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Simulation of aerosol distributions and radiative forcing for INDOEX: Regional climate impacts

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[1] The direct radiative forcing by aerosols over the Indian Ocean region is simulated for the Indian Ocean Experiment (INDOEX) Intensive Field Phase during Spring 1999. The forcing is calculated for the top-of-atmosphere (TOA), surface, and atmosphere by differencing shortwave fluxes computed with and without aerosols. The calculation includes the effects of sea-salt, sulfate, carbonaceous, and soil-dust aerosols. The aerosol distributions are obtained from a global aerosol simulation including assimilation of satellite retrievals of aerosol optical thickness (AOT). The time-dependent, three-dimensional aerosol distributions are derived with a chemical transport model driven with meteorological analyses for this period. The surface albedos are obtained from a land-surface model forced with an identical meteorological analysis and satellite-derived rainfall and insolation. These calculations are consistent with in situ observations of the surface insolation over the central Indian Ocean and with satellite measurements of the reflected shortwave radiation. The calculations show that the surface insolation under clear skies is reduced by as much as 40 W/m² over the Indian subcontinent by natural and anthropogenic aerosols. This reduction in insolation is accompanied by an increase in shortwave flux absorbed in the atmosphere by 25 W/m². The inclusion of clouds in the calculations changes the direct effect by less than 2 W/m² over the Indian subcontinent, although the reduction is much larger over China. The magnitude of the difference between all-sky and clear-sky forcing is quite sensitive to the three-dimensional spatial relationship between the aerosol and cloud fields, and other estimates of the difference for the INDOEX Intensive Field Phase are as large as 5 W/m². INDEX TERMS: 0305 Atmospheric Composition and Structure: Aerosols and particles (0345, 4801); 0345 Atmospheric Composition and Structure: Pollution—urban and regional (0305); 0368 Atmospheric Composition and Structure: Troposphere—constituent transport and chemistry; 3337 Meteorology and Atmospheric Dynamics: Numerical modeling and data assimilation


1. Introduction

[2] The Indian Ocean region includes some of the most densely populated countries on Earth. The pollution from fossil-fuel and biomass burning in these countries is already having a significant influence on the regional atmospheric chemistry. The anthropogenic aerosols released from this region are projected to become the dominant component of anthropogenic aerosols worldwide in the next 25 years [Nakicenovic and Swart, 2000, available at http://www.grida.no/climate/ipcc/emission/index.htm]. The Indian Ocean Experiment (INDOEX) is designed to measure the characteristics, distribution, and radiative effects of these aerosols. Measurements have been collected during the north-east winter monsoon each year from 1996–1999, with the main deployment of instrumentation during the Intensive Field Phase (IFP) in 1999. Several years of chemical, meteorological, and radiative data are available from the Kaashidhoo Climate Observatory (KCO, at 4.965°N, 73.466°E). In this paper, we use a model which is consistent with the field observations to derive...
The direct aerosol radiative forcing over the INDOEX region.

[1] The radiative forcing by aerosols over the Indian Ocean and Arabian Sea has been examined in several previous aerosol studies. Although there are numerous global forcing estimates for sulfate, carbonaceous, or soil-dust aerosols [e.g., Penner et al., 1992; Kiehl and Briegleb, 1993; Haywood and Ramaswamy, 1998], we summarize the studies that have examined forcing specifically in the Indian Ocean region. The first detailed calculations of the aerosol effects use data collected during the First GARP Global Experiment (FGGE) and summer Monsoon Experiment (MONEX) [Ackerman and Cox, 1982; Ellington and Serafino, 1984; Ackerman and Cox, 1987]. These studies focus exclusively on the forcing by soil-dust aerosols from the Saudi Arabian peninsula and Sahara Desert. The results show that the dust aerosols enhance the shortwave heating rates by up to 50% in the lower troposphere over some areas of the Arabian Sea [Ellington and Serafino, 1984; Ackerman and Cox, 1987]. The climatic effects of aerosols released by the Kuwait oil fires during the 1991 Gulf War have also been extensively modeled [Bakan et al., 1991; Ferretti and Visconti, 1993; Kao et al., 1994]. These are some of the first detailed calculations of the regional impact of anthropogenic aerosols, although of course the source of the aerosols was a transient phenomenon.

[2] The effects of anthropogenic aerosols on the climate of the Indian Ocean region has been modeled for individual aerosol species. Various studies have examined the impacts of sulfate aerosols on the Indian summer monsoon [Lal et al., 1995; Boucher et al., 1998], continental diurnal temperature cycle [Lal et al., 1996], and continental surface insolation [Venkataraman et al., 1999]. The surface radiative forcing by carbonaceous aerosols has been estimated by Reddy and Venkataraman [1999]. In contrast to these analyses, the present study includes the radiative forcing by all major natural and anthropogenic aerosol species. In addition, we focus specifically on characterizing the forcing and evaluating our estimates using observations from INDOEX.

[3] The most recent set of observations and modeling studies have been conducted as part of INDOEX [Ramanathan et al., 1996, available at http://www-indoex.ucsd.edu/publications/proposal]. The direct aerosol radiative forcing in the region has been rigorously characterized using surface observations [Jayaraman et al., 1998; Krishnamurti et al., 1998; Meywerk and Ramanathan, 1999; Conant, 2000; Jayaraman, 1999] combinations of surface and satellite data [Sathesh and Ramanathan, 2000; Ramanathan et al., 2001a], and Monte Carlo modeling [Podgorny et al., 2000]. These studies have demonstrated that the radiative forcing of anthropogenic aerosols extends over most of the Arabian Sea and Bay of Bengal and that the forcing is a significant term in the surface, atmospheric, and top-of-atmosphere (TOA) energy budgets. Additional experimental evidence and modeling of the forcing is discussed elsewhere in the INDOEX special issues.

[4] The present study should be viewed as an extension of the pioneering analysis by Ackerman and Cox [1987]. To the best of our knowledge, the work by Ackerman and Cox [1987] is the first calculation of the surface radiative energy budget for the Indian Ocean including realistic specification of land-surface albedos and the distributions, physical properties, and optical properties of aerosols and clouds. The enhancements in our calculation include time-dependent distributions of sea-salt, sulfate, carbonaceous, and soil-dust aerosols from a chemical transport model (section 2.1) and time-dependent surface albedos from a land-surface model (section 2.3). In contrast to Ackerman and Cox [1987], our calculations include the significant effects associated with anthropogenic aerosols. We also focus on the change in the surface and TOA fluxes (the “forcing”) associated with the introduction of these aerosols, while Ackerman and Cox [1987] address the magnitude and regional gradients in the total all-sky fluxes. They also focus on the summer monsoon, while we study winter conditions which allow for a larger anthropogenic effect. Most of the INDOEX observations have been collected from instruments on ships, islands, and coastal regions of India. The bulk radiative properties of the aerosols and the variation of aerosol forcing with aerosol optical depth from the model are compared to these in situ data. After evaluation of our radiative calculations against INDOEX observations, we calculate the forcing over the interiors of continental regions bounding the Indian Ocean. Our calculations therefore complement the INDOEX data over oceans and can be combined with these data to yield continuous maps of aerosol forcing over land and ocean [Ramanathan et al., 2001a].

[5] Two estimates of the forcing by aerosols are discussed. The forcing is first computed for a cloud-free atmosphere (section 3.1). The aerosol forcing is then evaluated including the radiative effects of clouds (section 3.2). As previous studies [Liao and Seinfeld, 1998; Ramanathan et al., 2001a] have shown, the aerosol radiative forcing under all-sky conditions can be quite different than the forcing under clear-sky conditions. These estimates of the forcing are compared with other calculations and observations for the region in the conclusion (section 4). Specifically, our estimate of the difference between the all-sky and clear-sky aerosol forcing is approximately 40% of the difference calculated by Ramanathan et al. [2001a]. The reasons for the disparity are examined in the conclusion.

2. Description of the Model

[6] The direct aerosol radiative forcing has been calculated over the Indian Ocean and adjacent continental regions between 40°–120°E and 30°S–30°N for January through March 1999. The forcing is defined as the difference in the net shortwave fluxes and shortwave heating rates computed with and without the effects of aerosols. This definition differs from that used by the Intergovernmental Panel on Climate Change (IPCC) [1995] since our forcing estimates include the effects of natural (nonanthropogenic) aerosols. The aerosol forcing has been calculated by applying a radiation model to the aerosol distributions simulated using a global aerosol assimilation model. The spectral land-surface albedos are obtained from a global land-surface model forced with a meteorological analysis and satellite-derived precipitation. Each of these components is described in greater detail below.

2.1. Aerosol Assimilation Model

[7] The model used to generate the time-dependent distributions of aerosols for input to the radiative transfer
calculations consists of a chemical transport model coupled to an assimilation package. This system is described in the work by Collins et al. [2001], and its simulation of the aerosol distributions during the INDOEX Intensive Field Phase is described in the work by Rasch et al. [2001]. The fidelity of the simulated long-range transport has assessed using aircraft data in the work by Clarke et al. [2001]. The basis of the system is the Model for Atmospheric Transport and ChEmistry (MATCH) [Rasch et al., 1997], a chemical transport model forced with meteorological fields from an external source. The aerosol species included in the model are sulfate, soil dust, black carbon (BC) and organic carbon (OC), and sea salt.

[10] In the current configuration of the model, the surface mass concentrations predicted by this diagnostic parameterization agree with measurements at the Maldives to within 10–15% [Rasch et al., 2001]. Since the sea salt contributes a small fraction of the direct radiative forcing (Section 3.1), the use of a more detailed sea-salt parameterization would not alter the conclusions of this study. The other aerosol species are computed from prognostic equations including surface sources, wet and dry deposition, and transport by resolved and subgrid-scale motion. The aerosols are assumed to be externally mixed. The dust is resolved in 4 size bins spanning 0.01–1 μm, 1–10 μm, 10–20 μm, and 20–50 μm, and each bin is represented by a lognormal distribution. Results from the model of Tegen and Lacis [1996], which includes multiple bins for submicron dust aerosols, show that characteristic lifetimes vary by only 5% over the submicron size range. Thus transporting all submicron mass together in one bin causes an insignificant bias compared to, for example, the uncertainty in the dust size distribution at the source. The other aerosols are represented by a single mass-mixing ratio for each species. The model saves the instantaneous aerosol mass-mixing ratios every 6 hours at the same horizontal and vertical resolution as the input meteorological fields.

[11] The formulation of the sulfur cycle is described in the works of Barth et al. [2000] and Rasch et al. [2000]. The emissions inventory for SO$_2$ is from Smith et al. [2001]. The sources for mineral dust are based upon the approach of Tegen and Fung [1994] with modifications to the soil mobilization to improve agreement with the Nimbus-7 absorptive-aerosol index from Herman et al. [1997]. The emissions of carbonaceous aerosols include contributions from biomass burning [Liousse et al., 1997], fossil fuel burning [Penner et al., 1993], and a source of natural organic aerosols resulting from terpene emissions (J. Penner, personal communication, 1999). During the INDOEX Intensive Field Phase, these data sets imply that the carbonaceous emissions for biomass burning, fossil fuels, and natural sources between 30N to 30S and 40E to 120E are 4.1×10$^7$, 1.4×10$^7$, and 2.3×10$^6$ kg/day, respectively. There the natural carbon particle emissions are approximately 4% of the total mass flux of carbonaceous aerosols.

[12] In this simulation we have adjusted the partitioning of the carbon emissions so that the ratio of BC to OC by mass is 1:2. The BC:OC ratio is very difficult to measure and is subject to large uncertainties [Huebert and Charlson, 2000]. The BC:OC ratio of 1:2 is intermediate between the ratios observed over the Indian subcontinent and the northern Indian Ocean by Novakov et al. [2000] during INDOEX. It is comparable to the ratios measured by Quinn et al. [2002] from the R/V Ron Brown and by Chowdhury et al. [2001] at KCO during the IFP. Our ratio is larger than the global-mean ratio of 1:7 to 1:10 used in other modeling studies [e.g., Liousse et al., 1996] but is in good agreement with the global-mean ratio of 2:5 obtained by Cooke et al. [1999]. The adoption of a single value is an oversimplification of both the geographical variation in the BC:OC ratio and the differences in the ratio for different combustion processes and fuels [Cooke et al., 1999]. In order to test the sensitivity of the results to this assumption, the calculations have been repeated with BC:OC ratios of 1:4 and 1:1.

[13] The assimilation is used to adjust the model for consistency with satellite-retrieved aerosol optical thicknesses (AOTs). The column-integrated AOTs over cloud-free ocean surfaces are obtained from AVHRR imagery recorded by NOAA polar-orbiting satellites. The retrievals are based upon the NOAA Pathfinder 2 aerosol algorithm [Stowe et al., 1997] and cloud-clearing procedure [Stowe et al., 1998]. In the algorithm, the aerosol is assumed to have optical characteristics similar to those of a marine aerosol [Stowe et al., 1997]. The assimilation is performed for each daytime satellite overpass at the model time step closest to the time of the overpass. The model AOTs are estimated from the aerosol mass-mixing ratios and relative humidities using an aerosol optics package appropriate for the 630 nm wavelength of the satellite retrievals. The mass-mixing ratios are adjusted by the assimilation procedure to minimize the differences between the modeled and retrieved AOTs using optimal interpolation [Lorenc, 1986; Levitt et al., 1998; Lamarque et al., 1999]. There are 14 satellite orbits per day, each of which yields a digital map of AOT extending roughly 1100 km in the east-west direction. These maps are interpolated onto the spatial grid of the aerosol model before assimilation. The retrievals from all AVHRR images collected during the ascending (afternoon) NOAA-14 orbits for January through March 1999 are input to the assimilation. The assimilation is equivalent to a source or sink of aerosol mass. Formally, the assimilation appears in the transport equations as an instantaneous source or sink term which is calculated after the other standard physical parameterizations for wet removal, dry deposition, etc. For the present model, the assimilation “source” is less than 20% of the physical sources of aerosol [Rasch et al., 2001]. The assimilation “sink” is less than any of the physical sinks.

[14] Like other aerosol retrieval schemes of its class, the errors in AOT from the NOAA Pathfinder algorithm can be large when the observed aerosol departs significantly from the properties adopted in the retrieval [Mishchenko and Travis, 1997]. Comparison of satellite retrievals based upon purely scattering optics with coincident estimates of AOT from surface sunphotometers in the INDOEX region indicates that these retrievals underestimate the AOT by approximately 30% [Rajeev et al., 2000]. The relative error is nearly constant during several weeks of observations. The NOAA Pathfinder retrieval is based upon optical properties for a remote marine aerosol [Stowe et al., 1997] with a size distribution similar to that used in the work of Rajeev et al., [2000]. Since the optics for maritime aerosols yield nearly...
estimates [1999. Our calculations of the direct aerosol forcing span [2001], we do not have the high-resolution data for January used to simulate the aerosol distributions for these forcing the chemical transport model with a single continuous different models, and since it is preferable to integrate
analysis and NCAR/NCEP reanalysis are based upon
the outgoing longwave flux is decreased by 2.6 W/m². The dry geometric radius is $r_g = 0.05 \mu m$ and the standard deviation is $\sigma_g = 2.03 \mu m$. The hygroscopic growth is computed using the Köhler curve, and the indices of refraction for the wetted sulfate aerosol are calculated as mass-weighted averages of the indices for pure H₂SO₄ and water. The optics for dust are the same as those used in the work of Collins et al. [2001]. The size distributions in each of the 4 dust size bins are lognormal with parameters given in Table 2. The index of refraction for dust is taken from measurements of Saharan dust [Patterson, 1981].

The optics for the sea-salt and carbonaceous aerosols are derived from the Optical Properties of Aerosols and Clouds (OPAC) data set [Hess et al., 1998]. Following Cooke et al. [1999], the OPAC optics for soot aerosols are used for black carbon, and the OPAC optics for water-soluble aerosols are used for organic carbon. OPAC treats soot as hydrophobic so the optics are not affected by ambient humidity. The optics for hydrophobic and hydrophilic organic carbon correspond to the optics for dry and wetted water-soluble aerosols, respectively. The optical properties depend upon the chemical and physical properties of the carbonaceous aerosols and can vary over a wide range [Liousse et al., 1996; Tegen et al., 1997]. The choice of optics for the carbon species is one of the largest sources of uncertainty in the forcing calculations. The sea-salt optics are treated using a mass-weighted combination of specific
extinctions for sub and super-micron sea salt from the OPAC data set. The corresponding single-scattering albedos \( \omega_0 \) and asymmetry parameters are calculated using standard formulae for combining optical properties of different atmospheric constituents [e.g., Cess, 1985]. The ratio of mass concentrations of super to submicron sea salt ranges from 6–16 [Haywood et al., 1999], and in our calculation we have assumed that the ratio is approximately 6. The resulting specific extinction at 70% relative humidity is intermediate between the values derived by Tegen et al. [1997] and Quinn and Coffman [1999]. The OPAC optics are interpolated onto the 19 spectral intervals used by the CRM using a high-resolution model calculation for the solar spectrum (W. Wiscombe, private communication, 2000). The hygroscopic growth of the extinction, asymmetry parameter, and single-scattering albedo in each spectral interval are fit as functions of relative humidity following Kiehl et al. [2000].

[20] The time series of the modeled column-average \( \omega_0 \) at the INDOEX KCO site during the INDOEX IFP is shown in Figure 3. The values of \( \omega_0 \) include all the aerosol species in the model and is evaluated at \( \lambda = 500 \) nm. The average of \( \omega_0 \) at KCO for the IFP is 0.877, which agrees with the average of estimates from surface and aircraft data to within ±0.02 [Satheesh and Ramanathan, 2000]. The experimental values include estimates from surface radio-

Figure 1. The total aerosol optical thickness from the chemical transport model averaged over January through March 1999. The AOT is estimated at a wavelength of 630 nm. Panel A: model run with assimilation of satellite AOT; B: difference in AOT from model with assimilation minus AOT from model without assimilation.
meters; in situ chemical and optical characterization from two ships, an aircraft, and fixed surface sites including KCO; multiwavelength lidar; and combinations of surface radiometers and Monte Carlo models [Ramanathan et al., 2001a]. There is a small trend toward decreasing values of $\omega_o$ during the experiment. During March 1999, the period of the largest concentrations of anthropogenic aerosols over KCO [Satheesh and Ramanathan, 2000], the daily mean $\omega_o$ is always less than 0.9. The only major aerosol species in the model with $\omega_o < 0.89$ at 500 nm is black carbon, and therefore the range of $\omega_o$ between 0.8 and 0.9 during March must result from a significant increase in the fraction of the total aerosol mass comprised of black carbon. In fact, the modeled mass of black carbon over KCO is 1.7 times larger in March compared to February. The change in the concentration of black carbon is coincident with a shift in large-scale flow regimes at the end of February [Rasch et al., 2001]. A map of $\omega_o$ averaged over the IFP is shown in Figure 4. The lowest values of $\omega_o$ occur over central India and south-east Asia, while the largest values occur over the southern Indian Ocean. The meridional gradient in $\omega_o$ is a consequence of the high concentrations of absorptive anthropogenic aerosols near regions with significant biomass burning and fossil-fuel consumption. In the 3-month mean, $\omega_o$ is less than 0.9 over the entire northern Indian Ocean, Arabian Sea, Bay of Bengal, and surrounding continental areas. As we will show in section 3.1, the low values of $\omega_o$ and large AOTs result in significant heating of the atmosphere over the northern Indian Ocean.

2.3. Land-Surface Model for Land-Surface Albedos

[21] Calculation of the direct radiative forcing over continental regions adjacent to the Indian Ocean requires

\[ \omega_o = \frac{S_{ATM}}{C_0 S_{0}} \]

Table 1. Daily Mean Shortwave Forcing Terms

<table>
<thead>
<tr>
<th>Symbol</th>
<th>Definition</th>
</tr>
</thead>
<tbody>
<tr>
<td>$S_{TOA}$</td>
<td>Forcing at TOA for clear-sky conditions</td>
</tr>
<tr>
<td>$S_k(TOA)$</td>
<td>Forcing at TOA for species $k$ (clear sky)</td>
</tr>
<tr>
<td>$S_o(TOA)$</td>
<td>Forcing at TOA with ocean albedos used everywhere</td>
</tr>
<tr>
<td>$S(0)$</td>
<td>Forcing at surface for clear-sky conditions</td>
</tr>
<tr>
<td>$S_k(0)$</td>
<td>Forcing at surface for species $k$ (clear sky)</td>
</tr>
<tr>
<td>$S_o(0)$</td>
<td>Forcing at surface with ocean albedos used everywhere</td>
</tr>
<tr>
<td>$S_{ATM}$</td>
<td>Forcing on atmosphere ($S_{TOA} - S(0)$)</td>
</tr>
<tr>
<td>$S_{ATM}$</td>
<td>Forcing on atmosphere with ocean albedos used everywhere</td>
</tr>
<tr>
<td>$Q$</td>
<td>Shortwave heating rate</td>
</tr>
<tr>
<td>$\tau$</td>
<td>AOT at $\lambda = 500$ nm</td>
</tr>
</tbody>
</table>

Table 2. Optical Properties for Soil-Dust Aerosols

<table>
<thead>
<tr>
<th>Size, $\mu$m</th>
<th>$r_m$ $^a$ $\mu$m</th>
<th>$\ln \sigma^b$</th>
<th>$\chi_c$ $^c$ m$^2$/g</th>
</tr>
</thead>
<tbody>
<tr>
<td>0.01 – 1</td>
<td>0.4</td>
<td>0.7884</td>
<td>1.303</td>
</tr>
<tr>
<td>1 – 10</td>
<td>2.0</td>
<td>0.7884</td>
<td>0.124</td>
</tr>
<tr>
<td>10 – 20</td>
<td>15.0</td>
<td>0.7884</td>
<td>0.040</td>
</tr>
<tr>
<td>20 – 50</td>
<td>25.0</td>
<td>0.7884</td>
<td>0.017</td>
</tr>
</tbody>
</table>

$^a$Geometric mean radius.
$^b$Width of log-normal distribution.
$^c$Optical extinction at $\lambda = 630$ nm.
estimates for the spectral surface albedos for diffuse and direct radiation. The CRM uses two visible albedos for diffuse and direct radiation at wavelengths less than 0.7 m\(\mu\)m and two near-infrared albedos for diffuse and direct radiation at wavelengths greater than 0.7 m\(\mu\)m. For ocean regions, the standard CCM albedo formula is used [Briegleb et al., 1986]. Albedos of land surfaces are much more heterogeneous, and unfortunately visible and near-infrared albedos were not measured directly over the region of interest during INDOEX. If measurements of TOA clear-sky fluxes are available, then these data can be used to estimate the broadband surface albedos with reasonable accuracy [Staylor and Wilber, 1990]. Additional assumptions are required to infer spectral land-surface albedos [Pinker and Laszlo, 1992]. However, only intermittent TOA flux data is available for 2 weeks during the INDOEX IFP. The spectral albedos for land surfaces have instead been calculated using the NCAR Land Surface Model (version 1) [Bonan, 1996]. The land-surface albedos are estimated from a radiative transfer code that treats the interaction of downwelling radiation with the vegetation canopy, bare soils, lakes, and rivers, and snow. LSM includes estimates of spectral albedos for several different types of bare soil, and it parameterizes the effects of soil moisture on these albedos. Seasonal changes in vegetation cover and type are obtained from an annually cyclic data set. The LSM calculates the land-surface albedos each hour at 1-degree resolution for January–March 1999.

[22] For this application, the LSM is integrated using a global, multiyear atmospheric forcing data set extracted from the NCAR/NCEP T62 meteorological reanalysis [Kalnay et al., 1996]. Satellite retrievals of precipitation [Xie and Arkin, 1997] and calculations of downwelling surface radiation based upon ISCCP cloud properties [Bishop et al., 1997] are used to correct known deficiencies in the reanalysis. The meteorological fields are time interpolated from the 6-hourly resolution of the reanalysis. The initial state of the LSM is the last instantaneous state of the model from an integration from 1979 through 1998 [Bonan et al., 2001]. For consistency with the solar geometry in CRM, the calculation of surface albedos is based upon the same ephemeris code [Michalsky, 1988; Spencer, 1989].

Figure 3. Time series of the simulated single-scattering albedo \(\omega_s\) of the aerosol in a 3 \(\times\) 3 grid-cell array centered on the INDOEX KCO site for January–March 1999. The column-average \(\omega_s\) is estimated at a wavelength of 500 nm. The averaging over model layers is weighted by the optical depth for each layer. Each point represents a daily mean for the 3 \(\times\) 3 array, and the vertical bars represent the spatial standard deviation within that array.

Figure 4. Simulated single-scattering albedo of aerosol at a wavelength of 500 nm for January–March 1999. Vertical averaging is weighted by the optical depth for each layer.
The shortwave surface albedos are shown in Figure 5. The albedos are derived from ratios of the corresponding time-mean downwelling and upwelling fluxes at each grid point in the model. The broadband albedo is less than 0.2 over the portions of the Indian subcontinent, south-east Asia, and eastern Africa included in the calculation. The only regions where the albedo exceeds 0.2 are over the Saudi Arabian peninsula and the snow covered regions of the Tibetan plateau and surrounding areas. Unfortunately it is not possible to evaluate the modeled surface albedos directly against observations. However, estimates of the surface albedo have been derived from measurements of the clear-sky TOA fluxes [Staylor and Wilber, 1990; Pinker and Laszlo, 1992]. Most of the estimates are based upon data from the NASA Earth Radiation Budget Experiment (ERBE) during 1985–1990 [Harrison et al., 1990]. The estimation of surface albedos from TOA fluxes requires a correction for the radiative effects of the atmosphere. In the INDOEX region, the correction process is complicated by the large aerosol loadings and heterogeneous aerosol composition (Figures 1A and 2). The shortwave albedo calculated with the LSM is compared against two estimates of the albedo from the NASA Surface Radiation Budget (SRB) data set for January–March 1986–1988 in Figure 6. The LSM albedos are larger than the Pinker and Laszlo [1992] values, but the albedos from LSM and Staylor and Wilber [1990] agree to within ±0.05 (Figure 6). Over southern China, the LSM albedos are larger than both empirical estimates. In general, the LSM is consistent with the SRB data to within the uncertainty in the techniques for deriving surface albedo from satellite fluxes. As we will demonstrate in section 3.1, errors in the LSM surface albedos are a secondary error term in the modeled direct aerosol forcing.

3. Forcing Calculations

3.1. Clear-Sky Radiative Forcing

The surface and satellite measurements collected during INDOEX can be used to evaluate the model calculation. The derivatives of the surface and TOA forcing with respect to AOT can be derived from simultaneous measurements of the fluxes and AOTs. In these estimates, the AOT is calculated at 500 nm. The modeled relationships between AOT and the surface and TOA forcing are shown in Figure 7. At TOA, the net flux decreases by \(-22 \text{ W/m}^2\) per unit increase in AOT, while at the surface, the net flux decreases by \(-73 \text{ W/m}^2\) per unit increase in AOT. These derivatives are only applicable over ocean surfaces for moderate values of the AOT. The corresponding derivatives have been calculated by Satheesh and Ramanathan [2000] from NASA CERES data at TOA and INDOEX instruments at the surface. The observationally derived derivatives are

\[
\frac{d S(\text{TOA})}{d\tau} = -25 \text{ W/m}^2
\]

\[
\frac{d S(0)}{d\tau} = -71.5 \pm 1.5 \text{ W/m}^2
\]

The modeled derivatives are consistent with the observed values to within 3 W/m² per unit optical depth, and our estimates are comparable to those derived from a radiative transfer model with \(\omega_o\) obtained from measured aerosol
optical properties [Satheesh and Ramanathan, 2000]. These results suggest that the model is consistent with the measured sensitivity of the TOA and surface forcing to changes in AOT. The sensitivity $dS(0)/d\tau$ of the surface forcing is approximately 3 times the sensitivity $dS(\text{TOA})/d\tau$ of the TOA forcing. This difference is characteristic of an aerosol which absorbs as well as scatters the incident sunlight. At TOA, the effects of absorption and scattering are subtractive, whereas at the surface the effects are additive. The result is that the surface forcing is much greater than the forcing at TOA for a given aerosol loading. The slopes for the two sensitivity experiments in which the BC:OC ratio is set to 1:4 and 1:1 are given in Table 3.

[25] The time-mean forcing for the TOA, surface, and atmosphere is shown in Figure 8. The correspondence between the forcing and AOT (Figure 1a) is particularly evident for $S(0)$ (Figure 8c). The regions with the largest absolute values of $S(\text{TOA})$ are located in the northern

Figure 6. Comparison of modeled and satellite-derived surface albedos for January–March averaged over 1986–1988. All data is mapped onto a 2.5 × 2.5 degree grid. Panel A: LSM calculation; B: surface albedo from Pinker and Laszlo [1992]; and C: surface albedo from Staylor and Wilber [1990].
The differences between the forcing computed with LSM and ocean albedos are shown in Figure 9. These differences are shown to demonstrate that the IFA-mean forcing calculations (Figure 8) over India and south-east Asia are not sensitive to large errors in the modeled surface albedo. With the exception of the deserts, positive values for the TOA and surface imply that the magnitude of the forcings are reduced when LSM albedos are used in place of ocean albedos. Positive values for the atmosphere indicate that the magnitude of the forcing is increased. The differences are identically 0 over ocean regions by definition and exceed 10 W/m$^2$ over the Saudi Arabian peninsula for the TOA and surface. In general, the largest relative changes in forcing associated with the land-sea albedo contrast occur in the TOA forcing. Over deserts, the absorption of shortwave radiation is increased by multiple reflections between the surface and dust aerosol. This process increases the average number of times the radiation traverses the layer of dust, which is moderately absorbing, and the absorptive gases and water vapor in the lower atmosphere.

The surface forcing by aerosols (Figure 8c) is $-35$ to $-40$ W/m$^2$ over central India, south-east Asia, and southern China. The maximum values near Myanmar are associated with the region of lowest $\omega_0$ (Figure 4). The forcing reduces the surface insolation by 15% over central India and by 17–20% over south-east Asia and southern China. For comparison, estimates of the climatological surface shortwave cloud forcing are plotted in Figure 10 [Staylor and Wilber, 1990; Pinker and Laszlo, 1992]. The aerosol surface forcing during the INDOEX IFP is more than twice the climatological surface cloud forcing over central India, Myanmar, and Thailand. The two forcings are similar in magnitude over the northern Arabian Sea and the Arabian Peninsula. The model results suggest that forcing by anthropogenic aerosols can exceed shortwave cloud forcing during the Indian winter monsoon.

The absorption by carbonaceous species results in atmospheric forcing that exceeds 25 W/m$^2$ over central India, south-east Asia, and southern China (Figure 8). The geographic patterns in $S$(ATM) are closely correlated with the AOT contributed by the black and organic carbon (Figure 2b). Since the shortwave heating rates are proportional to the shortwave absorption in the atmosphere, the changes in column-integrated heating caused by aerosols are proportional to $S$(ATM). The changes in the vertical profile of shortwave radiative heating rates are shown in Figure 11. The figure shows a meridional cross-section of the perturbations in $Q$ which includes the troposphere above the Maldives and the region around Bombay. The perturbations are calculated from averages of the daily mean heating rates during January through March 1999. The perturbations in $Q$ are less than 0.01 K/day south of 10$^\circ$S. In the northern hemisphere, the perturbation exceeds 0.7 K/day in the region around Bombay. This corresponds to an increase in the clear-sky heating rates in the lower troposphere by 130%. Approximately 90% of the increase in $Q$ is associated with the carbonaceous aerosols. Note that the model does not reproduce the double-layered vertical structure in aerosol profiles frequently observed during INDOEX [Ramanathan et al., 2001a]. As a result the modeled heating rates decrease monotonically with altitude.

Figure 7. Panel A: Variation in simulated TOA clear-sky net solar flux with 500 nm AOT in a 3 x 3 grid-cell array centered on the INDOEX KCO site. Each point represents a daily mean flux and AOT for February–March 1999. Solid line is the least-squares best fit to the model data. Panel B: Corresponding variation in simulated surface clear-sky net solar flux with 500 nm AOT.

Arabian Sea, the western coast of India downwind of Bombay, and the northern China Sea. There is a clear meridional gradient in $S$(TOA), with values ranging between $-3$ W/m$^2$ in the southern Indian Ocean to $-9$ W/m$^2$ in the northern ocean regions. There are also reductions in the forcing magnitude of 2 to 4 W/m$^2$ at the continental boundaries, and the forcing changes sign over the Arabian Peninsula. Over the desert regions, the introduction of dust aerosols increases the shortwave radiation absorbed by the land-atmosphere column. These large gradients in the forcing are associated with the discontinuities in surface albedo at the coastlines rather than discontinuities in the aerosol loading.

The effects of the land-surface albedos may be isolated by calculating the forcing using ocean albedos in place of LSM albedos over land surfaces. For an optically thin scattering aerosol, the TOA forcing should decrease quadratically with increasing surface albedo [Charlson et al., 1992].
The aerosol forcing can be decomposed into the forcings from individual species in order to assess the relative contributions of each aerosol type. To a good approximation, the total forcing is the sum of the forcings for each species. The relationship between the time-mean total forcing to the sum of forcings for carbon, dust, sulfate, and sea salt is shown in Figure 12. The difference between the total forcing at TOA and the sum of forcings for each species is 0.25 ± 0.32 W/m², and the corresponding difference at the surface is 0.30 ± 0.43 W/m².

The decomposition of the TOA forcing by species is shown in Figure 13. The largest forcing averaged over the INDOEX domain is contributed by the dust aerosols, and the smallest is contributed by sea salt. The fact that the forcing by dust is the largest component of the TOA forcing is related to the overestimation of dust concentrations in the model by factors of 2–3 [Rasch et al., 2001]. The forcings from sulfate and dust aerosols have opposite sign to the forcings from carbonaceous aerosols over all the land surfaces in the domain. The
The difference in sign can result in partial cancellation between the forcings by scattering and absorbing aerosols at TOA if the aerosols are co-located. The maximum values of the sulfate forcing occur over southern China, where the largest quantities of sulfate in the INDOEX region are generated by combustion of high-sulfur coal. The largest forcings on the atmosphere (Figure 14) and on the surface (Figure 15) are contributed by black and organic carbon. Over India, southeast Asia, and China, carbon contributes 85–90% of the atmospheric forcing and at least 60% of the surface forcing.

3.2. Direct Aerosol Forcing Including the Effects of Clouds

The direct aerosol forcing is affected by the presence of clouds. Clouds above the aerosols usually reduce the

Figure 11. Perturbation to clear-sky heating rates (in K/day) from introduction of aerosol averaged over 70–75°E. Heating rates are averaged over January–March 1999.
amount of sunlight interacting with the aerosols. High-altitude clouds also reduce the TOA aerosol forcing because of the nonlinear relationship between TOA albedo and total atmospheric optical depth. Clouds below the aerosols are equivalent to a highly reflective land surface and therefore reduce the TOA forcing (section 3.1). Since the relationship between transmission and total optical depth is also nonlinear, the low-altitude clouds generally reduce the surface forcing as well. Therefore the net result of including clouds in the calculations should be to decrease the direct aerosol radiative forcing at TOA and the surface. There can be additional effects from the indirect effects of aerosols on cloud radiative and physical properties [Ackerman et al., 2000], but we do not include the indirect effect in our calculations.

[32] The forcing including clouds is defined as the all-sky flux including aerosols minus the corresponding all-sky flux without aerosols. Based upon the spatial relationship between clouds and aerosols during the Indian winter monsoon, it is unlikely that the all-sky aerosol forcing could differ significantly from the clear-sky values. Long-term satellite and surface observations show that the amount of low-altitude clouds in the Bay of Bengal and the Arabian Sea is less than 30% during January to March [Bony et al., 2000]. The absence of low clouds is one of the reasons for choosing the winter monsoon season for INDOEX. Removal of continental aerosols by wet deposition in these areas reaches its minimum during the winter monsoon season.

[33] The calculation of all-sky aerosol forcing requires information on the three-dimensional distribution of condensed water and cloud amount. These fields are typically not available in operational meteorological analyses and are not included in the NCEP/NCAR reanalysis. The condensed water is derived from the reanalysis using “process reconstruction” [Rasch et al., 1997; Barth et al., 2000] using the prognostic cloud-water parameterization from CCM [Rasch and Kristjánsson, 1998]. The global-mean cloud liquid water predicted by this scheme in CCM is between 39–73% of the liquid water derived from satellite microwave data using several retrieval techniques [Rasch and Kristjánsson, 1998]. The cloud amount is diagnosed from the relative humidity field and other state variables using the standard cloud-amount parameterization from CCM [Hack, 1998]. The global and zonal-mean low and middle cloud amounts from the diagnostic parameterization fall within the ranges of the satellite and surface observations [Hack, 1998], although the high cloud amounts are 20% larger (absolute) than ISCCP retrievals. The distribution of low, middle, and high clouds and the total condensed water path derived using MATCH are shown in Figure 16. The maximum low and high cloud cover and condensed water are located at approximately 10°S, a region where the primary component of the AOT is sea salt (Figure 2). With the exception of southern China, the minimum cloud amounts and water paths occur between 10°N–30°N. The 6-hourly cloud and condensed water fields output by MATCH are interpolated to the hourly time resolution of the forcing calculations.

[34] The difference between the all-sky and clear-sky aerosol forcings are shown in Figure 17. The positive sign of the changes at TOA and the surface show that the magnitude of the aerosol forcing is reduced as expected. The changes over most of the INDOEX domain range between 0–4 W/m² and represent a small perturbation to the aerosol forcing in the northern Indian Ocean. The largest differences occur over southern China, where the forcing is reduced by as much as 20 W/m². This reduction is the consequence of 2 factors: the large sulfate loading in that region (Figure 2a), and the collocation of the sulfate and significant low cloud amount (Figure 16b). The absorption of radiation is reduced by up to 2 W/m² over the Indian subcontinent and Arabian peninsula, a change of less than 10% in the absorbed solar radiation. These calculations suggest that the effects of clouds on the direct forcing are a minor term in the insolation reaching land

**Figure 12.** Relationship of total shortwave forcing to sum of forcings by individual aerosol species. Panel A: TOA; and B: surface. The forcings are averages over January–March 1999 at each model grid point between 30°N–30°S, 40°E–120°E.

<table>
<thead>
<tr>
<th>BC/OC</th>
<th>dS(TOA)/dτ*</th>
<th>dS(0)/dτ*</th>
</tr>
</thead>
<tbody>
<tr>
<td>1:2</td>
<td>−22.2</td>
<td>−73.1</td>
</tr>
<tr>
<td>1:4</td>
<td>−26.6</td>
<td>−61.2</td>
</tr>
<tr>
<td>1:1</td>
<td>−17.3</td>
<td>−86.8</td>
</tr>
</tbody>
</table>

*Daily average flux (in W/m²) per unit AOT.
Figure 13. Decomposition of TOA shortwave radiative forcing shown in Figure 8 by aerosol species. Panel A: Sulfate; B: carbon; C: dust; and D: sea salt.

Figure 14. Decomposition of atmospheric shortwave radiative forcing shown in Figure 8 by aerosol species. Panel A: Sulfate; B: carbon; C: dust; and D: sea salt.
Figure 15. Decomposition of surface shortwave radiative forcing shown in Figure 8 by aerosol species. Panel A: Sulfate; B: carbon; C: dust; and D: sea salt.

Figure 16. Mean cloud properties used to compute effects of clouds on direct forcing during January–March 1999. Panel A: Low cloud amount (700 mb to surface, random overlap); B: midlevel cloud amount (400 mb to 700 mb); C: high cloud amount (10 mb to 400 mb); and D: grid-box averaged condensed water path (g/m²).
surface in the INDOEX region. Other studies [e.g., Ram-
anathan et al., 2001a] have obtained larger differences between the all-sky and clear-sky radiative forcing. The reasons for the disparity between the estimates are discussed in the conclusions.

4. Conclusions

[35] The direct aerosol radiative forcing has been computed for the INDOEX Intensive Field Phase during January through March 1999. The distribution of aerosols is obtained from an aerosol assimilation model, and the background atmospheric state is taken from the NCAR/NCEP meteorological analysis. The surface boundary conditions for the radiative calculations are obtained from a land-surface model forced with the same meteorological analysis and satellite retrievals of insolation and precipitation. This analysis extends the radiative budget in the Indian Ocean basin calculated by Ackerman and Cox [1987] to include the effects of anthropogenic aerosols.

Figure 17. Difference between direct aerosol forcing computed under all-sky minus clear-sky conditions for January–March 1999. Panel A: TOA; B: atmosphere; and C: surface.
[36] The radiative properties of the simulated aerosols are in good agreement with the in situ observations. The model reproduces the observed single-scattering albedo to within ±0.02, and it reproduces the measured sensitivities of the TOA and surface forcing to AOT to within 3 W/m² per unit \( \tau \). The model also simulates the chemical speciation of the aerosol measured at INDOEX surface sites to within 10–20% [Rasch et al., 2001]. Based upon the agreement with the observations, we have used the model to calculate the aerosol distributions and radiative effects over the Indian Ocean and surrounding continental regions.

[37] The modeled aerosols over India, southeast Asia, and southern China are composed primarily of sulfate and black and organic carbon. The combination of natural and anthropogenic aerosols reduce the surface insolation by up to 40 W/m² and increase the atmospheric shortwave absorption by up to 35 W/m². The aerosol surface forcing decreases the surface insolation by 15–20% over the continental areas and therefore significantly alters the surface energy budget. The increased shortwave absorption is accompanied by changes in the heating rate of up to 0.8 K/day. The maximum relative perturbation to the time-mean solar heating rates of up to 130% during the INDOEX IFP. Most of this perturbation is contributed by the effects of carbonaceous aerosols. The relative change in the heating rate from anthropogenic aerosols should be compared to the 50% change in \( Q \) calculated for the dust aerosols observed during MONEX [Ellingson and Serafinio, 1984; Ackerman and Cox, 1987]. The dynamical effects of this significant change in atmospheric heating are discussed in Ramanathan et al., [2001b].

[38] Using sensitivity studies, we have examined whether these results have a large dependence on the specification of surface albedos and on the presence of clouds. The largest uncertainties in the forcing related to surface albedos are associated with the desert regions in the Arabian peninsula. Over the Indian subcontinent and southeast Asia, the uncertainties in the forcing at TOA and surface are generally between 0–4 W/m² and represent a small relative error. The differences between the direct aerosol forcing computed with and without clouds are on the same order except over southern China. The direct forcing by aerosols over this region is reduced by approximately 50%, or 2 W/m², by the introduction of clouds. It is important to note that other calculations show that the IFA-mean difference between all-sky and clear-sky aerosol forcing is 5 ± 2 W/m² [Ramanathan et al., 2001a]. Extensive comparisons between the radiation codes used in the work of Ramanathan et al. [2001a] and this paper show that the instantaneous all-sky and clear-sky forcing agree to within 1–2 W/m² when the input cloud amounts, condensed water paths, and aerosol optical depths are identical. Further analysis has shown that the disparity in the estimates of all-sky aerosol forcing results from differences in the spatial relationships of the cloud and aerosol fields adopted in the work of Ramanathan et al. [2001a] and the present study.

[39] The large magnitude of the forcing by aerosols has several implications. First, the aerosol forcing is as large or larger than other heat sources and sinks, for example shortwave cloud forcing, which are currently included in the land-surface model and meteorological analysis system. Second, aerosol forcings of the magnitude observed during INDOEX could have a significant impact on the regional atmospheric and oceanic dynamics. Other studies using general circulation models [e.g., Ramanathan et al., 2001b] are underway to examine the atmospheric response. Studies of the coupled ocean-atmosphere response are needed to understand the effects on the regional climate system. Finally, it is clear that anthropogenic aerosols have significantly altered the radiative transfer over India and China. Projections of the aerosols released from these countries after further industrialization indicate that this region will be the dominant source of anthropogenic aerosols by 2025 [Nakicenovic and Swart, 2000]. Modeling studies of the aerosol forcing over the next century which are consistent with INDOEX observations are urgently needed for climate-change assessment.

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