it peaks at slightly above 1 km. Taken together, a vertically uniform profile from the surface to 700-hPa height is a reasonable approximation for an average profile.

The  $R \approx -0.6$  and  $R \approx -0.9$  experiments have a geographically and temporally varying *R*, largely because of varying cloudiness and surface albedo, and have the *R* values of roughly -0.6 and -0.9, respectively, if an area mean is taken over land. See Table 1 for the summary of all the experiments.

Figure 2 displays the daily mean of the prescribed forcing on the 29th Julian model day. Figure 2a shows the imposed heating rate in units of K day<sup>-1</sup> at the second lowest model level, and Fig. 2b shows the column-integrated  $F_A$  in units of W m<sup>-2</sup>. Figure 2c displays  $F_S$  for the R = -1.5 experiment. The CCM3 has a



FIG. 7. Land surface temperature change for the  $R \approx -0.9$  experiment. Superimposed are the streamlines for surface wind (wind at the lowest model level) change. Shading and contouring as in Fig. 6.

diurnal cycle of varying solar insolation, and the perturbation heating was also applied with a diurnal cycle. The forcing for the other months is adjusted for its magnitude, and Fig. 2d shows the time dependence of the imposed forcing adjustment. Forcing is applied every winter season repeatedly during the model integration. The daily and seasonal mean (January–March) aerosol radiative forcing in our study is largely consistent with estimates given in R01.

The atmospheric component of CCM3 was documented in Kiehl et al. (1996), and the land component in Bonan (1996). The atmospheric component of CCM3 we have used is a spectral model with a triangular truncation at wavenumber 42 (T42) and with 18 hybrid sigma pressure vertical levels. The 3D variables in the model outputs were linearly interpolated in  $\ln(p)$  to the following pressure levels: 925, 850, 775, 700, 600, 500, 400, 300, 250, 200, 150, 100, 70, 50, 30, and 10 hPa. The land surface component is much more advanced than the traditional "bucket" model, and essentially combines realistic radiative, ecological, and hydrologic processes (Bonan 1996). The model is run with a seasonally varying climatological SST, as generated with the observed 1950–79 SSTs.

The control CCM3 had 85 yr of simulated fields avail-

able, which we use for assessing the impact of the imposed aerosol forcing. We generated 60 yr of the  $R \approx -0.9$  experiment, 100 yr of the  $R \approx -0.6$  experiment, and 28 yr of the R = -1.5 experiment. The  $R \approx -0.9$  experiment also includes the archival of daily maximum and minimum temperature for 15 model years. All these runs were made with the same climatological SSTs and with the same land surface model. A lengthy integration was undertaken to ensure the statistical significance of the results presented here, relative to model internal variability. The variability internal to CCM3's atmosphere and land components is essentially the sum of pure atmospheric internal variability (i.e., unforced by boundary forcing change) and land surface forced variability.

To quantify the amplitude of CCM3's internal variability (i.e., variability not forced by SST change), we computed the precipitation anomaly and surface temperature anomaly relative to climatological seasonal cycle using 85 yr of control run, and then computed the standard error of the 85-yr mean (Fig. 3). The surface temperature standard error at each grid point in January– February (Fig. 3a) is less than 0.2 K in India. Our calculation of standard error assumes randomness of anomaly in each model year. We checked the power spectra



FIG. 8. The 775-hPa temperature change for the  $R \approx -0.9$  experiment. Shading and contouring are as in Fig. 6.

of the surface temperature anomaly and precipitation anomaly at a few selected points in our domain, and fortunately found them to have almost white spectrum features over periods greater than 1 yr (not shown). One hundred years of the  $R \approx -0.6$  experiment is practically comparable to 100 ensemble runs of yearlong integrations.

# **3.** Land surface and lower-troposphere temperature changes

Figure 4a displays the land area mean daily maximum and minimum surface temperature change in the  $R \approx$ -0.9 experiment, and Fig. 4b shows the daily mean change in each of the three experiments. Recall that the magnitude of the imposed negative surface forcing is smallest in the  $R \approx -0.6$  experiment and largest in the R = -1.5 experiment. Figure 4 illustrates that all the experiments produce surface cooling and the cooling amplitude increases quite linearly with -R. The peak surface cooling is about -0.5 K for the  $R \approx -0.9$ experiment (and -0.3 and -1.2 K for the other two experiments) and occurs during February. The land area mean of  $F_s$  is about -13 W m<sup>-2</sup> for the  $R \approx -0.9$ experiment from February to April. If the surface were to respond only through blackbody radiative emission, the corresponding surface temperature change would have been about -1.8 K. The actual changes were a few times smaller, due to the negative feedback effects of land-air coupling, discussed next.

In order to understand the processes that govern the magnitude of the surface cooling, we compare in Fig. 5 the area-averaged low-level temperature changes. Above the 1.5-km (P < 850 hPa) layer, the atmosphere warms in all the experiments (Fig. 5). Since the imposed atmospheric forcing  $F_A$  was the same for all the experiments, this indicates that the atmospheric temperature change is more dependent on atmospheric forcing than the surface forcing or surface temperature. This is not surprising, since general subsidence prevails over the subcontinent during the winter monsoon and radiativeconvective coupling between the surface and the atmosphere is weak, permitting the surface to cool and the atmosphere (z > 1.5 km) to warm. The near-surface temperature change (Fig. 5a) is simply a transition between the surface cooling and the atmospheric warming. The lower-atmospheric warming (Figs. 5c,d) is largest in April, in phase with the prescribed aerosol forcing (Fig. 2d), while the surface cooling is maximized in February. The surface cooling maximum precedes the atmospheric warming peak, perhaps because the wintertime has a more stable boundary layer and thus en-



FIG. 9. Precipitation change for the  $R \approx -0.9$  experiment: (a) averaged for the land and ocean areas both, and (b) for the land area, where  $F_A > 3$  W m<sup>-2</sup>. (c) The 60°-100°E mean precipitation change as a function of lat for each of the three experiments. In (c), the long dashed line (R = -1.5 expt), the short dashed ( $R \approx -0.6$ ), and solid line ( $R \approx -0.9$ ) are used. (d) The Feb-Apr averaged CCM3 precipitation climatology to compare to (c).

ables the surface to respond more to  $F_s$  than the springtime when the surface is climatologically warmer (Fig. 4c) and the boundary layer mixing is more active.

Reverting back to Figs. 4–5, it is seen that the nearsurface layer becomes more stable (decreasing lapse rate). As a result, the sensible heat flux from the surface decreases significantly (by 10 W m<sup>-2</sup> in March) and this decrease balances most of the solar flux reduction in the surface energy budget. The remaining balance of the  $F_s$ is compensated for by the decreased surface evaporation (about 3–4 W m<sup>-2</sup>) and a decrease in net (up minus down) longwave radiation (about 2–3 W m<sup>-2</sup>). Over the ocean regions, however, the imposed solar flux decrease is largely balanced by a decrease in evaporation (about 8 W m<sup>-2</sup> or 10% in January). The evaporation decrease over the ocean was determined largely by wind speed reductions. Figure 4a also shows the daily maximum and minimum temperature change in the  $R \approx -0.9$  experiment. Daily maximum and minimum temperatures both decrease, with a larger amplitude for daily maximum temperature. Thus, when the land boundary layer undergoes active mixing in the afternoon, such mixing will be significantly suppressed. In summary, the aerosol forcing significantly alters the air–surface interactions and the surface energy and moisture budget.

Figure 6 shows the surface temperature change pattern for each of the three experiments during February-April, and Fig. 7 depicts the evolutionary change from November to June for the  $R \approx -0.9$  experiment. In Figs. 6a-c, an ellipselike contour is superimposed to denote the area where  $F_A$  in Fig. 2b exceeds 3 W m<sup>-2</sup>. Statistically significant features outside of this contour must be driven dynamically by changes in mean circulation and transients, whereas features inside the contour reflect the direct impact of the aerosol radiative forcing. All the simulations produce cooling over most of South and Southeast Asia. In the  $R \approx -0.9$  and R  $\approx -0.6$  experiments (where *R* varies geographically), the maximum cooling occurs over the Indian subcontinent. The cooling for the  $R \approx -0.6$  experiment is smaller by a factor of about 1.5–2 (cf. Fig. 6a with Fig. 6b). In the R = -1.5 experiment, the temperature change is up to about -2.3 K.

The temperature change for  $R \approx -0.9$  experiment is described in more detail in Fig. 7, along with superimposed surface streamlines (depicting wind change) since temperature advection plays a role in driving the changes. The surface cooling peaks at about -1 to -1.5K during February, March, and April as would be expected from the seasonal variation of the imposed aerosol forcing (see Fig. 2d). The temperature advection anomaly seems to explain a large part of the geographical nonuniformity of the temperature change; for example, the little change over Southeast Asia in January– February is associated with the advection of warm maritime air. Note that there is a cyclonic feature around India. Section 5 discusses the dynamics underlying the large-scale circulation change.

The temperature change at 775 hPa for the  $R \approx -0.9$  experiment is shown in Fig. 8. The warming pattern, with a maximum over southern India and decreasing equatorward, is quite similar to the imposed heating profile (see Fig. 2). There is a significant warming northwest of the aerosol-forcing domain, due in large part to temperature advection. The overall atmospheric warming pattern opposes the northward decrease of climatological temperatures between the equator and 20°N, which has implications for the monsoonal circulation and precipitation as discussed next.

### 4. Precipitation

We will begin with the area-average precipitation changes shown in Fig. 9. As before, the averaging was



FIG. 10. Precipitation change pattern for the  $R \approx -0.9$  experiment. (a)–(d) Contour interval is 0.5 mm day<sup>-1</sup>, and values greater (less) than 1 contour interval are shaded dark (light).

taken over the area where  $F_A$  in Fig. 2b exceeds 3 W m<sup>-2</sup>. The area-average precipitation increases over both land and ocean, except during June and July. In terms of percentage (not shown), the increase (land and ocean together) is as high as 15% during April. The peak increase occurs during April (again land and ocean together as in Fig. 9a), in phase with the imposed forcing field (Fig. 2d) and the low-level temperature change field (Figs. 5c,d); whereas over land the peak occurs during the transition monsoon (between the NE and SW monsoon) period of May. We also note the intriguing feature of the decrease in June and July when the imposed aerosol forcing is zero. However, in terms of percentage the reduction is only about 1%-3% (for land as well as land plus ocean), compared with 6%-15% during April and May. Further discussion of the summer monsoon changes is deferred to the last section of this paper. As shown in the next section, the dynamical causes for the changes are linked to the spatial gradients in the imposed atmospheric forcing field, largely independent of surface cooling. Figure 9c shows that the precipitation change is indeed quite insensitive to the ratio *R* and depends mainly on the imposed atmospheric solar heating that is the same for all three experiments.

Recall that the evaporation from the surface decreased

over both the ocean and the land. Furthermore, the areaaverage precipitation increased over the haze region. This implies that the area-average precipitation outside the haze area must decrease to satisfy moisture and energy budgets. Indeed, the tropical average decrease exceeded 1% in February. Figure 10 shows the regional patterns of the changes for precipitation from November to June. The spatial pattern of the changes is somewhat related to the dominant modes of internal variability. The overall response has two key characteristics: A northward migration of the ITCZ during the wintertime bringing more rain north of the equator; and the suppression of precipitation over the Indonesian region and the Indo-Pacific warm pool extending from the eastern tropical Indian Ocean to the western tropical Pacific. Most of these features are found to be due to variation in the convective precipitation.

By April, extending into May and June, a larger area in the northern Indian Ocean undergoes a precipitation increase. The magnitude of the change is striking; over some areas the precipitation increase is more than 50% of CCM3's climatological value. Figure 11 shows the change in units of percentage, and this figure reveals more clearly regions of enhanced and suppressed precipitation. One region of the decreased precipitation is southwest



FIG. 11. Precipitation change pattern for the  $R \approx -0.9$  experiment in units of percentage.

Asia, which is a climatologically arid region with the wet season occurring during wintertime. The results over this region need to be further studied with more realistic regional models in order to make more accurate assessments. CCM3 simulates well the climatological precipitation in this area during winter. For example, the CCM3 precipitation averaged over  $25^{\circ}$ -40°N and  $50^{\circ}$ -65°E is 1.0 mm day<sup>-1</sup> during January–February, to be compared to the observed (Xie and Arkin 1997) precipitation of 1.1 mm day<sup>-1</sup>. The CCM3 precipitation during March–April is 1.2 mm day<sup>-1</sup>, and the observed precipitation is 1.0 mm day<sup>-1</sup> during the same period. The change over this area ( $25^{\circ}$ -40°N and  $50^{\circ}$ -65°E) is -0.29 mm day<sup>-1</sup> or about 30% in January–February.

By comparing with CCM3 precipitation climatology, we infer that these large changes are more due to shifting of the convection regime and less due to local amplification. The suppressed precipitation in the western equatorial Pacific (Figs. 9a,b) seems to result from a weakening of the Indian Walker circulation cell.

## 5. Dynamical linkages leading to the precipitation changes

The CCM3 responds to the aerosol forcing as follows: 1) low-level aerosol forcing warms the lower atmosphere; 2) low-level convergence moves toward the warmer low-level atmosphere, which fuels deep convection therein; 3) deep convection leads to an increase in deep diabatic heating; 4) the enhancement in deep heating drives further increases in the low-level convergence. The low-level convergence plays a prominent role in our proposed mechanism since it provides the moisture to sustain the precipitating clouds. The latent heat release is the dominating component of deep diabatic heating in the Tropics. Therefore, the low-level convergence enhances the deep heating. On the other hand, deep heating generates the low-level convergence and thus the low-level circulation (e.g., see Gill 1980),<sup>1</sup> thus completing a positive feedback loop between lowlevel convergence and deep heating. A quantitative analysis is given below to support the loop described above.

We first show the major changes in the deep heating field and the accompanying temperature changes. Figure 12 shows an altitude–longitude cross section along the  $0^{\circ}$ – $10^{\circ}$ N latitude belt for changes in temperature (Fig. 12a); sum of latent, sensible, and longwave radiative heating components (denoted as "diabatic without so-

<sup>&</sup>lt;sup>1</sup>Zhang and Krishnamurti (1996) demonstrated that the gross features of the tropical circulation are well simulated with Gill's dynamical model forced by the vertically integrated heating.



FIG. 12. Equatorial  $(0^{\circ}-10^{\circ}N)$  change cross section during Mar–Apr for the  $R \approx -0.9$  experiment: (a) temperature, (b) diabatic heating without solar heating rate, (c)  $-\omega$  (i.e., -dp/dt) and (d) solar heating rate. The y axis denotes pressure height in units of hPa. Values greater (less) than 1 contour interval are shaded dark (light).

lar" in Fig. 12b); the vertical pressure velocity field denoted by "-omega" in Fig. 12c; and the solar heating field (Fig. 12d). In spite of the fact that the imposed aerosol solar heating was confined to layers below 700 hPa, the entire troposphere is subject to a warming. This deep warming is due to the increase in diabatic heating shown in Fig. 12b. The increase in large-scale latent heating contributes most to the increases shown in Fig. 12b. The increase in the large-scale vertical velocity links the increase in latent heating to the increase in low-level convergence. This is because low-level convergence in the tropical latitudes will give rise to largescale lifting of air masses. The discussion thus far sets the stage for linking the aerosol forcing to low-level convergence, which is taken up next.

Aerosol forcing is linked to low-level convergence through a simple dynamical model proposed by Lindzen and Nigam (1987). Lindzen and Nigam (1987) linked the boundary layer temperature gradient to the low-level convergence. The Lindzen–Nigam model (hereafter LN model) is known to simulate the surface wind flow quite well in the central and eastern Pacific (Wang and Li 1993), and we find the LN model to explain well the low-level convergence pattern in the Indian Ocean in the context of CCM3.

First, focusing on the January-February period, we start with the time average field at  $60^{\circ}$ -70°E longitude. Figure 13a shows the 925-hPa temperature from the CCM3 control run, and Fig. 13b shows the simulated convergence at 925 hPa by CCM3. We apply the LN model to the temperature in Fig. 13a. The convergence predicted by this model is plotted in Fig. 13c. In the application of the simple model, the coefficients are as in Wang and Li (1993), except that the zonal Rayleigh coefficient is smaller than the meridional one following Deser's (1993) suggestion and 150 hPa is used as PBL thickness. The LN dynamics predicts within a few degrees the location of the maxima (at 5°N as opposed to 3°N by CCM3) and the minima (around 7°S). However, the convergence magnitude from the LN dynamics is rather small, because the LN model does not include the positive feedback between low-level convergence and deep heating. This positive feedback would deepen the trough and the ridge in the LN-predicted convergence. Furthermore we show the precipitation in Fig. 13b, which supports the contention that precipitation



FIG. 13. The 60°–70°E average: (a) 925-hPa temperature and (b) 925-hPa convergence and precipitation, from Jan to Feb CCM3 control run climatology, and (c) convergence calculated by LN model with temperature of (a). (d) Temperature and (e) 925-hPa convergence and precipitation change with  $R \approx -0.9$  experiment, and (f) the computer convergence change using the LN model with 925-hPa temperature change in (d). Note that the convergence calculated by the LN model in (c) and (f) is one order of magnitude smaller than that from the CCM3 run in (b) and (e).



FIG. 14. Area mean precipitation climatology from the CCM3 control run (dashed line) and from CPC merged analysis of precipitation (Xie and Arkin 1997).

(both location and intensity) is related to the location and the magnitude of the low-level convergence.

In order to test our hypothesis that  $F_A$ 's modification of the boundary layer temperature explains the precipitation change patterns (Fig. 10), the temperature change at 925 hPa (Fig. 13d) was extracted from the  $R \approx -0.9$ experiment, and was inserted into the LN model. The resulting latitudinal pattern of convergence change (Fig. 13f) is again in excellent agreement with the values extracted from the CCM3. Again, the magnitude is much smaller than the CCM3 convergence change. As in Fig. 13b, the changes in the precipitation are similar to the pattern of changes in the low-level convergence. In summary, the large-scale precipitation changes can be understood (qualitatively) from the changes in the boundary layer temperature gradients resulting from the aerosol forcing. From this, we would see that  $F_A$  widens the warmest low-level area northward (thus increasing area mean precipitation in Fig. 9a) since the imposed aerosol forcing in the atmosphere opposes the temperature decrease from the ITCZ. Note by comparing Figs. 13b and 13e, there is a northward shift of the convergence and precipitation maxima. This corroborates our earlier statement that the ITCZ shifts northward.

There is, however, one feature in the simulated changes that is particularly related to CCM3 deficiencies. Figure 9 showed that the maximum precipitation increase occurs in April and May. It is our speculation that this is most likely related to CCM3's tendency for triggering the SW monsoon earlier than observed. For example we compare the CCM3 simulated monthly precipitation with the observed climatology by Xie and Arkin (1997) in Fig. 14. The model climatological precipitation is generally larger than the observation, with the largest differences occurring in April and May. It is likely that the increased low-level convergence from the aerosol forcing accentuates this deficiency further.

Last, Fig. 12b shows the deep heating anomalies induced by INDOEX aerosols in the  $R \approx -0.9$  experiment. The magnitude of the haze-induced atmospheric heating stands out not only in the low levels but also in the deep atmosphere. Temperature changes (Fig. 12a) reveal a heterogeneous pattern with a peak warming around the 750-hPa level with lower warmings aloft and in the boundary layer. Furthermore, the warming has a minimum at the 600-hPa level. This interesting bimodal structure of the vertical temperature changes is reminiscent of Nigam et al.'s (2000) finding that ENSOcovariant temperature has a bimodal structure. One possible reason is that the temperature change induced by diabatic heating is reduced by the vertical motion associated adiabatic cooling, and is almost completely cancelled by the upward motion at 600 hPa-the midtropospheric layer. Since this issue is not the main thrust of this paper, we do not pursue this with more detailed diagnostic analysis.

### 6. Summary and discussion

We have sought to understand the regional climate change due to the late 1990s boreal wintertime south Asian haze. The spatial extent and the radiative forcing of these man-made aerosols have been documented recently during the Indian Ocean Experiment (INDOEX) field campaign (R01). The aerosol layer, largely confined between the 1000- and 700-hPa layers, contains as much as 10% (of fine particle mass) highly absorbing black carbon and covers a large region extending over most of the North Indian Ocean, and the land areas in South and Southeast Asia. The absorbing aerosol heats the low-level atmosphere by as much as 0.8 K day<sup>-1</sup> on a monthly mean basis, while both absorbing and scattering aerosols in concert reduce the net solar flux at the surface by as much as 20-40 W m<sup>-2</sup>.

The NCAR CCM3 was employed to simulate the regional climate change due to the low-level diabatic heating increase and the net surface radiative flux reduction. We have also explored the sensitivity of the climate change to a different absorption efficiency. Three numerical experiments were conducted and these differ only in the magnitude of the reduction of the surface radiation. The ratio *R* was defined as the surface aerosol forcing (a negative quantity) divided by the atmospheric forcing (a positive quantity). The  $R \approx -0.9$  experiment has an effective spatial and time average *R* of -0.9 over land, while the other two experiments have effectively -0.6, and exactly -1.5 for R over land. CCM3 coupled with an interactive land surface model was forced with a prescribed seasonal cycle of SSTs.

The principal findings are listed below.

- The February–April average land surface temperature decreases by about 1.5, 0.4, and 0.2 K, respectively, for the R = -1.5, -0.9, and -0.6 experiments. Air temperature increases throughout the low-level column are independent of the *R* value, except for the near-surface air changes that strongly depend upon *R*. The atmospheric warming above the near surface is nearly insensitive to the surface forcing, and the magnitude is about  $0.5 \sim 0.8$  K. The surface cooling is largest in February, and the low-level warming peaks in April.
- The aerosol forcing led to a surprisingly large dynamical response, which was insensitive to *R*. As we illustrated, the forcing weakened the north-south temperature gradient in the boundary layer north of the ITCZ, which in turn, enhanced the low-level convergence toward the warmer haze layer and moved the ITCZ northward. The increase in low-level convergence led to an increased convective rainfall and a latent heat release, which in turn led to a further increase in low-level convergence. This "positive feedback" between convergence and deep convective heating led to a large increase in precipitation over the haze area, accompanied by a decrease elsewhere, which was responsible for the simulated drought in southwest Asia.
- The surface cooling and atmospheric warming also results in suppressed evaporation from the ocean. Because of this decreased evaporation accompanied by the increase in precipitation over the haze area, the precipitation over the rest of the Tropics decreased, with a large fraction of this decrease concentrated over the Indonesian and the western Pacific warm pool region.

It is difficult to validate these simulations with observations for several reasons. For one, our model does not account for other changes such as the increase in greenhouse gases and natural coupled variabilities such as ENSO. The model also does not allow for the response of ocean temperatures. Nevertheless, the surface temperature from NCEP–NCAR reanalyses (Kalnay et al. 1996) does reveal a substantial cooling trend over the Indian region during wintertime (not shown). In verifying the rainfall change simulation, its observed interannual variability overwhelms any possible trend. Kiehl et al.'s (2000) study of the effects of the haze, which accounts for SST response through a slab ocean model, produces surface cooling (over land) that is consistent with the changes shown here (for the R = -0.9 and -1.5 experiments). The slab ocean model, however, produces much larger shifts in the ITCZ. At this stage of our understanding, we should consider the present set of CCM studies as merely illustrative of the potential effects of the haze on the regional climate. A fully coupled ocean-atmosphere GCM with realistic prescription of time-dependent greenhouse gases and aerosols is needed for further progress in this important problem.

The focus of the present study was restricted only to the dry season. The model results show that the dry season aerosols can also cause a small (1%-3%) reduction in rainfall during June and July (Figs. 9a,b). It is, however, premature to place confidence in the June– July results, because the wet season aerosol effects were not included. The next step is to introduce year-round aerosol forcing to the model for studying the year-round monsoon response. This step is important, since we need to address the fundamental issue in this region, that is, the impact of the haze on the SW summer monsoon.

Regarding the domain of influence, the climatic effect may not necessarily be confined to the Indian sector. Through the modification of the Indian Walker circulation cell, as implied by our experiments, the Pacific variability including ENSO can also be impacted. The INDOEX aerosol climate forcing has significant implications to global climate, as it increases troposphere-wide temperature (see Fig. 12a). The temperature change in the Tropics affects the subtropical jet stream and, thus, potentially affects the entire Northern Hemisphere.

Acknowledgments. This study was funded by the National Science Foundation (Grant ATM-0201946). The authors would like to thank Dr. I. Podgorny for Fig. 1, and Drs. S. Nigam, E. DeWeaver, and J. Hack for helpful discussions.

All the numerical experiments were conducted locally with two Beowulf cluster computers. In the running of CCM3 effectively on Beowulf cluster computer, several computer specialists provided help, notably J. Rosinsky of NCAR.

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