Impact of absorbing aerosol on precipitation: Dynamic aspects in association with convective available potential energy and convective parameterization closure and dependence on aerosol heating profile

Chul Eddy Chung and Guang J. Zhang
Center for Atmospheric Sciences, Scripps Institution of Oceanography, La Jolla, California, USA
Received 2 March 2004; revised 2 July 2004; accepted 14 July 2004; published 24 November 2004.

[1] The Indian Ocean Experiment (INDOEX), conducted from 1995 to 2000 to document aerosols in south Asia during winter monsoon season, revealed the existence of a layer of highly absorbing aerosols in the lower troposphere. The observed aerosol has one of the two distinctly different vertical distributions: (1) aerosols concentrated in the planetary boundary layer (PBL) below 1.5 km and (2) elevated aerosol profile peaking around 3 km. Here we provide the dynamical basis for understanding the direct effects of absorbing aerosols on the large-scale precipitation and the role of the aerosol vertical distribution. This was done through a series of the south Asian aerosol experiments with the National Center for Atmospheric Research Community Climate Model (CCM3), together with different convection parameterization closures, and the Community Atmospheric Model (CAM2). It is found that the land surface temperature underneath the aerosol layer is sensitive to the aerosol vertical distribution: The lifted layer of aerosols results in a significant cooling of the underlying land surface, while the PBL profile makes very little cooling. The mechanism of the aerosol effect on precipitation distribution is investigated by examining the correspondence between the aerosol heating profile, changes of precipitation, and the atmospheric convective instability. The direct aerosol heating of the near-surface air increases the convective available potential energy (CAPE), whereas the heating above the boundary layer decreases CAPE. Meanwhile, the regionally concentrated low-level aerosol heating tends to cause large-scale rising motion over time, which increases CAPE by decreasing the midlevel temperature. The net CAPE change is small for the lifted profile (i.e., profile elevated above PBL) because the CAPE increase by the midlevel cooling is counteracted by the CAPE decrease through the direct haze heating above the PBL. The precipitation increase averaged over the aerosol area is much larger when the PBL profile is used than when the lifted profile is used in the CCM3 with a CAPE-based convective parameterization closure. The sensitivity of the aerosol effect to convective parameterization closure is tested using a new closure, which is based on the environmental contribution to CAPE (CAPEe). It is shown that when this closure is used in CCM3, the precipitation increase averaged over the aerosol area is small regardless of the vertical profile. This is because the direct heating of either profile decreases CAPEe, opposing the CAPEe increase by the midlevel cooling.

INDEX TERMS:
0305 Atmospheric Composition and Structure: Aerosols and particles (0345, 4801); 1854 Hydrology: Precipitation (3345); 3314 Meteorology and Atmospheric Dynamics: Convective processes; 3337 Meteorology and Atmospheric Dynamics: Numerical modeling and data assimilation; 3374 Meteorology and Atmospheric Dynamics: Tropical meteorology; KEYWORDS: aerosol, precipitation, dynamics


1. Introduction

[2] The effect of aerosol on climate is a complex scientific problem. Most aerosols are short lived in the atmosphere and have spotty geographic distribution, regionally concentrated around the sources. While all aerosols reduce the solar radiation at the surface, carbonaceous aerosols and dust particles also absorb solar radiation and heat the atmosphere. On top of this, aerosols act as cloud condensation nuclei, and thus can have significant indirect climate effect through interactions with clouds.

[3] On regional scales, the aerosol radiative forcing can overwhelm the greenhouse gas radiative forcing. The Indian Ocean Experiment (INDOEX), conducted from 1995 to
2000, documented the characteristics, spatial extent, vertical distribution and temporal fluctuations of the south Asian aerosol using multiplatform observations [Ramanathan et al., 2001a]. It was found that a brown aerosol layer spread over most of the North Indian Ocean and the south/Southeast Asian continent during the dry season. The brownish color came from black carbon richness, which resulted in the single scattering albedo of about 0.9 both over land and in the open ocean. Black carbon, owing to its highly absorbing power, enhances the reduction of surface solar radiation due to sulfates, nitrates, and other scattering aerosols by a factor of 2 to 3 over the south Asian aerosol area [Satheesh and Ramanathan, 2000]. This absorbing aerosol directly reduces the surface net solar flux by as much as 20 W m\(^{-2}\) to 40 W m\(^{-2}\) on a monthly mean basis and heats the lowest 3 km atmosphere by as much as 0.4 to 0.8 K d. This gives a 50 to 100% enhancement of the solar heating in this layer [Ramanathan et al., 2001a]. In comparison, the greenhouse gas forcing is only +1.6 W m\(^{-2}\) in the atmosphere and +1.0 W m\(^{-2}\) at the surface over this area [Ramanathan et al., 2001b].

[4] Regionally concentrated absorbing aerosols alter the gradient of the absorbed solar radiation significantly, and are expected to impact the winter monsoon. Chung et al. [2002] estimated the effects of the south Asian aerosol on the surface temperature and regional rainfall using a climate model. A similar study by Menon et al. [2002] linked the black carbon aerosols to the regional trends in floods and droughts in China. These modeling studies were conducted using simplified vertical profiles of aerosol forcing. During INDOEX [Ramanathan et al., 2001a], two types of the aerosol vertical profiles were typically observed as shown in Figures 1a–1b: (1) aerosols heavily concentrated below 1.5 km (hereafter referred to as PBL profile; see Figure 1a) and (2) aerosols peaking around 3 km (hereafter referred to as lifted profile; see Figure 1b). Varying aerosol vertical profiles are observed elsewhere over the globe as well. During the Aerosols99 cruise between Virginia, USA and Cape Town, South Africa, aerosols were observed to peak below 1 km or above 2 km and extend to 4 km [Voss et al., 2001]. During the Aerosol Recirculation and Rainfall Experiment (ARREX 1999) and Southern African Regional Science Initiative (SAFARI 2000) dry season experiments, both elevated aerosol layer above the PBL and near-surface aerosol layer were observed by lidar in northeastern South Africa [Campbell et al., 2003]. In eastern China, the dust particles are found to extend to several kilometers and peak at varying heights from one observation to another [Zhou et al., 2002].

[5] Podgorny and Ramanathan [2001] demonstrated that the height of the aerosol concentration maximum significantly affects the aerosol radiative forcing at the TOA and the surface and the vertically integrated atmospheric forcing when low-level clouds are present. Questions arise. What are the effects of the vertical profiles of the aerosol forcing on the underlying surface temperature and large-scale precipitation pattern? Since precipitation in the tropics is mostly of convective nature, how are these effects related to convective representation in climate modeling studies? Current aerosol assimilation models have difficulties to simulate the elevated aerosol distribution like the profiles shown in Figure 1b. How does an inaccurate representation of the aerosol atmospheric forcing profile affect the accuracy of the model results? Understanding the dynamics of the aerosol forcing—temperature and precipitation linkage would provide useful insight to these questions.

[6] This study explores the relationship between aerosol profile, precipitation changes and the atmospheric convective instability, and uses the relationship to address the above question. It provides a dynamic basis for understanding the direct radiative effect of absorbing aerosols on large-scale precipitation by linking them to fundamental aspects of convective parameterization. Specifically, we examine the correspondence between aerosol forcing profiles, changes of precipitation and convective available potential energy (CAPE) in the atmosphere and model convection closure schemes. The investigation is conducted by prescribing aerosol forcing over south Asia and the northern Indian Ocean on the basis of the highly absorbing haze data from INDOEX.

2. Data and Climate Models

2.1. Aerosol Forcing

[7] The aerosol radiative forcing is defined as the effect of aerosol, both natural and anthropogenic, on the radiative fluxes at the top of the atmosphere (TOA), at the surface and on the absorption of solar (and longwave in the case of dust and sea salt particles) radiation within the atmosphere. The aerosol effect can be included in a model in at least two ways. As with Menon et al. [2002], one could specify the aerosol optical properties including the single scattering albedo, and let the model radiation code generate the aerosol radiative forcing. Chung et al. [2002] and Chung and Ramanathan [2003], on the other hand, obtained the atmospheric and surface aerosol forcing on the basis of INDOEX observations and directly inserted the forcing into a climate model. The first approach has been widely used but is not guaranteed to generate an accurate aerosol radiative forcing due to approximated radiation algorithms and deficient low-level cloud simulation. After a detailed comparison of the two approaches, Chung et al. [2002] took the latter approach in their study. Here we adopt their approach and insert into the climate models an idealized, yet realistic aerosol forcing derived from the observations.

[8] As shown in Figures 1a and 1b, distinctly different vertical profiles of aerosol forcing are observed in INDOEX. To incorporate such aerosol profiles in our modeling study, we use an idealized, vertically uniform structure to mimic the observed vertical distributions. In each numerical experiment, we prescribe one of the three idealized atmospheric aerosol heating profiles as shown in Figure 1c. The prescribed aerosol forcing profile has the same vertical structure at all the grid points in the south Asian haze area in each experiment. The heating rate shown here represents that from the grid point of maximum heating (its horizontal variation is shown in Figure 2a). The magnitude of the heating rate in our estimate is based on Podgorny and Ramanathan’s [2001] study. Their study incorporated the INDOEX observations (especially, aerosol optical depth, single scattering albedo, and vertical profile) and ran the MACR (Monte Carlo Aerosol Cloud Radiation) model. In Figure 1c, one profile has constant atmospheric heating from the surface to 850 hPa and zero above (PBL profile, corresponding to Figure 1a), another has constant heating from 800 hPa to 670 hPa and zero elsewhere.
(lifted profile, corresponding to Figure 1b), and the third profile has constant heating from the surface to 700 hPa (uniform profile). The uniform profile is identical to the “Shrunk Haze Forcing” used by Chung and Ramanathan [2003]. For each experiment, the prescribed surface aerosol forcing is equal to the reduction of solar radiation through absorption and scattering due to the presence of aerosol, and is set equal to the vertically integrated atmospheric forcing multiplied by a factor, which is $1.5$ in the case of clear sky and increases with cloudiness, as with Chung and Ramanathan [2003] for the Shrunk Haze Forcing. This multiplication factor varies with time and space as a function of the model predicted cloudiness, and is based on the INDOEX observations [e.g., Satheesh and Ramanathan, 2000] and Monte Carlo Aerosol Cloud Radiation model calculations of the south Asian haze forcing under various cloudiness conditions by Podgorny and Ramanathan [2001].

The aerosol forcing in the atmosphere and at the surface was prescribed in the numerical experiments. Figure 2a shows the geographic distribution of the prescribed atmospheric forcing for the uniform profile experiment in units of K/day. The same geographical distribution holds for other experiments as well since the specified forcings differ only in the vertical structure. The horizontal pattern and the heating rates are identical to the Shrunk Haze Forcing of Chung and Ramanathan [2003], and are idealized from the satellite observations of Li and Ramanathan [2002]. Figures 2b–2d show the vertically integrated atmospheric forcing in the three experiments. Since the thickness of the heating layer for the uniform profile is about twice as large as the other two cases, the vertically integrated forcing (in W/m²) for the uniform profile is also roughly twice as large. Similarly, the surface forcing in this case is approximately twice as large as in the other two experiments as well.

The seasonal evolution of the forcing is the same as that of Chung et al. [2002] and Chung and Ramanathan [2003]: from June to September there is no aerosol forcing and from October to May, the prescribed aerosol forcing gradually increases to a maximum in March and then decreases to zero by the end of May. The models were integrated with the climatological seasonal cycle of SSTs, generated from the observed 1950–79 SSTs, and the interactive land surface model. We conducted 42 model years of integration for the uniform forcing experiment, and contrasted its mean with the mean from 85 years of the control run without the aerosol forcing. For all the other experiments, the model was integrated for about 22–23 years. We believe that averaging the output over 20 model years or more produces statistically significant results.
in assessing the south Asian aerosol effects on regional temperature and rainfall.

2.2. Community Climate Model (CCM3) and Convection Closures

The Community Climate Model (CCM3) is used for most of the numerical experiments in this study. The atmospheric component of CCM3 was documented by Kiehl et al. [1996], and the land component of Bonan [1996]. The atmospheric component of CCM3 is a spectral model with a triangular truncation at wave number 42 (T42) and with 18 hybrid sigma pressure vertical levels. Over the ocean, 6 layers are located below 690 hPa.

Using CCM3, Chung et al. [2002] showed that the aerosol forcing has a strong impact on the tropical rainfall distribution. Since tropical precipitation is of convective nature, interaction between aerosol forcing and convective parameterization deserves further investigation. CCM3 uses the convective parameterization of Zhang and McFarlane [1995]. The scheme assumes that when the atmosphere is convectively unstable, convection responds to remove the excess convective available potential energy (CAPE) at an exponential rate with a relaxation time of about 2 hours. This type of CAPE-based closure is used in many GCMs. Several variants of the Arakawa-Schubert scheme [Arakawa and Schubert, 1974] also used such a closure [e.g., Moorthi and Suarez, 1992; Sud and Walker, 1999]. Recently, Zhang [2002, 2003] found that the generation of CAPE by the tropospheric large-scale processes above the boundary layer is well balanced by the free tropospheric contribution to the removal of CAPE by convection, and proposed that such a balance of CAPE generation and consumption can be used as a closure for convection. The contribution to CAPE from the large-scale processes above the boundary layer is referred to as the environmental contribution to CAPE (CAPEe) in this paper, since to a parcel lifted from within the boundary layer the processes above the layer happen in its “environment”. Both the original Zhang and McFarlane [1995] scheme and the revised scheme with this new closure are used in this study.

2.3. Community Atmospheric Model (CAM2)

The Community Atmosphere Model (CAM2) is a new version of the National Center for Atmospheric Research
NCAR model, updated from CCM3. Like CCM3, CAM2 is also a spectral model at T42 resolution in the horizontal, but it has 26 hybrid sigma pressure vertical levels. Both CCM3 and CAM2 use the Zhang and McFarlane [1995] convection scheme. There are also important changes. CAM2 includes a more advanced land surface model. The major upgrades from CCM3 in model physics are a prognostic parameterization of condensed cloud water [Rasch and Kristjánsson, 1998], inclusion of evaporation of convective precipitation, a new thermodynamic package for sea ice, an explicit representation of fractional land and sea ice coverage, a new treatment of geometrical cloud overlap in the radiation calculation, and a new parameterization for the longwave absorptivity and emissivity of water vapor (W. D. Collins et al., Description of the NCAR Community Atmosphere Model (CAM2), NCAR technical report, 171 pp., 2002, available at http://www.ccsm.ucar.edu/models/atm-cam/). As mentioned earlier, CCM3 is used for most of the experiments in this study. CAM2 are used only to demonstrate that similar results and conclusions to those from CCM3 are expected in CAM2 as well.

3. Surface Responses

Figure 3 displays the January–March land surface temperature changes in response to each of the three aerosol forcing profiles. Recall that the sea surface temperature is prescribed in all of the numerical experiments in our study. Thus aerosols have no materialized effect on the ocean surface. Since the south Asian aerosol forcing is prescribed from October to May during each of the integration years, using fixed SSTs is not a bad approximation except for April–August period. Our analysis of the preliminary coupled experiments shows that SST changes by the south Asian aerosol are slow and significant from April to July/August. In the current study employing uncoupled models, we look at the January–March mean only. The aerosol impacts before January are too small to be of any further interest to us.

As Figure 3 shows, the PBL profile decreases the underlying surface temperature very marginally while the uniform profile and lifted profile drive similar amplitude of cooling. The land surface temperature change averaged over India is −0.78 K (uniform profile), −0.23 K (PBL profile), and −0.81 K (lifted profile). It is interesting to note that the lifted profile produces as much cooling as the uniform profile despite the fact that the surface solar radiation reduction in the uniform profile is about twice as large as in the lifted profile. Clearly, the lifted profile is far more effective in reducing the underlying surface temperature than is the PBL profile.

The reason for the effective surface cooling by the lifted profile is straightforward: it is the result of the balance between the near-surface warming effect and the surface cooling effect. The reduction of surface solar radiation due to the aerosol effect cools the land surface. On the other hand, the heating of the atmosphere by the absorbing aerosols warms the air in the layer where the heating takes place. The heat exchange between the land surface and the atmosphere above it through boundary layer mixing and surface turbulent fluxes reduces both the surface radiative cooling and the atmospheric warming. Comparing the cases of uniform profile with PBL profile, the atmospheric radiative heating rate within the PBL is about the same while the reduction of the surface solar radiation is about half in the PBL profile as in the uniform profile case. As a result, the extra reduction of solar radiation on the surface in the uniform profile case leads to larger decrease of surface temperature compared to the PBL profile. The lifted profile has the radiative warming above the PBL, thus feedbacks through turbulent fluxes to counteract the surface radiative cooling are very small, leading to large cooling of the land surface even though the solar radiation reduction is only
about half of that in the uniform profile case. Chung et al. (2002) also examined the surface cooling by the south Asian haze. In their investigation, the uniform profile of the atmosphere heating and different strengths of the surface forcing were used to demonstrate the sensitivity of the surface temperature to the ratio of the atmospheric heating rate to the surface forcing magnitude.

4. Precipitation Change

4.1. Salient Features

We begin with the precipitation changes due to the uniform aerosol profile in CCM3, CCM3 with the new convection closure and CAM2, as shown in Figure 4. Again, all the panels pertain to the January–March mean as in section 3A. Over the aerosol forcing area denoted by a dashed line in each panel, the CCM3 and CAM2 precipitation changes (Figures 4a and 4c) appear qualitatively similar in comparison with the changes in CCM3 with the new closure (Figure 4b). The precipitation increases in CCM3 and CAM2 are located over the area of maximum aerosol forcing. It is important to note that CCM3 and CAM2 have similar precipitation climatology over the aerosol area during this season (not shown) and both use CAPE-based convection closure. On the other hand, the precipitation increase in the CCM3 with the new convection closure has a much smaller area coverage, although the general patterns of increasing precipitation in the north and decreasing precipitation in the south are similar in all the three runs. The precipitation change averaged over the aerosol area is 0.293 and 0.078 mm/d for CCM3, and CCM3 with the new closure, respectively.

Figure 5 shows the CCM3 precipitation changes due to the effects of the PBL and lifted profiles. The effect of the PBL profile (Figure 5a) is similar to that of the uniform profile (Figure 4a), while the precipitation increases resulting from the lifted aerosol profile are much weaker and their maximum is located far to the south. The average precipitation change over the aerosol area is 0.236 mm/d for the PBL profile and 0.078 mm/d for the lifted profile. Recall that the lifted profile was as effective in cooling the underlying surface as the uniform profile in section 3A. Conversely, the PBL profile increases the precipitation as much as the uniform profile. As far as the aerosol area averaged precipitation change is concerned, the uniform and PBL profiles in CCM3, and the uniform profile in CAM2 prevail over the uniform profile in CCM3 with the new closure or the lifted profile in the CCM3.

4.2. Dynamics Behind the Changes

As shown above, precipitation changes due to aerosol forcing are sensitive to the aerosol profile and convective parameterization closure. The role of aerosol forcing is to modify the thermodynamic structure of the atmosphere, while the role of convective parameterization is to determine how convection, thus precipitation, responds to such thermodynamic changes. In CAPE-based closures, convection is determined by the amount of CAPE in the atmosphere. In a convectively unstable atmosphere (i.e., when CAPE is greater than zero), CAPE increases when the temperature or moisture in the boundary layer increases, or the upper layer temperature decreases. In a stable atmosphere (i.e., when CAPE is zero), changes of temperature or moisture do not affect CAPE unless the changes are large enough to make the atmosphere unstable. CAPE is nonnegative and is set to zero if the atmosphere is stable. Thus changes of CAPE depend not only on the temperature perturbations, say, resulting from the aerosol forcing, but also on the basic thermodynamic state of the atmosphere. Keeping this in mind, we first examine the direct response of CAPE to aerosol forcing in Figure 6. In calculating the CAPE changes in Figures 6b–6e, the imposed aerosol
heating in the atmosphere in units of K/d was taken as the low-level temperature increase in units of K, and this temperature perturbation was added to the temperature field of the mean state from the model simulation to compute the resultant CAPE change. A focus on this immediate CAPE change by the aerosol heating is justified, because other processes that affect the CAPE are at longer timescales. One of such other processes is the land surface cooling, which leads to near-surface air cooling through turbulent fluxes and then affects the CAPE. Another is the large-scale rising motion due to the aerosol heating over a large area, the dynamics of which is that proposed by Lindzen and Nigam [1987]. In the actual model integration, all these processes play a role in changing CAPE and producing the precipitation changes shown in Figures 4–5. However, we first focus on the immediate and direct CAPE response to aerosol forcing.

[20] The most salient feature of Figure 6 is that the uniform and PBL profiles in CCM3 increase the CAPE significantly (Figures 6b–6c) while the lifted profile actually decreases the CAPE, though with roughly one-third of the magnitude (Figure 6d). The CAPE increase from the uniform profile in the CAM2 (Figure 6e) is comparable to that in CCM3. The reason for a negative CAPE change by the lifted profile is that the positive temperature perturbation due to warming above the boundary layer affects the environmental temperature, but not the parcel’s temperature since the parcel is lifted from the boundary layer. When the atmosphere is warmer above the boundary layer, the lifted parcel gains less buoyancy and the negative CAPE change reflects this. Note that the positive CAPE changes in Figures 6b, 6c, and 6e are much larger than the negative CAPE change in Figure 6d. This is because CAPE fluctua-

Figure 5. January–March precipitation change in CCM3 by (a) PBL profile aerosol and (b) lifted profile.

Figure 6. Low-level aerosol heating and CAPE change by it. (a) Imposed aerosol forcing at the second model lowest level. (b)–(e) CAPE changes by the low-level aerosol heating. (f) Environmental CAPE (CAPEe) change by the uniform profile aerosol forcing. In calculating the CAPE/CAPEe change, the aerosol heating in units of K/d was plugged in as a temperature change for a day. Specifically, the temperature changes from the aerosol heating and the background model climatology were both hourly diurnal cycles averaged over February in the CAPE/CAPEe change computation. Hourly treatment was needed in view of the nonlinear nature of CAPE.
tions are predominantly controlled by changes in the thermodynamic properties of the boundary layer air [McBride and Frank, 1999; Zhang, 2002; Donner and Phillips, 2003]. These studies show that over 90% of the observed variations of CAPE in both the tropics and midlatitudes can be explained by the boundary layer equivalent potential temperature changes. In Figures 6b, 6c, and 6e, aerosol heating directly affects the boundary layer air whereas in Figure 6d it does not. Another noticeable feature of Figure 6 is that the horizontal pattern of CAPE change does not exactly follow that of the aerosol heating. The CAPE change reaches maximum around 5°–10°N (Figures 6b–6e) while the heating is largest from 15° to 0°N (Figure 6a). This is because the basic state of the model atmosphere north of 15°N is convectively stable and becomes unstable equatorward.

Figure 6f shows the environmental contribution to CAPE change (CAPEe), instead of the total CAPE change, for the uniform profile and the CCM3 with the new convection closure. Similar to Figure 6d, since the aerosol heating warms the environment above the boundary layer, the environmental contribution to CAPE change is negative. On the other hand, unlike Figure 6d, which is also the total CAPE change (the parcel’s contribution to CAPE change in this case is zero), the parcel’s contribution (not shown) is positive because of warming of the near-surface air. However, this part of the total CAPE change is retained in the atmosphere and does not directly enter the convective parameterization with the new closure, as the new closure only considers the balance of CAPE changes due to the large-scale processes and convection above the boundary layer.

[22] A closer comparison between the immediate CAPE change by the aerosol heating and the precipitation change is made in Figure 7 to deepen insight. In Figure 7, the zonal average over 85°–90°E of precipitation change is displayed (in dashed line) together with the CAPE change averaged over the same longitude band (in solid line with solid circle markers). In the cases of the uniform and PBL profiles in CCM3 when the CAPE increase due to aerosol heating is large (Figures 7a and 7b), the precipitation maximum is located where the CAPE increase is largest. This suggests that the precipitation changes in the full model integration can be largely explained by the immediate CAPE response to aerosol heating. In the cases of the lifted profile in CCM3 and the uniform profile in CCM3 with the new closure (Figures 7c and 7d), the correspondence between the model precipitation changes and CAPE changes is less obvious. This implies that in addition to the immediate CAPE response, other processes induced indirectly by the aerosol forcing are also important in leading to the simulated precipitation changes.

[23] Such processes include the large-scale rising motion induced by the low-level aerosol heating, surface cooling from reduced solar heating, changes of clouds and possibly others. However, aerosol indirect effect on cloud micro-
physics is not considered in this study. Thus any changes in condensational heating related to cloud changes must be from secondary responses of clouds to the dynamics and thermodynamic changes. Surface cooling can affect the boundary layer temperature through turbulent heat exchanges. This is true over land. However, since the atmosphere is already stable over most of the land areas, further stabilization by cooling the near-surface air does not seem to have much effect on convection and precipitation; note that most of the precipitation changes shown in Figure 7 are south of 20°N over the Indian Ocean and the Bay of Bengal. Over the oceans, since our experiments use the “atmosphere only” model with prescribed SSTs, the surface cooling mechanism is not materialized. Thus the most likely additional process that needs to be considered is the low-level convergence, and the resulting rising motion, in association with the aerosol heating. A large-scale rising motion decreases the free tropospheric temperature and increases the CAPE. The relevant dynamics starts with the fact that the low-level heating would introduce a horizontal gradient of the low-level temperature and low-level thickness. Lindzen and Nigam [1987] suggest that temperature gradients in the PBL can result in pressure gradients hydrostatically. Such pressure gradients generate winds within the PBL, leading to mass convergence in the warm area. The Lindzen-Nigam model equation is:

\[
\varepsilon u - fv = -\frac{\partial \phi}{\partial x} + \frac{gH_0 \partial T}{2T_0 \partial x}, \quad \varepsilon v + fu = -\frac{\partial \phi}{\partial y} + \frac{gH_0 \partial T}{2T_0 \partial y}
\]

\[
\mu u + 2H_0 \frac{\partial H_0}{\partial x} \left( \frac{\partial u}{\partial x} + \frac{\partial v}{\partial y} \right) = 0,
\]

where \(H_0\) is average layer thickness, \(\phi\) is layer top height, \(H\) is density scale height, \(T\) is layer mean temperature, \(\mu\) is the Newtonian cooling coefficient, and \(\varepsilon\) is the Rayleigh friction coefficient.

[24] The Lindzen-Nigam mechanism well explains the boundary layer wind features in the central/eastern Pacific from the boundary temperature gradient [Wang and Li, 1993]. Similar arguments can be applied to the low-level temperature gradients created by the aerosol heating in our study. Here the aerosol heating, regardless of its vertical structure, will create a mass convergence in the forcing layer and a rising motion above it. Figure 8 shows the anomalous divergence produced by the lifted profile through the Lindzen-Nigam dynamics. In solving the Lindzen-Nigam dynamics equation [Lindzen and Nigam, 1987; Wang and Li, 1993] for the lifted profile, the layer thickness was set to 130 hPa, and the Newtonian cooling coefficient was set to 10^{-3} s^{-1}. As Figure 8 shows, the lifted profile creates a large-scale convergence centered around the aerosol forcing maximum. The upward motion associated with this large-scale convergence will cool the layers in the free atmosphere and increase CAPE, compensating for the immediate CAPE decrease by the aerosol heating. The application of the Lindzen-Nigam dynamics to the PBL profiles yields very similar results. This may explain why the precipitation increased over a significant portion of the aerosol area (Figures 5b and 7c) despite the estimated immediate CAPE decrease in response to the lifted profile (Figure 6d) that should decrease the precipitation. In the case of the PBL profile, the positive immediate CAPE change is amplified by the CAPE increase through the midlayer cooling, leading to much bigger precipitation increases (Figure 5a).

[25] The precipitation changes outside the forcing area seem to be a large-scale adjustment of the atmosphere to the perturbed convection within the forcing area. The adjustment would depend largely on the model climate.

5. Discussion

[26] In this study, we have attempted to explain how the absorbing aerosol forcing impacts the underlying surface temperature and the large-scale precipitation, and how this impact depends on the aerosol vertical profile and the model convective parameterization. The aerosol forcing in our numerical experiments was derived from the INDOEX observation, which revealed two typical vertical profiles. Our study demonstrated a strong efficacy of the lifted profile for cooling the surface over the PBL profile. On the contrary, the precipitation is increased a few times more by the PBL profile than by the lifted profile in CCM3 when CAPE-based convection closure is used.

[27] We also explored the aerosol effects on precipitation in CCM3 using a new convection closure, which is based on the balance between CAPE generation and consumption above the boundary layer. With this new closure, the precipitation increase is insignificant regardless of the aerosol vertical structure. This is because the low-level warming of any structure leads to an immediate decrease in CAPE and opposes the CAPE increase from upward motion dynamically induced by the low-level warming. Clearly, the aerosol effect is highly sensitive to the vertical distribution of aerosols and to convective parameterization in the model.

[28] The surface temperature ties in with the precipitation through its effect on the near-surface temperature and
evaporation from the surface. For example, cooling the surface reduces the evaporation from the surface and therefore slows down the hydrological cycle. This evaporation feedback is expected to be on much longer timescales (~several months), and will likely be the dominant influence on the precipitation in summer when most aerosols have been washed away by the monsoon rainfall and yet the underlying surface (especially ocean) is still cold because of the accumulative cooling effects of the dry season aerosols. In this slow process way, the lifted profile may be more effective than the PBL profile in changing the rainfall. This study limited the analysis to the dry season precipitation.

[29] The validation of the precipitation change simulation is very challenging and beyond the scope of this study. First of all, we only coped with the South Asian aerosol effects, neglecting the aerosol elsewhere. Second, the precipitation changes, as simulated here, are mainly over the ocean for which a long-term precipitation product does not exist. Last, in reality there have been many other climate forcings and changes, such as El Niño-like Pacific warming [Trenberth and Hoar, 1996], and the greenhouse gas increase. These other factors obscure the effects of absorbing aerosols. We are planning on a future study to validate the absorbing aerosol effects on the precipitation as more observations become available.

[30] Acknowledgments. We are deeply indebted to V. Ramanathan for his interest in the work and for his valuable comments. This work was supported by NSF grants ATM-0201946 to V. Ramanathan (CEC) and ATM-0204798 (GIZ). The reviewers’ constructive comments have helped improve the manuscript greatly.

References


Zhang, G. J., and N. A. McFarlane (1995), Sensitivity of climate simulations to the parameterization of cumulus convection in the Canadian Climate Centre general circulation model, Atmos. Ocean, 33, 407–446.