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Geoengineering climate by stratospheric sulfur injections: Earth system vulnerability to technological failure

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Abstract We use a coupled climate–carbon cycle model of intermediate complexity to investigate scenarios of stratospheric sulfur injections as a measure to compensate for CO_2-induced global warming. The baseline scenario includes the burning of 5,000 GtC of fossil fuels. A full compensation of CO_2-induced warming requires a load of about 13 MtS in the stratosphere at the peak of atmospheric CO_2 concentration. Keeping global warming below 2°C reduces this load to 9 MtS. Compensation of CO_2 forcing by stratospheric aerosols leads to a global reduction in precipitation, warmer winters in the high northern latitudes and cooler summers over northern hemisphere landmasses. The average surface ocean pH decreases by 0.7, reducing the calcifying ability of marine organisms. Because of the millennial persistence of the fossil fuel CO_2 in the atmosphere, high levels of stratospheric aerosol loading would have to continue for thousands of years until CO_2 was removed from the atmosphere. A termination of stratospheric aerosol loading results in abrupt global warming of up to 5°C within several decades, a vulnerability of the Earth system to technological failure.

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1 Introduction

Fossil fuel emissions are growing. Atmospheric CO_2 concentration is projected to increase to 540–970 ppmv (parts per million by volume) by the end of the 21st century (IPCC 2001). The worst-case scenario would be to burn all the available fossil fuel within the next several centuries, increasing atmospheric CO_2 concentration up to 1,400–1,800 ppmv. Globally-averaged annual surface air temperature could be warmer than today by 5°C for thousands of years (Archer and Brovkin 2008; Ridgwell and Hargreaves 2007), leading to the collapse of the major ice sheets and sea level rise ultimately of tens of meters. To avoid extreme climate change, current policy debates focus on a reduction in carbon emissions through increased energy efficiency and a shift from fossil fuels to renewable sources of energy, such as wind, solar, and geothermal energy, and biofuels, possibly complemented by large-scale use of nuclear energy. The ambitious plans of the European Union include a reduction in CO_2 emissions from their 1990 level by 20% by 2020. This is a challenging target, given that EU emissions in the year 2004 were 4.4% above the level for 1990, far from the Kyoto goal of a 10% reduction by the year 2012. At the same time, emissions are growing in developing countries not limited by Annex I of the Kyoto agreement, such as China and India. In the absence of visible progress in emission reductions, climate scientists have begun to discuss emergency approaches to cool the planet back to a secure level (Keith 2008).

Such climate modification options had already been proposed by Budyko (1977), who called the possibilities to inject sulfate aerosols into the stratosphere "artificial volcanos". As has happened after large volcanic eruptions, stratospheric aerosols could cool the Earth's surface by reducing the solar radiation reaching the lower atmosphere. This and other proposals for deliberate climate modifications (Marchetti 1977) raised serious concerns that our knowledge about the response of weather patterns to geoengineering is not yet adequate and that a forecast of long-term climate change with some "acceptable insurance" should be produced before any large-scale climate modification scheme comes into operation (Kellogg and Schneider 1974). Schneider (1996, 2001) warned about unexpected surprises in regional climate responses and pointed out the high probability that climate control would collapse on the centennial time scale necessary for the removal of anthropogenic CO_2 from the atmosphere. Recently, this proposal has been taken up by Crutzen (2006) who suggested stratospheric aerosol emissions as a sort of "emergency brake" to prevent the climate system from going out of control. Wigley (2006) compared effect of stratospheric aerosol emissions with volcanic eruptions and pointed out that to cool the climate down to the pre-industrial level, an emission on a scale of the Pinatubo eruption would be needed every second year. Other geo-engineering options include stabilization of the atmospheric CO_2 concentration through pumping liquid CO_2 into the deep ocean, land geological reservoirs (e.g., Hoffert et al. 2002), or accelerating the carbonation of rock minerals (Lackner 2003). Some approaches suggest solar deflectors placed in space (Early 1989; Hoffert et al. 2002). These measures would modify the climate in different ways and have long-term consequences for the land and marine biosphere, which it is important to test with Earth system models.

Concerning aerosol injection, model simulations by Govindasamy and Caldeira (2000) showed that, at a first approximation, the radiative forcing of CO_2 could be compensated by a reduction in solar radiation. A reduction in radiation would also mean a decrease in photosynthetically active radiation, but together with the increase in CO_2 concentration this would mean a net increase in the global plant productivity relative to present-day conditions (Govindasamy et al. 2002).

Here we use an Earth System model of intermediate complexity (EMIC), CLIMBER-2, to investigate the effect of compensating growing atmospheric CO_2 concentrations by injecting sulfur aerosols into the stratosphere. EMICs usually have a coarser spatial and temporal resolution than general circulation models, but they are computationally efficient and often include more climate system components than GCMs (Claussen et al. 2002). That makes EMICs ideally suited for the fast calculation of the long-term consequences of CO_2 mitigation measures (Lenton et al. 2006; Plattner et al. 2008).

We calculate the scenario of emission into the stratosphere of a sulfate aerosol homogeneously distributed over the globe at a rate sufficient to compensate for the global mean near-surface temperature change caused by the release of 5,000 GtC of fossil fuel, which is considered a conservative estimate of fossil fuel reserves (Kvenvolden 2002). In these scenarios, aerosol forcing compensates for any temperature increase above a certain threshold: 0°, 2°, or 4°C. We run our coupled climate–carbon cycle model for 10,000 years to evaluate the long-term dynamics of climate change, CO_2, and compensating sulfate forcing. We analyze the regional patterns of climate change at the CO_2 concentration peak (around year 2300) and illustrate an inherent risk of albedo enhancement with a scenario of a sudden technological breakdown of aerosol loading to the atmosphere. We comment on the economic costs of aerosol emissions in Section 4.

2 Methods

2.1 Climate–carbon cycle model

CLIMBER-2.3 includes a 2.5-dimensional statistical–dynamical atmosphere module with a coarse spatial resolution of 10° in latitude and 51° in longitude (Ganopolski et al. 2001; Petoukhov et al. 2000). The ocean component has three zonally averaged basins with a latitudinal resolution of 2.5° and 20 unequal vertical levels. A zonally averaged sea-ice module predicts ice thickness and concentration and includes ice advection. The model includes a terrestrial biosphere model, an oceanic biogeochemistry model, and a phosphate-limited model for marine biota (Brovkin et al. 2002). A deep-ocean carbonate sediment model (Archer 1991) calculates the carbonate compensation necessary to account for a long-term CO_2 uptake by the ocean. In response to a doubling of the atmospheric CO_2 concentration, the mean annual global surface air temperature and precipitation increase by 2.6°C and 9%, respectively. These values are within an uncertainty range of general circulation models (Petoukhov et al. 2005).

2.2 CO_2 scenarios

The CO_2 emission scenario is taken from (Archer and Brovkin 2008). This idealized scenario resembles the historical trend of fossil fuel burning from year 1700 until 2000. It assumes an emission of 5,000 GtC with 90% of this amount released during the period 2000–2300 (Fig. 1). A peak emission rate of 25 GtC/year is reached in year 2150. Although the amount of 5,000 GtC seems large, especially in comparison with total historical emissions of about 300 GtC, this scenario implies less fossil fuel burning in the twenty-first century than assumed by the IPCC SRES A2 scenario (Fig. 1, red line).

In all simulations, anthropogenic CO_2 emissions are added to the atmosphere and then interactively redistributed between the ocean and atmosphere in accordance with atmosphere–sea gas exchange. In these simulations, we neglected the land–atmosphere carbon flux for several reasons. At present and in the near future, the land carbon uptake substantially reduces the airborne fraction of fossil fuel emissions (Bala et al. 2005), but on a millennial time scale and for 5,000 GtC of fossil fuel emission, the amount of carbon uptake or release from the land biosphere will be small compared to the uptake of CO_2 by the ocean. Besides, we do not possess a prognostic model of land-use change or a scenario of land-use change after 2100. In addition, the role of land carbon in the global carbon balance in the future is quite uncertain. Simulations performed with different models suggest that land can be either a sink or a source of greenhouse gases depending on the soil respiration response to temperature change (Cramer et al. 2001; Friedlingstein et al. 2006) and land-use dynamics (Sitch et al. 2005). Hence we focus here on the ocean uptake and do not account for changes in land carbon storages to a first approximation. A weathering flux of $CaCO_3$ to the ocean (which balances the deep-ocean carbonate sedimentation in the pre-industrial simulation) was kept constant.

Radiative forcing of atmospheric CO_2 is accounted for in all simulations. We have not considered here radiative forcing of greenhouse gases other than CO_2, or the forcing of tropospheric sulfate aerosols. At present, negative radiative forcing of tropospheric aerosols roughly compensates the positive forcing of non-CO_2 greenhouse gases, such as CH_4 or N_2O, but in the future a reduction in tropospheric aerosol emissions could lead to an additional warming effect. This would require even stronger compensation by stratospheric aerosol emissions than in our simulations.

A second scenario of CO_2 dynamics is a stabilization of global temperature at 2°C threshold with CO_2 forcing only. This scenario was chosen to compare pure CO_2 forcing with combined CO_2 and aerosol forcings which keep the global temperature at the same threshold.

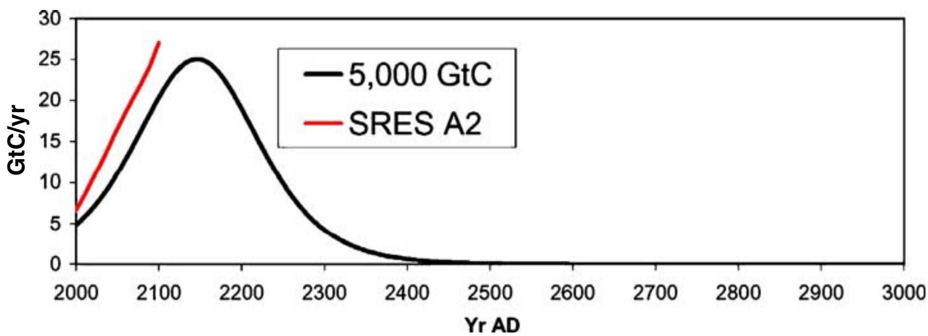

Fig. 1 Scenario of anthropogenic fossil fuel release of 5,000 GtC used in all simulations (*black line*), after (Archer and Brovkin 2008). Fossil fuel emissions from IPCC SRES A2 scenario for years 2000–2100 (*red line*)

2.3 Stratospheric aerosol forcing

The compensating stratospheric aerosol emission scenario is calculated so that the resultant change in the global mean near-surface temperature due to the combined CO_2 and aerosol effects, T_g, does not exceed a target temperature $T_g^* = T_g^0 + \Delta T_g$:

$$T_g \leq T_g^0 + \Delta T_g \tag{1}$$

where T_g^0 is the pre-industrial annual mean surface air temperature, and ΔT_g is a target change in temperature (taken as 0°, 2°, and 4°C).

Calculation of the radiative effect of essentially non-absorptive stratospheric aerosol is conducted according to the following algorithm (Charlson et al. 1991). The solar constant I_0 in the model is replaced by

$$\tilde{I}_0(t,\varphi) = I_0\left(1 - A^{\text{str}}(t,\varphi)\right) \tag{2}$$

where

$$A^{\text{str}}(t,\varphi) = \beta\left[1 - \exp\left\{-\delta_a^{\text{str}}(t)\sec\xi(t,\varphi)\right\}\right] \tag{3}$$

is albedo of the stratospheric sulfate aerosol dependent on latitude φ and time t. Thus we account for the effect of multiple reflections between the stratospheric sulfate aerosol layer and the underlying atmosphere and surface. In (3), $\sec\xi(t,\varphi)$ is the secant of the solar zenith angle $\xi(t,\varphi)$, $\beta = 0.23$ is the upward scattered fraction for the stratospheric sulfate aerosol particles, and $\delta_a^{\text{str}}(t)$is the stratospheric sulfate aerosol optical thickness. Here we assume, in line with Charlson et al. (1991) and Chylek and Wong (1995), that the upscattering fraction β can be assigned a constant value, which corresponds to an average solar zenith angle. The optical thickness is calculated as

$$\delta_a^{\text{str}}(t) = \sigma_a^{\text{str}} B_a^{\text{str}}(t), \tag{4}$$

where σ_a^{str}=3.5 m^2/g is the value of the mass-scattering cross-section specified for the stratospheric sulfate aerosol and $B_a^{\text{str}}(t)$ is the stratospheric sulfate aerosol loading (in g/m^2).

We assume that stratospheric sulfate aerosol is homogeneously distributed over the globe. The dynamics of the atmospheric load is calculated as a relaxation towards target temperature in the case that CO_2-induced warming overshoots the target:

$$\frac{\partial B_a^{\text{str}}(t)}{\partial t} = \begin{cases} k\left(T_g(t) - T_g^*\right), & \text{if } T_g(t) \geq T_g^* \text{ or } \quad B_a^{\text{str}}(t) > 0 \\ 0, & \text{if } T_g(t) < T_g^* \text{ and } \; B_a^{\text{str}}(t) = 0 \end{cases}, \tag{5}$$

where k (temperature sensitivity to aerosol emission) is a control parameter which determines how quickly the temperature $T_g(t)$ approaches the target temperature T_g^*. A value of $k = 0.01$ g/(m^{2}s K) was used for keeping the temperature near the target with precision $\varepsilon = 0.1$°C, fulfilling the condition $T_g(t) < T_g^* + \varepsilon$.

The rate of anthropogenic emission of the sulfate aerosol into the stratospheric layer, $E_a^{\text{str}}(t)$, was calculated in accordance with the balance equation

$$\frac{\partial B_a^{\text{str}}(t)}{\partial t} = E_a^{\text{str}}(t) - \frac{B_a^{\text{str}}(t)}{\tau_a^{\text{str}}}, \tag{6}$$

where τ_a^{str} is the residence time for the stratospheric sulfate aerosol. Substituting equation (5) into (6) gives the emission rate as

$$E_a^{str}(t) = \frac{\partial B_a^{str}(t)}{\partial t} + \frac{B_a^{str}(t)}{\tau_a^{str}} = \max\left(k\left(T_g(t) - T_g^*\right) + \frac{B_a^{str}(t)}{\tau_a^{str}}, 0\right). \qquad (7)$$

2.4 Numerical experiments

The background simulation CTRL includes the release of 5,000 GtC and no compensation for CO_2-induced warming. Simulations AER0, AER2, and AER4 denote the aerosol-induced compensation of global temperature above thresholds of 0°, 2°, and 4°C, respectively, in accordance with Eqs. 1–7. Simulation 2°CO2 is a simulation with pure CO_2 forcing which keeps global annual mean temperature increase at the 2°C level. Run INS2 is the same as AER2 but with compensation for radiative CO_2 forcing achieved through a reduction in the solar constant. The insolation forcing differs from the aerosol forcing because the latter includes the effect of multiple reflections between the stratospheric aerosol layer and land surfaces (as well as clouds) which is increasing from equator to poles. Therefore, the effective aerosol albedo $A^{str}(t, \varphi)$ depends on the latitude in the AER2 simulation. In the INS2 simulation, the aerosol effect is absent and the solar irradiation is equally reduced at each latitude. Simulation AER2CEASE follows AER2 till the year 2300 then sulfur emissions are set to zero. This scenario imitates an abrupt termination of aerosol compensation due to technological failure, a catastrophic event, or the inability to continue the sulfur emissions due to lack of resources.

3 Results

The temperature increase in the CTRL scenario reaches 2°C and 4°C in 2070 and 2130, respectively (Fig. 2a). A maximum value of about 7°C appears around the year 2350. Afterwards, the global temperature stays almost constant despite a decrease in the CO_2 concentration from a maximum of 1900 ppmv in the year 2320 to a value of 1,570 ppmv in year 3000 (Fig. 2b). This slow response of temperature is due to the long turnover time of the deep ocean where warming reaches its maximum in about year 3500 (not shown).

Compensation of CO_2-induced warming by stratospheric aerosols leads to stabilization of global temperature at levels of 0, 2, and 4°C (Fig. 2a) and even to a reduction in atmospheric CO_2 concentration because of the increasing CO_2 solubility in the ocean (Friedlingstein et al. 2006; Scheffer et al. 2006). The feedback leads to a difference of 200 ppmv (ca. 15%) between CTRL and AER0 simulations in the year 3000. An increase in carbon uptake in simulations with stratospheric aerosol injections due to an absence of temperature-driven suppression of carbon sinks was recently reported by Matthews and Caldeira (2007).

Stratospheric sulfur loads calculated in accordance with Eqs. 1–5 reach their maxima around year 2350 at about 13, 9, and 5 MtS for 0, 2, and 4°C thresholds, respectively (Fig. 2c and Table 1). Crutzen (2006) calculated a load of 5.3 MtS for mitigating a radiative forcing of doubling of the CO_2 concentration. In the AER0 scenario, a load of 12.7 MtS in the year 2300 compensates for an approximately

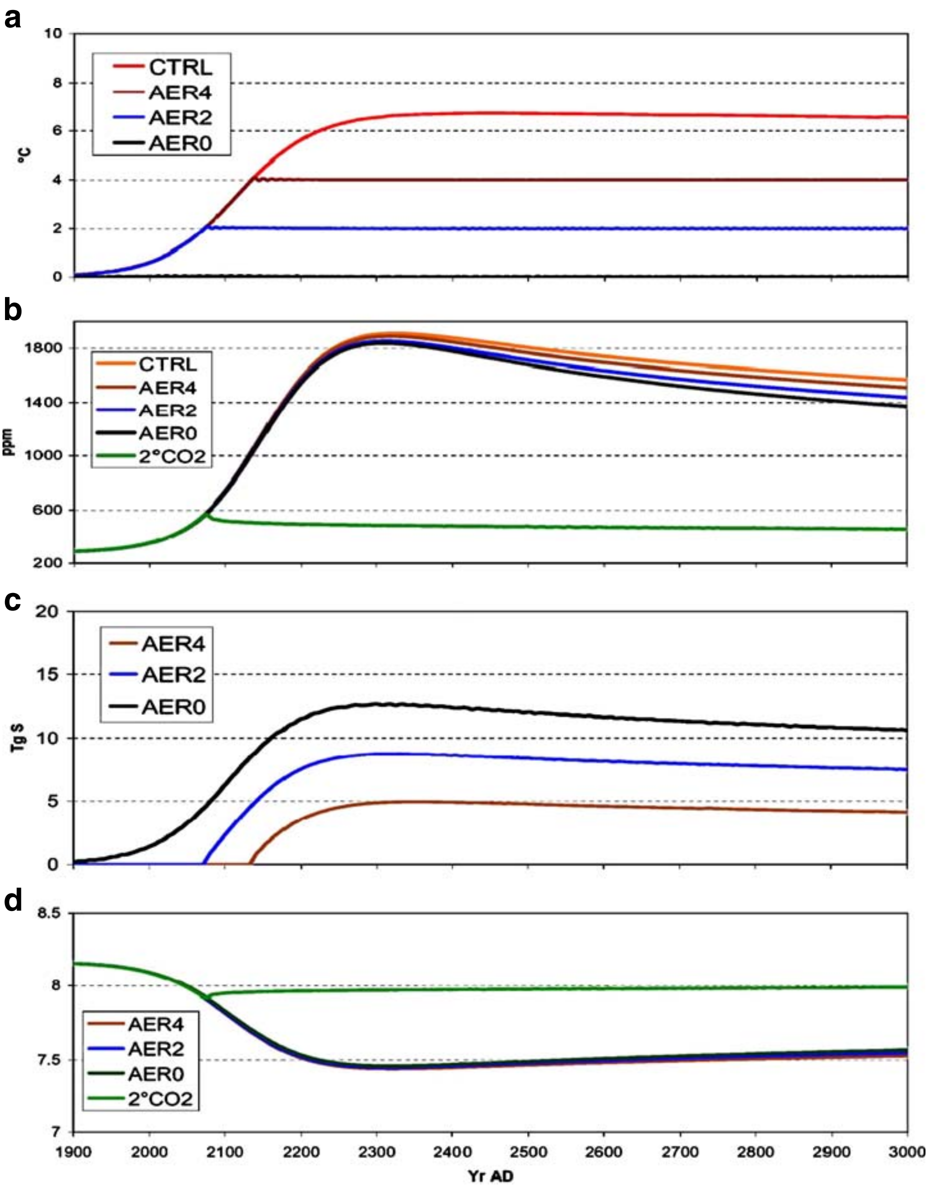

Fig. 2 Simulated changes in global mean annual surface air temperature in °C (**a**), atmospheric CO_2 concentration in ppmv (**b**), 10-year average of stratospheric sulfur load in MtS (**c**), and global mean surface ocean pH (**d**)

6.5-fold increase in the atmospheric CO_2 concentration from 280 to 1,836 ppmv. Assuming a logarithmic dependence of the radiative forcing on the CO_2 concentration, the 5.3 MtS load for the doubling of the CO_2 level would correspond to a 14.3 MtS load for the 6.5-fold increase in CO_2. Since the load in the AER0 scenario compensates not for the radiative forcing of CO_2 but for a surface air warming

Table 1 Simulation results for years 2100, 2300, 3000, and 10,000 A.D.

Simulation	Atmospheric CO_2, Ppm	Sulfur load, Tg S	Global ΔT[1], °C	Global ΔP[2], %	Surface ocean pH	Oceanic uptake, GtC/year
Year 2100						
2°CO2	513	0	2	8	7.95	1.7
AER0	732	6.2	0	−1	7.83	5.4
AER2	736	2.3	2	8	7.82	5.3
AER4	740	0	2.8	12	7.82	5.0
Year 2300						
2°CO2	487	0	2	8	7.97	0.7
AER0	1836	12.7	0	−3	7.46	3.9
AER2	1852	8.7	2	7	7.45	3.7
AER4	1886	4.9	4	17	7.44	3.3
Year 3000						
2°CO2	459	0	2	8	8.01	0.3
AER0	1372	10.6	0	−2	7.57	0.9
AER2	1439	7.5	2	7	7.55	0.8
AER4	1512	4.1	4	17	7.53	0.8
Year 10,000						
2°CO2	438	0	2	8	7.99	0.05
AER0	864	7.2	0	−1	7.77	0.03
AER2	894	4.6	2	7	7.76	0.03
AER4	933	1.8	4	17	7.74	0.03

which is not yet equilibrated in the year 2300 because of the thermal inertia of the ocean, our estimate of the load comes very close to the estimate calculated by Crutzen (2006).

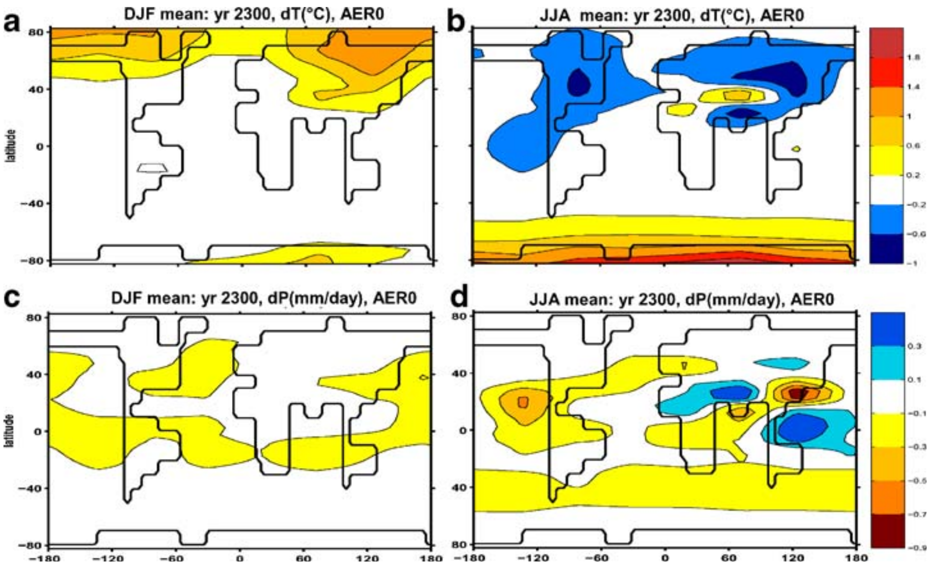

Fig. 3 Climate change in year 2300 relative to pre-industrial, simulation AER0. Shown are 10-year averaged changes in surface air temperature, °C, for December–January–February (**a**) and June–July–August (**b**). (**c**) and (**d**) are for changes in precipitation, in mm/day

The loads decrease after the peak around 2350, following the CO_2 dynamics as they compensate for the radiative forcing of excessive CO_2. However, even in the year 10,000 almost 8,000 years after the load peak, the loads are reduced only to 40–60% of their maxima (Table 1) because of the millennial lifetime of fossil fuel CO_2 in the atmosphere (Archer and Brovkin 2008). During the first thousand years after the end of the CO_2 emissions, an oceanic carbon uptake is constrained by CO_2 mixing in the deep ocean and dissolution of carbonate sediments on the ocean floor. On a longer time scale, the oceanic uptake is regulated by an imbalance between terrestrial weathering and carbonate sedimentation (the carbonate compensation mechanism, see Archer et al. 1997; Broecker and Peng 1987). In the year 10,000, carbonate compensation results in a small but not negligible carbon uptake (<0.1 GtC/year, Table 1). Carbonate compensation leads to a slow increase in the total oceanic alkalinity, a lowering of atmospheric CO_2 concentration, and a slow increase in oceanic pH from their minima around year 2300 (Fig. 2d and Table 1).

Regional patterns of the difference between aerosol and CO_2 forcings are shown in Fig. 3. In the AER0 simulation, globally averaged annual changes in air surface temperature are zero, but regional and seasonal temperature changes are substantial. During winter, polar regions are warmer by up to 2°C in both the northern (Fig. 3a)

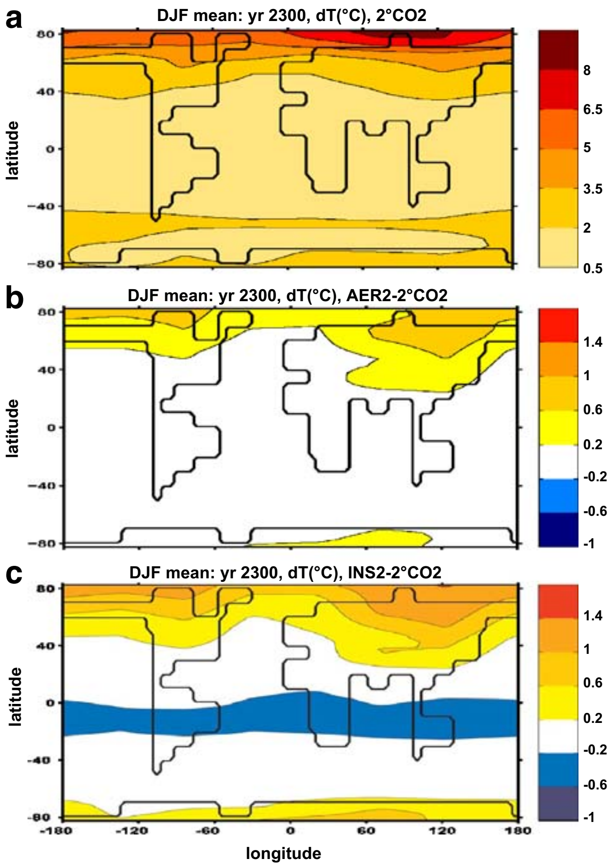

Fig. 4 Changes in surface air temperature (°C) in year 2300 relative to pre-industrial, simulation 2°CO2 (**a**). Difference in surface air temperature between simulations AER2 and 2°CO2 in year 2300 (**b**). Difference in surface air temperature between simulations INS2 and 2°CO2 in year 2300 (**c**). Shown are 10-year averaged values for December–January–February

and southern (Fig. 3b) hemispheres. The aerosol forcing is almost absent in high latitudes during winter time when the solar radiation flux is very small, while the longwave radiative CO_2 forcing is active during all seasons. In summer, negative radiative forcing of aerosols leads to about 1°C cooling over North America and Eurasia (Fig. 3b). At the same time, a warming of 0.5–1°C is simulated over the Tibetan plateau and North Africa. These regions have a relatively high surface albedo and adding an aerosol layer does not make their planetary albedo higher but lower. Aerosol over bright surfaces increases the multiple scattering of light between the surface and the atmosphere thereby enhancing the absorption of solar radiation. As a result, the air temperature over the regions with high surface albedo increases (Fig. 3b). In monsoon regions in East and South-East Asia, cooling leads to a reduction in precipitation. This is caused by a reduction in monsoon intensity associated with surface cooling and a reduced temperature gradient between ocean and land. In southern Europe and subtropical North America, summer aridity is increased as well. Precipitation is lowered over tropical land masses during all seasons (Fig. 3c–d), except for tropical Asian regions during June–August (Fig. 3d).

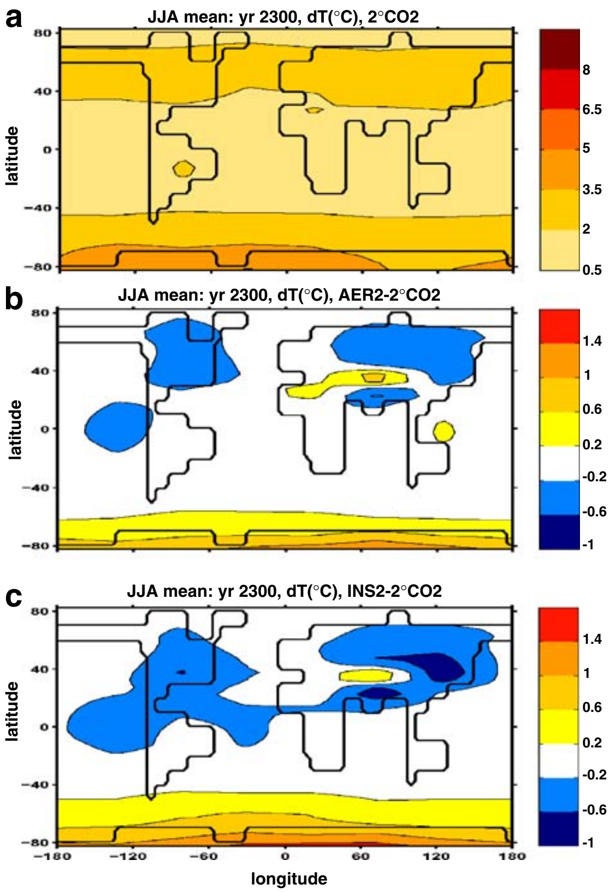

Fig. 5 The same as Fig. 4 but for June–July–August

The AER0 scenario with complete compensation of CO_2-induced warming requires a major effort as today's climate is about 1°C warmer than during the pre-industrial period. Stabilization of global warming at 2°C is more feasible. In the 2°CO2 simulation, climate is stabilized by reduction of CO_2 emissions, keeping the climate at the 2°C warming threshold. This warming is not homogeneous. An increase in temperature in the winter polar regions is very pronounced (up to 8°C in the northern hemisphere, Fig. 4a), while warming in tropics and subtropics is below 2°C in all seasons (Figs. 4a and 5a). This is a typical fingerprint of CO_2 forcing (IPCC 2001). Superposition of CO_2 and aerosol forcings has its typical geographical patterns as well. In the AER2 simulation, winter warming in polar regions is stronger by about 1°C than in the 2°CO2 simulation because of the higher CO_2 concentration (Figs. 4b and 5b). During summer, surface air over northern hemisphere landmasses is cooler by about 0.5°C because of the negative radiative forcing of aerosols with the exception of the Tibetan plateau and North Africa regions which have relatively high surface albedo.

In the INS2 simulation, excessive CO_2 forcing is compensated by interactive adjustment of solar irradiance (4% reduction in year 2300). Geographical patterns of temperature change due to insolation (Figs. 4c and 5c) are similar to temperature patterns of aerosol forcing (Figs. 4b and 5b) but more pronounced. Compensation with insolation reduction causes a slight cooling (about 0.3°C) in the tropical regions and cooling of up to 1°C over Eurasia and North America during summer (Fig. 5c).

A scenario of sudden technological breakdown AER2CEASE assumes the AER2 scenario until 2300 and with no sulfate emissions hereafter. The atmospheric sulfur load declines to zero within several years (Fig. 6a) and global temperature increases immediately (Fig. 6c). The abrupt warming reaches 4°C in a few decades and asymp-

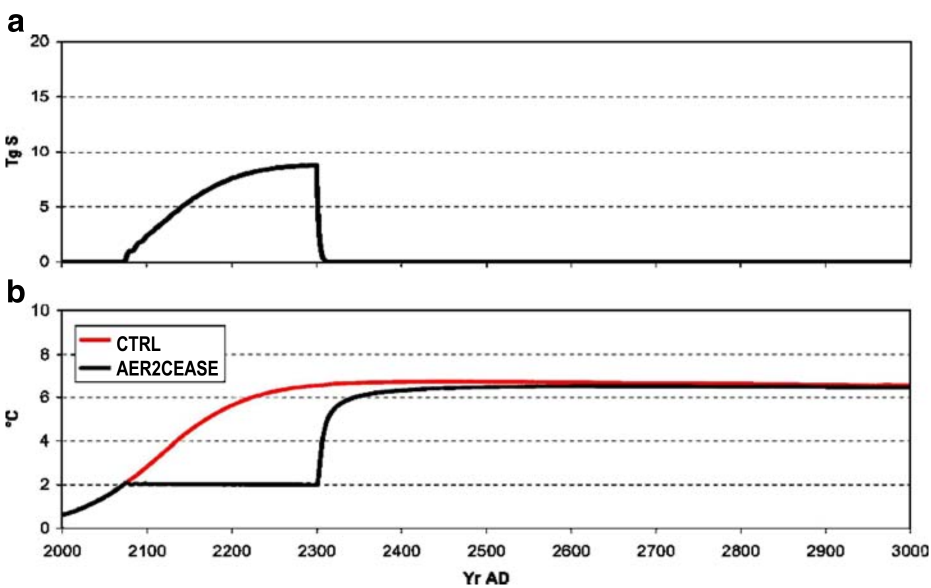

Fig. 6 Dynamics of stratospheric sulfur load in MtS (**a**) and changes in mean surface air temperature in °C in simulations AER2CEASE and CTRL (**b**)

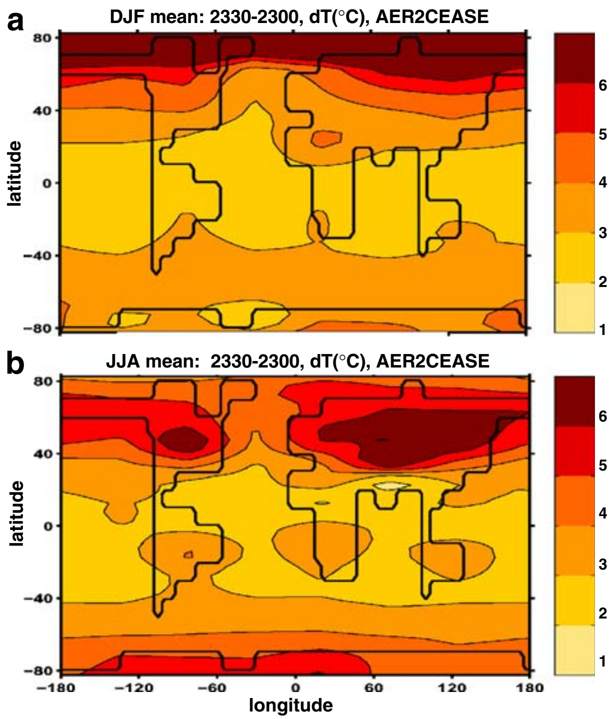

Fig. 7 Difference in surface air temperature (°C) between years 2330 and 2300, simulation AER2CEASE. Shown are 10-year averages for December–January–February (**a**) and June–July–August (**b**)

totically approaches the CTRL scenario hereafter. In 30 years after the emissions breakdown, the Arctic region is 6–10°C warmer in winter (Fig. 7a) while northern landmasses are about 6°C warmer in summer (Fig. 7b). This warming would be much more rapid than one of the most abrupt and extreme global warming events recorded in geologic history, the Paleocene–Eocene Thermal Maximum (PETM) event about 55 Myr ago, when sea surface temperatures rose between 5 and 8°C over a period of a few thousand years (Zachos et al. 2001). In the AER2CEASE simulation, a warming of similar magnitude occurs within few decades. An unprecedented abruptness of climate change as a consequence of a failure in geoengineering was stressed recently by Matthews and Caldeira (2007) who pointed out that the warming rates in this case could be up to 20 times greater than present-day rates.

4 Discussion

We use a geographically explicit climate–carbon cycle model to calculate sulfur load profiles that keep the temperature at a set of target levels (Fig. 2c). In the case of full compensation for radiative forcing of anthropogenic CO_2, the sulfur load is 13 MtS at maximum. Assuming a lifetime of stratospheric aerosols of about 2 years (Hansen et al. 1992; Minnis et al. 1993), required stratospheric emissions are about 7 MtS/year. This amount seems to be quite affordable in comparison with global sulfur production of 66 MtS/year in 2006 albeit most of current sulfur production is a result of processing of fossil fuels (USGS 2007).

Trenberth and Dai (2007) analyzed observed changes in the hydrological cycle after the Pinatubo eruption of 1991. They found a substantial decrease in precipitation over land (0.07 mm/day in annual terms) including a drought in South-East Asia. This is in agreement with the AER0 simulation in which annual precipitation over land declined by 0.05 mm/day and the most pronounced decrease in precipitation was simulated during summer over South Asian monsoon regions (Fig. 3d).

In 1992, a distinguished panel of the American National Academy of Sciences estimated that launching 1 t of sulfur at a height of 20 km by means of naval rifles would cost about 20,000 US$ (NAS 1992). The sulfur could be packed into 1 t shells, so that 7 MtS could be propelled by 700 naval rifles firing about 30 times a day (other options could rely on suitably equipped airplanes or rockets). Total costs would be about 140 billion US$—less than 0.5% of today's Global Domestic Product (GDP). GDP would increase due to economic growth, while propulsion costs would decrease due to learning by doing (see Teller et al. 1997 for prospects of cost reductions by a factor of 10 to 1,000). Aerosol injection may have to continue for millennia due to the very long period of anthropogenic CO_2 persistence in the atmosphere (Archer and Brovkin 2008; Bengtsson 2006). But to the extent to which it makes sense to assess long-term socio-technical trajectories in terms of GDP, total costs amount to less than 0.1% of world GDP—as long as collateral risks can be excluded.

In this respect, our results support the previous assessments by Crutzen (2006) and Wigley (2006) that stratospheric sulfur injections might be a feasible emergency solution for cooling the planet. Of course, other impacts of sulfur emissions, such as their effect on stratospheric ozone and the environmental consequences of sulfur depositions should be vigorously tested, but this is beyond the scope of our model. What needs to be stressed here is that sulfur emissions would have to continue for millennia unless later generations find a secure way to remove CO_2 from the atmosphere. It is clear that this would require a degree of geopolitical stability that is not very likely even for the next decades, let alone millennia.

A critical consequence of climate geoengineering is a possibility of extremely rapid warming in case the emissions are abruptly interrupted, as in the AER2CEASE scenario. Within a few decades, winter warming in the polar regions exceeds 10°C, and summer warming in the northern temperate latitudes will be about 6°C, from an initial controlled climate that was 2°C warmer than the pre-industrial climate (Fig. 7). Such a rapid warming would be a disaster for ecosystems, especially fragile ones in the polar regions. The amplification of global warming through emissions of methane released from thawed permafrost regions and, later, from methane hydrates stored on the continental slopes in the ocean, would seem to be unavoidable. Coming generations would have to live with the danger of this "Sword of Damocles" scenario, the abruptness of which has no precedent in the geologic history of climate.

When trying to assess the consequences of possible failures of geoengineering, the difficulties in even assigning probability measures to such failures amplify the reasons for a risk-adverse agent to avoid such risks (Weitzman 2007). Moreover, when assessing these kinds of risks, monetary units are not an appropriate metric (Jaeger et al. 2008). In the past, experiences of dreadful suffering like World War I did not affect global GDP growth, because GDP can grow by producing weapons as much as by producing food. The ethical reasons for avoiding war, however, cannot simply be traded against opportunities for increased consumption—they are a case of lexical preferences, where one kind of preference has precedence over another kind. Therefore, expressing the suffering caused by World War I in monetary units would become not only arbitrary, but actually misleading. An analogous situation arises when considering a geopolitical crisis in a world depending on steady aerosol injection to counter global warming.

Moreover, limiting the temperature increase to 2°C in the AER2 scenario will not stop the long-term increase in sea level which could be about 20–30 m in 10 to 30 thousand years (Archer 2007). The resulting drowning of basically all present-day coastal cities would be a barbaric act few would dare to envisage, even if it would have no impact on future economic growth. Basic patterns of biodiversity, cultural heritage, and many dimensions of human life give rise to lexical preferences much as the problems of war discussed in the previous paragraph.

Reducing temperatures back to the pre-industrial level in the AER0 scenario will solve the problem of sea level rise, but it will reduce precipitation almost everywhere, including the densely populated monsoon regions in Asia (Trenberth and Dai 2007). This suppressing effect of aerosols on the monsoon is already quite remarkable (Cramer 2006) and can be higher in the future (Mudur 1995; Zickfeld et al. 2005). And ocean pH will decrease by 0.7 units leading to drastic consequences for marine carbonating species, especially in the polar regions (Cao et al. 2007; Riebesell et al. 2000).

In our analysis, we did not account for ice sheets dynamics, land use changes, atmospheric chemistry, or non-CO_2 greenhouse gases. The approach we are using here is limited in many aspects. In particular, the atmospheric model has only one stratospheric layer. We understand that our model is far from being perfect and that is it obviously necessary to continue with the development of climate system models which could account for stratospheric dynamics, atmospheric chemistry, and other important aspects of the climatic response to sulfate injections into the stratosphere. However, we think that our model provides physical insights into the first-order response of the climate–carbon cycle system to the CO_2 and aerosol forcings. Similarity to the results of Govindasamy and Caldeira (2000), Wigley (2006), and Matthews and Caldeira (2007) supports us in this conclusion.

In summary, we have found that climate change mitigation using stratospheric aerosol emissions is associated with high risks which will persist for centuries and even millennia. Therefore, this geoengineering option cannot be seen as a solution to the problem of human-made climate change. Assessment of this option and its consideration as a sort of emergency brake in case climate change becomes too dangerous must not distract the scientific mainstream from searching for sustainable approaches to diminishing economic dependence on fossil fuels (Lawrence 2006; Pacala and Socolow 2004; Schneider 1996). There is a long way to go, but this is the only secure way to avoid the high risks of dangerous anthropogenic climate change in the future.

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