The Climate Response to Stratospheric Sulfate Injections and Implications for Addressing Climate Emergencies

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ABSTRACT

Stratospheric sulfate aerosol injection has been proposed to counteract anthropogenic greenhouse gas warming and prevent regional climate emergencies. Global warming is projected to be largest in the polar regions, where consequences to climate change could be emergent, but where the climate response to global warming is also most uncertain. The Community Climate System Model, version 3, is used to evaluate simulations with enhanced CO₂ and prescribed stratospheric sulfate to investigate the effects on regional climate. To further explore the sensitivity of these regions to ocean dynamics, a suite of simulations with and without ocean dynamics is run.

The authors find that, when global average warming is roughly canceled by aerosols, temperature changes in the polar regions are still 20%–50% of the changes in a warmed world. Atmospheric circulation anomalies are also not canceled, which affects the regional climate response. It is also found that agreement between simulations with and without ocean dynamics is poorest in the high latitudes. The polar climate is determined by processes that are highly parameterized in climate models. Thus, one should expect that the projected climate response to geoengineering will be at least as uncertain in these regions as it is to increasing greenhouse gases. In the context of climate emergencies, such as melting arctic sea ice and polar ice sheets and a food crisis due to a heated tropics, the authors find that, while it may be possible to avoid tropical climate crises, preventing polar climate emergencies is not certain. A coordinated effort across modeling centers is required to generate a more robust depiction of a geoengineered climate.

1. Introduction

a. Motivation

The Intergovernmental Panel on Climate Change Fourth Assessment Report (IPCC AR4) projects global and annual mean warming of 1.7°C to 4.4°C this century under the A1B emissions scenario (Meehl et al. 2007). Warming in the northern high latitudes is projected to be 1.5–4.5 times the global mean values in global climate models (Holland and Bitz 2003). Even under stabilized emissions or cessation of emissions, the planet will continue to warm due to the gases that have already been emitted (Matthews and Caldeira 2008; Ramanathan and Feng 2008; Solomon et al. 2009), with a very real possibility that committed warming is equal to or greater than “dangerous” levels (Ramanathan and Feng 2008) that some claim may have catastrophic consequences (Hansen et al. 2007). Blackstock et al. (2009) define climate emergencies as “those circumstances where severe consequences of climate change occur too rapidly to be significantly averted by even immediate mitigation efforts.” Such emergencies would include, for example, the loss of habitat for polar bears, displaced arctic ecosystems, thawing permafrost, rapid sea level rise due to melting Greenland and West Antarctic ice sheets, or a large reduction in crop production due to temperature changes in the tropics.

Recent Arctic warming and record summer sea ice area minimums have spurred expressions of concern for, and investigations into, the fate of sea-ice-dependent polar bears (Regehr et al. 2010), arctic ecosystems (Grebmeier et al. 2006), permafrost (Lawrence et al. 2008), and the way of life of local communities (Hinzman et al. 2005 and references therein). Each of these systems depends on either sea ice area (e.g., for hunting, resting, breeding; Moore and Huntington 2008) or subfreezing surface temperatures over land (e.g., permafrost and Greenland).
Maintaining surface temperatures and preserving sea ice are thought to be necessary to avoid threatening such systems.

On the other side of the globe, the world’s greatest ice sheets are found in West Antarctica and the Antarctic Peninsula, storing huge amounts of water that could potentially raise sea level by many meters. There is a very real danger that land ice calving and/or melting could accelerate and cause greater sea level rise than is anticipated from thermal expansion of seawater alone. The catalysts for such events are higher air temperatures, and more importantly, warmer ocean waters sloshing up against the ice sheet outlets that melt outlet glaciers and ice shelves and could destabilize grounding lines (Oppenheimer 1998; Schoof 2007). The most in-peril ice sheets flow into the Ross and Weddell Seas (Oppenheimer 1998) in West Antarctica, which has been shown to be warming currently (Steig et al. 2009) and losing ice mass (Chen et al. 2009; Velicogna 2009).

Whereas the greatest warming is projected to occur in the polar regions, the tropics show relatively modest temperature changes under increasing CO2. Nonetheless, ecosystems in the tropics may be among those most affected by a changing climate: small amplitude climate variability in the tropics, combined with a tightly coupled ocean–atmosphere system, means that even small climate changes can have important consequences for living systems, whose evolution was built on a narrow climatic conditions have lower physiological thresholds and, thus, are put under more stress for a given warming than those from more climatically variable regions (Tewksbury et al. (2008) and references therein). Specifically, crops grown in the tropics, providing livelihood and sustenance to billions of people, abide by similar laws, so even small climate changes in the tropics can be detrimental. Global warming will cause temperature and precipitation to surpass optimal growing conditions, adversely affecting ecosystems, agriculture, and food security for billions (Battisti and Naylor 2009).

Should climate evolve as models predict, and severe consequences emerge, swift action may become necessary. However, even if all anthropogenic emissions of greenhouse gases ceased, the planet would continue to warm, possibly by a significant amount (Armour and Roe 2011; Hare and Meinshausen 2006). Thus the only solution in such a scenario would be to counteract the rise in temperature by some other means. Accordingly, the domain of feasible solutions to the global warming problem has expanded from adaptation and mitigation by greenhouse gas emissions reductions to include geoengineering: the deliberate modification of the earth’s radiative budget so as to stop the climate change due to increasing anthropogenic greenhouse gases. Geoengineering has evolved from a topic of intermittent discourse between scholars (via publications), to news media and the blogosphere. Recently, major world governments and important scientific societies—such as the Royal Society of the United Kingdom (Shepherd and Rayner 2009), the American Meteorological Society (American Meteorological Society 2009), and the U.S. Government Accountability Office—have made formal statements and issued reports on the topic. Exploration of implementation and deployment technologies, in some form or another, is currently being undertaken (Blackstock et al. 2009), even to the point of pending patent applications (Intellectual Ventures Laboratory, Stratoshield White Paper, available online at http://intellectualventureslab.com/wp-content/uploads/2009/10/Stratoshield-white-paper-300dpi.pdf). Our current understanding of the effect geoengineering will have on the climate system, especially on a regional scale, is not sufficient to rule out unfavorable consequences (Robock et al. 2010), however, and calls for more research have been made (American Meteorological Society 2009; Hegerl and Solomon 2009; Keith et al. 2010).

Numerous schemes have been proposed to alleviate the warming due to anthropogenic emissions of greenhouse gases. These schemes fall into two groups: those that alter the sources and sinks of carbon in the earth system, and those that alter the albedo of the planet. Several implementation schemes have been proposed to accomplish albedo management. These proposals include placing reflective mirrors in space, seeding clouds to make them brighter, and injection of sulfate aerosols or its precursors into the stratosphere. Each of these schemes would work by allowing less shortwave energy to reach the surface of the planet, thereby reducing surface temperature. It is unknown which of these ideas may be most effective in alleviating temperature rise. However, sulfate injection is the leading contender because it is inexpensive to implement, uses existing technology (Robock et al. 2009), and it would be quick to affect surface temperatures if commenced, as well as quick to terminate (Matthews and Caldeira 2007; Robock et al. 2008; Browkin et al. 2009). These factors place stratospheric sulfate injections on top of the list of relatively realistic solutions that could be deployed in the near future and is the guiding reason for our choice to simulate these injections in a GCM and study its effects on the model’s regional climate.

b. Background

Many of the geoengineering modeling studies to date evaluate the impact of aerosols or sunshade technology by uniformly reducing the solar constant in a model.
Govindasamy and Caldeira (2000) and Govindasamy et al. (2003) showed in their studies that uniformly reducing the solar constant in an atmosphere general circulation model (AGCM) coupled to a slab ocean sufficiently canceled the global mean warming due to doubling and quadrupling CO₂, respectively. They found a greater cooling in the tropics compared to the poles and a reduction in seasonal amplitude of temperature. In the context of the polar emergencies discussed in the section 1a, these studies suffer from the exclusion of ocean and sea ice dynamics. Lunt et al. (2008) performed a similar study, but with an AGCM coupled to a full ocean GCM, albeit at low resolution, wherein they quadrupled CO₂ and reduced the solar constant. They noted a reduced pole-to-equator temperature gradient, reduced temperature seasonality, and a reduced intensity of hydrologic cycle compared with a preindustrial control. Matthews and Caldeira (2007) investigated the transient response of the climate to insolation reduction in a model with an interactive carbon cycle, made up of an energy–moisture balance atmosphere, dynamic ocean, and dynamic–thermodynamic sea ice. Their study emphasized the fast response time of the climate to turning on and off geoengineering. All of the above studies are limited for the purpose of understanding the effects of stratospheric aerosols by use of a forcing (reduced solar constant) that has a very different spatial structure than would be realized by stratospheric aerosol injections, which we will show also has important implications for the response of the climate system to the net forcing of increased carbon dioxide plus aerosols.

The combination of imposed forcings is not necessarily, a priori, expected to result in stabilized climate on a regional scale for three reasons. First, stratospheric sulfate forcing, such as is prescribed in our experiments, does not have the same properties as a forcing from increased carbon dioxide because the former primarily acts on shortwave radiation and the latter primarily on longwave radiation. Thus, due to lack of sunlight, the efficacy of sulfate aerosols in the polar regions may be diminished. Second, studies have shown that modifications to shortwave versus longwave radiation affect temperature and precipitation differently (Allen and Ingram 2002; Bala et al. 2008). A perfect stabilization of surface temperature by solar radiation management necessarily excludes perfect stabilization of precipitation (both regionally and in the global average) because of the differing energetic properties of the forcing agents. Third, the spatial distribution of stratospheric sulfate aerosol versus carbon dioxide is not identical, with carbon dioxide being well mixed in the troposphere and a sulfate layer limited to the lower stratosphere. The latter effect, we will show, has profound implications for the response of the climate—especially for the effectiveness of geoengineering to avoid the two polar emergencies that we consider here. Even if the negating effect of a sulfate layer was perfect, there is also some question as to just how feasible it is to tune to the correct amount of sulfate in the real world, where time scales of adjustment are spatially varying and large natural variability will obscure the response of the earth system to changes in forcing.

Modeling studies of the response of the climate system to geoengineering have become more realistic by including simulation of aerosol injection into the stratosphere (Rasch et al. 2008a; Robock et al. 2008; Jones et al. 2010). Rasch et al. 2008a simulated geoengineering by injecting sulfur dioxide into the stratosphere, in such a manner that the quantity of particles would provide enough (globally averaged) negative radiative forcing to counteract the positive radiative forcing due to a doubling of carbon dioxide. They found that the size distribution of the aerosols affected the efficacy of the cooling—specifically smaller sizes were more effective.

Robock et al. (2008) were the first to utilize a climate model with a dynamical ocean and sea ice to investigate the transient response to geoengineering with stratospheric sulfur injections. They simulated both tropical and arctic instantaneous injections and found that the effect of arctic injections, whose purpose was to recover sea ice extent, was not restricted to the Arctic region but extended south to 30°N. Their model results also displayed a weakening of the Asian and Africa summer monsoons. Though both Rasch et al. (2008a) and Robock et al. (2008) use a more realistic forcing, in the context of the climate emergencies in the polar regions, both of these studies lack key ingredients: the model used by Rasch et al. (2008a) did not include sea ice dynamics or ocean dynamics, while the model employed by Robock et al. (2008) is very insensitive to either greenhouse gases or stratospheric aerosol (Jones et al. 2010) and the ocean resolution used is extremely coarse. Ammann et al. (2010) conduct fully coupled atmosphere–ocean model simulations that transiently counteract greenhouse warming (via the IPCC A2 scenario) with either a solar reduction or a stratospheric aerosol layer derived from the 1991 Mt. Pinatubo eruption and focus on the regional effectiveness of the combined radiative forcing. They find that the net forcing induces enhanced atmospheric zonal circulation anomalies, which contribute to residual Arctic warming.

The objective of this study is to investigate whether the climate emergencies can be avoided through solar management by injecting sulfate aerosols into the stratosphere. To our knowledge, this is one of the first attempts to maintain global mean surface temperature transiently, in a fully coupled GCM with a more realistic forcing (see also Ammann et al. 2010), and the only
attempt to illuminate the role of ocean dynamics in determining the regional response. We do this by ramping up carbon dioxide concentration concurrently with stratospheric aerosol concentration, in our case with a prescribed sulfate burden, in a climate model with full ocean dynamics. We then compare these experiments to experiments performed without full ocean dynamics to illuminate some of the uncertainties in projecting a potential future geoengineered world that are particularly relevant to the climate emergencies. We will focus on three main regions of concern: the Arctic circle, the West Antarctic ice sheet region (including the Antarctic Peninsula), and the tropics. The paper is organized as follows. Section 2 describes the global climate model and experiment design. Results are described in section 3. We first discuss the transient simulations and broad global results. We further consider the vertical structure of the atmospheric response to stratospheric aerosols and to increased carbon dioxide and then discuss regional surface climate response to these forcings in the context of the three climate emergencies. We expound upon the uncertainties and discuss the broader implications of our results in section 4. Conclusions are presented in section 5.

2. Model and simulations

We perform our experiments using the National Center for Atmospheric Research (NCAR) Community Climate System Model, version 3 (CCSM3) (Collins et al. 2006), which has components for atmosphere, ocean, land, and sea ice. We run each simulation without full ocean dynamics to determine the effect of ocean dynamics on the climatic response to geoengineering (see Table 1). The slab ocean utilized is a modified version of the more common slab ocean model with motionless sea ice. Our version has the complete CCSM3 thermodynamic–dynamic sea ice model, and we refer to it as the Dynamic Sea Ice Slab Ocean Model (DISOM). This model was introduced and used by Holland et al. (2006) and Bitz et al. (2006). The full atmosphere–ocean general circulation model configuration of CCSM3 is called OGCM in this paper. The ocean heat flux convergence (OHFC) prescribed in the DISOM simulations is derived from the surface flux and ocean heat storage climatology of the full OGCM from a 1990s CCSM3 control so that the mean state of the DISOM and OGCM control simulations are the same. We use the ocean component (i.e., DISOM and OGCM) to differentiate between the model configurations when referring to simulations (see Table 1).

We conduct experiments with various carbon dioxide concentrations, some in combination with geoengineering. We do this using both configurations of CCSM3, (i) DISOM (equilibrium) and (ii) OGCM (transient), so that we conduct a total of eight simulations, listed in Table 1. For each model configuration we have a control run (annually periodic external forcing from 1990s levels, with CO2 = 355 ppm and other greenhouse gases set to 1990 levels), an increased CO2 run (co2), a stratospheric sulfate-only run (aero), and a “net” run that has both increased CO2 and a sulfate layer (geoco2).

The forcings are applied instantaneously in the DISOM experiments, and then we run the model to equilibrium (a minimum of 40 years). We analyze the last 40 years of the DISOM control and geoco2 runs and the last 20 years of the co2 and aero runs. The CCSM3 OGCM control and co2 runs were obtained from NCAR (Collins et al. 2006). Carbon dioxide concentration is ramped

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### Table 1. Experimental details. The run names can be understood in the following way: “control” is the control run, “aero” has a prescribed sulfate layer that is annually periodic in DISOM and ramped in OGCM, and “co2” has carbon dioxide doubled in DISOM and ramped in OGCM. DISOM is the Dynamic Sea Ice Slab Ocean Model and OGCM has both dynamic sea ice and ocean. Columns specify the experiment name, type of ocean component, carbon dioxide concentration, and annual mean total atmospheric burden of sulfate, in teragrams (10^12g) of sulfur equivalent. The value of 1 × CO2 is 355 ppm.

<table>
<thead>
<tr>
<th>Expt</th>
<th>Ocean type</th>
<th>CO2 concentration</th>
<th>Sulfate aerosol burden(TgS)</th>
</tr>
</thead>
<tbody>
<tr>
<td>control</td>
<td>DISOM</td>
<td>1 × CO2</td>
<td>0</td>
</tr>
<tr>
<td>aero</td>
<td>1 × CO2</td>
<td>8</td>
<td></td>
</tr>
<tr>
<td>co2</td>
<td>2 × CO2</td>
<td>0</td>
<td></td>
</tr>
<tr>
<td>geoco2</td>
<td>2 × CO2</td>
<td>8</td>
<td></td>
</tr>
<tr>
<td>control</td>
<td>OGCM</td>
<td>1 × CO2</td>
<td>0</td>
</tr>
<tr>
<td>aero</td>
<td>1 × CO2</td>
<td>Linear ramp from 0</td>
<td></td>
</tr>
<tr>
<td>co2</td>
<td>1% ramp from 1990 concentration</td>
<td>0</td>
<td></td>
</tr>
<tr>
<td>geoco2</td>
<td>1% ramp from 1990 concentration</td>
<td>Linear ramp from 0</td>
<td></td>
</tr>
</tbody>
</table>
at 1% yr\(^{-1}\) from the 1990 level. We ramp the sulfate burden linearly from zero so that the amount prescribed at any given time provides a global average negative radiative forcing that approximately equals the positive radiative forcing of the carbon dioxide. In the case of geoco2, we integrated the model until the carbon dioxide reached four times modern concentrations of CO\(_2\) and the sulfate burden reached 16 teragrams of sulfur equivalent (TgS). Figure 1a depicts the years for which means are computed for all results, unless stated otherwise in the text, red and blue dashed: 70–89, and black and green dashed: 60–99.

The sulfate forcing, or imposed “geoengineered layer,” is a prescribed burden of sulfate (SO\(_4\)) in the stratosphere and has a monthly climatology repeating annually. The annually and zonally averaged sulfate concentration multiplied by layer thickness at the time of CO\(_2\) doubling is shown in Fig. 2, which corresponds to a total annual mean burden of 8 TgS. By prescribing the aerosol distribution, we ignore a major additional source of uncertainty in our study: the chemistry of sulfate formation and its transport. However, these processes were taken into account in the generation of the SO\(_4\) climatology. The SO\(_4\) climatology is taken from the results of a model study by Rasch et al. (2008a), whereby they continuously injected a prescribed size distribution of SO\(_2\) (sulfur dioxide) into the stratosphere at an altitude of 25 km from 10\(^{\circ}\)N to 10\(^{\circ}\)S, where it was transported by winds and interacted chemically. They used a prescribed size distribution with a dry mode radius, standard deviation, and effective radius values of 0.05, 2.03, 0.17 \(\mu\)m, respectively, which is meant to simulate a volcanic-like size distribution. Once in the stratosphere, the SO\(_2\) oxidizes to form sulfate aerosol, which is transported and removed via wet and dry deposition.

In the study by Rasch et al., the volcanically sized aerosol distribution did not fully cancel the warming due to doubled carbon dioxide. Because we prescribe the aerosol distribution, we have scaled up the sulfate climatology by the same fraction at each latitude and height in the atmosphere, to better cancel the equilibrium warming under the 2 \(\times\) CO\(_2\) scenario experiment in DISOM. This results in an annual mean prescribed burden of sulfur equivalent in our simulations of 8 TgS (to counteract 2 \(\times\) CO\(_2\)) compared with 5.9 TgS in Rasch et al. (2008a). It has been shown that there may be some limitation to the effectiveness of sulfate aerosols when the microphysics of sulfur dioxide injection and sulfate aerosol formation are taken into account, such that the burden required to cancel a doubling of CO\(_2\), for instance, would be greater than what is estimated in our

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**Fig. 1.** Time series of global-mean annual-mean (a) surface temperature (°C), (b) precipitation (mm day\(^{-1}\)), and (c) TOA net flux (positive downward, W m\(^{-2}\)) for OGCM experiments. Ramping of sulfate and/or carbon dioxide begins in year 10 in the figure. The 1870s preindustrial control is included for reference (black, dashed line). The boxes indicate the years of averaging for all results, unless stated otherwise in the text, red and blue dashed: 70–89, and black and green dashed: 60–99.

**Fig. 2.** The annual-mean zonal mean sulfate concentration multiplied by layer thickness (kg m\(^{-2}\)) at 2 \(\times\) CO\(_2\). The corresponding global total burden is 8 TgS.
study (Heckendorn et al. 2009), and some have suggested that directly injecting sulfuric acid vapor may improve the mass to radiative forcing ratio Pierce et al. (2010). These complications are ignored here, as we prescribe sulfate aerosols with a particular distribution and specified optical properties.

3. Results

Our suite of experiments shows the extent to which global and annual mean warming from rising CO₂ can be offset by placing sulfate aerosols with particular optical properties and spatial distribution in the stratosphere. We first show global-mean, annual-mean results and annual mean spatial maps as a baseline. We then turn our focus to specific regions, namely, the Arctic, West Antarctica and the Antarctic Peninsula, and the tropics, to examine the results in the context of climate emergences.

a. Global

Table 2 lists globally, annual averaged values of temperature, precipitation, and sea ice area and volume from the set of simulations. It is no surprise that the equilibrium temperature change of the DISOM geoco2 case relative to the DISOM control is near zero (Table 2) because we adjusted the concentration of aerosols specified in the DISOM geoco2 run through several iterations. It is more remarkable that the transient warming in the OGCM forced by ramping CO₂ at the rate of 1% yr⁻¹, shown in Fig. 1a, can be effectively canceled up to about the 70th year after forcing commencement (the time of CO₂ doubling) by linearly ramping sulfate aerosol concentration in the stratosphere.

For reference, Figs. 3a and 3b show the spatial maps of annual average surface temperature change between the DISOM and OGCM co2 and control simulations and Figs. 3c and 3d show the companion maps for geoco2. In the annual mean, the presence of a sulfate aerosol layer is able to cancel surface temperature rises due to increased CO₂ nearly everywhere but the Arctic, which will be discussed in more detail in the Arctic subsection.

The OGCM exhibits a slight global mean warming at the end of the analysis period (see Fig. 1a). Hence, we also compute the linear temperature trend spatially for the period before the global mean temperature diverges from the control (years 11–80). Figure 4 shows the magnitude of the temperature change extrapolated to year 80 (the midpoint of the analysis period), computed using the linear trend of annual mean global mean surface temperature for years 11–80. The pattern that emerges matches that seen in Fig. 3d, indicating that the spatial pattern of response is fundamental to the combination of increasing CO₂ and increasing sulfate layer burden, and is not influenced by the existence of a residual global mean warming (0.08°C in the 40-yr average).

When aerosol concentrations are designed to cancel global warming, they do not also cancel global mean precipitation changes (Bala et al. 2008; Robock et al. 2008; Ricke et al. 2010). Indeed, placing sulfate aerosols in the stratosphere reduces precipitation more than the precipitation increases on average from raising CO₂. Thus, the globally averaged precipitation rate declines by between 1% and 2% at the time of CO₂ doubling for both transient and equilibrium geoco2 cases relative to their controls (see Table 2). Figure 1b shows the change in precipitation with time for the OGCM simulations. Although the globally averaged surface temperature stays nearly constant for the first 80 years of the geoco2 experiment, precipitation slowly declines with increasing sulfate burden. This result supports the theory put forth in Allen and Ingram (2002), whereby longwave and shortwave radiation affect precipitation and temperature differently.

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### Table 2. Global annual-mean (top) values and (bottom) differences (Δ) from respective control runs. OGCM means are calculated for the 40 years surrounding the time of CO₂ doubling in the control and geoco2 runs. The OGCM co2 and Aero cases are 20-yr means surrounding the time of CO₂ doubling. Surface temperature (K), precipitation rate (mm day⁻¹), total Northern Hemisphere (NH) and total Southern Hemisphere (SH) sea ice area (SIA) (10¹³ m²), total NH sea ice volume, total SH sea ice volume (SIVol) (10¹³ m³).

<table>
<thead>
<tr>
<th>Model</th>
<th>Expt</th>
<th>Surface temp</th>
<th>Precipitation rate</th>
<th>SIA NH, SH</th>
<th>SIVol NH, SH</th>
</tr>
</thead>
<tbody>
<tr>
<td>DISOM</td>
<td>control</td>
<td>288.1</td>
<td>2.80</td>
<td>1.11, 1.23</td>
<td>2.75, 2.07</td>
</tr>
<tr>
<td>OGCM</td>
<td>control</td>
<td>288.0</td>
<td>2.79</td>
<td>1.12, 1.37</td>
<td>2.74, 2.45</td>
</tr>
</tbody>
</table>

<table>
<thead>
<tr>
<th>Model</th>
<th>Expt</th>
<th>ΔSurface temp</th>
<th>%Δ Precipitation</th>
<th>%Δ SIA NH, SH</th>
<th>%Δ SIVol NH, SH</th>
</tr>
</thead>
<tbody>
<tr>
<td>DISOM</td>
<td>geoco2</td>
<td>0.09</td>
<td>1.47</td>
<td>3.89, 0.10</td>
<td>9.59, 1.98</td>
</tr>
<tr>
<td>OGCM</td>
<td>geoco2</td>
<td>0.08</td>
<td>1.43</td>
<td>2.17, 2.21</td>
<td>3.12, 3.09</td>
</tr>
<tr>
<td>DISOM</td>
<td>aero</td>
<td>−3.36</td>
<td>8.54</td>
<td>25.5, 135</td>
<td>128.6, 123.6</td>
</tr>
<tr>
<td>OGCM</td>
<td>aero</td>
<td>−1.27</td>
<td>0.10</td>
<td>15.4, 0.19</td>
<td>65.9, 3.03</td>
</tr>
<tr>
<td>DISOM</td>
<td>co2</td>
<td>2.70</td>
<td>5.21</td>
<td>28.8, 45.3</td>
<td>52.0, 22.4</td>
</tr>
<tr>
<td>OGCM</td>
<td>co2</td>
<td>1.38</td>
<td>2.15</td>
<td>25.8, 10.3</td>
<td>57.3, 15.7</td>
</tr>
</tbody>
</table>
Figure 1c displays the top of atmosphere (TOA) net flux anomaly from the control simulation for the OGCM co2, aero, and geoco2. The OGCM geoco2 has an annual mean TOA imbalance anomaly of 0.06 W m\(^{-2}\) during our analysis period, while the DISOM geoco2 has an imbalance of less than 0.01 W m\(^{-2}\), indicating that the geoco2 simulations are well balanced and the radiative forcing from the imposed sulfate layer successfully counterbalances that from increased CO\(_2\) during the analysis period.

b. Vertical structure

The global average forcing at the top of the atmosphere in the geoco2 experiments is effectively zero until the time of CO\(_2\) doubling, but there are important spatial differences, particularly in the vertical. We discuss and show the DISOM results of the vertical and zonal mean temperature in this section as an example; however, the OGCM has a very similar vertical temperature response to the combined CO\(_2\) and aerosol forcing. Raising CO\(_2\) causes tropospheric warming and slight to no cooling in the lower stratosphere (Fig. 5a). The sulfate aerosol concentration is at a maximum over the tropics, where the original injection of sulfur dioxide in the Rasch et al. (2008a) study was located, which causes an increase in absorption of solar and infrared radiation there compared to the control climate. By virtue of this spatial distribution, the sulfate aerosol produces a local warming maximum in the lower stratosphere over the tropical region (Fig. 5b). The net result of doubled CO\(_2\) and a sulfate layer on zonal mean temperature is to leave the troposphere much like the 1990s control (Fig. 5c). Yet in the stratosphere, the cooling due to increased carbon dioxide does little to abate the tropical stratosphere.
sulfate-driven warming. These nonnegligible changes in the vertical structure of temperature in the atmosphere cause noticeable differences to the zonal-mean wind field.

Figures 5d–f display the vertical structure changes in zonal-mean zonal wind in the DISOM perturbation experiments. In the annual mean, enhanced CO2 forces the Southern Hemisphere (SH) polar stratospheric vortex to shift equatorward, while the Northern Hemisphere (NH) polar stratospheric vortex displays a broad, weak enhancement in the upper atmosphere (Fig. 5d). The subtropical tropospheric jets and zonal-mean surface winds change little due to doubled CO2. In contrast, forcing by sulfate aerosols alone causes a clear poleward shift of the stratospheric and tropospheric polar vortex in the SH and a strengthening of the stratospheric polar vortex in the NH (Fig. 5e). The net result of the combined forcings in geoco2 looks similar to that of the sum of the two separate forcings. In the NH the polar vortex is strengthened even more in geoco2 than in co2 (an increase in mean zonal wind of about 30% at the peak location compared to 20% in co2), and this strengthening is especially apparent in December–February (DJF) (not shown) when the strengthening also extends down to the surface. In the SH the net result is an equatorward shift of the stratospheric polar vortex and a poleward shift in the jet in the troposphere. The zonal mean temperature and wind response patterns look very similar in the OGCM for the geoco2 case. Thus, the addition of sulfates does not counteract the circulation anomalies due to increased CO2. We will see in the following sections that these upper-level differences are indeed manifested at the surface as changes in climate (although the surface wind response tends to be weaker in OGCM than DISOM).

c. The Arctic

Figures 6a and 6b display the change in the summer, June–August (JJA), surface temperature in the geoco2 simulations as compared to the control integrations for the DISOM and the OGCM, respectively. Also noted on the figure is the location of the sea ice edge, defined as the region within which there is a 15% or greater sea ice concentration, for geoco2 (dashed) and control (solid). As expected, summer temperatures over sea ice remain unchanged, as ice keeps the air temperature at about 0°C in summer. However, northern land surfaces are overcooled in both models, at some locations by over 1°C, with the notable exception of warming in Greenland. Although the sea ice edge in the Arctic is nearly unchanged, the NH sea ice volume is reduced by 10.0% and 2.8% for DISOM and OGCM, respectively, with the greatest thinning of sea ice found around Greenland in the DISOM and East Siberian Sea in the OGCM (not shown, but the pattern is very similar to DJF with slightly greater magnitude changes—see Figs. 6e–f). Additionally, there are reductions in sea ice concentration (not shown) of up to 10% in the marginal ice zones, especially in DISOM.

Figures 6c and 6d show the change in surface temperature in Northern Hemisphere winter (December–February). Contrary to JJA, the DJF surface temperature response in geoco2 displays a broad residual warming throughout much of the Arctic. This is expected, since the impact of aerosols is diminished in the polar night and the warming due to increased CO2 is not compensated. This warming results in increased water vapor and an increase in surface evaporation throughout the high latitudes that further contributes to the surface warming by an enhanced greenhouse effect. However, there is spatial structure to these temperature changes that cannot be due to the well-mixed CO2, as both model configurations exhibit enhanced regional warming of 2°–4°C over northern Eurasia. Although this is cooler than a 2 × CO2 world, temperature differences are still up to 50% of the warming expected from 2 × CO2 here, with the broadest regional warming occurring in the DISOM model (cf. Figs. 3a,b with Figs. 6c,d). The sea ice in DJF is decreased in both area (along sea ice margins, not shown) and volume of 3.5% and 8.8% respectively in the DISOM and 2.1% and 2.8% in the OGCM, although the sea ice extent is nearly unchanged. Figures 6e–f
display the associated pattern of sea ice thickness change, exhibiting more widespread and greater thinning in the DISOM than OGCM.

Compared with the control climate, both the DISOM and the OGCM geoco2 simulations have enhanced winter west to southwesterly winds at 950 mb over northern Eurasia and the northern Atlantic Ocean and Nordic seas, with the OGCM 950-mb wind enhancement mostly limited to the Nordic seas (Fig. 7). These increased winds are the near-surface expression of the upper-level polar vortex, which is enhanced in winter as described earlier. The 950-mb zonal wind changes are statistically significant everywhere: the magnitude is about 1 m s$^{-1}$ or greater. The circulation anomalies enhance the advection of climatologically warmer and moister marine and lower-latitude air into northern Europe and Asia ($v' \cdot \nabla T$ and $v' \cdot \nabla q$) and help to explain the robust pattern of surface warming over Eurasia. The anomalous moisture transport increases downwelling longwave at the surface directly and also by providing additional water vapor for latent heating of the atmosphere. In the OGCM, Eurasia is also warmed by increased sensible heat advection associated with the reduction in the sea ice thickness and in the area covered by sea ice to the north of Eurasia.

The pattern of atmospheric circulation change and surface warming is familiar as the postvolcanic eruption winter response in which the climate exhibits a positive Arctic Oscillation (AO) phase due to a strengthened polar vortex (Robock 2000; Stenchikov et al. 2002; Shindell et al. 2004). However, the pattern cannot uniquely be attributed to the stratospheric aerosols in this case. The sulfate alone induces strengthened westerlies at the surface, most strongly only over northern Eurasia in the aero simulations (not shown), which implies that much of the AO response pattern elsewhere is due to increased CO$_2$, as the co2 case also exhibits surface circulation changes of the same sign as a positive AO (not shown). This suggests that the sulfate layer does not counteract CO$_2$-induced circulation changes; rather it nudges the circulation in the high northern latitudes in the same way as does increasing CO$_2$. This anomalous circulation plays a dominant role in structuring the pattern of temperature response in northern Eurasia through temperature

FIG. 5. Zonal-mean, annual-mean (a)–(c) temperature and (d)–(f) zonal wind in the DISOM simulations. Contours are the control; in colors are differences between the perturbed experiment and the control experiment. The thick black line indicates the zero line; dashed is negative temperature or easterly wind anomaly and solid is positive temperature or westerly wind anomaly. The contour interval is 1.0°C or 8 m s$^{-1}$. 
and moisture advection, which, in addition to local feedbacks, further amplifies downwelling longwave radiation in the region. The anomalous winds also exert a wind stress on the ocean that affects ocean circulation in the OGCM geoco2 simulation (see Fig. 8b). Note that the zonal wind stress changes exhibited in Fig. 8 are comparable between the co2 (8a) and geoco2 (8b) simulations.

The spatial extent of polar residual winter surface warming in the OGCM simulation is much smaller than in the DISOM. In particular, the geoco2 OGCM exhibits cooler SSTs near the United Kingdom and south of Greenland (where the DISOM does not), near regions of deep-water formation. This cooling in the North Atlantic in the OGCM (which exists in the annual mean as well) can only be due to changes in ocean heat transport—the only difference between the two model configurations. In Fig. 9, we show annual-mean, Atlantic and Arctic Ocean zonal mean potential temperature and Atlantic meridional overturning circulation (AMOC) anomalies in the NH for the geoco2 and co2 OGCM simulations. The lower panel of Fig. 9b shows the Atlantic and Arctic Ocean warming from increased CO2, which in some places extends down below 2 km. The lower portion of the lower panel in Fig. 9d shows the change in the
AMOC due to increased CO₂, which is weakened everywhere. In the geoco2 simulation, the sulfate aerosols have managed to cool the ocean everywhere (note the reduced color scale), and particularly north of 50°N (upper portion of the upper panel of Fig 9b). The effect of the combination of sulfates and increased CO₂ on the AMOC is slightly more complex. The AMOC is weakened poleward of 30°N above 1 km in depth. The net result of the ocean temperature and circulation changes under geoengineering is a reduction in northward heat transport in the NH, shown in Fig. 10 along with the co2 and aero anomalies. Thus, in the geoco2 scenario there is less residual warming and thinning of sea ice in the OGCM than in the DISOM model.

In summary, arctic climate changes induced by increasing CO₂ are not perfectly canceled by the injection of stratospheric sulfate aerosols, especially in winter due to the ineffectiveness of the sulfate aerosols when there is no sunlight. There is a consistent warming signal over northern Europe and Asia in geoco2 DJF in the DISOM.
and OGCM that is enhanced by the intensified near-surface westerlies over Europe and Asia. Changes in the North Atlantic caused by the net forcings of sulfate aerosol and increased CO₂ are the same sign as, but weaker than, changes occurring under increased CO₂ alone. The residual surface winds give rise to circulation and heat transport changes in the ocean. As a result, the North Atlantic Ocean in the OGCM geoco2 experiment exhibits cooler surface temperatures and slightly thicker sea ice regionally around Greenland and in the neighboring Arctic when compared to the DISOM, thus better offsetting winter surface warming from increased CO₂.

d. The Antarctic

As in the Arctic, there is residual winter, June–August (JJA), warming over Antarctica from the combined sulfate aerosol and CO₂ forcing (Figs. 11a,b). Both DISOM and the OGCM have residual warming on and around the Antarctic Peninsula; however, the surface warming is much more focused along the Antarctic Peninsula in the DISOM, whereas it is widespread in the OGCM, extending eastward past the Weddell Sea and along the Antarctic shore to almost 130°E. It is these regions where we focus our analysis. The DISOM and the OGCM exhibit temperature changes of opposite sign over East Antarctica. Atmospheric circulation differences in DISOM and OGCM, and associated ocean dynamical responses in OGCM, are responsible for these differences in the surface temperature and sea ice responses between the DISOM and OGCM experiments.

The DISOM geoco2 case has strengthened westerlies over the entire Southern Ocean (Fig. 11c), which is a manifestation of the poleward shift in the subtropical jet (Fig. 5f). The changes in atmospheric circulation lead to advection of climatological warm air from the north by the anomalous north–northwesterlies just west of the Antarctic Peninsula in the DISOM, causing warming over the Bellingshausen Sea and the peninsula. Additionally, the wind stress causes sea ice to be transported away from the east coast of the peninsula into the Weddell Sea where it accumulates. This creates the pattern of thinner ice adjacent to thicker ice in Fig. 11c. This thinner ice allows more heat from the ocean to be conducted through the ice to the air throughout the winter season. Then the thinner ice is advected eastward by the mean westerlies, generating the warming maximum along the east side of the peninsula and extended warmth to the northeast.

The responses in Antarctic winter to the combined aerosol and CO₂ forcing in the OGCM model is different from that in the DISOM model in two ways. First, coupled feedbacks in the tropical Pacific cause changes in the mean state that affect the tropospheric winds in the SH via atmospheric teleconnections (also see

Fig. 9. Annual-mean OGCM Atlantic Ocean zonal mean potential temperature (°C) for (a) control and (b) perturbation experiment differences. Atlantic Ocean meridional overturning circulation (Sv) for (c) control and (d) perturbation experiment differences. Note the smaller color scale for geoco2 as compared to co2.
section 3e). Second, the subsequent changes in wind stress on the Southern Ocean force changes in the ocean circulation that greatly affect the upper-ocean temperature and thus sea ice thickness. We break the Antarctic response into two main regions, depicted in Fig. 11d. Region A lies west of the Antarctic Peninsula and encompasses the Bellingshausen and Amundsen Seas. Region B lies east of the peninsula and is focused on the Weddell Sea but extends east to roughly 70°E.

The zonal surface wind and wind stress anomalies that are evident in the DISOM geoco2 case are also found in the OGCM, but they are overwhelmed by a much larger eddy contribution in JJA. In particular, in region A an anomalous high pressure region (seen as anticyclonic wind stress circulation in Fig. 11d) exists in austral winter. The geopotential height anomalies in the OGCM extend in the vertical, are nearly equivalent barotropic, and display a clear Rossby wave train signal emanating from the tropics (not shown). A recent study (Ding et al. 2011) has shown that such a high pressure anomaly in the Bellingshausen and Amundsen Seas is induced when tropical sea surface temperatures are prescribed in a GCM to have a warm anomaly in the central Pacific Ocean in JJA. As is described in the next section, the OGCM geoco2 case does indeed display a warm anomaly in the central Pacific, albeit small, but similar in magnitude and positioning as in the Ding et al. (2011) study (0.2°–0.3°C in the OGCM compared with 0.2°–0.4°C in Ding et al.). Thus, the warm anomalies over the western Antarctic seas and peninsula, and most importantly the Ross ice shelf, are due to anomalous atmospheric advection of mean temperature, and are consistent with tropical SST changes generating a Rossby wave train that reaches the shores of Antarctica.

Region B exhibits more expansive thinning of sea ice in the OGCM than in the DISOM everywhere from 70°W to 60°E (see Figs. 11c,d). Figure 12 displays OGCM geoco2 ocean potential temperature differences at various depths in SH winter, along with a cross section of the zonal average ocean temperature difference within the longitude sector demarcated in the figure. North of about 60°S, there is warming at most depths as the nearly vertical isotherms have shifted southward. South of roughly 60°S in the Weddell Sea there is residual warming above about 200-m depth and cooling below, suggesting that there is anomalous upward heat transport in that region. To diagnose the source of heat near the surface that thins sea ice in the Weddell Sea, be it the atmosphere or ocean, we next examine ocean temperature and heat transport anomalies and perform an energy budget analysis.

We divide the Weddell Sea basin, defined by the markings in Fig. 12b with a northern edge at 60°S, into upper-layer (0–200 m) and bottom-layer (200–5000 m) boxes. We then compute the vertical energy flux between the layers as a residual after taking into account the top surface fluxes, the horizontal fluxes, and the temperature tendency in the upper layer. The upper layer displays a positive temperature tendency, but a horizontal heat divergence and reduced top surface heat flux into the layer. Thus, energy conservation demands that there is an anomalous upward flux of heat into the upper 0–200-m layer from below. We find a value of 1.3 W m⁻² upward flux into the upper layer from below. The increase in upper-layer heat in the Weddell in the geoco2 OGCM simulation is due to the poleward shift of the Antarctic Circumpolar Current (ACC, not shown). This poleward shift of the ACC in the geoco2 OGCM simulation (not shown) appears as an enhancement in zonal circulation at the northern edge of the Weddell Sea, a greater Ekman transport northward, and hence an increase in the divergence and upwelling of warmer circumpolar deep waters south of the shifted ACC. South of 55°S in the Weddell Sea, isotherms are shallower in the top 500 m and are generally displaced southward in the geoco2 run, compared to the control run, at the approximate latitude of the ACC in this region (see Fig. 12a). These changes are due to the increased zonal wind stress on the ocean (shown in Fig. 8b) that occurs in the annual mean (but is obscured in Fig. 11d because of the large eddy component evident in JJA). Notably, the geoco2 zonal wind stress anomalies around Antarctica, shown in Fig. 8b, are similar in magnitude and pattern.
to those in an increased CO2 world (Fig. 8a). Hence, as in the Arctic, CO2-induced circulation changes are not canceled by the inclusion of sulfate aerosols.

e. The tropics and subtropics

Figure 13 displays the tropical and subtropical (30°S–30°N) seasonal surface temperature changes in the geoco2 simulations as compared to respective controls for DISOM and OGCM. Temperature changes within this region for all models and seasons are smaller than 1°C and usually less than 0.5°C. In fact, much of the temperature change in the tropics under geoengineering, especially on land, is not significantly different from the control climate at the 95% level, as judged by a Student’s t test. However, there is a sizable cooling over the equatorial Pacific in DJF and JJA in DISOM, which has an effect on tropical precipitation.

Figure 14 shows the corresponding images for precipitation. As discussed earlier, in both DISOM and OGCM there is a net drying due to the combined forcing of aerosols and CO2, although there is interesting and important spatial structure in the precipitation changes. Over the oceans, the cooled regions are usually drier regions (in general, the regions of statistically significant precipitation change coincide with regions of statistically significant temperature change). The control simulation of CCSM3 features a double intertropical convergence zone (ITCZ) over the Pacific Ocean: there are two branches of high precipitation straddling the dry equator (Collins et al. 2006) and both of our DISOM and OGCM control simulations display this behavior. Thus the changes in precipitation in geoco2 in the DISOM translate to a widening and strengthening of the dry tongue along the equator. It is noteworthy that the sign of the precipitation change along the equator in the central tropical Pacific in the OGCM, although not statistically significant, is opposite to DISOM, in JJA and particularly in DJF. The local precipitation maximum here is consistent with the local warming signal in the same location. This feature also appears as a local warming maximum in the co2 runs (not shown). As discussed in the previous section, this feature in OGCM geoco2 is associated with the generation of a teleconnection pattern in the southern Pacific Ocean in JJA that produces the anomalous high pressure west of the Antarctic Peninsula and thus the residual warming seen in the Ross and Amundsen Sea regions due to
anomalous atmospheric temperature advection (see Figs. 11b,d). Ocean dynamics is clearly very important to the response of the tropical Pacific SST and precipitation over the ocean.

In terms of monsoonal precipitation, the DISOM and OGCM model results in Fig. 14 clearly indicate an increase in summer precipitation over western India. East Asian coastal locations show a decrease in summer precipitation, and there are hints of an African summer monsoon reduction as well. However, the results are not significant in these regions. This pattern of precipitation response differs from Robock et al. (2008), which has a weakened Southeast Asian monsoon and no increase in Indian summer precipitation. In fact, across published sulfate injection studies, there is little agreement in the regional pattern of precipitation changes (Robock et al. 2008; Rasch et al. 2008b; Jones et al. 2010).

\[ f. Beyond \ a \ 2 \times \ CO_2 \ geoengineered \ world \]

We extended the OGCM geoco2 simulation past the point of CO2 doubling to investigate whether the sulfate layer is still able to counteract increased CO2 at higher levels. Starting from doubled CO2 in the geoco2 run, CO2 is increased 1% yr\(^{-1}\) and the annual mean sulfate burden is linearly increased until the concentration of CO2 is four times 1990 levels and the aerosol burden is 16 TgS. The green line in Fig. 1a displays the time series of global-mean, annual-mean surface temperature. Beyond the time of CO2 doubling (year 80 in Fig. 1), the effectiveness of the aerosol to offset the warming decreases: the global average temperature increases by 0.3° by the time CO2 triples and by 0.6°C by the time it quadruples. Also, by 3 × CO2, the precipitation has declined 1.8% and at 4 × CO2, 2.2%. Not surprisingly, the TOA radiative balance slowly becomes more positive as well. Enhanced albedo feedbacks could play a role, as there are slight reductions in surface albedo and cloud cover in the geoco2 (not shown), but these decreases do not display stronger rates of reduction after year 80. The increasing net TOA flux is thus likely due to a slow saturation in sulfate scattering leading to the appearance of nonlinearities in the burden to sulfate aerosol radiative forcing ratio at high sulfate burdens.
The studies of Matthews and Caldeira (2007), Robock et al. (2008), and Jones et al. (2010) indicate that the response time of the global mean surface temperature to solar radiation management is relatively fast, and an abrupt termination of aerosol injections would cause a rapid rise in temperature back to where it would have been had no geoengineering been implemented. We performed one additional simulation where the sulfate layer was abruptly shut off at the time of CO2 tripling, to confirm previous work investigating the effects of a sudden termination of sulfate loading. The orange line in Fig. 1a indicates that a sudden termination of geo-engineering leads to a rapid temperature rise back to what it would have been had such measures never been performed. In this case, the rough estimate is that the global mean temperature would rise 2°C in a matter of 20 years. Hence, the rate of temperature rise is greatly increased when compared to a scenario that never implements geoengineering.

4. Discussion

Our results show that climate change under stratospheric aerosols and increased carbon dioxide is smaller than under increased CO2 alone. However, maintaining the modern climate is not possible: the combined forcings result in residual changes in the annual averaged climate, in the seasonality of temperature and precipitation, and in regional patterns of atmospheric and oceanic circulation. Many of these residual differences result because the aerosol layer is not able to counterbalance the circulation anomalies induced under increased CO2. Moreover, avoiding polar climate emergencies is not a certainty, in part because CO2 forcing continues to operate throughout winter there, whereas the sulfate layer is less effective.

We evaluated the role of ocean dynamics by comparing experiments with a climate model coupled to a full ocean general circulation model and to a slab ocean. Under the joint forcing of increased carbon dioxide and an aerosol layer, surface temperature in northern Eurasia is cooled in summer, yet exhibits residual warming in winter. The residual climate changes in the Arctic are partially muted by ocean dynamical feedbacks that reduce the amount of poleward ocean heat transport into the North Atlantic Ocean, limiting the thinning of sea ice around Greenland and keeping SSTs cooler than when the ocean dynamics is prescribed. The repercussions of
increased winter surface temperature in northern Eurasia go beyond reduced sea ice. Arctic marine mammals in general are not equipped to adapt swiftly to climate changes (Moore and Huntington 2008) and thus their well-being may be compromised by such a residual warming.

With forcing by both increased CO₂ and sulfate aerosols, Antarctica exhibits overcooling on the continent in summer (not shown) and residual warming on and around land surfaces in winter. Surface wind changes drive changes in ocean circulation and act to amplify the surface warming in winter around Antarctica. Anomalous upward ocean heat flux in the Weddell region results in slightly greater upper-ocean temperatures as well. This residual upper-ocean and surface air warming in and around the ice sheet exit regions does nothing to allay the potential for the West Antarctic ice sheets to become unstable due to increased melting, especially at the base of the ice shelves (Oppenheimer 1998; Meehl et al. 2007; Thoma et al. 2008; Jenkins et al. 2010). It is this instability (tipping point) that causes concern about rapid sea level rise (Notz 2009).

We find the highest effectiveness of geoengineering for our climate emergencies is in the tropics. Except in some places over the ocean, the temperature and precipitation differences under increased carbon dioxide and a sulfate layer are small. This suggests that it may be possible to avoid serious food security problems and deleterious impacts on tropical organisms, so long as the reduction in surface shortwave flux does not cause adverse impacts on crop yields. Furthermore, models tend to agree more in projections of future warming in the tropical regions (see Fig. 15b), providing some confidence that these tropical projections of surface temperature would be robust to the choice of model used to evaluate the impact of a geoengineering scenario. However, the inclusion of ocean dynamics is crucial, as even the sign of the equatorial Pacific Ocean temperature and precipitation response to increased CO₂ and sulfate burden depends on ocean dynamics. This affects the circulation response in the Southern Hemisphere through teleconnections, which greatly affect the surface air and ocean temperatures that bathe the ice shelves of West Antarctica and the Antarctic Peninsula (Ding et al. 2011).

**FIG. 14.** The difference in (a),(b) mean JJA and (c),(d) mean DJF precipitation rate (mm day⁻¹) between the geoco2 runs and their corresponding control runs for (a),(c) DISOM and (b),(d) OGCM. Dots indicate regions of significant precipitation change at the 95% level using a Student’s t test.
There is considerable uncertainty in other aspects of stratospheric aerosol injections that are not related to climate response per se: for instance, the creation of the sulfate aerosol layer itself. Heckendorn et al. (2009) have found that when using a 3D chemistry–climate model with a 2D aerosol model to simulate sulfur dioxide injection into the stratosphere, the aerosol sizes grow larger than expected. The result is an increase in particle sedimentation and an ensuing nonlinear relationship between aerosol burden and injection rate, resulting in even more aerosol being necessary to stabilize the global average temperature. Others have found that increasing the sulfate burden in the stratosphere could delay the recovery of the ozone hole by between 30 and 70 years due to the increased surface area that the sulfate aerosol provides to catalyze ozone destruction reactions (Tilmes et al. 2008).

Model uncertainties are amplified in the sensitive high latitudes (Randall et al. 2007; DeWeaver 2007). These uncertainties result in large differences in ocean and ice variables among the IPCC AR4 models forced with a business-as-usual greenhouse gas ramping scenario (Meehl et al. 2007) and contribute to the larger spread in polar climates among Coupled Model Intercomparison Project 2 (CMIP2), which are due to wide ranges of ocean heat transport, mean state of sea ice, and cloud cover variables (Holland and Bitz 2003; Holland et al. 2001; Bitz 2008). Figure 15a displays the intermodel average of the annual-average surface temperature difference between years 2080 and 2099 for the IPCC AR4 A1B emissions scenario, while Fig. 15b displays the standard deviation in the ensemble average change in annual average surface temperature. The greatest differences in the projected warming are found in the Arctic Ocean and Southern Ocean, particularly in the Ross and Weddell Seas under areas of current sea ice cover. These are the particular regions that may experience a climate emergency.

Thus, putting our geoengineering simulations in the context of the surface temperature spread in IPCC AR4 models, it is likely that the projected polar responses to geoengineering will be highly sensitive to the choice of the climate model. Accordingly, we endorse a call, put forth by Kravitz et al. (2011), for modeling centers to unite and execute a suite of coordinated, IPCC-style geoengineering simulations so as to sort out robust and nonrobust responses to geoengineering. However, as climate models fail to sample the long, low-probability tail of very high future warming that is ubiquitous in estimates of climate sensitivity (Randall et al. 2007; Knutti and Hegerl 2008; Roe and Baker 2007), they may also fail to sample the full response to geoengineering.

5. Conclusions

This study fills a gap in research on stratospheric aerosol injections by 1) imposing more realistic forcing scenarios, 2) executing geoengineering simulations using a standard resolution fully coupled atmosphere–ocean–sea ice model with state-of-the-art sea ice physics, and 3) comparing the result with a mixed layer ocean model with thermodynamic–dynamic sea ice to illuminate the role of ocean dynamics. Importantly, we have also focused on regions that have the potential to experience
a climate emergency. We have executed a suite of experiments to simulate stratospheric aerosol injections in a high CO$_2$ world (assuming that the delivery system will deliver sulfate aerosols with specified optical properties). In general, and in keeping with previous modeling work, the climatic effects of an aerosol layer plus doubled carbon dioxide are smaller than in a world with only doubled carbon dioxide. We have shown, however, that on seasonal time scales and regional spatial scales, stratospheric sulfate does not necessarily cancel all the effects of increased CO$_2$, especially circulation. In particular, there are still substantial climate changes in the very regions where climate emergencies may drive societies to geoengineer. Unfortunately, these are also the regions that suffer the greatest uncertainty in the response to forcing, due to strong coupling between the atmosphere, ocean, and sea ice and to deficiencies in the parameterizations of unresolved physics in the models (in particular, ocean mixing, sea ice rheology, atmospheric boundary layer processes, and clouds). Thus, one cannot rule out the possibility of geoengineering failing to avoid polar climate emergencies.

Countless other issues abound, both climatic and nonclimatic, including the ignorance of other consequences of increased carbon dioxide, such as ocean acidification. The likelihood of a climate surprise occurring due to geoengineering is high because research into geoengineering is still nascent, unintended consequences are a certainty, and the uncertainties of geoengineering are layered on top of those of global warming, complicating them. The question remains as to whether the apparent global warming abatement geoengineering may provide outweighs the (i) risk of unforeseen consequences being worse than predicted, (ii) risk of altogether unforeseen negative consequences, (iii) risk of failure in international cooperation, (iv) risk of failure of the chosen geoengineering mechanisms, leading to rapid temperature rise, and (v) risk of choosing winners and losers in the climate battle. It is our opinion that it would be imprudent to believe that the risk of unintended consequences is small enough to consider geoengineering a solution at this time. More research is required, and a coordinated modeling effort is a logical first step.

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