The impact of natural versus anthropogenic aerosols on atmospheric circulation and clouds in the Community Atmosphere Model

Robert J. Allen$^1$ and Steven C. Sherwood$^2$
$^1$University of California Irvine, Irvine, CA
$^2$University of New South Wales, Sydney, Australia

Submitted to: ?

*Corresponding author address:* Robert J. Allen, University of California Irvine, 240R Rowland Hall, Irvine, CA 92697.
Email: rjallen@uci.edu
Abstract

The Community Atmosphere Model, using both climatologically varying sea surface temperatures (SSTs) and a slab ocean model (SOM), is used to investigate the equilibrium response of the direct effects of anthropogenic and natural aerosols on climate, focusing on circulation and clouds. Aerosol forcing is prescribed based on a recently developed satellite-based aerosol data set. We find that aerosols are capable of significantly affecting local and remote atmospheric circulation and that absorbing versus reflecting aerosols leads to opposite circulation changes. At low latitudes, anthropogenic aerosols, which primarily warm the Northern Hemisphere (NH), strengthen (weaken) the mean meridional mass circulation during December-January-February (June-July-August). This is associated with a northward shift of the Intertropical Convergence Zone. Natural aerosols, which primarily reflect solar radiation and cool the NH, have the opposite effect. These changes are largest in the SOM experiments. Aerosols also have a remote impact on DJF NH mid- and high-latitude circulation, forcing changes similar to the Arctic Oscillation, but only in the prescribed SST experiments. Anthropogenic aerosols lead to an increase (decrease) in zonal wind near 60(30)°N, high-latitude surface (stratospheric) warming (cooling), and reduced high-latitude surface pressure. Natural aerosols have the opposite effect. These changes are associated with changes in the vertical propagation of wave activity, which drives wave-mean flow interaction in the lower stratosphere. We also find anthropogenic aerosols—especially during June-July-August—lead to a significant semi-direct cloud effect, with increasing (decreasing) low-level cloud at low (mid) latitudes. These results suggest aerosol type enhances variability of both tropical and extratropical circulation and accurate specification of aerosols is necessary to simulate low and high latitude climate change.
1 Introduction

Aerosols are important to the climate system because they affect the radiative balance of the planet. First, they scatter and absorb solar radiation, and therefore contribute to atmospheric solar heating and surface cooling (Ramanathan et al., 2001). This is referred to as the aerosol direct effect. Because absorbing aerosols heat the atmosphere, they can also decrease relative humidity and evaporate clouds (Ackerman et al., 2000). This is referred to as the semi-direct effect and can cause additional heating of the surface. Aerosols also have an impact on cloud microphysical properties (and hence, the radiative properties). Because aerosols act as cloud condensation nuclei (CCN), an increase in aerosols leads to an increase in CCN, which in turn leads to an increase in cloud droplet number density, a reduction in cloud droplet effective radius and ultimately, brighter clouds (Twomey, 1977). This is referred to as the first aerosol indirect effect, which enhances reflection of solar radiation back to space by clouds (and also reduces the solar radiation reaching the ground). The smaller cloud droplets also promote suppression of precipitation. This is referred to as the second indirect effect and leads to an increase in cloud lifetime (Albrecht, 1989).

Several studies have documented the climatic impacts of reflecting (e.g. sulfate) aerosols. For example, Roeckner et al. (1999) use a coupled general circulation model (GCM) with a tropospheric sulfur cycle and find that increasing anthropogenic sulfate aerosols mitigate greenhouse warming by reflecting shortwave radiation. They also find the intensity of the global hydrological cycle becomes weaker if direct and indirect sulfate aerosol effects are included (possibly due to reduced transfer of sensible and latent heat from the surface to the atmosphere). Moreover, the simulated temperature trends during the last half of the 20th century are in better agreement with observations if both greenhouse gases and sulfate aerosols are included (e.g. Mitchell et al., 1995).

Sulfate aerosols also affect large-scale atmospheric circulation, particularly in the tropics, causing a southward shift of the Intertropical Convergence Zone (ITCZ) (Williams et al., 2001). This shift is attributed to a reduced inter-hemispheric temperature gradient since reflecting aerosols predominately cool the NH. Rotstayn and Lohmann (2002) suggests that this shift may have contributed to the Sahelian drought of the 1970’s and 80’s. Similarly, Cox et al. (2008) finds that sulfate aerosols alter the meridional SST gradient in the equatorial Atlantic, which affects Amazonian precipita-
tion; future aerosol reductions imply increasing risk of Amazonian drought.

Reflecting aerosols are also capable of affecting high-latitude circulation. Several studies (e.g. Shindell et al., 2004; Robock, 2000) have shown that tropical volcanic aerosols (which are primarily composed of sulfate) affect NH high-latitude winter climate, causing a positive Arctic Oscillation (AO) anomaly. The conventional mechanism explaining this effect is heating of the low latitude stratosphere (and ozone loss at high latitudes, which cools the polar lower stratosphere) enhances the meridional temperature gradient and strengthens the westerly winds near the tropopause. The enhanced westerlies then propagate down to the surface via wave-mean flow interactions, creating a surface temperature response pattern typical of the AO. Stenchikov et al. (2002) also finds a positive AO response results from low latitude cooling of the troposphere (since volcanic aerosols predominantly reflect solar radiation).

Recent research has focused on the role absorbing aerosols have on climate, especially in the South Asian region, where emissions of fossil fuel black carbon (BC) increased by $\sim 6$ fold since 1930 (Ramanathan et al., 2005). Several studies have associated absorbing aerosols to weakening of the Indian monsoon (Meehl et al., 2008; Ramanathan et al., 2007, 2005). This weakening is caused by reduced evaporation over the Indian Ocean, a reduced land-sea temperature contrast and meridional SST gradient in the northern Indian Ocean, and increased atmospheric stability. Similarly, Randles and Ramaswamy (2008) find that increases in only scattering aerosols weakens the monsoonal circulation and inhibits precipitation. However, at higher-extinction optical depths, low-level convergence and enhanced vertical velocity overcome the stabilizing effects of absorbing aerosols and enhance the monsoonal circulation. This conclusion is similar to an observational study by Bollasina et al. (2008), who find that excessive aerosols in May leads to reduced cloud amount and precipitation, increased surface shortwave radiation, and land surface warming. This leads to an enhanced June/July monsoon. Several other regional impacts of absorbing aerosols exist, including a meridional shift of Asian rainfall (Menon et al., 2002), Himalayan glacier retreat (Ramanathan et al., 2007), and reduction of Arctic sea ice (Flanner et al., 2007).

A few studies have identified remote impacts of regional absorbing aerosol forcing. For example, Kim et al. (2006) conduct simulations with the NASA Global Modeling and Assimilation Office finite-volume general circulation model (fvGCM) using aerosol forcing functions derived from the Goddard
Ozone Chemistry Aerosol Radiation and Transport model (GOCART). They find that absorbing aerosols (dust and black carbon) excite a planetary-scale teleconnection pattern in sea level pressure, temperature, and geopotential height spanning north Africa through Eurasia to the north Pacific. They also show that the surface temperature signature associated with the aerosol-induced teleconnection is very similar to the spatial pattern of observed long-term trend in surface temperature over Eurasia.

Wang (2007) uses an interactive aerosol-climate model based on equilibrium simulations with CCM3 and a slab ocean model to examine the direct radiative forcing of BC aerosols. He finds that BC is able to cause a significant change in tropical convective precipitation ranging from the Pacific and Indian Ocean to the Atlantic Ocean, with a strengthened (weakened) Hadley cell in the Northern (Southern) Hemisphere. This change occurs often well away from emission centers, demonstrating a remote climate impact.

Similarly, Chung and Seinfeld (2005) find that direct forcing of anthropogenic BC leads to a northward shift of the ITCZ, increasing (decreasing) precipitation between 0-20°N (0-20°S). This change in precipitation is enhanced if BC is assumed to be internally mixed with sulfate (which leads to more warming). These results are similar to Roberts and Jones (2004) and Yoshimori and Broccoli (2008).

Absorbing aerosols may also affect high-latitude circulation. Chung and Ramanathan (2003) investigate the influence of the interannual variability of the SE Asian haze on global climate. Two CCM3 experiments are conducted, representing two extreme locations of the forcing, 1. extended haze forcing (EHF), in which the haze extends to 10°S, and 2. shrunk haze forcing (SHF), in which the haze is confined north of the equator. Over India, where the forcing is concentrated, the two experiments produce similar climate changes. The remote impacts of the two experiments, however, are significantly different, with the EHF experiment yielding zonal wind changes similar to those associated with the positive AO (the SHF experiment yields negligible changes). The SE Asian haze may therefore partially explain the observed increase in AO variability (Feldstein, 2002).

The purpose of this study is to investigate the low- and high-latitude circulation changes resulting from anthropogenic (primarily absorbing) and natural (primarily reflecting) aerosols. Aerosol forcing is prescribed using a satellite based aerosol data set (Chung et al., 2005). As such, only the direct radiative effects of aerosols are considered. Furthermore, the aerosol forcing is not allowed to change in response to meteorological (e.g. wind, precipita-
tion) changes. However, this approach may lead to more accurate absorbing aerosol induced circulation changes, since the absorbing aerosol forcing in most GCMs is significantly underestimated (Sato et al., 2003; Ramanathan and Carmichael, 2008). To help determine the robustness of the results, experiments are conducted using both climatological SSTs and a slab ocean model.

This paper is organized as follows. Section 2 discusses the different approaches of incorporating aerosol forcing into a climate model, the satellite data-set used in this analysis, and experiment design. Section 3 presents the results, including circulation changes at low latitudes (e.g. mean meridional mass circulation changes) and mid/high-latitudes (zonal wind changes and the Arctic Oscillation). The effects of anthropogenic aerosols on clouds are also discussed. Conclusions follow in Section 4.

2 Experimental Design

2.1 Incorporation of Aerosols

The aerosol effect can be included in a climate model in at least three ways. In the most sophisticated approach, the distribution and amount of aerosols in the atmosphere are predicted based on emissions (e.g. Wang, 2007). Additional physical processes, such as transport and mixing, dry deposition, and wet removal are also included. The calculation of these processes utilizes the predicted winds, temperature, air density and pressure, cloud cover, and precipitation by the climate dynamics model. This interactive approach, therefore, allows aerosols to adapt to rapidly changing circulation, precipitation, etc.

Alternatively, one could use an off-line (prescribed) approach, as is done in this study. For example, the aerosol optical properties (based on observations) can be specified; the model radiation code then generates the aerosol radiative forcing (e.g. Menon et al., 2002). One could also specify the atmospheric and surface aerosol forcing using observations (Ramanathan et al., 2007). Chung (2006) shows that the off-line approach, despite not allowing aerosols to interact with the climate, yields similar results (i.e. changes in precipitation) as the on-line approach.

The primary reason for using the prescribed approach is because the absorbing aerosol forcing (primarily BC) in most GCMs—which use the in-
teractive approach is underestimating by a factor of 2-4 (Sato et al., 2003). This underestimation is due to a number of factors, including neglect of the internally mixed state of BC with other aerosols, neglect of biomass burning (about 40% of total BC emission) and BC concentrations that peak too close to the surface (Ramanathan and Carmichael, 2008).

2.2 Satellite Derived Aerosol Forcing

We use the satellite-based aerosol forcing of Chung et al. (2005), which is based on the Moderate Resolution Imaging Spectroradiometer (MODIS), with gaps filled using the Georgia Tech/Goddard Global Ozone Chemistry Aerosol Radiation and Transport (GOCART) model. Ground based observations from the AErosol RObotic NETwork (AERONET) are used to examine uncertainties in the satellite data. The anthropogenic fraction of the aerosol forcing is obtained by differencing the natural and total aerosol forcing. The natural forcing is obtained using 1. the GOCART model to determine the natural fraction of observed aerosol optical depths (AOD); and 2. the MODIS fine-mode fraction (FMF) of AOD, which is the ratio of AOD from larger aerosol particles to the total MODIS AOD (since anthropogenic aerosols are mostly sub-micron sized). These parameters are then introduced into the Monte-Carlo Aerosol Cloud Radiation (MACR) model with satellite climatology of clouds to derive 1. the reduction of solar radiation at the surface ($F_{SFC}$); and 2. the increase in atmospheric solar absorption ($F_{ATM}$) due to aerosols for 2000-2003.

The uncertainty in this surface and atmospheric aerosol forcing (global annual mean) is 10-20%. However, there is also uncertainty in distinguishing natural versus anthropogenic aerosols. For example, MODIS retrievals of FMF have been found to be systematically larger than in situ measurements by 0.2 in the Aerosol Characterization Experiment (ACE)-Asia region (Anderson et al., 2005). This underestimation of retrieved particle size distributions is consistent with an analysis of the Puerto Rico Dust Experiment (PRIDE) data set (Levy et al., 2003).

2.3 Experiment Details

CAM is run with both a data ocean model (DOM) (i.e. climatological seasonal cycle of SSTs) and a slab ocean-thermodynamic sea ice model (SOM) at T42 resolution ($\sim 2.8^\circ \times 2.8^\circ$) and 26 vertical levels. With the SOM,
Table 1: Definition of the three CAM aerosol experiments and signals (differences between experiments). Each experiment was run with a DOM and SOM.

<table>
<thead>
<tr>
<th>Experiment</th>
<th>Aerosols Used</th>
</tr>
</thead>
<tbody>
<tr>
<td>NULL</td>
<td>none</td>
</tr>
<tr>
<td>NATURAL</td>
<td>natural</td>
</tr>
<tr>
<td>ALL</td>
<td>anthropogenic + natural</td>
</tr>
</tbody>
</table>

<table>
<thead>
<tr>
<th>Signal</th>
<th>Experiments Used</th>
</tr>
</thead>
<tbody>
<tr>
<td>NAT</td>
<td>NATURAL-NULL</td>
</tr>
<tr>
<td>ALL</td>
<td>ALL-NULL</td>
</tr>
<tr>
<td>ANTHRO</td>
<td>ALL-NATURAL</td>
</tr>
</tbody>
</table>

the ocean mixed layer contains an internal heat source (Q flux) representing seasonal deep water exchange and horizontal ocean heat transport. SOM (DOM) experiments are run for 150 (100) years, the last 100 (70) of which are used in this analysis, when no significant trend in TOA net energy flux existed.

Monthly aerosol forcings (F_{ATM} and F_{SFC})–as estimated by Chung et al. (2005)–are incorporated into the CAM radiation module. Aerosol absorption is uniformly distributed throughout the lowest \( \sim 3 \) km of the atmosphere, in accord with observation (Ramanathan et al., 2001). Although aerosol forcing is almost independent of zenith angle (\( \theta \)) when \( \theta \) is small, aerosol forcing \( \rightarrow 0 \) as \( \theta \rightarrow 90^\circ \). Thus, the added aerosol forcing is multiplied by a scaling factor that depends on \( \theta \) (Chung, 2006).

Table 1 lists the CAM experiments performed, as well as the five signals (i.e. differences between experiments) that will be discussed. The anthropogenic + natural experiment is compared to two control runs, the natural experiment (to get the anthropogenic signal, designated as ANTHRO) and a control run with no aerosols (to get the total aerosol signal, designated as ALL). The natural aerosol signal (designated as NAT) is obtained by comparing the natural experiment with the no aerosol control run. CAM default aerosol are removed by setting the aerosol mass mixing ratio for the five aerosol species (carbonaceous, dust, sea salt, sulfur, and volcanic) to zero.

Figure 1 shows the prescribed zonal mean DJF and JJA atmospheric heating (F_{ATM}) and reduction in surface solar radiation (F_{SFC}) for the three
Figure 1: DJF (top) and JJA (bottom) zonal mean reduction in surface solar radiation ($F_{SFC}$; left) and atmospheric heating ($F_{ATM}$; right) for each experiment. Units are W m$^{-2}$.
experiments. The forcing is largest in the NH, where the major sources of aerosols exist. There is also a clear seasonal signal, with larger forcing during each hemisphere’s summer, when solar radiation is maximum. ANTHRO $F_{ATM}$ is much larger than that for NAT, due to a greater proportion of absorbing aerosols.

During DJF, the maximum forcing occurs near 10°N, primarily due to anthropogenic aerosols from central Africa (due to biomass burning) and India (due to fossil-fuel burning) and natural aerosols (dust) from central Africa. ANTHRO $F_{ATM}$ ($F_{SFC}$) peaks near 6(-7) W m$^{-2}$. The NAT forcing is substantially less, with $F_{ATM}$ ($F_{SFC}$) peaking near 0.5(-3) W m$^{-2}$. Because ANTHRO $F_{ATM}$ is nearly equal in magnitude to $F_{SFC}$, ANTHRO is comprised of a high proportion of absorbing aerosols. However, because NAT $F_{ATM}$ is only $\sim 20\%$ of $F_{SFC}$, a low proportion of absorbing aerosols (i.e. a high proportion of reflecting aerosols) comprise NAT.

During JJA, ANTHRO $F_{ATM}$ for the entire NH is at least 3 W m$^{-2}$. Two distinct maxima exist at 5°S and 40°N, with $F_{ATM}$ approaching 7 and 8.5 W m$^{-2}$, respectively. These maxima are due to aerosols from southern Africa and southeast Asia/China. NAT has substantially less heating, with most of the NH $F_{ATM}$ between 0.5-1 W m$^{-2}$, peaking near 15°N at 1.2 W m$^{-2}$. India, the Middle East, the Gobi Desert, and parts of China contribute most to NAT, again primarily due to dust. In terms of $F_{SFC}$, the disparity between anthropogenic and natural aerosol is much less. ANTHRO (NAT) minimizes at $\sim -9(-5)$ W m$^{-2}$ near 40(15)°N. Similar to DJF, ANTHRO (NAT) JJA possess a much higher (lower) proportion of absorbing versus reflecting aerosols. This is true year-round.

3 Results
Table 2: Summary of global (GL) and Northern Hemisphere (NH) annual (ANN), Dec-Jan-Feb (DJF), and Jun-Jul-Aug (JJA) mean differences for selected climate variables. Significance is denoted by bold (≥ 90%); * (≥ 95%) and ** (≥ 99%). Units for temperature, cloud, precipitation and flux variables are K, %, mm day\(^{-1}\) and W m\(^{-2}\), respectively.

<table>
<thead>
<tr>
<th></th>
<th>ANTHRO SOM</th>
<th>NH</th>
<th>NAT SOM</th>
<th>NH</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>DJF</td>
<td>JJA</td>
<td>ANN</td>
<td>DJF</td>
</tr>
<tr>
<td>T(_{sf/c})</td>
<td>0.08</td>
<td>0.08</td>
<td>0.08</td>
<td>0.14</td>
</tr>
<tr>
<td>T(_{sf/c} - \text{ind})</td>
<td>-0.02</td>
<td>-0.05*</td>
<td>-0.02</td>
<td>-0.003</td>
</tr>
<tr>
<td>T(_{700})</td>
<td>0.18**</td>
<td>0.20**</td>
<td>0.20**</td>
<td>0.24**</td>
</tr>
<tr>
<td>T(_{700} - \text{ind})</td>
<td>0.20**</td>
<td>0.32**</td>
<td>0.23**</td>
<td>0.26**</td>
</tr>
<tr>
<td>CLOW</td>
<td>0.06</td>
<td>0.10**</td>
<td>0.07**</td>
<td>0.05</td>
</tr>
<tr>
<td>CMED</td>
<td>-0.28**</td>
<td>-0.26**</td>
<td>-0.22**</td>
<td>-0.15**</td>
</tr>
<tr>
<td>CHI</td>
<td>0.03</td>
<td>0.10**</td>
<td>0.04</td>
<td>0.38**</td>
</tr>
<tr>
<td>PRECC</td>
<td>-0.02**</td>
<td>-0.02**</td>
<td>-0.02**</td>
<td>0.02**</td>
</tr>
<tr>
<td>PRECL</td>
<td>-0.008**</td>
<td>-0.02**</td>
<td>-0.01**</td>
<td>-0.006</td>
</tr>
<tr>
<td>SHFLX</td>
<td>-0.60**</td>
<td>-1.1**</td>
<td>-0.91**</td>
<td>-0.89**</td>
</tr>
<tr>
<td>FSNS</td>
<td>-1.5**</td>
<td>-2.7**</td>
<td>-2.2**</td>
<td>-2.3**</td>
</tr>
</tbody>
</table>

<table>
<thead>
<tr>
<th></th>
<th>ANTHRO DOM</th>
<th>NH</th>
<th>NAT DOM</th>
<th>NH</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>DJF</td>
<td>JJA</td>
<td>ANN</td>
<td>DJF</td>
</tr>
<tr>
<td>T(_{sf/c})</td>
<td>-0.04</td>
<td>-0.004</td>
<td>-0.02</td>
<td>-0.07</td>
</tr>
<tr>
<td>T(_{sf/c} - \text{ind})</td>
<td>-0.13**</td>
<td>-0.10**</td>
<td>-0.10*</td>
<td>-0.19**</td>
</tr>
<tr>
<td>T(_{700})</td>
<td>0.10**</td>
<td>0.20**</td>
<td>0.13**</td>
<td>0.09**</td>
</tr>
<tr>
<td>T(_{700} - \text{ind})</td>
<td>0.12**</td>
<td>0.27**</td>
<td>0.18**</td>
<td>0.13**</td>
</tr>
<tr>
<td>CLOW</td>
<td>0.22**</td>
<td>0.26**</td>
<td>0.17</td>
<td>0.19</td>
</tr>
<tr>
<td>CMED</td>
<td>-0.19**</td>
<td>-0.14**</td>
<td>-0.17**</td>
<td>-0.07</td>
</tr>
<tr>
<td>CHI</td>
<td>0.12**</td>
<td>0.15**</td>
<td>0.09**</td>
<td>0.30**</td>
</tr>
<tr>
<td>PRECC</td>
<td>-0.02**</td>
<td>-0.03*</td>
<td>-0.08</td>
<td>-0.02</td>
</tr>
<tr>
<td>PRECL</td>
<td>-0.004</td>
<td>-0.01**</td>
<td>-0.007**</td>
<td>0.001</td>
</tr>
<tr>
<td>SHFLX</td>
<td>-0.80**</td>
<td>-1.2**</td>
<td>-0.95**</td>
<td>-0.93**</td>
</tr>
<tr>
<td>FSNS</td>
<td>-1.8**</td>
<td>-3.0**</td>
<td>-2.4**</td>
<td>-2.5**</td>
</tr>
</tbody>
</table>

Here T\(_{sf/c}\) = surface temperature, T\(_{700}\) = 700 hPa temperature, CLOW = low-level cloud fraction, CMED = medium-level cloud fraction, CHI = high-level cloud fraction, PRECC = convective precipitation rate, PRECL = large-scale precipitation rate, LHFLX = surface latent heat flux, SHFLX = surface sensible heat flux, FSNS = net surface solar radiation.
3.1 Vertical Temperature Structure

Figure 2 shows the vertical cross section of temperature change ($\Delta T$). For DJF, both ANTHRO and NAT DOM possess a warming troposphere. Low latitudes warm the most, broadly consistent with the location of maximum aerosol heating (Fig. 1). ANTHRO DOM, however, possess a NH tropospheric warming maximum near $40^\circ$N. Because this maximum is about $30^\circ$N of the maximum aerosol heating, it is driven by changes in the large-scale circulation (see Section 3.2.2). In the stratosphere, NAT also warms despite no directly applied stratospheric aerosol forcing. Opposite changes exist at high NH latitudes, with the Arctic stratosphere cooling (warming) for ANTHRO (NAT). $\Delta T$ is generally largest in the NH, where maximum aerosol forcing exists. However, significant DJF warming occurs over Antarctica for both ANTHROs, when solar radiation is most intense. The remaining discussion will focus on the NH.

The DJF ANTHRO SOM $\Delta T$ is similar to that for ANTHRO DOM, with a warming troposphere. Similarly, the maximum NH warming occurs near $40^\circ$N, poleward of the maximum $F_{ATM}$. Substantial warming also occurs at high NH latitudes ($> 80^\circ$N). For NAT SOM—unlike the other signals—the NH troposphere cools significantly, consistent with the large reduction of surface solar radiation, relative to atmospheric heating. NAT SOM cools whereas NAT DOM warms because $2/3$ of the DOM surface (i.e. the ocean) is constrained and not allowed to cool in response to the aerosol forcing.

The changes for JJA are generally larger than those for DJF, especially in the NH because of the larger aerosol forcing. ANTHRO DOM and SOM are similar, with tropospheric warming, peaking in the NH high latitudes near 850 hPa at $\sim$5K. This is partially due to significant snow and ice melt (not shown), which leads to a reduction in albedo, increased absorbed solar radiation and additional warming via the ice-albedo feedback. Despite an applied surface forcing of -2 W m$^{-2}$ (Fig. 1) at high NH latitudes, the equilibrium change in net short wave radiation at the surface is nearly 10 W m$^{-2}$ for both ANTHRO SOM and DOM. This is consistent with other studies—indeed, independent of forcing type, the Arctic generally changes most due to the ice-albedo feedback.

For NAT SOM, the NH troposphere cools significantly during JJA. This cooling is larger than that during DJF because of the larger (i.e. more negative) $F_{SFC}$. NAT DOM also cools poleward of $30^\circ$N, but warms at low latitudes. Again, this disparity between DOM and SOM signals is because
Figure 2: DJF (top four panels) and JJA (bottom four panels) temperature change. Symbols represent significance (assessed with a standard t-test using the pooled variance) at the 90% (diamond), 95% (cross) and 99% (dot) confidence level. Units are K.
the DOM cannot directly respond to the reduced surface shortwave radiation over the oceans. The stratosphere for both NAT DOM and SOM warm, similar to DJF. Similar effects—where temperature changes in the troposphere are accompanied by changes of the opposite sign in the stratosphere—have occurred in other climate models (Hansen and Coauthors, 2005, 2007). Although the exact mechanism is not known, the stratospheric warming may be related to changes in the Brewer-Dobson Circulation.

Compared to other absorbing aerosol studies (Chung and Seinfeld, 2005; Yoshimori and Broccoli, 2008), the vertical structure of warming for ANTHRO SOM is similar, but doesn’t extend as far vertically. Chung and Seinfeld (2005) find maximum BC-induced warming (internally mixed aerosols) near 400 hPa at 30°N, especially during JJA, of about 1.2 K. The corresponding warming in this study is ~0.4 K. The lower warming in this analysis is likely because ANTHRO SOM is forced with not only absorbing (i.e. BC) aerosols, but other anthropogenic aerosols, like sulfate, that tend to cool the surface and atmosphere. This is further illustrated in Table 2, which shows a global (NH) annual mean ΔT_{sfc} of 0.08(0.17) K, compared to Chung and Seinfeld’s ΔT_{sfc} of 0.37(0.54) K. However, ANTHRO SOM ΔT_{sfc} is similar to Wang (2004), who assumes externally mixed BC aerosols, at 0.09 K for the global annual mean.

3.2 Circulation

3.2.1 Low Latitudes

Figure 3 shows the change in the mean meridional mass circulation (ΔMMC). For both DJF and JJA, ΔMMC is largest and most significant for the SOM experiments. This is consistent with constrained SSTs in the DOM experiments. Furthermore, the changes are opposite for ANTHRO and NAT. Do you think we should include the mean MMC for the control experiment? If so, that’s 4 more plots since I would need the control MMC for no aerosols SOM and DOM and natural DOM and DOM. I didn’t include these because that’s a lot of plots on one figure. And if we include the control for MMC, we probably should include it for U and PRECC and wherever else we indicate a shift/meridional displacement in a climate variable.

The DJF MMC for ANTHRO (NAT) SOM weakens (strengthens). The weakening also exists in ANTHRO DOM, but the signal is less significant.
Opposite changes occur in JJA, where the ANTHRO (NAT) SOM MMC strengthens (weakens). The MMC for NAT DOM also weakens, but the signal is smaller. These results are consistent with other aerosol studies focusing on the direct effects of BC aerosols (Wang, 2007, 2004; Chung and Seinfeld, 2005; Yoshimori and Broccoli, 2008; Roberts and Jones, 2004), and the direct (Yoshimori and Broccoli, 2008) and indirect (Rotstayn and Ryan, 2000; Williams et al., 2001) effects of sulfate aerosols.

The change in MMC is associated with the change in inter-hemispheric temperature difference (Mantsis and Clement, 2009; Zhang and Delworth, 2005). Because most of the aerosol forcing is north of the equator, the associated \( \Delta T \) is larger in the NH. During DJF (JJA), ANTHRO aerosols—which warm the NH troposphere (Fig. 2)—weaken (strengthen) the inter-hemispheric temperature difference. For ANTHRO SOM, the NH annual mean \( \Delta T_{700} \) is 0.31 K, but only 0.09 K for the SH. This causes the DJF (JJA) Hadley cell to weaken and contract (strengthen and expand), decreasing (increasing) the cross-equatorial heat transport from the warmer to the colder hemisphere. Because NAT aerosols cool the NH troposphere (Fig. 2), they have the opposite effect. For NAT SOM, the NH annual mean \( \Delta T_{700} \) is -0.21, compared to -0.07 in the SH.

This change in tropical circulation is associated with changes in several other fields, including a shift of the Intertropical Convergence Zone (ITCZ) and the associated rising air, convective precipitation, and clouds (Section 3.3). For ANTHRO (NAT), the ITCZ shifts northward (southward) in both seasons. This is illustrated in Figure 4, which shows the SOM change in convective precipitation (PRECC). \( \Delta \text{PRECC} \) is largest in DJF and over the Pacific and Atlantic Oceans, where ANTHRO (NAT) SOM yields a northward (southward) displacement of convective precipitation.

For ALL (not shown), \( \Delta \text{MMC} \) is small, but the pattern is similar to ANTHRO; ALL therefore leads to a weakening (strengthening) of the MMC in DJF (JJA) and northward movement of the ITCZ. This is consistent with aerosol studies looking at the direct (Chen et al., 2007) and direct and indirect (Jones et al., 2007) effects of anthropogenic (sulfate aerosols, BC, organic aerosols, etc.) aerosols, but inconsistent with others (Kristjansson et al., 2005).
Figure 3: DJF (top four panels) and JJA (bottom four panels) mean meridional mass circulation change. Negative (positive) values represent counterclockwise (clockwise) circulation change. Symbols represent significance (assessed with a standard t-test using the pooled variance) at the 90% (diamond), 95% (cross) and 99% (dot) confidence level. Units are $10^{10}$ kg s$^{-1}$.
Figure 4: DJF (top) and JJA (bottom) convective precipitation change for ANTHRO and NAT SOM. Symbols represent significance at the 90% (diamond), 95% (cross) and 99% (dot) confidence level. Units are mm day$^{-1}$. 
3.2.2 Mid- and High-Latitudes and AO Implications

Figure 5 shows the change in zonal wind (ΔU) for DJF and JJA. Similar to the low-latitude change in MMC, ANTHRO and NAT generally have opposite patterns of ΔU, with both SOM and DOM yielding similar patterns of change. Unlike ΔMMC, where the DOM changes were small, both SOM and DOM possess significant ΔU.

The SOM experiments show a meridional displacement of the NH subtropical jet. During both seasons, the ANTHRO (NAT) SOM jet shifts poleward (equatorward), similar to the corresponding shift of the ITCZ. This shift is consistent with the thermal wind equation and the mid-latitude ΔT. During DJF, the maximum tropospheric warming (cooling) is centered near 40°N (Fig. 2) for ANTHRO (NAT). This implies strengthening (weakening) of the jet to the north (south) for ANTHRO SOM. NAT is opposite, with weakening (strengthening) of the jet to the north (south). During JJA, the maximum warming (cooling) for ANTHRO (NAT) SOM occurs at NH high latitudes. This results in a general weakening (strengthening) of the NH meridional temperature gradient (∇T) for ANTHRO (NAT) SOM, consistent with the weakening (strengthening) of the equatorward flank of the jet.

The meridional displacement of the jet is quantified by finding the latitude of the maximum zonal wind for each season-year at 150 hPa. Taking the difference of the mean jet location (experiment-control) yields the jet displacement, as shown in Table 3. For the NH, ANTHRO SOM possess significant poleward displacement during ANN and JJA of ~0.3°. DJF shows negligible movement because the decrease in U occurs near the center of the jet, whereas the increase occurs nearly 20° poleward. Opposite to ANTHRO, NAT SOM shows equatorward displacement, significant for all three seasons, equal to ~0.3°. Small displacements occur in the SH, where the aerosol forcing is weaker and temperature changes are smaller. Similar results are obtained for other upper tropospheric/lower stratospheric pressure levels, as well as if the average of the three latitudes of maximum U are used.

The DOM ΔU is similar, but generally weaker (stronger) in the sub tropics (high-latitudes), with larger changes in the stratosphere. During JJA, ANTHRO DOM shows a general, diffuse weakening of U in the NH, consistent with the large warming at high NH latitudes and a reduced ∇T. A significant increase occurs in the SH, centered on 50°S. This is consistent with the cooling near 60°S. NAT DOM shows an increase near 30°N, where
Figure 5: DJF (top four panels) and JJA (bottom four panels) zonal wind change. Symbols represent significance at the 90% (diamond), 95% (cross) and 99% (dot) confidence level. Units are m s$^{-1}$. 

17
Table 3: Displacement of the NH (top) and SH (bottom) subtropical jet stream at 150 hPa for the SOM experiments. Significance is denoted by bold (≥ 90%); * (≥ 95%) and ** (≥99%). Units are degrees latitude.

<table>
<thead>
<tr>
<th></th>
<th>Signal</th>
<th>ANN</th>
<th>DJF</th>
<th>JJA</th>
</tr>
</thead>
<tbody>
<tr>
<td>NH</td>
<td>ANTHRO SOM</td>
<td>0.24</td>
<td>-0.02</td>
<td>0.30</td>
</tr>
<tr>
<td></td>
<td>NAT SOM</td>
<td>-0.27**</td>
<td>-0.2</td>
<td>-0.38</td>
</tr>
<tr>
<td>SH</td>
<td>ANTHRO SOM</td>
<td>-0.01</td>
<td>0.07</td>
<td>-0.36</td>
</tr>
<tr>
<td></td>
<td>NAT SOM</td>
<td>-0.09</td>
<td>0.16</td>
<td>0.16</td>
</tr>
</tbody>
</table>

low-latitude warming gives way to mid-latitude cooling.

During DJF, ANTHRO DOM exhibits a significant increase (decrease) centered near 60(30)°N, with the enhanced zonal winds extending throughout the troposphere and stratosphere. NAT shows an opposite, but less significant pattern of ∆U. These changes are reminiscent of the positive and negative AO (Thompson and Wallace, 2000), respectively, and are associated with ∆T and the propagation of wave activity. The ANTHRO changes are similar to Chung and Ramanathan (2003) who find that Indian aerosols yield zonal wind changes consistent with the positive AO.

Figure 6 shows the DJF change in the Eliassen-Palm (EP) flux and flux divergence (Holton, 1992). The changes are largest for the DOM experiments, with generally opposite changes for NAT and ANTHRO. NAT (ANTHRO) shows upward and poleward (downward and equatorward) wave energy propagation, peaking between 40-60°N. This is consistent with a decrease (increase) in wave refraction. The average NH percent change in wave refraction (i.e. the ratio of the meridional and vertical EP flux) is 18% and -6.5% (10.2% and -3.7%) for ANTHRO and NAT DOM (SOM), respectively. Convergence (divergence) of NAT (ANTHRO) EP flux occurs above 200 hPa and poleward of 40°N as waves deposit more (less) westerly momentum, inducing a torque that decelerates (accelerates) the winds. Although SOM changes are generally in the same direction as their DOM counterpart, they are confined to the troposphere and smaller in magnitude. This is consistent with other studies that have found DOM experiments—which limit the tropospheric response—yield an amplified stratospheric response (Rind et al., 2005), and increase the ability of troposphere forcing to affect stratospheric variability (Braesicke and Pyle, 2004).
Figure 6: DJF EP flux (m² s⁻²) and EP flux divergence (10⁻⁶ m s⁻²) change. Contour interval is \{-6,-4,-2,0,2,4,6\}.
Figure 7: DJF sea level pressure change. Units are hPa.

The decreased (increased) wave drag is also associated with decreased (increased) poleward transport of air in the middle stratosphere and decreased (increased) subsidence in the lower stratosphere, which hydrostatically leads to decreased (increased) mass and $P_{SL}$ at high latitudes. This is shown in Figure 7, where $\Delta P_{SL}$ is negative (positive) poleward of $50^\circ N$, peaking at -2.0(1.8) hPa for ANTHRO (NAT) DOM. The reduced (increased) subsidence also leads to adiabatic expansion (compression), decreasing (increasing) temperatures in the lower stratosphere at high latitudes (Fig. 2). By thermal wind balance, this increased (reduced) meridional temperature gradient implies an increase (decrease) in the strength of the westerlies in the mid-stratosphere (Fig. 5). This strengthened (weakened) shear results in more (less) refraction of planetary wave energy. Since planetary waves are propagating up from the surface, there is more (less) refraction by the increased (decreased) vertical shear below the area of increased (reduced) zonal wind. More (less) refraction of planetary wave energy at the lower boundary of the wind anomaly leads to wave divergence (convergence) (Fig. 6) and an acceleration (deceleration) of the zonal wind in that area. Over time, the wind anomaly propagates downward from the mid-stratosphere to the troposphere (Haynes et al., 1991; Shindell et al., 2001; Song and Robinson,
The EP flux changes are the result of tropospheric temperature changes. Both ANTHRO and NAT DOM possess a warming tropical troposphere due to the imposed aerosol heating. However, ANTHRO (NAT) warms most (cools) near 40°N, resulting in a reduced (increased) meridional $\nabla T$. Figure 8 shows the change in geopotential height, temperature and horizontal winds at 500 hPa for both DOM signals. The aerosol heating induces local circulation changes in the mid-latitudes, concentrated into four centers near 40°N: the central Pacific Ocean, North America, the eastern Atlantic Ocean and northeast China/Japan (these centers also exist in the SOM experiments). The latter two teleconnections bear resemblance to those during MAM as found by Kim et al. (2006). For ANTHRO (NAT), these centers are associated with higher (lower) geopotential heights, clockwise (counterclockwise) circulation and warmer (colder) temperatures. The warmer (colder) temperatures are associated with warm (cold) air advection, as well as sinking (rising) air that adiabatically warms (cools) (not shown). This warming (cooling) in the mid-latitudes yields a decreased (increased) $\nabla T$, which is associated with reduced (increased) zonal mean energy and amplitudes of tropospheric planetary wave, and decreased (increased) wave flux activity into the lower stratosphere. These results are similar to Stenchikov et al. (2002) who found the tropospheric cooling effect of volcanic aerosols decreases $\nabla T$ between 30-60°N, resulting in decreased vertical wave activity flux and circulation changes consistent with the positive AO.

### 3.3 Clouds

Table 2 shows several significant changes—in terms of global and NH seasonal means—in low (CLOW), medium (CMED) and high (CHI) clouds. Both ANTHROs yield similar cloud changes, with increasing high and low clouds. For ANTHRO SOM, NH high cloud fraction increases for DJF (0.38%), JJA (0.45%) and ANN (0.32%). This increase in high cloud cover is consistent with other absorbing aerosol studies (Chung and Seinfeld, 2005; Yoshimori and Broccoli, 2008; Hansen et al., 1997; Wang, 2004), but inconsistent with others (Roberts and Jones, 2004; Cook and Highwood, 2004; Chen et al., 2007). Similarly, low clouds significantly increase during JJA, both globally (0.10%) and particularly in the NH (0.24%). For ANTHRO DOM, the increase in CLOW is more widespread, with all seasons possessing a significant increase. This agrees with Wang (2004), who also finds a general increase.
Figure 8: Geopotential height (left panel) and temperature (right panel) change at 500 hPa. Horizontal wind change (m s$^{-1}$) is included with $\Delta T$. Symbols represent significance at the 90% (diamond), 95% (cross) and 99% (dot) confidence level. Units are m and K, respectively.
(0.14-0.21%) in global annual mean CLOW using CAM. Both NAT SOM and DOM yield a significant decrease in high clouds for all seasons, opposite to ANTHRO. However, CLOW increases in both NATs (especially SOM), similar to the ANTHRO signals, with the largest increase during DJF. CMED significantly decreases during all seasons for both ANTHROs and both NATs.

Figure 9 shows the zonal mean change in cloud cover. The DJF changes—especially CMED and CHI—are largest at low latitudes and are associated with changes in the mean meridional circulation (Fig. 3). Recall that ANTHRO (NAT) SOM is associated with a northward (southward) displacement of the ITCZ. This is further reflected by increases (decreases) in CMED and CHI near 10°N for ANTHRO (NAT) and opposite changes at 10°S. This DJF tropical signal is less significant with the DOM experiments, since the ocean surface cannot respond to the aerosol forcing.

The remaining discussion will focus on JJA ANTHRO SOM changes. The discussion is generally applicable to ANTHRO DOM.

3.3.1 High Clouds

During JJA, a tropical signal similar to that during DJF is still present, but it is smaller in magnitude. Outside the tropics, in the middle and high NH latitudes, several significant changes exist. CHI increases throughout the entire NH, especially poleward of 50°N. Figure 10 shows that this increase is primarily over land and is associated with an increase in convective mass flux (CMF) and relative humidity (RH) at 300 hPa. This suggests an increase in convection—despite the increased atmospheric stability due to enhanced lower-tropospheric warming relative to that at the surface (global annual mean ∆T_{sfc} of 0.08 K versus ∆T_{700} of 0.20 K for ANTHRO SOM)—is the main reason for the increase in CHI. The pattern of ∆CHI is also consistent with the pattern of convective precipitation (PRECC) change (Fig. 4), with areas of increasing (decreasing) CHI corresponding to areas of increasing (decreasing) PRECC. Similar results exist for ANTHRO DOM (not shown). Interestingly, both NAT signals—especially SOM—yield decreases in CHI, which are associated with decreases in CMF (now shown). Because natural aerosols cool the surface more than the atmosphere (global annual mean ∆T_{sfc} of -0.23 K versus ∆T_{700} of -0.14 K for NAT SOM), they stabilize the atmosphere, which reduces convection.

The increase in CHI may be due to CAM’s convective scheme, which may be unable to tolerate any increase in convective available potential en-
Figure 9: DJF (left panel) and JJA (right panel) change in low (CLOW), medium (CMED) and high (CHI) cloud cover. Symbols represent significance at the 90% (diamond), 95% (cross) and 99% (dot) confidence level. Units are %.
ergy (CAPE) (it has a CAPE-based closure and trigger). Using the single column version of CAM (SCAM), when energy is added to the boundary layer, the model responds with more deep convection; the added energy can then radiate from the upper troposphere (Sherwood, unpublished). Similar experiments with the Laboratoire de Meteorologie Dynamique (LMD) climate model yield similar results. Is this model’s convection scheme also based on CAPE? I cannot find much info on this model on-line. It does appear to be one of those used for the AR4

3.3.2 Low and Medium Clouds

CLOW increases at low- and mid-latitudes, between 20°S and 50°N, and decreases at higher NH latitudes, between 50-80°N. Figure 10 shows that the low-latitude increase occurs over both land and ocean, especially near Africa. However, a distinct land-sea contrast exists at higher NH latitudes, with a decrease (increase) over land (ocean). Areas of decreased (increased) CLOW correspond to areas of decreased (increased) RH at 700 hPa, especially over land. The changes in CLOW are consistent with a significant decrease in JJA large-scale (stable) precipitation (PRECL) poleward of 30°N, primarily over land (not shown). Unlike CHI, there is little correspondence between CLOW and CMF (at 700 hPa). Although both NATs – especially SOM – also yield increased CLOW, a similar land-sea contrast does not exist.

CMED decreases over most of the NH, except poleward of 70°N. Similar to CLOW, a land-sea contrast exists (but it is less pronounced), with CMED decreasing over both land and ocean, with the largest decrease over land.

The decrease in CLOW and CMED at mid-latitudes is consistent with similar studies (Hansen et al., 1997; Cook and Highwood, 2004). This has been called the semi-direct effect, where absorbing aerosols heat the atmosphere column, reduce RH and hence, cloud cover. However, this reasoning neglects changes in circulation. If the aerosol heating is able to induce diabatic circulations, water vapor will be imported into the heated regions, possibly leading to more shallow and/or deep convection. If not, the heating will just stabilize the column and suppress clouds. This may be the reason for the difference in tropical and mid-latitude low cloud behavior: In the tropics the Rossby radius is big and heating over tropical land masses (e.g. Africa) will drive circulations that enhance cloudiness locally. Subsidence over the oceans is also increased, which may enhance marine stratus. Thus, low-latitude clouds over both land and ocean increase. At high latitudes,
Figure 10: Change in CLOW and CHI (top); and the corresponding change in RH (middle) and CMF (bottom) at 700 and 300 hPa, respectively. Units are %, % and kg (m$^2$·day)$^{-1}$.  

\[26\]
however, the Rossby radius is small and the heating tends to get balanced by a
geostrophic adjustment, so the net result is stabilization of the column, a
reduction in RH, and less cloud. This decrease is largest on land because the
increase in stability is largest. For ANTRHO SOM, the difference between
the JJA NH mean $\Delta T_{sf_c}$ and $\Delta T_{700}$ is 0.27 K. Over land, the corresponding
difference is 0.43 K (Table 2), which suggests a larger increase in stability
over land. I do not really find this circulation effect in Hoskins and Karoly, 1981. Is there a different reference?

I attempted to quantify the semi-direct effect using an approach
similar to Wang 04, by looking at the differences in net solar flux
between all-sky and clear sky conditions (at the surface and TOA). However, I realized that I never incorporate the aerosol forcing into
the clear sky radiative flux calculation. I don’t even have this data—it’s a separate data set that Eddy never gave me. Plus, I’d have to
re-run everything, which would take forever and something I don’t
have time to really do. So I am unable to quantify the semi-direct
effect, which is a bummer (especially since I already calculated
everything and compiled a nice table).

4 Conclusions

The Community Atmosphere Model is used to investigate the equilibrium
climate response to anthropogenic and natural aerosols, focusing on atmo-
spheric circulation and cloud changes. The aerosol direct effect (i.e. ab-
sorption and reflection of solar radiation) is incorporated directly into the
CAM radiation module, based on a recent satellite aerosol data set (Chung
et al., 2005). To help determine the robustness of the results, experiments
are performed with both climatological SSTs and a slab ocean model. We
find that aerosols affect low- and mid/high latitude circulation, with natural
(mostly reflecting) versus anthropogenic (mostly absorbing) aerosols yielding
opposite circulation changes. We also find anthropogenic aerosols lead to an
unconventional semi-direct effect, with increasing low and high cloud.

At low latitudes, anthropogenic (natural) aerosols lead to a northward
(southward) shift of the ITCZ. For anthropogenic aerosols, this is associated
with a weakening (strengthening) of the mean meridional mass circulation
during DJF (JJA), which is due to the larger aerosol burden in the NH, which
weaken (strengthen) the inter-hemispheric temperature gradient during DJF.
Natural aerosols have the opposite effect because the NH troposphere cools more than the SH. $\triangle$MMC is smaller in the DOM experiments because SSTs are prescribed, resulting in a smaller inter-hemispheric temperature gradient change (i.e. ANTHRO DOM (NAT) leads to less NH warming (cooling)). This suggests that aerosols affect the variability of precipitation at low-latitudes, consistent with other studies for the Amazon (Cox et al., 2008) and the Sahel (Rotstayn and Lohmann, 2002). The shift in the ITCZ is also associated with meridional shifts of the subtropical jets. For ANTHRO (NAT) SOM, the jets move poleward (equatorward), with the most significant movement in the NH (about 0.3°).

Although much uncertainty exists, global emissions of BC have generally increased over the latter half of the 20th century (although they have fallen somewhat since 1990) (Novakov et al., 2003; Ito and Penner, 2005; Bond et al., 2007). Global emissions of sulfate aerosols, however, have been declining since the 1970s (Smith et al., 2004; van Aardenne et al., 2001). This contrast in absorbing versus reflecting aerosol trend suggests that aerosols may be related to observed changes in the Hadley cell (which has expanded) and poleward jet displacement (Seidel et al., 2008). The observed movement, however, is larger than reported here at 1.0 to 2.4° per 25 years.

At mid- and high-latitudes, anthropogenic (natural) aerosols lead to circulation changes similar to the positive (negative) phase of the AO, with increasing (decreasing) zonal winds near 60°N, increasing (decreasing) high-latitude stratospheric temperatures and decreasing (increasing) sea-level pressure. These changes, however, are only present in the DOM experiments. These AO-like changes result from altered tropospheric temperature gradients, which affect the vertical propagation of wave activity. ANTHRO (NAT) possess an increase (decrease) in $\nabla T$ between low and mid-latitudes which yields a decrease (increase) in vertically propagating wave activity, resulting in acceleration (deceleration) of the stratospheric zonal winds. This wind anomaly eventually propagates back down through the troposphere (Haynes et al., 1991; Shindell et al., 2001; Song and Robinson, 2004; Stenchikov et al., 2002), where it manifests itself at the surface as sea-level pressure and temperature anomalies.

The ability of aerosols to force AO-like circulation changes may help to explain recent increases in the variability of the AO (Feldstein, 2002). Given the aforementioned trends in BC and sulfate, aerosols may also help explain the positive phase persistence of the AO during the 1990’s. These results also suggest that deficient absorbing aerosol forcing may be one reason why GCMs...
underestimate the trend and variability of AO-like circulations (Hurrell et al., 2004; Gillett, 2005).

Aerosols are also capable of affecting climate through their effects on clouds. Although the indirect aerosol effects are not considered here, we find a significant semi-direct effect for both ANTHRO DOM and SOM. During JJA, low clouds generally increase, with a global (NH) mean increase of 0.10% (0.24%) for ANTHRO SOM and 0.26% (0.46%) for ANTHRO DOM. The pattern of change is more complex, however, with increases at low latitudes and decreases at mid/high latitudes, where a strong land-sea contrast also exists (larger decrease over land). This is an unconventional semi-direct effect, where absorbing aerosols primarily affect stability and circulation. At low latitudes, aerosol heating leads to circulation anomalies that import vapor over land and increase subsidence over ocean, thus favoring clouds. The increase in low clouds over the ocean, however, may be unique to CAM since marine stratus is parameterized based on stability at low levels. At mid/high latitudes, aerosol heating leads to stabilization, a decrease in RH, and less low cloud cover. The overall increase in low cloud implies a negative semi-direct effect (which cools the planet, opposite to the direct warming effect), similar to (Wang, 2004), who also found an increase in low cloud cover using CAM. However, the significant decreases in mid-level clouds may suggest otherwise.

Similar to low clouds, high clouds also increase (especially in the NH) for ANTHRO SOM and DOM. This increase is associated with increased convection, despite the stabilizing effect of aerosols. This may be an unrealistic effect due to CAM’s convection scheme, which is based on CAPE. However, CHI decreases in NAT SOM and DOM, and this decrease is associated with reduced convective mass flux. In recent cloud resolving model studies coupled with a radiation scheme and land surface model (Fan et al., 2008; Jiang and Feingold, 2006), aerosol radiative effects on convective clouds led to a decrease in cloud fraction, because the atmosphere is stabilized and convection is reduced. These effects are exacerbated with increasing aerosol absorption (Fan et al., 2008), which is opposite to what we find here.

**Steve’s cloud aliasing?**

These results show that accurate specification of aerosol forcing is necessary for simulation of both clouds and circulation, and therefore, climate change.
5 Acknowledgments

This study was funded by grants X and Y. We thank Chul Eddy Chung for kindly providing his aerosol data set and for other useful suggestions; and Charlie Zender for providing computing resources at UC Irvine. Part of this manuscript contributed to RJA’s PhD thesis.

References


